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**THE START-PERRANPORTH ZONE-
TRANSPRESSIONAL REACTIVATION ACROSS A
MAJOR BASEMENT FAULT IN THE VARISCAN
OROGEN OF S.W. ENGLAND**

by
Simon Andrew Steele

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**A thesis submitted in partial fulfilment of the degree of Doctor of Philosophy at
the Department of Geological Sciences, University of Durham.**

1994



- 1 MAY 1995

To Mum and Dad

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1994

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ABSTRACT

The Start-Perranporth Zone is one of a number of E-W trending zones in S.W. England, which are characterised by an anomalous and structurally complex deformation history, and which are thought to reflect the influence of pre-existing basin architecture. The SPZ straddles the Start Complex in S. Devon, and is approximately coincident with the northern margin of the Gramscatho Basin in S. Cornwall. It appears to coincide with significant sedimentological, geochemical and metamorphic transitions, and may mark the site of a pre- to Early Devonian terrane boundary. This terrane boundary may have formed the northern margin to a series of small possibly transtensional basins, including the Start and Gramscatho Basins, in which thick successions accumulated prior to inversion during the Variscan orogeny. The pelitic sequences in these basins (Gramscatho Group sandstones, Start greyschists) are geochemically similar to one another, and to other Rhenohercynian basinal sequences in mainland Europe. Both the Gramscatho and Start basins are characterised by the presence of incipient ocean crust (Lizard ophiolite, Start greenschists), with a strongly depleted N-type MORB signature and evidence of ridge-related sub-oceanic early deformation.

The interlayered green and grey schists of the Start Complex are separated from the shallow marine Meadfoot shales to the north by a steep north dipping normal fault, the Start Boundary Fault, which bears evidence of a long-lived movement history. This fault is intimately associated with large volumes of highly altered and replaced basic intrusives, and appears to be the surface manifestation of the basin bounding fault at depth. Approaching the SBF, the strain intensifies, primary folds tighten, the primary cleavage steepens to sub-vertical and mineral stretching lineations switch from SSE plunging (sub-parallel to the Variscan transport direction) to sub-horizontal approximately E-W trending. Immediately adjacent to the SBF, sheath folds occur, suggesting very high along strike shear strains. Small scale structures, e.g. shear bands, refold relationships, etc. consistently indicate that dextral simple shear is important during Variscan shortening. Similar, though somewhat more cryptic, evidence for dextral shear is also seen in the L. Devonian shales north of the SBF.

In S. Cornwall there is a similar focusing of high strain along the northern Gramscatho margin, with a tightening of folds, a backsteepening of the primary cleavage, and the development of overprinting late crenulations. Primary stretching lineations lie E-W. There is no evidence for sheath folding on either coast, although broad phyllonite zones bearing dextrally asymmetric quartz augen provide evidence of long-lived dextral shear. Many of the high strain fabrics on the east coast are absent, probably faulted out along a major NW-SE dextral strike-slip fault (the Pentewan Fault).

The small scale structural evidence along this zone consistently indicates that dextral transpression was the dominant deformation mechanism during Variscan orogenesis. The structural transitions are also suggestive of fault buttressing, e.g. secondary backfolding, backthrusting, etc. and it appears that the ~E-W trending basin bounding fault acted as an oblique buttress to the NNW directed Variscan nappes, the high angle obliquity of this collision inducing dextral transpression in the shortening cover sequence. This fault buttressing mechanism readily accounts for all of the observed anomalous small scale structures, and the marked along strike persistence of the anomalous zone.

CONTENTS

Copyright	iii
Acknowledgements	iv
Abstract	v
Contents	vi

Chapter 1: Basin inversion in orogens, transpressional tectonics and structural nomenclature

1.1 Basin inversion in orogens	1
1.2 Transpression	2
1.2.1 Introduction	2
1.2.2 Modelling transpressive strains	2
1.2.3 Kinematic partitioning of transpressive strains	8
1.2.4 Folding in transpressive regimes	11
1.2.5 Transpression at fault and shear zone bends	16
1.2.6 Summary	18
1.3 Complex fold geometries	18
1.3.1 Fold interference patterns	18
1.3.2 Curvilinear fold axes	19
1.4 Shear criteria	21
1.4.1 Ductile shear criteria	21
1.4.2 Brittle shear criteria	25
1.5 Vergence	29
1.5.1 Fold vergence	29
1.5.2 Cleavage vergence	29
1.6 Facing	30
1.6.1 Areas of refolding	32
1.6.2 Areas of sheath folding	32
1.7 Aims of thesis	33

Chapter 2: The stratigraphic and structural evolution of S.W. England

2.1 The Variscan orogen	34
2.1.1 Introduction	34
2.1.2 Southwest England	36
2.2 The stratigraphic evolution of S.W. England	37
2.2.1 The pre-Devonian basement	37
2.2.2 Upper Palaeozoic stratigraphy of S.W. England	38

2.3 The tectonic evolution of Southwest England	54
2.3.1 Introduction	54
2.3.2 Early work	55
2.3.3 Recent tectonic models	57
2.4 Geophysical evidence and its bearing on tectonic models	64
2.5 Summary	65

Chapter 3: The lithology and geochemistry of the Start Complex and adjacent regions, South Devon.

3.1 Introduction	67
3.2 Geochemistry and petrology of the Start Complex	69
3.2.1 The Greenschists	69
3.2.2 The Greyschists	78
3.2.3 Intermediate lithologies	86
3.2.4 Offshore extent of the Start Complex	86
3.2.5 Summary	87
3.3 The Lower Devonian succession	88
3.3.1 Introduction	88
3.3.2 The Dartmouth Beds	89
3.3.3 The Meadfoot Beds	90
3.4 The intrusives	94
3.4.1 Pre-orogenic intrusives	94
3.4.2 Post-orogenic intrusives	96
3.5 Discussion	98

Chapter 4: The structural evolution of the Start Complex and adjacent regions

4.1 Introduction	101
4.2 The Start Complex	101
4.2.1 Introduction	101
4.2.2 Previous deformation chronology	103
4.2.3 The minor structures of the greyschists	105
4.2.4 The minor structures of the greenschists	118
4.2.5 The microstructure	126
4.2.6 Large scale structure of the Start Complex	128
4.3 The Start Boundary Fault Zone	132
4.3.1 Introduction	132
4.3.2 Greenstraight	133
4.3.3 Salcombe Estuary	139

4.3.4 Inner and Outer Hope	144
4.3.5 The petrology and microstructure of the fault rocks	148
4.4 The Devonian sequence	151
4.4.1 Introduction	151
4.4.2 Previous deformation chronology	151
4.4.3 The west coast section	153
4.4.4. The east coast section	157
4.4.5 The microstructure	162
4.4.6 Summary	164
4.5 Structural correlations across the SBF, and its regional significance	164
4.6 Discussion and evidence for transpression	167
4.6.1 Origin of east-west lineations	167
4.6.2 Relationship between F2 folds and the early lineations in the Start Complex	170
4.6.3 F3 fold development in the Start Complex	171
4.6.4 Sheath folds at the Start Boundary Fault: geometry and significance	173
4.6.5 Shear bands and extensional fractures	175
4.6.6 Late faults	182
4.7 Summary	184
Chapter 5: The structural evolution of the northern Gramscatho margin	
5.1 Introduction	186
5.2 Sedimentology of the northern Gramscatho margin	190
5.2.1 Introduction	190
5.2.2 Previous work	190
5.2.3 The Gramscatho Beds	191
5.2.4 The Meadfoot Beds	195
5.2.5 Summary	199
5.3 Petrology of the igneous rocks	199
5.3.1 Introduction	199
5.3.2 Pre-orogenic intrusives	200
5.3.3 Post-orogenic intrusives	201
5.3.4 Summary	204
5.4 The structure of the eastern coastal sections	204
5.4.1 Introduction	204
5.4.2 Previous work	205
5.4.3 Mevagissey Bay	206
5.4.4 St. Austell Bay	215
5.5 The structure of the western coastal sections	223

5.5.1 Introduction	223
5.5.2 Previous work	223
5.5.3 The primary structure-Towan Head to Holywell Bay	225
5.5.4 The high strain (D2) zone-Holywell Bay to Perran Bay	230
5.5.5 Discussion	238
5.6 Conclusions and discussion of the evidence for dextral transpression	239
5.6.1 The northern margin of the Gramscatho Basin	239
5.6.2 Evidence for transpression	244
5.6.3 Post-orogenic extensional collapse	250
Chapter 6: Conclusions: The evolution of the Start-Perranporth Zone	
6.1 Introduction	253
6.2 The Start Complex	253
6.3 The northern Gramscatho margin	255
6.4 The evolution of the Start-Perranporth Zone	256
References	259
Appendix 1: Analytical geochemistry	274
Appendix 2: Simple constructions in transpression zones	281

CHAPTER 1

BASIN INVERSION IN OROGENS, TRANSPRESSIONAL TECTONICS AND STRUCTURAL NOMENCLATURE

1.1 Basin Inversion in Orogens

In recent years, the concept of basin inversion has been increasingly applied to the restacking of highly stretched continental margins within orogenic zones, with examples from the Alps (Butler, 1989; Hayward & Graham, 1989), the Subvariscan foldbelt of South Wales (Gayer & Jones, 1989; Powell, 1989) and the Rhenohercynian of SW England (Pamplin, 1988; Hartley & Warr, 1990; Warr, 1991).

Williams *et al.* (1989) define basin inversion as the process which occurs when basin controlling extensional faults reverse their movement during compressional tectonics, and to varying degrees, basins are turned inside out to become positive features. Basin stratigraphy developed before, during and after extensional fault movement may be described as pre-, syn- and post-rift sequences.

The control exerted by these early extensional faults on the compressional geometries and kinematic patterns during orogenic deformation and basin inversion is being increasingly recognised from a large number of mountain belts (Hayward & Graham, 1989). The pre-existing extensional faults may be responsible for triggering ramp generation and can complicate both the trend and sequence of thrusting. One important effect that these faults have on the cover sequence, is described from the Alps by Hayward and Graham (1989). They note that adjacent to old extensional faults the syn-rift sediments exhibit an intensification of internal deformation (folds and cleavage), the implication being that the detachment at the base of the cover sequence is transformed into strain in the vicinity of the 'blockage' represented by the old extensional faults. In addition, backthrusting, which represents the backwards expulsion of the syn-rift sequence, can be seen to develop. The 'blockage' is referred to as a 'buffer-stop', or more commonly as a 'buttress' (Butler, 1989) and the process whereby there is a marked intensification and localisation of strain adjacent to a former basin bounding fault, is referred to as 'buttressing'. However, it is reasonable to expect a non-orthogonal relationship between the orientation of these faults, and the direction of maximum shortening, and it is highly likely that localised transpression zones will arise in the cover sequence as a consequence of oblique buttressing.



1.2 Transpression

1.2.1 Introduction

The term transpression was first used by Harland (1971) to describe the deformation that arose from oblique plate convergence in Caledonian Spitsbergen. Harland (1971) defined transpression, in terms of stress, as those regimes which operate in zones of oblique compression. This definition, however, implies an association between transpression and oblique plate collision, which clearly is not the exclusive tectonic setting for transpression zones. Also, given the complexity of strain in natural examples of transpression zones, the evaluation of stress directions is very difficult. Sanderson and Marchini (1984) redefined transpression as '...a wrench or transcurrent shear accompanied by horizontal shortening across, and vertical lengthening along, the shear plane'. This definition is non-specific to any one plate-tectonic setting, and describes the finite strain state within a transpressional environment.

1.2.2 Modelling Transpressive Strains

Sanderson and Marchini (1984) produced a general mathematical model for transpression by factorising the deformation into its pure shear and simple shear components. In their model, they assume that the transpression zone is laterally confined (i.e. there is no extrusion of material at its ends) and that there is no volume loss from the system. Thus, in order to conserve volume, the shortening across the zone must be compensated for by vertical thickening. Figure 1.1 shows the transpression geometry and defines the parameters α^{-1} and γ , where α^{-1} specifies the shortening across the zone (strictly it is the ratio of the deformed to original width of the zone), α specifies the vertical stretch and γ is the shear strain parallel to the zone.

Finite Transpressive Strain

The above factorization allows the strain to be specified in terms of two parameters, α^{-1} and γ (Sanderson & Marchini, 1984). By assigning different values to these parameters, the finite strain can be assessed for different degrees of shortening and shear strain (see Sanderson & Marchini, 1984 for method). The values of the principal strain axes are plotted onto a contoured logarithmic Flinn diagram to give a finite deformation plot (Fig. 1.2). Figure 1.2 clearly shows that the shape of the strain ellipsoid varies with α^{-1} : when $\alpha^{-1} < 1$, k values less than 1 result; when $\alpha^{-1} = 1$ then $k = 1$ and when $\alpha^{-1} > 1$ then k values greater than 1 are produced. Thus, transpressional

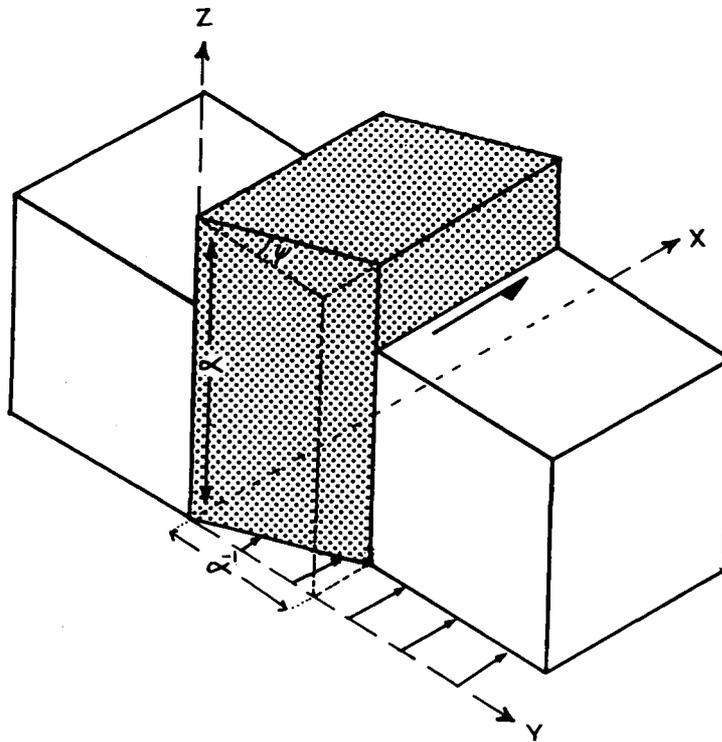


Figure 1.1 Transpression geometry, showing deformation of a unit cube by shortening parallel to Y-axis and shear parallel to X-axis. Volume is conserved by lengthening parallel to Z-axis (vertical). After Sanderson & Marchini (1984).

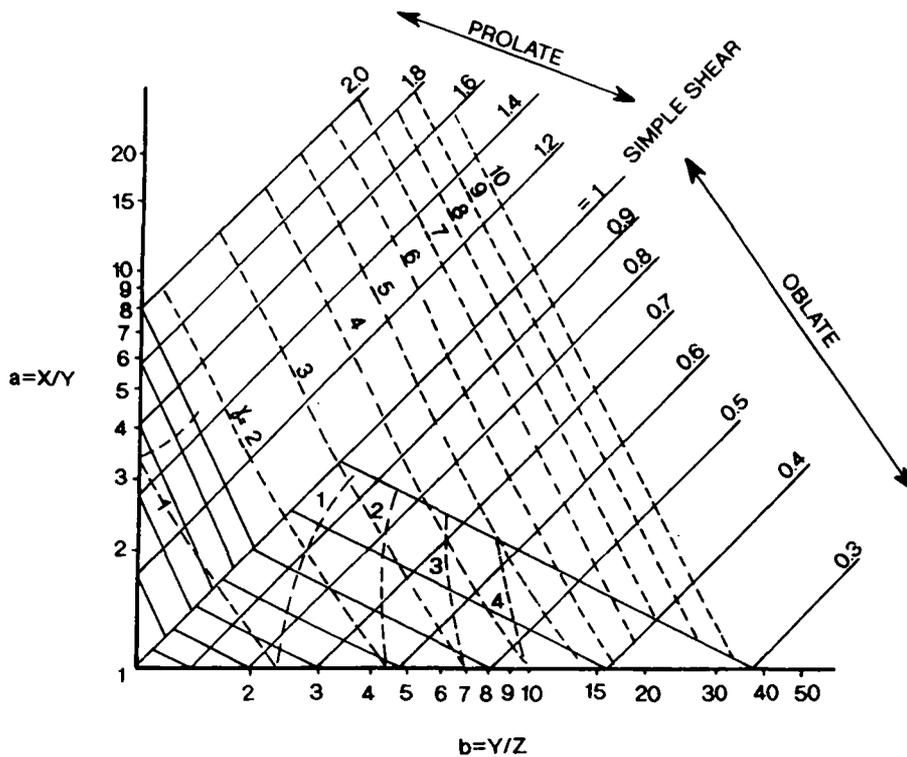


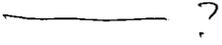
Figure 1.2 Flinn diagram showing axial ratios $a=X/Y$ and $b=Y/Z$ produced by transpressional model for various values of α^{-1} (continuous lines) and γ (dashed line). After Sanderson & Marchini (1984).

shear zones contain S(L)-dominated fabrics and transtension zones L(S)-dominated fabrics (Sanderson & Marchini, 1984; Fossen & Tikoff, 1993). Although Figure 1.2 gives the shape of the finite strain ellipsoid, it does not show variations in the orientation of the principal strain axes, one of which must be vertical. For simple transcurrent shear ($\alpha^{-1}=1$), the Y-axis is vertical. For transpression ($\alpha^{-1}<1$) either X or Y may be vertical, and thus the XY plane (cleavage?) is always vertical, lying at an angle θ to the zone boundary (Fig. 1.3). For transtension ($\alpha^{-1}>1$) either Z or Y may be vertical, and thus the XY plane 'switches' between the vertical and the horizontal. The concept of switching of the principal strain axes, manifest as stretching lineations in deformation zones in the case of the X-axis, is fundamental to transpressional tectonics. In Figure 1.2, the X axis switches from vertical to horizontal with increasing shear strain (γ), where the solid lines 'bounce' off the b-axis of the Flinn plot. Such switching is also evident in Figure 1.3. Sanderson & Marchini (1984) make it quite clear that both Figures 1.2 & 1.3 describe the finite strain fields, and it is important not to assume that the lines on these diagrams represent true deformation paths, since any finite strain state may be reached by an infinite number of deformation paths. In order to fully understand transpression zone kinematics, further assumptions must be made about the deformation paths.

Incremental Transpressive Strains

Two special cases were considered by Sanderson and Marchini (1984) in order to approximate deformation paths.

i) Constant Incremental Strain

By letting $\gamma \rightarrow 0$ and $\alpha^{-1} \rightarrow 1$, the incremental strain can be approximated (see Sanderson & Marchini, 1984 for methodology), and the deformation path may be modelled. These show that, as with the models above, the finite strain axes may swap along some paths of constant incremental strain, and thus the maximum finite stretch may be normal to the maximum incremental stretch. The relationship between those structures related to finite strain (cleavage, stretching lineations, etc.) and those related to incremental strain (fractures, fibres, etc.) may thus be very complex. Also, the incremental strain axes will be parallel to the principal stresses and thus the influence of α^{-1} on the orientation of compressional (folds, thrusts, etc.) and extensional (veins, dykes, normal faults, etc.) structures within a transpression zone can be assessed if one assumes an elastic, isotropic rheology (Fig. 1.4). McCoss (1986) provides a simple geometrical construction to determine the relative magnitudes of the principal incremental strain axes, assuming the constant volume, homogeneous strain model of Sanderson and Marchini (1984; Fig. 1.1). This involves .

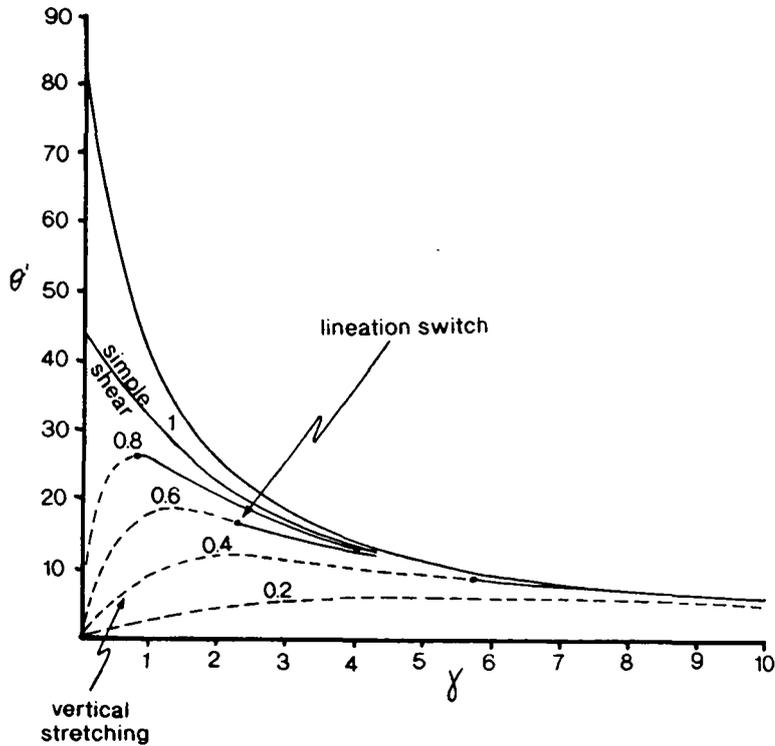


Figure 1.3 Plot of orientation of long axis of strain ellipse in horizontal plane (θ'). Continuous lines indicate X-axis horizontal, dashed lines indicate X-axis vertical. Note for α^{-1} values between 0.8 and 0.4 stretching axis (X) switches from vertical to horizontal with increasing γ . After Sanderson & Marchini (1984).

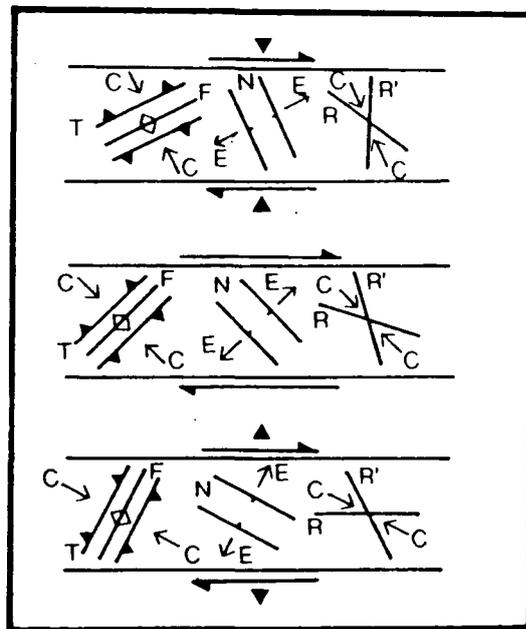


Figure 1.4 Diagram to show the initiation orientations of fractures and associated structures for a Simple Transpression model (top), compared with simple shear (centre) and transension (bottom). C, 'compression axis' (σ_1); E, 'extension axis' (σ_3); N, normal faults; T, thrust faults; R, R', conjugate Riedel shears or wrench faults; V, veins, dykes or extension fractures; F, fold axes. After Sanderson & Marchini (1984).

measuring the angle (A) between \underline{S} (the direction of the zone boundary displacement*) and the zone normal (Fig. 1.5). The incremental relative magnitudes of the principal axes and the ellipsoidal shape are a function of A , and are outlined in Figure 1.6. In Figure 1.6, eight precise values of A can be recognised, which bound distinctly different tectonic regimes ($A=0^\circ, 180^\circ$ and the sinistral and dextral systems where $A=70.5^\circ, 90^\circ$ & 109.5° ; see also Table 1 in Appendix 2). These critical field boundaries can be extended into the finite strain field, since most of the critical angles remain fairly stable with increasing strain. The exceptions are $A=70.5^\circ$ (ASTP or axially symmetric transpression angle) and $A=109.5^\circ$ (ASTT or axially symmetric transtension angle) where the boundaries migrate slowly towards 90° with increasing strain. This migration has two important consequences on the kinematics and deformation paths of transpression zones. Firstly, if the direction of displacement is constant, A will be fixed, and a switching of principal strain axes can only occur within the zone defined by $70.5^\circ < A < 109.5^\circ$, and then only at high strains, unless A lies very close to the ASTP angle. Secondly, the stability of a regime with a constant motion vector suggests that if axis switching does occur, it is more likely to be due to changes in the displacement direction relative to the zone boundary. Multiple switching occurring within a zone of superimposed incremental strains may suggest that the relative displacement direction is varying slightly about the critical ASTP angle (McCoss, 1986).

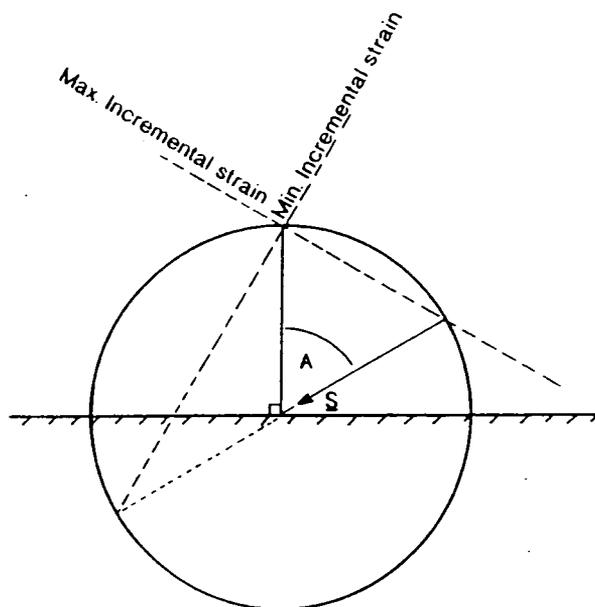


Figure 1.5 The geometrical construction for ASTP transpression. Modified after McCoss (1986: see Appendix 2 for methodology of construction).

* In the case of constant incremental strain \underline{S} represents the zone boundary displacement and is underlined to distinguish it from S , an increment of shortening in Simple Transpression.

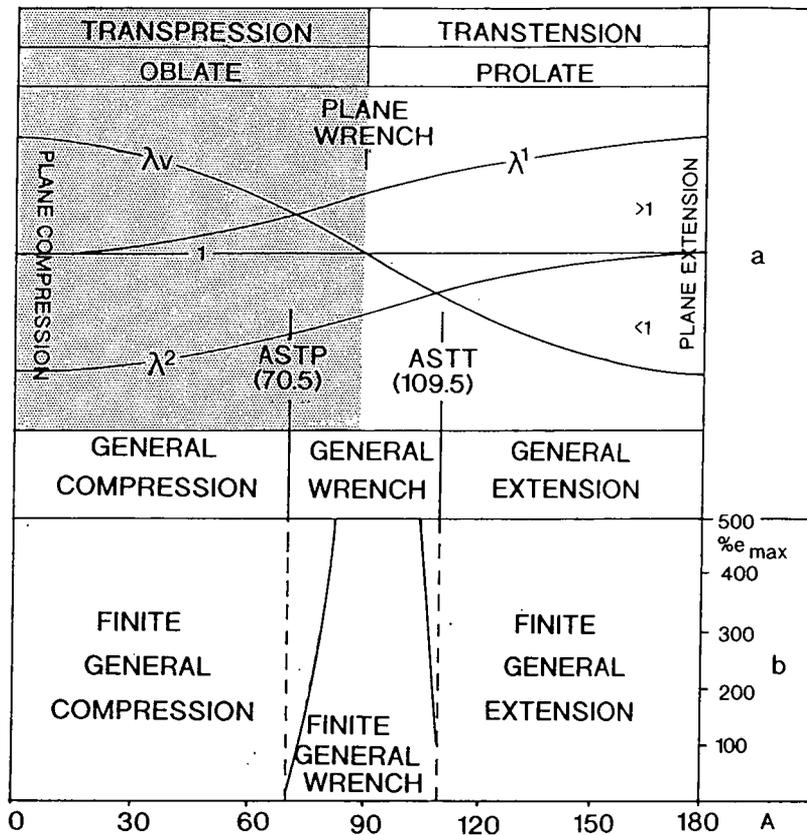


Figure 1.6 A graphical solution for determining the characteristics of (a) the incremental strain ellipsoid and (b) the finite strain ellipsoid. The orientations where axially symmetric transpression and transtension occur are shown by ASTP and ASTT, respectively. λ_v is the vertical principal quadratic elongation. $\lambda_1 > \lambda_2$ are horizontal principal strains. e_{max} is the maximum finite extension. Shaded quadrant highlights the transpressional regime. Note that lineation switching occurs at ASTP where λ_v crosses λ_1 . Modified after McCoss (1986).

ii) Simple Transpression

Deformation paths are probably determined by two main factors, the external boundary displacements of the system and the internal rheological variations of which layering is an important geological example (Sanderson & Marchini, 1984). Harland (1971) suggested a scenario, which he called simple transpression, in which the deformation can be quantified in terms of the boundary conditions resulting from two rigid boundaries approaching one another obliquely (Fig. 1.7). Assuming that the deforming material is isotropic, then the finite strain can be determined at various increments of shortening (S), and hence the deformation path may be specified. Thus the parameters α^{-1} and γ can be expressed in terms of S for any given value of β (Fig. 1.8), where β is the angle between the relative movement vector and the rigid boundary. The deformation paths that arise (Fig. 1.8) are similar to those in the constant incremental strain model, with axis swapping where β is small. This is similar to the modelling of McCoss (1986) which showed that lineation switching occurs when the angle between \underline{S} and the zone boundary (i.e. $90-A$) is low. If assumptions are made about the boundary conditions and strain distribution, modelling of the transpressive strains that

arise from oblique continental collision could be attempted (Sanderson & Marchini, 1984).

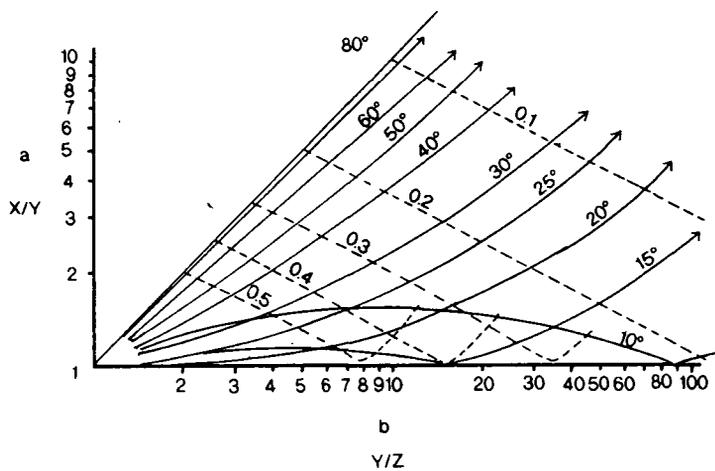
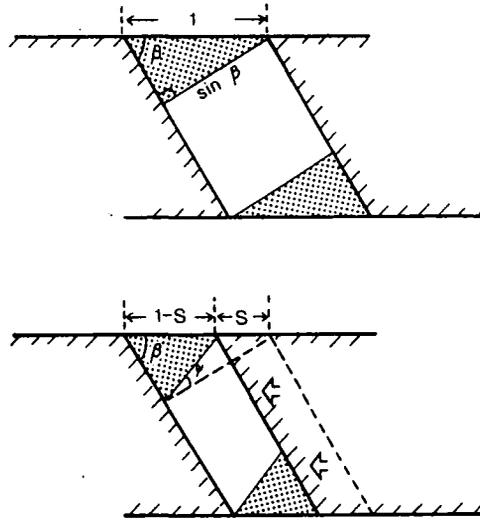


Figure 1.7 (Top) Simple Transpression model involving movement of rigid blocks (diagonal shading), defining S and B. Transpressive deformation occurs in stippled zone.

Figure 1.8 (Bottom) Strain paths for Simple Transpression model. Continuous lines are strain paths for labelled B angles, dashed lines indicate amount of shortening across the zone (α^{-1}), $a=X/Y$, $b=Y/Z$. Both after Sanderson & Marchini (1984).

1.2.3 Kinematic Partitioning of Transpressive Strain

In recent years, studies of actual transpression zones have demonstrated that the deformation, strain and kinematics are typically partitioned (Harland, 1971; Woodcock *et al.* 1988; Holdsworth & Strachan, 1991; Sanderson *et al.* 1991). The modelling of transpressive strain, presented above, assumes deformation of an isotropic medium, and, as natural examples of deformation zones are rarely homogeneous, the effects of pre-existing anisotropy and rheological variations will clearly be profound and must be taken into consideration.

Woodcock *et al.* (1988) define strain or kinematic partitioning as resulting from the resolution of the oblique relative plate motion vector into its two constituent components of displacement, perpendicular to, and parallel to, the plate margin. Thus shortening structures form parallel to the plate margin and strike-slip structures form along the deformation zone. Two types of strain partitioning are recognised and defined by the chronological relationships of the margin parallel, and margin normal displacements. These are: i) Spatial strain partitioning and ii) Temporal strain partitioning and for the purposes of definition will be treated separately, although it is clear that they are end members of a continuum that exists in natural examples of transpression zones.

Spatial Strain Partitioning

'Spatial strain partitioning occurs when the resolved components of the oblique relative motion vector are manifest as synchronous, spatially distinct domains of deformation' (Curtis, 1993).

The modelling of constant incremental and simple transpressive strains described in the previous sections (see 1.1.2) does not consider the synchronous development of zone normal shortening, and zone parallel strike-slip structures. Instead the deformation paths predict a gradational change in the finite strain ellipsoid relative to the increasing shear strain, with lineation switching occurring at specific points for certain predictable deformation paths (Figs 1.2 & 1.3). However, Holdsworth and Strachan (1991) and Strachan *et al.* (1992) provide persuasive evidence for the coeval formation of spatially partitioned strain within NE Greenland, suggesting that the possible presence of a pre-Caledonian shear zone architecture may, in part, be responsible for the deformation style. Harland (1971) was the first to appreciate that pre-existing crustal architecture could cause a kinematic partitioning, noting that '...some preceding fracture pattern or fabric of the lithosphere.' governed the siting of strike-slip movements within a transpression zone. The influence of pre-existing basement structure has also been invoked by Woodcock *et al.* (1988) to explain the structure and kinematics observed across north and west Wales, the suggestion being that even the partitioned component of shortening can be further partitioned into zones of pure and simple shear. An obvious present day example of spatial strain partitioning occurs along the Alpine Fault in New Zealand, where a 45mm a^{-1} oblique convergence vector resolves into 40mm a^{-1} parallel to the fault zone and 22mm a^{-1} normal to it. The Alpine Fault Zone consists of discrete domains of oblique overthrusting linked by strike-slip faults (Norris *et al.* 1990).

Harland (1971) suggested that as shortening progressed in a transpression zone, the layering would be rotated into steeply dipping or sub-vertical zones. Incompetent material in these elongate vertical zones would then effectively act as

zones of weakness, readily accommodating strike-slip movements, and thus the compression itself could generate a means for shear zones to develop. However, although these vertical anisotropies lead to a spatial distribution of strain, their evolution is time dependant, leading to the concept of temporal strain partitioning.

Temporal Strain Partitioning

'Temporal strain partitioning describes the change in deformation style, from one resolved component of the oblique relative motion vector to the other, with respect to time.' — ?

Unlike spatial partitioning, a time dependant change in strain is predicted in the models of Sanderson and Marchini (1984) and McCoss (1986), and depends on the specific deformation path followed, i.e. it depends on the values of γ and α^{-1} (see Fig. 1.2). In these cases the orientation of the X-axis of the finite strain ellipsoid switches from the vertical to the horizontal (Sanderson & Marchini, 1984), or vice-versa (McCoss, 1986) at high values of shear strain (γ). As mentioned earlier, the progressive development of vertical anisotropies (e.g. slaty cleavage) or the rotation of incompetent layers into the vertical by tight folding (Harland, 1971), can result in the introduction of strike-slip component of transpressive strain. However, it is debatable whether the switch in the finite extension direction, or the change in kinematics, may be considered an example of *partitioning*, i.e. the strain is not actually partitioning, it is simply changing from one style to another.

Ellis and Watkinson (1987), working on oblique subduction zones, offer an example of such time dependant strain or kinematic interchanging. They suggest that many collisional zones are characterised by two deformation phases. The first, early deformation develops during progressive metamorphism and is associated with significant orogen parallel extension (stretching lineations, sheath folds, etc.), which may reflect the relative plate motion. The later deformation occurs after the metamorphic peak as the footwall imbricates into the hangingwall during subduction. Stretching lineations suggest that a fundamental kinematic transition occurs, during the imbrication (possibly related to a transition in the buoyancy forces), with later structures no longer directly representing the relative plate convergence. Thus, a switch from orogen parallel, to orogen normal kinematics occurs with time.

It appears that a change in the bulk rheological behaviour of the deforming medium with time is a fundamental aspect of this partitioning/switching. Fritz and Neubauer (1993) describe transpressive deformation in the southern Bohemian Massif, where lower level ductile deformation was accommodated in a large reverse shear zone. With increased exhumation, more brittle behaviour led to a partitioning of strain into discrete fault zones, with plate boundary parallel and plate boundary normal components.

Another possible cause of temporal strain partitioning/switching is a change in the incremental strain, due to either a change in the relative plate motion vector, or to a change in the value of α^{-1} as a result of irregularities along the deformation zone.

1.2.4 Folding in Transpressive Regimes

Initiation and Rotation of Folds

In transcurrent simple shear, the incremental shortening direction lies at 45° to the shear plane (Fig. 1.4) suggesting that folds would be expected to initiate normal to this direction. However documented examples of en-échelon folds commonly form angles of $<45^\circ$ to the fault zone (Moody & Hill, 1956). Although, in a simple shear environment, fold axes rotate towards the shear plane with increasing strain, a high shear strain ($\gamma > 2$) is necessary to reduce the angle from 45° to 22.5° , this implying a shortening of c 60% (Sanderson & Marchini, 1984) which greatly exceeds values reported from natural examples (Moody & Hill, 1956). A transpressive model, however, allows fold initiation at angles $<45^\circ$, Figure 1.9 showing the angle of initiation for different values of incremental strain. Having established the various initiation angles Sanderson and Marchini (1984) then modelled the progressive deformation of folds in a transpressive regime. They showed that although early folds passively rotate and tighten, any newly formed folds should still initiate at a fixed angle to the zone. This results in a pattern of folding in which the early major folds lie at low angles to the zone whilst later minor folds on their shallow dipping limbs or hinge zones lie at higher angles to the zone. The relationship between the major and minor folds of a single generation, and the relationship between folds of different generations may thus be used to determine the shear sense of the transpressive zone (Sanderson & Marchini, 1984).

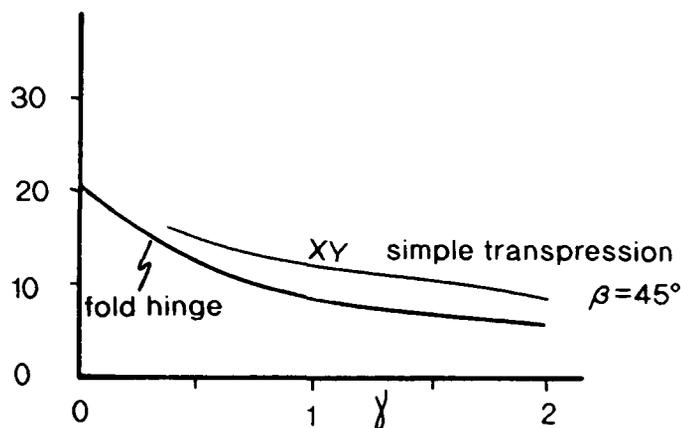


Figure 1.9 Initiation angle of fold axis within a simple transpressive model, assuming that the fold develops as a material line during the first increments of strain, and then rotates passively. After Soper (1986).

Fossen and Tikoff (1993) make some important modifications to the Sanderson and Marchini (1984) model, showing that horizontal material lines (fold axes, lineations, etc.) remain horizontal during transpression/transension, but rotate at different rates and to different points during deformation. A horizontal line with an initial orientation of 90° to the shear direction rotates in a similar way under simple shear and transpression, although horizontal lines with different orientations rotate more slowly under a transpressive regime. Transtension makes all horizontal lines rotate more slowly when compared to simple shear, and they rotate towards a horizontal asymptotic line which makes an angle ϕ with the X-axis (Fig. 1.10). The value of ϕ is dependant on W_k (W_k =kinematic vorticity number; $W_k=0$ for simple shear and $W_k=1$ for pure shear), such that in Fig. 1.10 $\phi=24^\circ$ for $W_k=0.75$ (Fossen & Tikoff, 1993). A more important observation is the behaviour of plunging linear markers during transpression/transension. All inclined lines rotate more slowly during transpression/transension than during simple shear. For transtension the inclined lines rotate towards the point noted above (i.e. at an angle ϕ to the X-axis), whilst during transpression the inclined line rotates away from the horizontal towards the vertical (Fig. 1.10), but only when the X-axis is vertical. If the X-axis is horizontal in transpression, the inclined lines would rotate towards the horizontal. The stereograms in Figure 1.10 show that fold hinges that behave in a passive manner can never become parallel to the shear direction in transtensional shear zones, and only if perfectly horizontal in transpressive shear zones (Fossen & Tikoff, 1993).

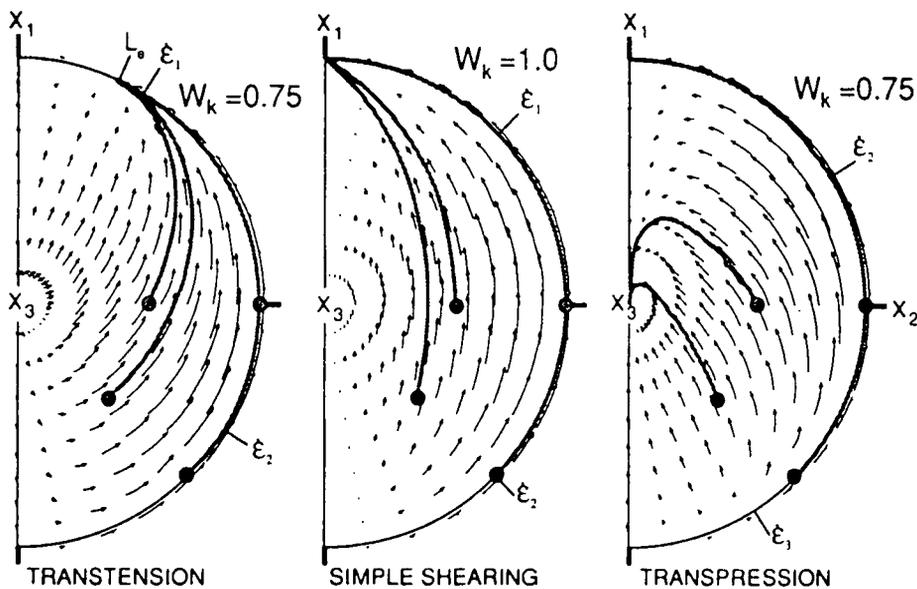


Figure 1.10 Stereographic illustration of the progressive rotation of passive line markers for transpression-transtension ($W_k=0.75$) and simple shearing (wrenching). Arrows indicate the result of simple shear strains of 0.25 combined simultaneously with $k=1.1165456$ (transpression), $k=1.1165456$ (transtension), and $k=1$ (simple shear) to perform the rotations. The lengths of the arrows therefore indicate the rates of rotation in the different fields of the stereograms. After Fossen & Tikoff (1993).

Cleavage Transected Folds in Transpressive Regimes

Transected folds are folds with a contemporaneous, non-axial planar cleavage (Pratt & Fitches, 1993). Powell (1974) defined them as 'folds in which a cleavage surface can be traced from one limb, across the axial plane surface, to the other limb'. They are commonly recognised in the field by the divergence of cleavage-bedding intersection lineations on the opposite limbs of folds (Johnson, 1991). Cylindrical folds with an axial planar cleavage have cleavage-bedding intersections parallel to the fold axis in all parts of the fold. In transected folds, however, both the direction and amount of plunge of the lineations may change across the axial plane (Pratt & Fitches, 1993; Fig. 1.11).

Two angular components of transection were identified by Borrodaile (1978):

1. Δ = the minimum angle between the cleavage and the fold axis (measured in the plane that contains the fold axis and the pole to cleavage).
2. d = the angle between traces of cleavage and the axial plane measured on the fold profile plane.

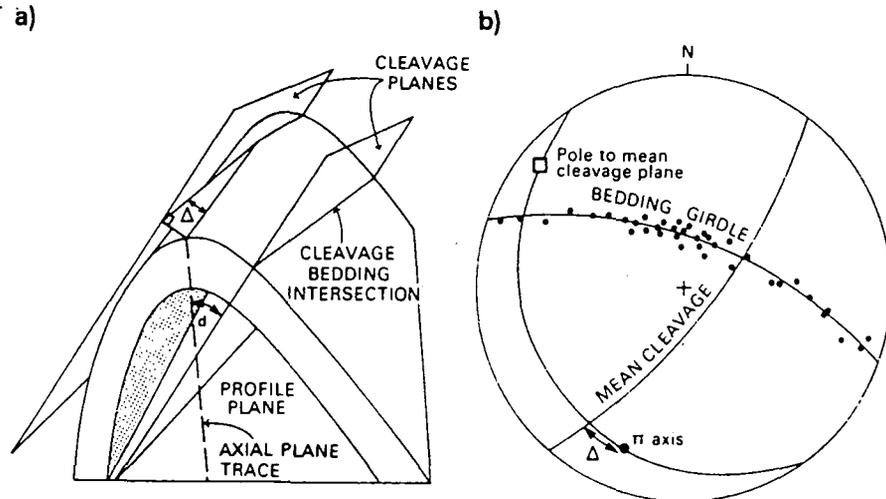


Figure 1.11 (a) Transected fold (an extreme example) showing Δ and d transection angles. The cleavage-bedding intersections plunge in opposite directions on the two fold limbs. (b) stereogram illustrating the Δ angle. After Pratt & Fitches (1993).

In recent years, cleavage transection has become synonymous with transpression, and, in particular, clockwise transection with sinistral transpression in the British Caledonides (Soper & Hutton, 1984; Soper *et al.* 1987; Woodcock *et al.* 1988). Clockwise transected folds are thought to develop during sinistral transpression because most folds initiate before the pressure solution stage of cleavage development (Ramsay, 1967; Borrodaile, 1978; Soper, 1986). This is especially true of rigid layers within ductile envelopes, which will begin to buckle after very little layer parallel shortening and long before cleavage development (Soper, 1986; Pratt & Fitches, 1993). Thus, the relative timing of cleavage and fold development appears critical to

understanding the sense of strike-slip in transpression, as is the behaviour of the fold once initiated.

However, in modelling transpression, neither Harland (1971) nor Sanderson and Marchini (1984) make any association between transpression and folds with transected cleavage. Transpression, like simple shear, does not necessarily provide an immediate theoretical explanation for folds transected by synchronous cleavage (Treagus & Treagus, 1992).

A persuasive case for a link between sinistral transpression and clockwise transection came from Soper (1986) who argued that the domainal anastomosing cleavage common to many lithologies of the British Caledonides, results from a series of pressure solution cleavage increments. He argued that in sinistral transpression, early cleavage surfaces are passively rotated, and that later increments are superimposed in a clockwise sense. This model was used to account for transection angles (Δ) up to 10° . However, Soper (1986), following Sanderson and Marchini (1984), considered fold hinges to be incremental strain markers that become passively rotated with further increments of strain i.e. the hinge does not remain parallel to the finite XY plane during transpression. Treagus and Treagus (1992) and Curtis (1993) noted that a fold would not be visible after the first increments of strain, and only after considerable layer parallel shortening (c50%) and fold development by an active hinge migration mechanism (i.e. the fold hinge tracks the XY plane of the finite strain ellipsoid), does the fold hinge become fixed by considerable curvature. Only thereafter is it passively rotated. Treagus and Treagus (1992) suggest that the fold axis will only lie slightly closer to the zone margin than the XY plane, and thus transpressional deformation of horizontal layers cannot account for folds transected by as much as 10° . Pratt and Fitches (1993), on the other hand, cite examples from the Lower Palaeozoic Welsh Basin and show that line-length balancing indicates that these folds were established after only small values of shortening (5-25%). They suggest that the beds began to buckle into long wavelength folds after only minor layer-parallel shortening (long before the onset of pressure solution cleavage) and subsequently were passively rotated to give large Δ values.

All of the above models assume horizontal beds. Treagus and Treagus (1992) modelled the deformation of inclined, layered systems and were able to demonstrate a variety of transection relationships, controlled by the obliquity of strike to the zone margin. Two cases were considered using sinistral transpression as an example: (a) dipping layers that strike parallel to the zone margin, and (b) dipping layers that strike oblique to the zone margin.

In the first case, the trace of the XY plane (cleavage) forms anticlockwise of the fold axis which develops within the various dipping layers, opposite to the

relationship commonly associated with sinistral transpression. The transection angle (Δ) also increases in proportion to the amount of layer dip (Fig. 1.12a). In the second case, inclined layers striking oblique to the plate margin display a wide range of transection senses and angles (Fig. 1.12b). The orientation of the layering to the XY plane is critical, dividing fields of anticlockwise, axial planar and clockwise transection (Fig. 1.12c). It is therefore theoretically possible to get a complete variation of relationships, although the first case of simple dipping layers is generally more likely to be the normal situation for the closure of a sedimentary basin (Treagus & Treagus, 1992).

It would appear that the theoretical evidence suggests that the sense of cleavage transection should not be used to exclusively infer the sense of shear in transpression zones, and that additional information e.g. supporting kinematic data, may also be necessary. That said, the consistency of the sense of cleavage transection in the British Caledonides is often cited as compelling evidence for its use in determining the sense of shear in transpression zones (Soper, 1986; Treagus & Treagus, 1992).

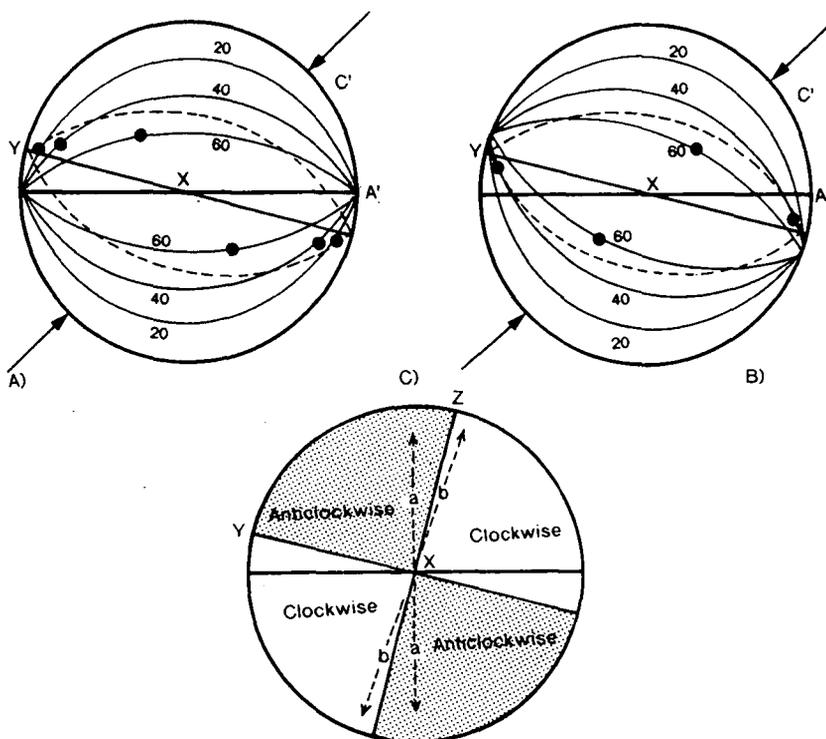


Figure 1.12 The effect of inclined and oblique layering on cleavage transection, with respect to a sinistral transpressional deformation. Transpression vector is bold arrow, transpression zone margin (a) east-west. Solid line is XY plane, broken curves circular sections. (a) planes striking parallel to zone margin, dips 20°, 40° and 60°. Fold axes (solid dots) are cut anticlockwise by the XY plane. (b) Planes striking 20° to the zone margin, otherwise as (a). Fold axes are transected clockwise. (c) Fields for which bedding poles would have anticlockwise transection of fold axes by XY plane (ACL, shaded), and clockwise (CL, blank). Positions for examples in (a) and (b) indicated by broken lines, labelled a, b. After Treagus & Treagus (1992).

1.2.5 Transpression at Fault and Shear Zone Bends

In wrench-dominated tectonics, departures from parallel-sided deformation zones, such as offsets or changes in zone width, will introduce added complexity to the fault zone. Offsets are fundamental features of strike-slip faults, with the offset sense determining whether zones are compressive ($\alpha^{-1} < 1$; Fig. 1.13a) or dilational ($\alpha^{-1} > 1$; Fig. 1.13b). If the sense of offset is the same as the sense of fault slip (i.e. right offset along dextral fault) then crustal extension, subsidence and the formation of a pull-apart basin occurs; i.e. transtension. Where the offset sense is opposed to the shear sense pressure ridges or push-ups occur; i.e. transpression (Sanderson & Marchini, 1984; Aydin & Nur, 1985).

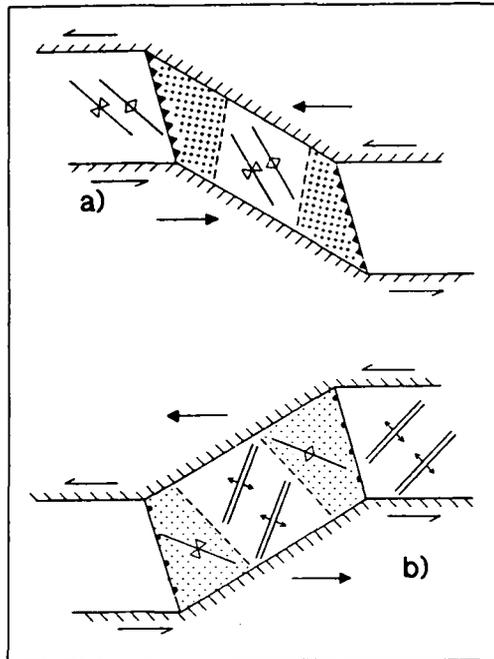


Figure 1.13 Upper diagram shows generation of Simple Transpression (a) in a region of offset of a shear zone (i.e. simple shear). The upper rigid block (diagonal shading) moves left relative to the lower one, generating the area of transpression (dots and crosses). The unshaded parts of the zone undergo simple shear. Orientations of folds inside and outside of the transpression region are shown. Lower diagram shows Transtension (b) at a dilational offset. After Sanderson & Marchini (1984).

Dilational Offsets

The formation of pull-apart basins (Fig. 1.13b) at fault releasing bends or dilational offsets has been much studied and general patterns of geometry, structure and sedimentation are recognised, e.g. Mann *et al.* (1983). Of particular relevance to this thesis is the style of basin floor and the sedimentation of pull-aparts.

Pull-apart basins may be divided into two groups on the basis of their style of floor. One group form true rifts that extend at depth to hot rocks of the upper mantle, similar to those found above an oceanic spreading ridge. These basins are formed by extreme attenuation of the upper crust during extension, which is followed by the forceful upwelling of hot mantle. These basins lack true basement, with the older

sediments lying above a volcanic sill and dyke complex, which intrudes the older sediments (Kerr & Kidwell, 1991).

The second group floor along a detachment or décollement surface and may be further subdivided into those that detach against flat lying tectonic surfaces, and those that bottom out unconformably along older basement (Cheadle *et al.* 1986).

In terms of sedimentation, many basins are typified by high sedimentation rates, rapid facies changes, abrupt thickening of sedimentary sequences over short distances, numerous unconformities reflecting syntectonic sedimentation, and the presence of locally derived skewed fan bodies of fault margin breccia facies (Hempton, 1983; Hempton & Dunne, 1984). The most distinct stratigraphic feature of pull-apart basins is the extreme thickness of onlapping sedimentary sequences relative to their area, which is thought to reflect the migration of the depocentre by means of syndepositional strike-slip faulting (Crowell, 1982).

Compressive Offsets

At compressive offsets, transpression produces folds at high angles to the overall zone and the compression is accommodated by crustal thickening. Boundaries between the transpressive region and the rest of the shear zone develop as high angle reverse, or oblique-slip faults (Sanderson & Marchini, 1984). These uplifted regions are areas of intense strain and are commonly manifested as flower structures (Wilcox *et al.* 1973, Fig. 1.14). Such structures, typified by the Mecca Hills along the San Andreas fault (Sylvester, 1991), usually display spatial strain partitioning, and commonly comprise a central vertical zone which rises upward and outward into a series of convex-up, oblique thrust faults.

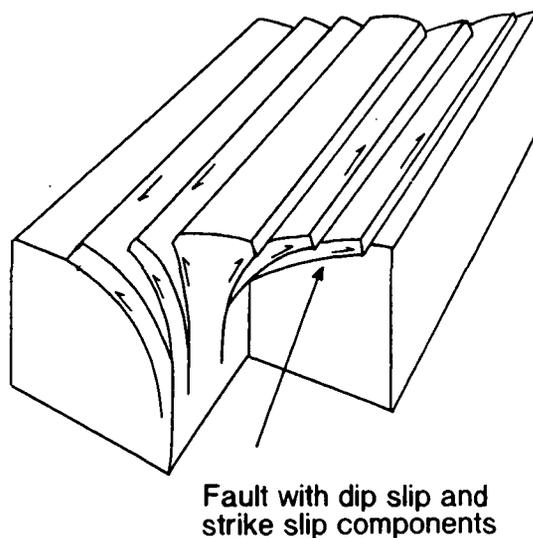


Figure 1.14 Flower or palm tree structure with sideways oblique overthrust faults developed along a left-handed transpressional strike-slip fault. After Ramsay & Huber (1987).

1.2.6 Summary of the Characteristics of Transpression Zones

Transpression is a three-dimensional or triaxial deformation that can be resolved into a combination of pure shear and simple shear, and provides a model with which to interpret structural features within a fault-bounded zone of deformation (Sanderson & Marchini, 1984). Crustal shortening ($\alpha > 1$) and wrench type shear produce:

- a) a flattening or oblate strain ($k < 1$).
- b) steep cleavages and a stretching lineation which may be either vertical or horizontal.
- c) compressional structures e.g. folds and thrusts at small oblique angles to the zone.
- d) extensional structures e.g. normal faults, dykes and veins at high angles to the zone.
- e) crustal thickening and vertical uplift.

1.3 Complex Fold Geometries

1.3.1 Fold Interference Patterns

Polyphase deformation is characterised by more than one generation of folds, often forming in different orientations. These structures will successively overprint older fold sets producing characteristic interference patterns in outcrop. Ramsay and Huber (1987), following Ramsay (1967), recognise four basic end members of a continuous series of fold interference patterns for two phases of cylindrical folding, F1 and F2.

i) Type 0: Termed "zero" type by Ramsay and Huber (1987), this pattern arises when both the F1 and F2 axes and axial planes are parallel to one another (Fig. 1.15a). In this case the F1 axial plane remains unfolded. However this style of interference produces none of the geological features that are generally characteristic of fold superposition, and the resulting three dimensional geometry is practically identical to that of fold structures produced during a single phase of deformation.

ii) Type 1: The "egg-box" or "dome and basin" pattern arises when both F1 and F2 fold axes and axial planes are orthogonal, or at high angles to one another (Fig. 1.15b). The F1 axial plane remains unfolded. Ramsay and Huber (1987) note that in zones of high finite strain, "sheath" or "eyed" folds may resemble the basin and dome pattern, but suggest that two phase patterns may be distinguished as "dome and basin" patterns show a high degree of geometric organisation, and usually two orthogonal fabrics.

iii) Type 2: The "mushroom" or "angel-wing" pattern occurs when F2 fold axes lie at high angles to the F1 axes within the F1 axial surface, and thus, the F1 and F2 axes are not coaxial (Fig. 1.15c). In this case the F1 axial planes are folded. This pattern differs from the completely closed forms of Type 1 in that some of the fold closures are attached by stalks.

iv) **Type 3:** The "hook" pattern is due to coaxial refolding where the F1 and F2 axial planes are at a high angle to each other (Fig. 1.15d). The F1 fold axial planes are refolded. The limbs of the first folds which originally converged together or diverged away from the first fold axial surfaces, will continue to do so after refolding. Closed types of patterns, described above, do not generally occur with this style of refold.

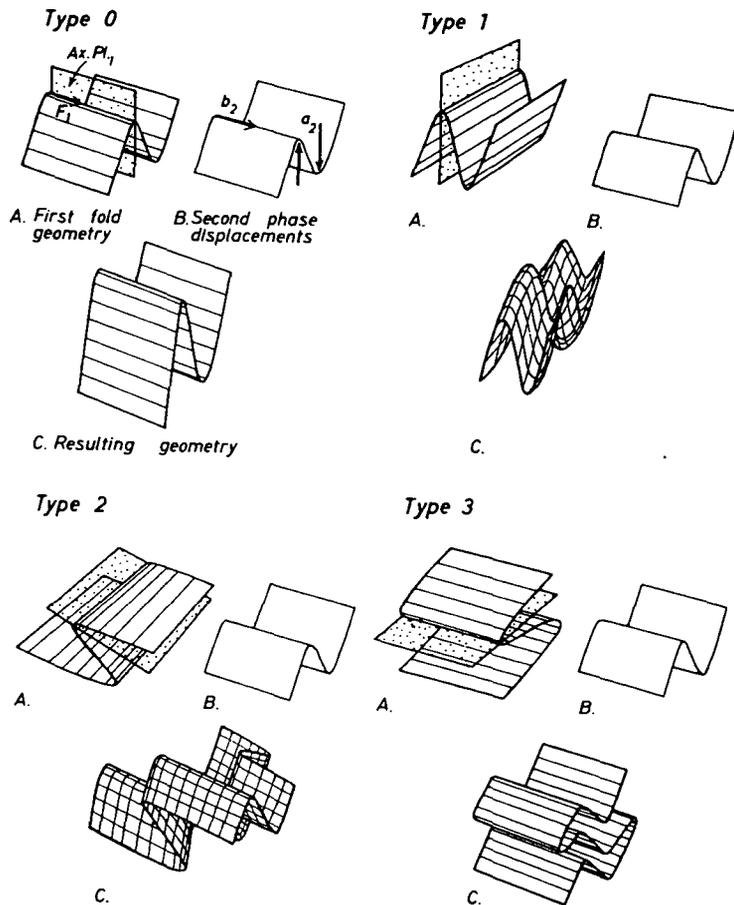


Figure 1.15 The four principal types of three dimensional fold forms arising by the superposition of shear folds on pre-existing fold forms. After Ramsay & Huber (1987).

1.3.2 Curvilinear Fold Axes

All natural fold axes are curvilinear to some degree and this may arise as a consequence of the mode of fold initiation (Dubey & Cobbold, 1977), and/or be due to fold initiation in inclined layers (Treagus & Treagus, 1981). Dubey and Cobbold (1977), using empirical modelling of flexural slip folds, showed that non-cylindrical folds result from: a) folds initiating at non-cylindrical deflections; b) hinge lengthening being slower than fold amplification, and c) fold interference. The first two factors generally produce periclinal folds, whilst the third factor produces folds with significantly warped axial surfaces and markedly curved hinges. Treagus and Treagus

(1981) showed that oblique layers in plane strain can also give rise to folds with axes which are oblique to the XY plane and generally develop *en-échelon* periclinal folds. With increasing fold axis obliquity a considerable variation in fold axis azimuth and plunge can occur.

Sheath Folds

In recent years, several terms have been coined to describe highly non-cylindrical folds whose axes are curvilinear within their axial plane. Ramsay (1962a) provides the first description, calling these structures "eyed" folds, whilst Dearman (1969), working on highly deformed rocks of the Lizard thrust zone, preferred the term "tergiversate" fold (Lat. *tergum*-the back, *vers/vertere*-to turn). The expression 'sheath fold' was introduced by Carreras *et al.* (1977), and subsequently firmly established in the literature by Cobbold and Quinquis (1980), who presented three models for sheath fold development based on empirical results. However, sheath folds remained loosely defined and not clearly distinguished from other non-cylindrical folds until Ramsay and Huber (1987) defined these structures as folds whose hinge line variation was greater than 90°.

In many areas, there appears to be a connection between shear zones (e.g. in nappe complexes) and the occurrence of sheath folds (Ramsay & Huber, 1987). Many authors relate these phenomena and suggest that the mechanism of sheath fold formation is related to simple shear deformation. However, these structures may also arise in other high strain, deformation regimes such as pure shear, either plane strain or constrictional, or three dimensional differential (i.e. non-planar) flow (Skjerna, 1989).

The mechanism which has gained broad acceptance over the last few years is Model 1 of Cobbold and Quinquis (1980), with modifications by Ramsay (1980). This involves subjecting a multilayer to homogeneous simple shear, with the shearing plane parallel to the sheet dip of the layering. This results in any initially symmetrical 'deflections' becoming strongly asymmetric, the sense of asymmetry reflecting the overall shear sense. Ramsay (1980) suggested that these 'initial irregularities' were syn-shear, derived slightly non-cylindrical folds (i.e. showing slight variations in their axial plunges), which had their main axial direction at right angles to the shear direction. With increasing strain the hinge line becomes more strongly curvilinear within its axial plane and the fold, therefore, becomes more non-cylindrical (Fig. 1.16). The final fold shape depends on both the initial 'deflection' geometry and the total magnitude of strain, and is thus, by definition, independent of the deformation history. At high shear strains ($\gamma > 10$) most 'deflections' of geological realistic proportions become sheath-like (Cobbold & Quinquis, 1980).

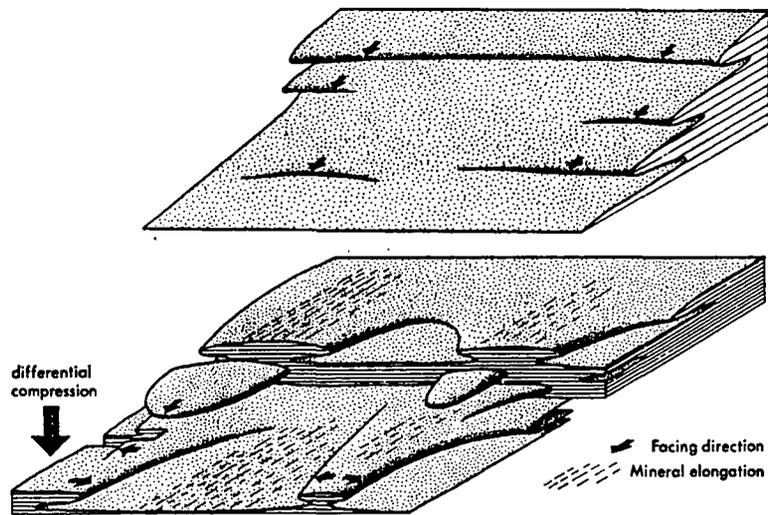


Figure 1.16 A multilayer undergoing homogeneous simple shear resulting in initially symmetrical deflections becoming highly asymmetric. Fold hinge lines become markedly curvilinear within the axial surface to give sheath folds. Note (a) the double vergence closures in the shaded face and (b) the parallelism between the mineral extension lineation and the sheath fold X-axis. After Dearman (1969).

1.4 Shear Criteria

1.4.1 Ductile Shear Criteria

The simplest model of a shear zone is that of a highly strained rock between planar and parallel sided boundaries, outside which the rock remains unaffected by the shear deformation (Ramsay & Graham, 1970). When the boundary of a shear zone can be observed, the geometry of structures formed by the strain gradient can almost always be used to unambiguously define the shear sense e.g. changes in finite strain state, sigmoidal form of schistosity, orientation of folded and boudinaged competent layers (e.g. dykes, etc.), fracture openings and the geometry of *én-echelon* vein arrays (Ramsay & Huber, 1987). However, shear zones can be very wide, and it may be impossible to observe the zone boundary and offset marker horizons may be scarce or absent. In these cases, the small scale structures may be used to deduce the displacement sense (Simpson & Schmidt, 1983). These are as follows:

S-C band structure

The term S-C fabric was introduced by Berthé *et al.* (1979), who described the evolution of mylonites in an orthogneiss deformed in the S. Armorican Shear Zone. They recognised two types of foliation in the gneiss: S-surfaces related to the accumulation of finite strain (i.e. cleavage surfaces parallel to the XY plane), and C-surfaces related to localised high shear strain (C stands for *cisaillement* or shear; Fig. 1.17A). The C-plane, in the sense of Berthé *et al.* (1979), lies parallel to the shear zone

boundary. Subsequent work has shown that the shear planes may develop at a small angle to the shear zone boundary, these being termed shear bands (Simpson & Schmidt, 1983) or extensional crenulation cleavage (Platt & Vissers, 1980). Simpson and Schmidt (1983) also recognised much less common, but distinct antithetic shears (Fig. 1.18).

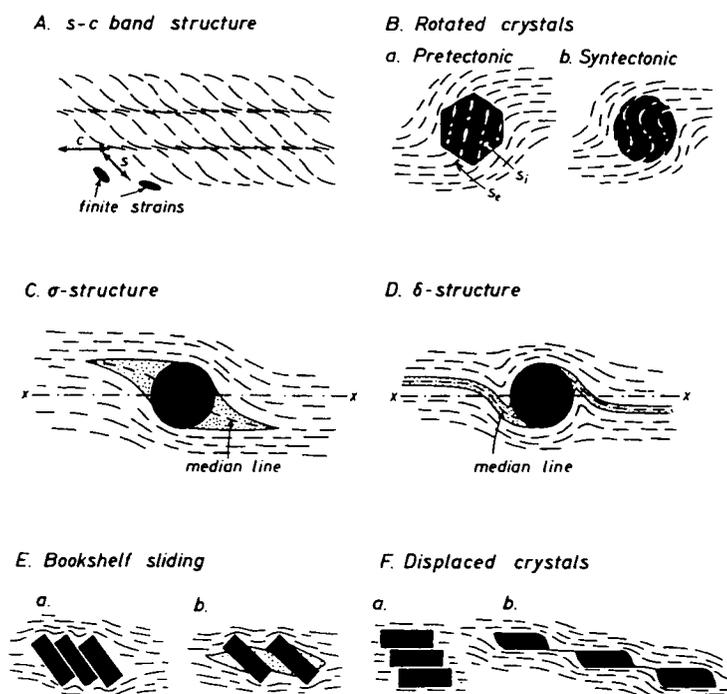


Figure 1.17 Criteria used to determine shear sense in shear zones. All have sinistral shear sense. After Ramsay & Huber (1987).

Berthé *et al.* (1979) suggest that an important characteristic of S-C fabrics is that the S- and C- surfaces develop simultaneously. The question of timing was addressed by Platt and Vissers (1980), who showed that, in general, it is difficult to specify exact timing relationships between S- and C- surfaces, except possibly where a substantial metamorphic change is demonstrable between the periods of S- and C- surface formation. Lister and Snoke (1984) defined two types of S-C fabrics: Type I and Type II. Type I S-C fabrics are essentially those described by Berthé *et al.* (1979) from deformed granitoids, and involve narrow zones of intense shear strain which cut across the (mylonitic) foliation. Lister and Snoke (1984) suggest a variety of scenarios for Type I S-C fabric development, ranging from synchronous foliation and shear banding, through to a pre-existing fabric being transected by considerably later, narrow shear bands.

The second type of S-C fabric (Type II) recognised by Lister and Snoke (1984) occurs predominantly in quartz-mica rocks involved in zones of intense non-

structures are characteristic of shear zones where the recrystallization rates are higher than the rotation rates (Ramsay & Huber, 1987).

Passchier and Simpson (1986) further subdivide these structures into two types:

σ_a -type: In this case the porphyroclast lies isolated in a homogeneous matrix. If the recrystallised tails are long, stair-stepping is easily detected and the tails trend sub-parallel to the main foliation. In this respect the fabric is similar to the Type II S-C mylonite fabrics described by Lister and Snoke (1984), and their mica 'fish' may be regarded as a special type of σ_a -type porphyroclast (Passchier & Simpson, 1986; Fig. 1.19a)

σ_b -type: In mylonites with a heterogeneous fabric due to shear band foliation, σ_b -type porphyroclasts are associated with C-planes (Berthé *et al.* 1979), i.e. narrow zones of high shear strain oblique to the main foliation (S-planes). The C-planes tend to enclose large porphyroclasts and dynamically recrystallised material from the porphyroclast grain margin is deflected along them (Passchier & Simpson, 1986; Fig. 1.19b).

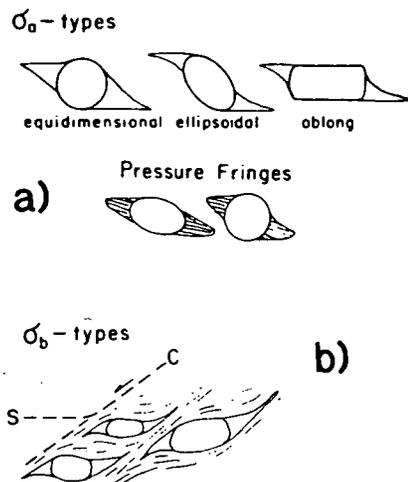


Figure 1.19 Classification of porphyroclasts. (a) σ_a -Type porphyroclast lies isolated in a homogeneous matrix and may be equidimensional, ellipsoidal or oblong. (b) σ_b -Type porphyroclast occurs in mylonites with a heterogeneous fabric (S-C mylonites) and is associated with the C-planes. All have sinistral shear sense. After Passchier & Simpson (1986).

δ -structure

This structure is very similar to the σ -structure, but differs in that the recrystallised pressure shadow tail is strongly rotated by the porphyroclast (Fig. 1.17D). The term δ -structure describes the curved, and often embayed nature, of the generally narrow tail. In this case the median line crosses the trend of the schistosity. This type of structure is particularly characteristic of regions of high shear strains where the recrystallisation rates are lower than the rotation rates (Passchier & Simpson, 1986).

Bookshelf sliding

This phenomenon occurs where individual crystals fracture during shear, with the individual parts being rotated in the direction of shear (Fig. 1.17E:a). This results in a contrary sense of shear being set up between the individual fragments. With increasing shear the fragments may separate and recrystallised pressure shadows may develop between them (Ramsay & Huber, 1987; Fig. 1.17E:b).

Displaced crystals

This only occurs in crystals with a well developed crystal cleavage. If these are orientated with this cleavage sub-parallel to the shear plane, internal detachment and gliding along these surfaces occurs (Fig. 1.17F:a). In high strain zones it is not uncommon to find individual crystal fragments connected by zonal films of phyllosilicates (Fig. 1.17F:b).

1.4.2 Brittle Shear Criteria

Determining the sense of fault movement from fault surfaces is essential in the study of individual fault kinematics, especially when the sense of slip cannot be determined directly from offset geological structures. Study of the minor structures (sense of movement criteria) associated with slickensides and slickenlines, the presence of secondary structures associated with faults, and the adjacent wallrock deformation, can all be used to infer the relative sense of movement on brittle faults.

Slickenline striations

The morphology and cause of lineations on a slickenside fault surface may be varied, and is usually dependant on lithology. Figure 1.20 illustrates several types of slickenside lineations, and their sense of movement interpretation.

i) Asperity ploughing (Means, 1987)

This is represented by furrows, grooves or scratches on slickenside surfaces, which are produced by resistant protuberances on one side of a pair of surfaces moving relative to one another, scoring a groove in the opposite block. The grooves terminate in the final ploughing element position, the end of which points toward the movement of the missing fault-wall. Excavation of the groove may be accomplished by brittle or ductile, i.e. pressure solution, processes (Fig. 1.20a).

ii) Crystallization linked to steps along the fault surface

These steps generally form perpendicular to the striation on the fault surface, the risers to which face the direction of movement of the missing block (Fig. 1.20b). These accretion steps (Petit, 1987) produce potential voids that can either be filled by fibrous crystals of calcite (in the case of limestones) during slow dilation of the voids (Durney & Ramsay, 1973), or blocky euhedral calcite precipitation in rapidly opened voids, i.e. microseismic events. Individual accretion steps can display a combination of these crystal growth styles, indicating a change between aseismic and seismic motion.

iii) Erosional sheltering

When a surface containing hard asperities becomes eroded by fault movement, elongate ridges are preserved behind the hard particle parallel to the slip direction. The leading face of this hard particle may display minor pressure solution effects, especially within carbonate rocks (Fig. 1.20c).

iv) Slickolites (Bretz, 1940)

Slickolites are similar to stylolites in being dissolution surfaces. However, unlike stylolites, the spikes and columns of slickolites point subparallel to the slip surface, and parallel to the direction of movement. The solution spikes and columns observed on a slickolite surface point in the relative motion direction of the block containing them (Fig. 1.20d).

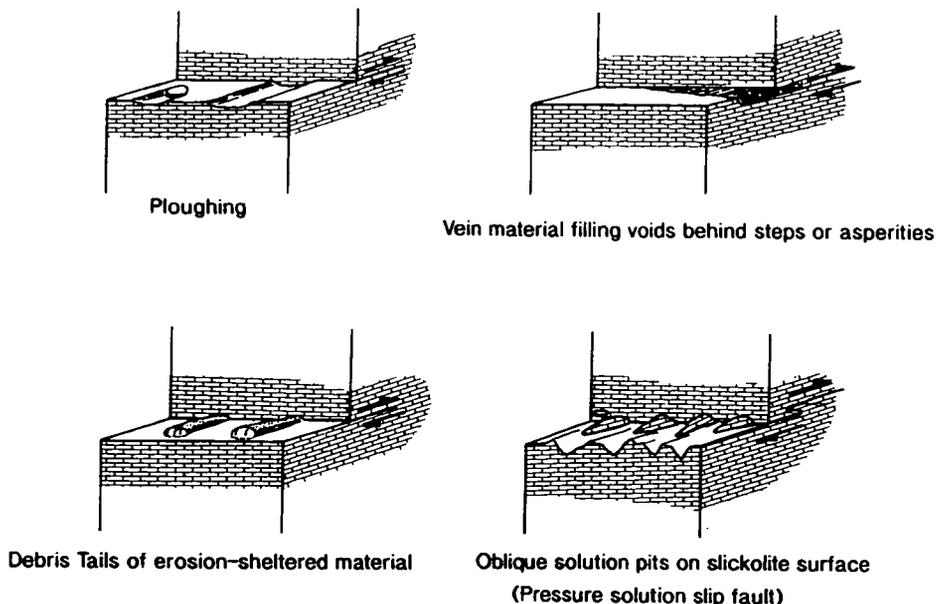


Figure 1.20 Types of slip parallel movement indicators found on slickensides. (a) Asperity ploughing. (b) Accretion steps and fibres. (c) Tails or ridges of erosion sheltered material on the down-slip side of hard asperities. (d) Oblique spikes on slickolite surfaces. Modified after Means (1987).

Sense of movement structures involving secondary fractures

An important feature commonly observed along fault planes is the presence of secondary repeated fractures of the same type, that possess a regular angular relationship with the mean fault plane. These structures were reviewed by Petit (1987), whose nomenclature is employed unless otherwise stated. This terminology is based mainly on Riedel-type experimental nomenclature, to describe the geometrical position of the fractures, but does not imply that the fractures can be explained mechanically by Riedel experiments.

These secondary fractures can be divided into three groups based on fracture type (striated or non-striated), and orientation: a) *group T* includes repetitive tension fractures and no shear fractures; b) *group R* all synthetic shear fractures of R orientation, and c) *group P* all types of secondary shears of P orientation. Within groups R and P morphological differences can be distinguished by use of a suffix. If only R or P secondary shear fractures are present, the suffix is O; if the main or mean fault plane is completely striated, M is used; and if non-striated secondary fractures are present then T is used (Fig. 1.21a-c).

i) T criteria

The characteristic of T criteria movement sense indicators is that the mean fault plane is fully striated, and intersected by fractures that are dominantly perpendicular to the striations, which are themselves non-striated. These 'tensile' fractures make an angle of 30° to 90° to the main fault plane, can be open or filled, and their intersections with the main fault either planar or curved. In the latter case, the fractures tend to be perpendicular to the main fault, forming a crescent shape, the 'horns' of which point in the direction of movement of the missing block (Fig. 1.21a). These directly correspond to 'crescentic fractures' associated with glacier striations.

ii) R criteria

R criteria indicators are characterised by a mean fault plane which is joined by repeated secondary striated fractures, which describe a small angle to the fault wall, and have a Riedel shear orientation. Again, their intersection with the main fault is virtually perpendicular to the slip direction (Fig. 1.21b).

a) RO (R only) type: There is no mean striated fault plane. The R shears are very closely and regularly spaced, displaying a slight striation. The fault displays a serrated profile, due to the intersection of R and R' secondary shears. This type is uncommon.

b) RM type: The main fault plane is fully striated, with irregularly distributed R shears of various size intersecting the fault surface.

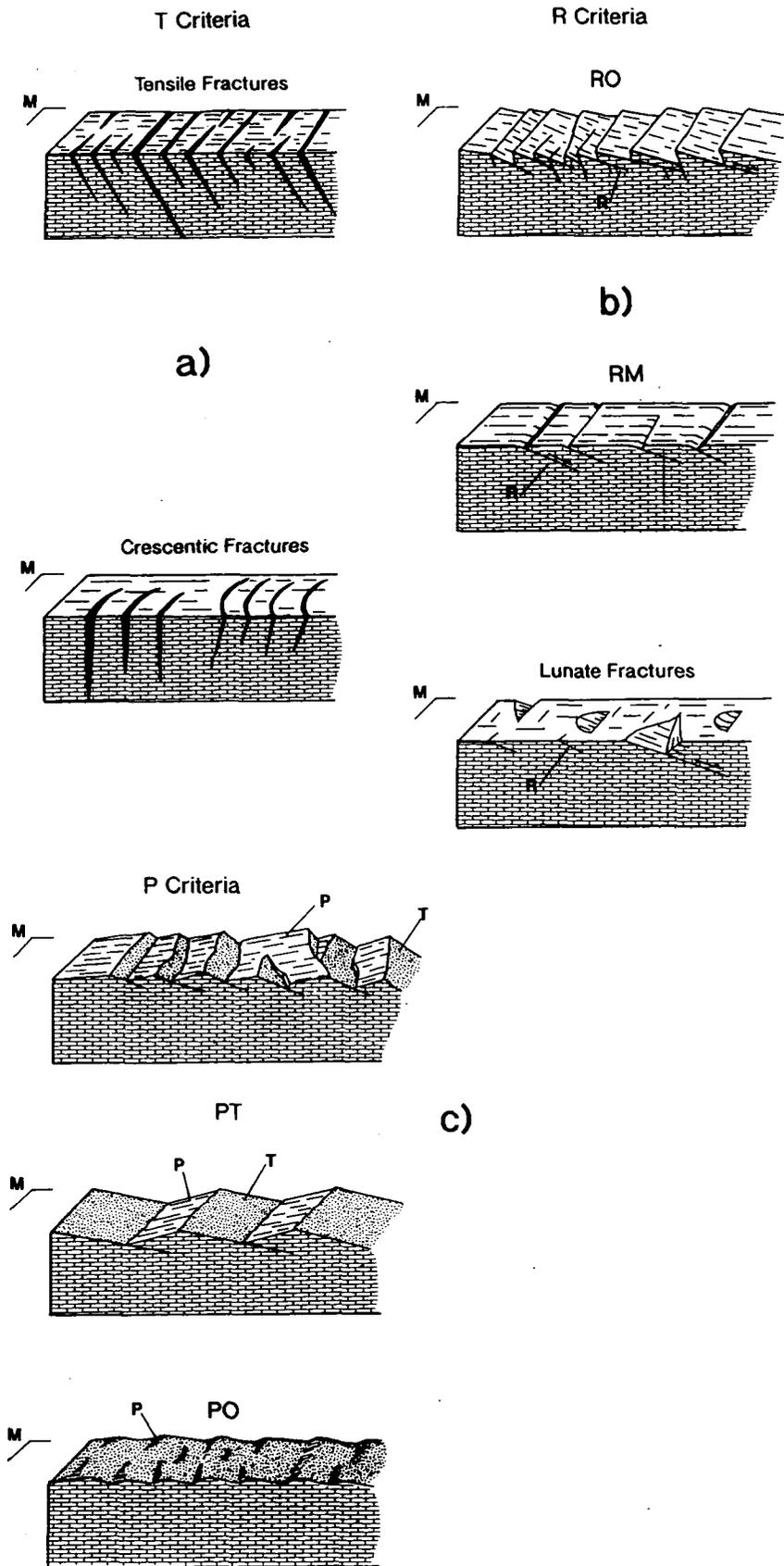


Figure 1.21 (a)-(c) Main types of criteria for sense of motion determination based on repetitive secondary fractures. (a) T-criteria. (b) R-criteria. (c) P-criteria. See text for details. Modified after Petit (1987).

iii) P criteria:

Characteristically, the fault plane is always incompletely striated, but where developed, it appears on the side of asperites facing the movement of the missing block (Fig. 1.21c). The formation of this structure implies very small slips, as otherwise the striation would have become more extensive and could have led to a fully striated slickenside. Again, this category can be split into two types:

a) PT type: Planar non-striated surfaces clearly dip into the fault-wall at a small angle, whereas the P shears are well developed and strongly striated. Some of the more strongly striated P surfaces show shallow steps, which always ascend in the direction of the missing block.

b) PO (P only) type: The non-striated (protected) surfaces of the asperites (lee side) do not project into the fault-wall. The P shears display only a weak striation.

1.5 Vergence

The concept of vergence was introduced by Stille (1924) and has been extensively used since, although its precise meaning has been much debated (e.g. Roberts, 1974; Bell, 1981). Stille (1924) used the word "vergenz" to describe the directional sense of overturning of minor folds and therefore implies that the folds have been transported. Bell (1981), in keeping with modern usage, defines vergence only in terms of geometrical relationships with no implication of movement. Two types of vergence are recognised:

1.5.1 Fold vergence

The principal use of minor fold vergence is to locate major fold axes. Geometrically, asymmetric minor folds can be considered to have short limbs which have been rotated from a position now preserved by the longer limbs (Bell, 1981). Vergence of asymmetric folds was defined by Roberts (1974) as 'the horizontal direction, within the plane of the fold profile, towards which the upper component of...rotation is directed' (Fig. 1.22a). Although the use of an azimuth to define the fold vergence has advantages over the use 'S' (sinistral) or 'Z' (dextral) prefixes, since it is independent of fold plunge variations, the idea of sinistral and dextral vergence is still applied to vertically plunging folds (Bell, 1981: Fig. 1.22b).

1.5.2 Cleavage vergence

Roberts (1974) noted that vergence 'can be extended to describe bedding-cleavage relationships where bedding is not folded' but failed to define this concept. Bell (1981)

defined cleavage vergence as the horizontal direction, within the plane normal to the fabric intersection lineation, towards which a younger fabric needs to be rotated (through the acute angle) so that it becomes parallel to the older fabric (Fig. 1.22c). The earlier fabric may commonly be bedding, but if the vergence of a later cleavage is being considered, it must be related to the immediately preceding fabric. Cleavage vergence, like fold vergence, indicates the direction in which an antiform may be encountered (Bell, 1981).

The concepts of fold and cleavage vergence are distinct from, and should not be confused with, the concepts of fold and cleavage facing (as illustrated in Figs 1.23a & 1.23b).

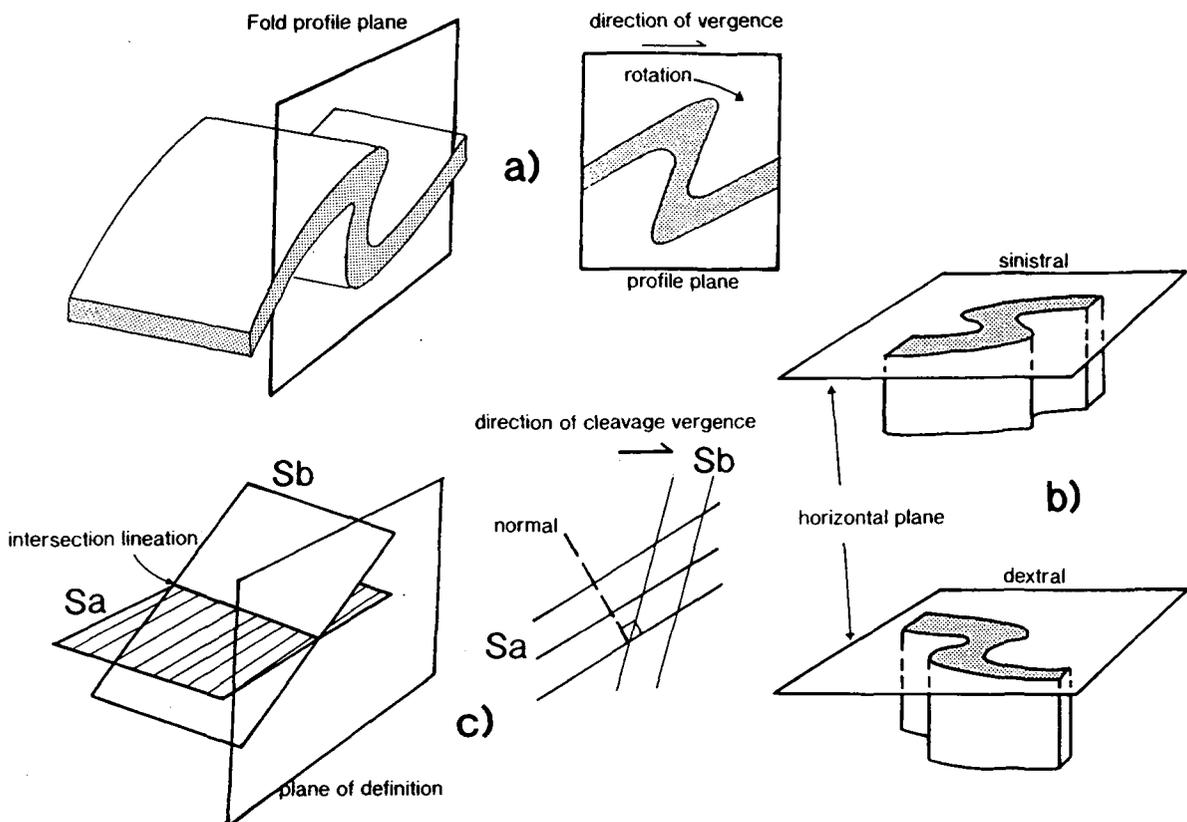


Figure 1.22 (a)-(c) Fold and cleavage vergence. (a) Fold vergence defined by the direction of the upper component of rotation within the fold profile plane. (b) Asymmetry of vertically plunging folds, which may have sinistral (top) or dextral (bottom) vergence. (c) Cleavage vergence defined by the direction of rotation of the younger fabric into parallelism with the older (Sa-older, Sb-younger). Modified after Bell (1981).

1.6 Facing

The concept of facing provides an accurate means of relating tectonic structure to the polarity of stratigraphic successions (Holdsworth, 1988). The term was first used by Shackleton (1957) in his description of folds and cleavages from the Highland Border. He defined fold facing (Fig 1.23a) as follows: 'A fold faces in a direction normal to its

axis, along the axial plane, and towards the younger beds. This coincides with the direction towards which the beds face at the hinge'. He also defined cleavage facing as the direction within the cleavage plane, normal to the bedding/cleavage intersection, towards the younger beds. The concept of facing was expanded to include faults by Lisle (1985), who defined fault facing as the direction, normal to the bedding plane intersection, along the fault plane and towards the younger beds.

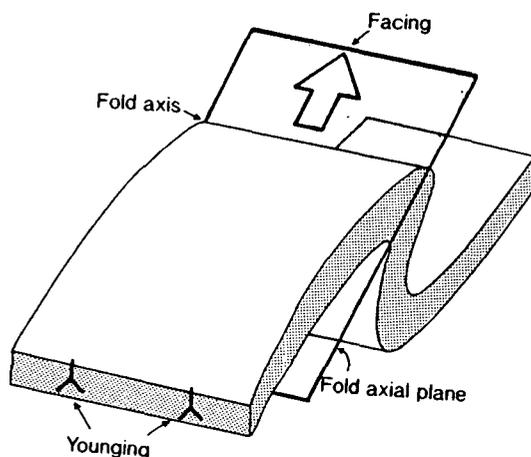


Figure 1.23 Fold facing defined as the direction of younging of the beds, normal to the bedding/cleavage intersection, within the axial plane.

Holdsworth (1988) noted that facing, in common with all linear fabrics, possesses both an inclination and an azimuth, but, as it is a polar rather than an axial lineation, a horizontal reference plane is used and the terms upward-, downward- and neutral-facing, are employed to qualify the facing azimuth. Holdsworth (1988) provided a simple construction for the stereographic analysis of facing, in which a reference plane (cleavage, fault or axial plane) is plotted on a stereonet along with the fold axis or bedding intersection (Fig. 1.24a). The line containing the facing direction lies within the reference plane at 90° to the fold axis or bedding intersection. The approximate field facing i.e. upward, downward etc., is then applied to the stereonet. Downward-facing directions are recorded by the lower hemisphere intersection, which is read off as a plunge (Fig. 1.24a). Upward-facing directions are upper hemisphere intersections, which have an equal and opposite azimuth and inclination to that of the line of facing (Fig. 1.24b). Neutral facing directions have an azimuth, but no inclination, and it is easy to decide which intersection of the line of facing on the primitive is appropriate from field observations (Fig. 1.24c).

The above technique can be applied to a wide variety of geological problems where facing is an important factor e.g. areas of refolding and areas of sheath folding (Holdsworth, 1988).

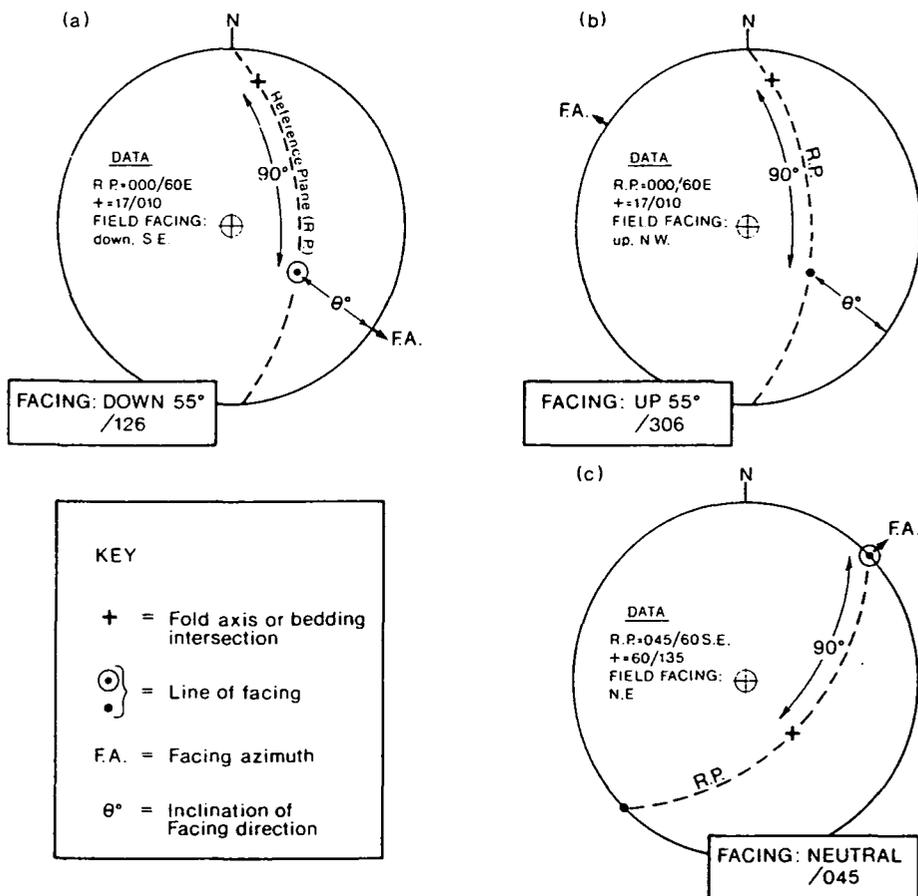


Figure 1.24 Stereographic projections to determine facing. The line of facing lies within the reference plane at 90° to the fold axis or bedding intersection. Solid circle and circle with dot symbol denote upward- and downward-facing lines respectively. An arrow outside the primitive shows the facing azimuth. After Holdsworth (1988).

1.6.1 Areas of refolding

Bell (1981) details the facing variations that occur in two different examples of refolding: coaxial and orthogonal. In coaxial refolding (Fig. 1.25a) F1 folds (major and minor) change their facing across F2 traces. In the absence of F2 minor folds, facing changes alone may be used, in this case, to define F2 axial traces. A significant feature of orthogonal refolding (Fig. 1.25b) is that F1 folds do not change their facing around F2 axial traces. This highlights the fact that vergence changes alone do not supply sufficient information to locate fold axes consistently, with both vergence and facing having to be taken into account.

1.6.2 Areas of sheath folding

As discussed above (section 1.3), sheath folds arise in areas of high strain, due to the modification of folds with initially curvilinear axes. It is usual to define the original facing direction of a sheath fold as the direction of younging along the extensional lineation (i.e. parallel to mineral lineation in Fig. 1.16). Clearly, those sheath folds

which develop in vertical shear zones will display bimodal facing, with c. 50% of the folds possessing a *primary* downward facing. Thus, in high strain areas, care must be used in interpreting fold geometry, vergence and facing.

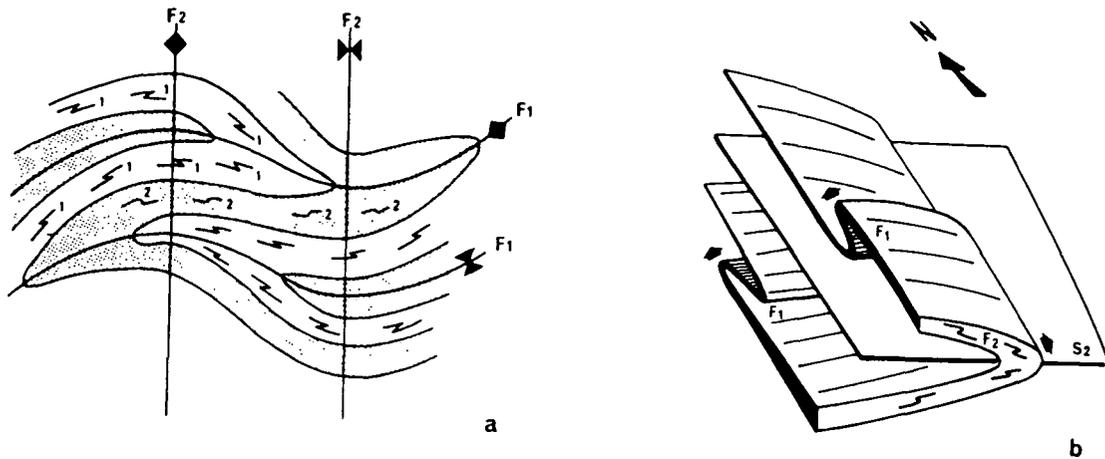


Figure 1.25 (a) Coaxial refolding. Minor F_1 folds do not change their vergence across F_2 axes, but do change their facing direction. (b) Orthogonal refolding. Both F_1 and F_2 minor folds change their vergence, but not their facing, across F_2 fold axes. Arrows indicate facing direction. After Bell(1981).

1.7 Aims of Thesis

This project aims to detail the structure, sedimentology and geochemistry of two along-strike areas within S. Devon and Cornwall, which are characterised by regionally anomalous structure and which mark the site of significant sedimentological, geochemical and metamorphic transitions. The Start Complex, which was entirely remapped in the course of this project, is described and discussed in Chapters 3 (sedimentology, petrology and geochemistry) and 4 (structure and regional setting). Its proposed along-strike equivalent, which straddles the northern margin of the Gramscatho Basin in S. Cornwall, is described and discussed in Chapter 5.

CHAPTER 2

THE STRATIGRAPHIC AND STRUCTURAL EVOLUTION OF S.W. ENGLAND

2.1 The Variscan Orogen

2.1.1 Introduction

From the Early Silurian to Early Devonian, the major continental masses of Laurentia/Greenland, Fennoscandia/Baltica and the microcontinental terranes of east and west Avalonia collided, closing the Early Palaeozoic ocean Iapetus, and forming the North Atlantic Caledonides. To the south of the newly formed Laurussian landmass lay the proto-Tethys or Rheic Ocean; beyond it was Gondwana. Continent-continent collision between Gondwana and Laurussia, which closed the Rheic Ocean, initiated during the late Viséan to Namurian and formed the supercontinent of Pangea (Zeigler, 1982; Soper & Hutton, 1984; Soper *et al.* 1992; Fig 2.1). This collision caused the Variscan Orogeny.

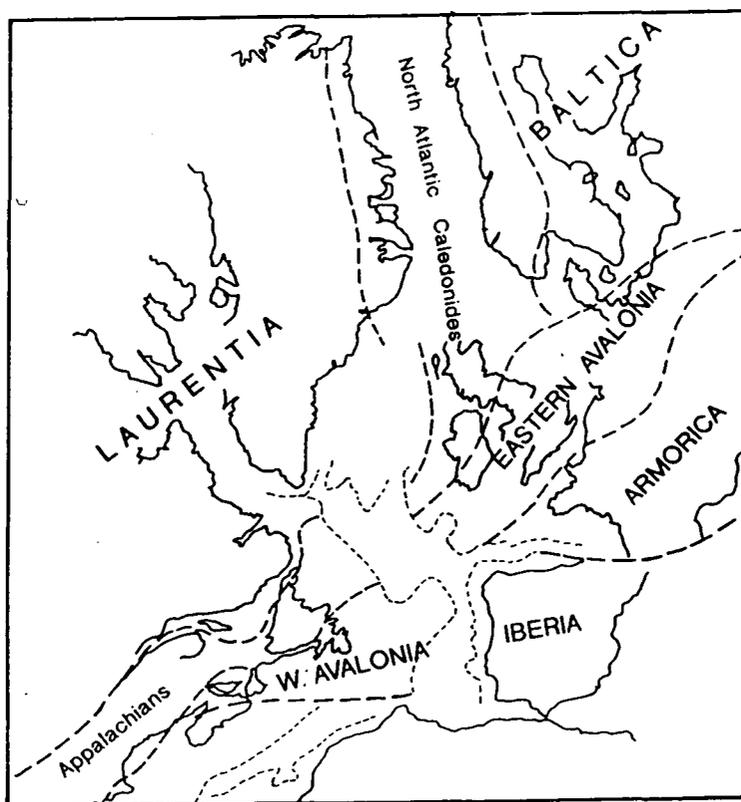


Fig. 2.1a A pre-Atlantic reconstruction of the Laurentia, Baltica and Gondwana continents, showing the positions of the east and west Avalonian micro-continental terranes (after Soper & Woodcock, 1990).

The British Isles lie at the northern edge of the Variscan orogen, forming one part of a linear WSW-ENE fold belt that includes the Ouachita belt in Texas, the Alleghenian belt in the Appalachians and the Mauritanian belt in NE Africa (Anderton *et al.* 1979). South and east of the British Isles, Variscan inliers of variably deformed Precambrian and Palaeozoic rocks occupy considerable tracts of western and central Europe, continuing into Caucasus and central Asia (Rast, 1983).

Kossmat (1927) subdivided the European Varicides into 3 E-W trending tectonostratigraphic zones, separated by major thrusts (Fig 2.2). Although this zonation has since been largely substantiated and refined by petrological, radiometric, structural and sedimentary data (Franke & Engel, 1982), the exact boundaries of the zones are still open to debate (Matthews, 1984).

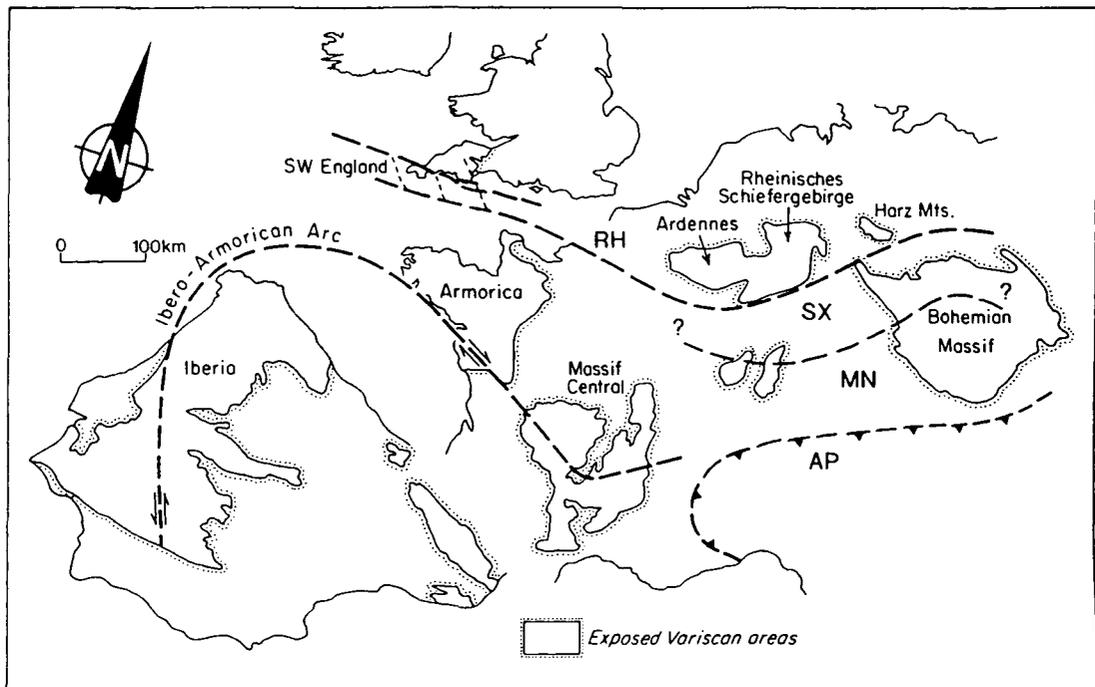


Figure 2.2 Tectonostratigraphic zones of the European Variscides. Present day exposed Variscan areas are indicated. RH, Rhenohercynian Zone; SX, Saxothuringian Zone; MN, Moldanubian Zone. AP, Alpine Front. Modified after Barnes & Andrews (1986).

SW England is traditionally regarded as lying in the northern part of the Rhenohercynian zone, with the rest of northern Britain and Ireland forming the Sub-Variscan foreland. Matthews (1977,1984) contended that there is only limited stratigraphic continuity between SW England and the Rhenohercynian Zone, and that, in terms of structure and petrology, a correlation with the Saxothuringian Zone would be equally valid. Badham (1982) noted the importance of long-lived strike-slip movements in the evolution of the Varicides. He stated that no continuity is likely

across Europe, since it is impossible to constrain the significant lateral displacements in this area, and is thus difficult to restore the dissected components. However, much detailed evidence has begun to emerge recently to suggest that there was sedimentary and tectonic continuity between the Variscan sequences in Devon and Cornwall, and those of the Rhenohercynian Zone in northern Europe. This includes similarities in the stratigraphic record (Holder & Leveridge, 1986a), similar age and provenance of flysch sedimentation (Engel & Franke, 1983), foreland younging of isotopic dates (i.e. cooling ages, Dodson & Rex, 1971; Ahrendt *et al.* 1983) and coaxial polyphase deformation (Rathey & Sanderson, 1982; Weber & Behr, 1983).

2.1.2 Southwest England

A map of southwest England shows an apparently simple outcrop pattern with east-west trending belts of Devonian strata flanking a broad outcrop of Carboniferous rocks, the 'Culm synclinorium' (Fig 2.3). The Devonian rocks are bordered to the south by the metamorphic and igneous complexes of Start Point (the Start Schists) and the Lizard peninsula (the Lizard ophiolite). The Upper Palaeozoic succession is intruded by a series of Permian granite bosses, linked at depth to form a large batholith. In eastern Devon, a gently tilted Permian and Mesozoic sedimentary cover unconformably overlies the Palaeozoic succession (Hobson & Sanderson, 1983).

The earliest work in southwest England was carried out by Sir Henry de la Beche, who published his "Report on the Geology of Cornwall, Devon and west Somerset" in 1839. There was considerable controversy regarding the age of his "grauwacke group" and the overlying "Carbonaceous group" which resulted in Sedgwick and Murchison defining the Devonian system using faunal criteria. The official revision of De la Beche's maps was commenced by Ussher in 1870 in the Wellington district, who then worked westward throughout the 1870's to encompass Exeter, Torquay, Bolt/Start and ultimately worked through Ivybridge to St Austell (Edmonds *et al.* 1975). Subsequently, a considerable amount of research focused on S.W. England (see Edmonds *et al.* 1975 for a detailed history of research), but it was only relatively recently that the true structural complexity of S.W. England was recognised. The work of Hendriks (1937) is of particular importance in this respect, because she was probably the first person to point out the regional significance of the complex small-scale structures, showing how they are indicators of larger scale processes and structure.

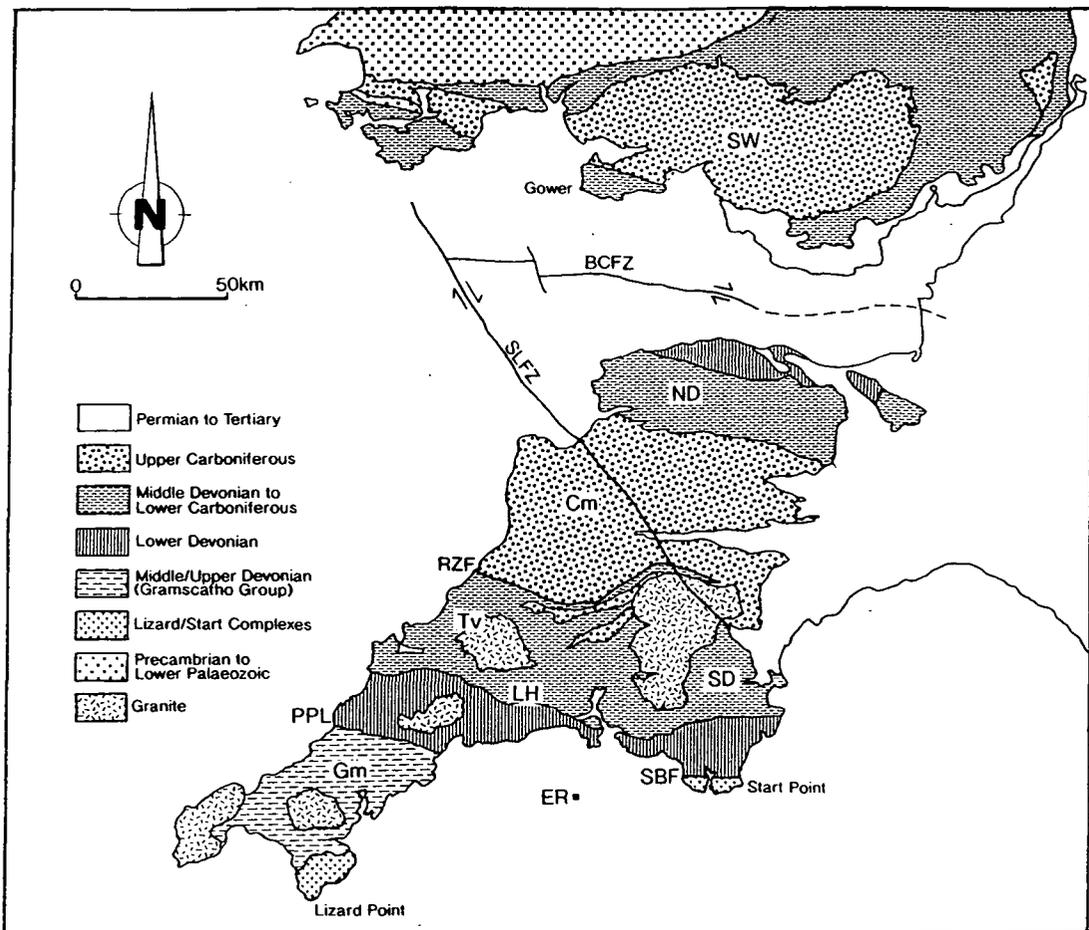


Figure 2.3 Simplified general geological map of SW Britain showing the distribution of basins, highs and major faults. Gm, Gramscatho Basin; Tv, Trevone Basin; SD, South Devon Basin; Cm, Culm Basin; ND, North Devon Basin; SW, South Wales Basin; LH, Liskeard High; SBF, Start Boundary Fault; PPL, Perranporth-Pentewan Line; RZF, Rusey Fault Zone; SLFZ, Sticklepath Lustleigh Fault Zone; BCFZ, Bristol Channel Fault Zone; ER, Eddystone Reef. Inferred sense of Variscan displacement is indicated on the Sticklepath Lustleigh Fault Zone and the Bristol Channel Fault Zone. Modified after Hartley & Warr, 1990.

2.2 The stratigraphic evolution of SW England

2.2.1 The pre-Devonian basement

The Pre-Devonian basement of SW England is uncertain (Goode & Merriman, 1987; Cope, 1987). Limited borehole and seismic reflection profiles indicate a Proterozoic volcanosedimentary succession overlain by Lower Palaeozoic sediments, that probably represent the eastern Avalonian basement (Fresney & Taylor, 1980). The oldest fossiliferous rocks in SW England are quartzites and limestones of probable Ordovician and Silurian ages that occur as exotic olistoliths in the Upper Devonian Verran Series, between Roseland Peninsula and Verran Bay. They are thought to have been eroded from the front of a complex northerly overthrusting pile which included the Lizard,

Lower Palaeozoic and metamorphic basement nappes (Leveridge, 1974; Barnes, 1983). The occurrence of similar rocks to the north and south, e.g. the Rudwig Conglomerate of Pembrokeshire, the Llanishen Conglomerate near Cardiff and the Armorican Quartzite of Brittany, might indicate that much of the Cornubian peninsula is underlain by these lithologies (Chadwick *et al.* 1983). Intermediate to acidic xenoliths of cataclastic granitoids within Devonian volcanics of the Gramscatho Basin may represent fragments of early Cadomian plutonics, similar to those observed in the Armorican Massif. These inclusions suggest that some ensialic crustal material may lie at depth beneath the Land's End area (Goode & Merriman, 1987). Shackleton (1984) noted the effect of the regional basement on the deforming cover sequence and suggests that anomalous Variscan structural trends in south Cornwall reflect the influence of the Cadomian basement.

The garnetiferous granite gneiss of the Eddystone Reef (see Fig 2.3) is a rare example of the exposed regional basement. Holdsworth (1989a) suggested that it represents an outcrop of the Normannian High (see 2.3.3) and considers it to have formed within the Mid-Silurian to Early Devonian European Caledonides. It records a K-Ar isotopic age of 375 ± 17 Ma (Miller & Green, 1961), indicating that it was last deformed towards the end of the Devonian. Doody and Brooks (1986), using SWESE (South-West England Seismic Experiment) data, interpreted the Start Complex as an onshore remnant of the Pre-Devonian basement, linking it to the Eddystone reef along a single basal refractor, thought to represent the top of the crystalline basement. Edwards *et al.* (1989) cast doubt on the validity of this basal refractor and demonstrated that it lies well within the cover sequence. Holder and Leveridge (1986a) disclosed unpublished BGS borehole data from Plymouth bay where Start-type schists are overlain by Eddystone gneisses. The interpretation of the refraction data proposed by Doody and Brooks (1986) could reflect the physical characteristics of the rocks, but is not necessarily representative of their age (Evans, 1990). Field studies of the Start Complex suggest that it is a metavolcanic/pelitic sequence of probable Devonian age, although this remains unproven in the absence of reliable radiometric or faunal data (Floyd *et al.* 1993a; see 3.2.5).

2.2.2 Upper Palaeozoic stratigraphy of S.W. England

It is generally agreed that, with the exception of certain parts of south Cornwall, Devonian sedimentation was initiated on continental crust forming the southern extension of the Old Red Sandstone Continent (Selwood, 1990). Thick continental sedimentation extended across much of southern Britain towards an ocean lying further south. A progressive northward onlap of marine sediments began in the late Seigenian

and was accompanied by sedimentary basin development because, whilst the effects of Variscan deformation are locally severe, it is still possible to recognise areas of enhanced Upper Palaeozoic sedimentary thickness which were probably controlled by differential subsidence of the continental shelf (Selwood, 1990; Shail, 1992; Fig. 2.3). It is probable that the subsidence was controlled by basement faults responding to lithospheric extension and/or transtension occurring during the Devonian and Early Carboniferous (Reading, 1975; Leeder, 1976; Badham, 1982; Barnes & Andrews, 1986). With the onset of orogenesis in the mid-to-late Carboniferous, lithosphere loading by nappe emplacement may have been an important subsidence process. Basins seem to have progressively evolved from south to north ahead of a northerly migrating deformation front.

Lower Devonian Shelf Sedimentation:

i) Dartmouth Group (Gedenian?-Mid Seigenian)

During the Lower Devonian most of SW Britain was covered by alluvial plains bringing detritus southwards from the Welsh Caledonides (Bluck *et al.* 1988). This proposal is based on faunal and lithological similarities between the Lower Old Red Sandstone of S. Wales (Allen, 1979) and the Dartmouth Group in S. Devon and Cornwall although, to date, no direct evidence has emerged to confirm that these successions are components of a linked depositional system.

The Dartmouth Group has been described by Dineley (1966), Seago and Chapman (1988), and Smith and Humphreys (1991). It represents the oldest stratigraphic unit in SW England, yielding Seigenian fossils, and although its base is not observed, Dineley (1986) reported traquairaspids in South Devon, suggesting the succession ranges well down into the Gedinnian (Bluck *et al.* 1988; see Fig 2.9). Hendricks (1951) indicated that the Dartmouth Slates could be subdivided into lithological units, but did not establish a formal stratigraphy. Dineley (1966) divided the Dartmouth Beds into four mappable formations with younging towards the northwest. These are: the Wembury Siltstones (youngest); the Yealm Formation; the Scobbiscombe Sandstones; and the Warren Sandstones (oldest). Hobson (1976b) showed that younging was towards the southeast while Seago and Chapman (1988) further divided the Wembury siltstones into the Renney Rocks Formation and the Wembury Formation, and omitted the Scobbiscombe Sandstone as it was one of many thin sandstone units.

Sequences are mudstone dominated (80-90%) but include sandstones and local pebbly mudstones (Smith & Humphreys, 1991). Smith and Humphreys (1991) considered these intervals of pebbly mudstones to be the result of syn-sedimentary

faulting related to basin foundering and their suggestion of an active tectonic regime contrasts with the conventional view of a tectonically quiescent Dartmouth Group (e.g. Dineley, 1966). Contemporaneous bimodal volcanicity indicates the probable initiation of continental rifting (Durrance, 1985).

Dineley (1966) suggested an alluvial setting for the Dartmouth Group, basing this on comparisons with Devonian sediments in Wales and on the presence of fish and plant remains. Allen (1979), in his review of Devonian sedimentation, agreed with a distal alluvial setting. Dineley's (1966) intimation that phases of coastal plain sedimentation occurred in tidal flats and lagoons is important because it implies that the Dartmouth Group represents the furthest southward advance of the coastal plains of alluviation of the ORS continent (Anderton *et al.* 1979). Evans (1981) reported a brachiopod fauna of Middle-Late Seigenian age in a marine band in east Cornwall and suggested that Meadfoot-type conditions (i.e. shallow marine) were locally developed in the Dartmouth Group. However, Smith and Humphreys (1991) showed that marine influences were subordinate during deposition of the Dartmouth Group, and that the above marine fauna came from a single horizon stratigraphically high up in the sequence, in a facies transitional to the Meadfoot Group. The marine nature of the overlying Meadfoot Group does not necessarily imply a marginal marine setting for the Dartmouth Group because the two groups are not coeval (Dineley, 1966).

Both the Dartmouth Group and the succeeding Meadfoot Beds are observed across the whole of SW England in what Hobson (1976a) termed the Dartmouth Antiform, but which Coward and McClay (1983), in a recent reinterpretation, considered to be a thrust structure.

ii) Meadfoot Group (Seigenian-Upper Emsian)

In unfaulted sections, the Meadfoot Group is observed to conformably overlie the Dartmouth Group (Hobson, 1976a; see Fig 2.9) and appears to reflect a widespread Seigenian transgression when shallow marine conditions replaced continental sedimentation. Bluck *et al.* (1988) suggested that the main transgression event was diachronous, as evidenced by the variable age ranges of brachiopod faunas recorded at the Dartmouth/Meadfoot boundary across Devon and South Cornwall (e.g. Evans, 1981). The Meadfoot Group may be divided into two formations; the Meadfoot Beds and the Staddon Grits.

The Meadfoot Beds are mudstone-dominated, deposited in fairly shallow water, but not in the intertidal zone as there is no evidence of exposure (Richter, 1967). Evans (1985), using brachiopod associations, concurred with a subtidal setting. In contrast, Selwood and Durrance (1982) proposed a tidal flat and lagoonal

environment. Pound (1983), whilst agreeing with the shallow marine setting of the former two authors, considered a storm-dominated shelf setting as more appropriate since evidence for tidal activity is unequivocal.

The Staddon Grits (Harwood, 1976; Evans, 1985) are a lenticular unfossiliferous facies of late Emsian age at the top of the Meadfoot Group which thicken westwards into Cornwall (Dineley, 1966). Although information on the depositional setting is limited, Harwood (1976) proposed a nearshore bar setting, whilst Selwood and Durrance (1982) preferred an offshore bar setting with incoming sediment being supplied by rivers reworking a Dartmouth Beds alluvial plains facies. Pound (1983) noted that the petrographic assemblage of the Staddon Grits is significantly different to that of the Dartmouth Beds and suggests that the bulk of the Staddon Grits represent a deltaic incursion of sediment derived from fault blocks to the north. Burton and Tanner (1986) describe the Staddon Grits from the Liskeard area and note that the presence of a prolific fauna of filter feeders is not suggestive of deltaic deposition. Their evidence favours the offshore bar setting proposed by Selwood and Durrance (1982).

During the uppermost Emsian times there is evidence of a deepening of marine conditions across much of the area. The Jennycliff Slates, which conformably overly the Staddon Grits, were deposited on a storm dominated shelf below wave base (Pound, 1983).

Basin Formation:

With continued deepening, probably the result of differential subsidence, the former Lower Devonian shelf area appears to have broken up into a series of basins and rises, which subsequently went on to develop their own stratigraphic successions (Selwood, 1990). The Gramscatho basin, however, probably originated immediately seaward of the shelf and, consequently, has a different evolutionary history to the intrashelf basins further north.

i) Gramscatho Basin (Eifelian-Fammenian)

This is the most southerly of the Rhenohercynian Zone basins in W England and, in recent years has been the most extensively studied. Two major lithostratigraphic divisions outcrop in this basin; the Gramscatho Group and the Mylor Slate (Figs. 2.4, 2.5). Although it has long been recognised that this succession exhibits characteristics of a deep water flysch sequence (Hendriks, 1937), the relationships between some of the facies types and associations are still open to interpretation (Shail, 1989).

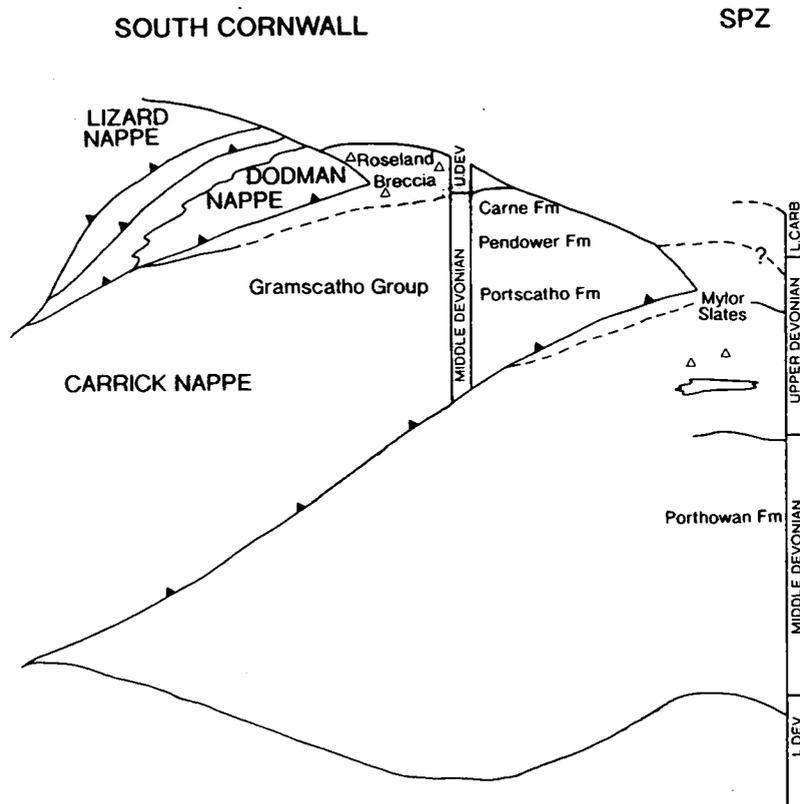


Figure 2.4 Diagram illustrating the stratigraphy of the Gramscatho basin. Modified after Bluck *et al.* (1988). SPZ represents the approximate position of the Start-Perranporth Zone.

Following the early work of Hill and McAlister (1906) it had generally been accepted that the Mylor Slate Formation was the older stratigraphic unit and occupied the core of a major fold, the 'Truro Antiform', whilst the Gramscatho Group was exposed on the NW and SE flanks. The Portscatho Formation (see Fig. 2.5) had long been ascribed a probable Middle Devonian age on the basis of widespread *Dadoxylon* plant fragments (Lang, 1929), whilst conodont data indicated that the Pendower Formation was Eifelian (Sadler, 1974; Leveridge, 1974). However the more recently obtained palynomorph data of Turner *et al.* (1979) indicated that the Mylor Slate Formation, at least in part, was Famennian in age and thus younger than the overlying Portscatho succession.

In order to account for this relationship Leveridge *et al.* (1984) postulated the existence of the Carrick Thrust (Fig 2.6), a major thrust which was envisaged to have transported the Portscatho Formation and other southerly allochthonous units of the Gramscatho Group northwards over the Mylor Slate Formation and Porthowan Formation. Evidence for thrusting came indirectly from offshore seismic reflection profiles which indicated the existence of southerly dipping reflectors within basement immediately south of the peninsula (Day & Edwards, 1983). More direct evidence came from a re-interpretation of olistostromes at the southern Mylor-Portscatho

boundary, which were considered to have formed contemporaneously with thrusting (Leveridge & Holder, 1985). Thus the outcrop pattern can now best be interpreted as a northern parautochthonous unit, in which the Gramscatho Group youngs upwards into the Mylor Slate Formation, that has been over-riden by a series of NNW transported thrust nappes (Fig. 2.4).

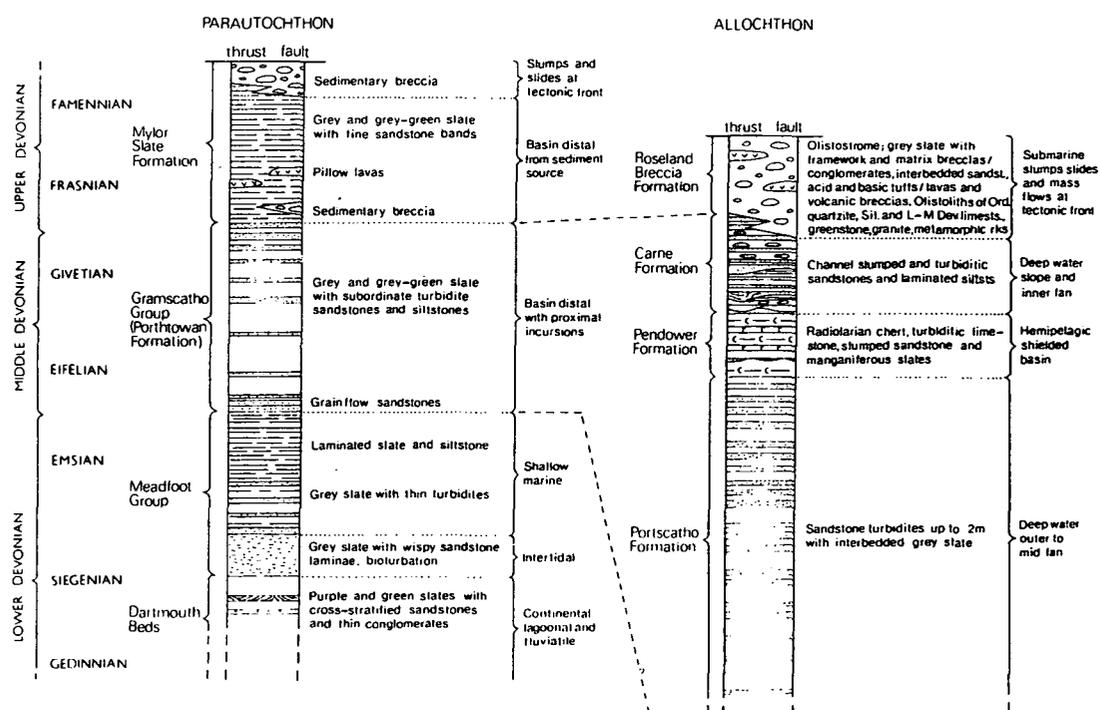


Figure 2.5 The lithostratigraphy of S. Cornwall. After Holder & Leveridge (1986b).

Palaeontological control is poor, with faunal ages for the Gramscatho Group ranging from Eifelian (Sadler, 1973) to Frasnian/Famennian (LeGall *et al.* 1985). Dating of the underlying Meadfoot Group and the overlying Mylor Slate Formation broadly limits it to the Middle Devonian. Wilkinson and Knight (1989) identified late Famennian palynomorphs within the Portscatho Formation to the south of the Carrick Thrust and thus it would appear that, although the Gramscatho Group is regionally overlain by the Famennian Mylor Slate, the recent palynomorph studies suggest that they may be partly contemporaneous (Shail, 1989).

In the parautochthon, the junction of the Meadfoot Group with the overlying Gramscatho Group appears transitional with thick bedded and massive coarse greywackes of Gramscatho affinities interbedded with laminated silts typical of the Meadfoot Group. These sandstones are not of the deltaic facies, as are the Staddon Grits in the Meadfoot Group to the north (Pound, 1983) but may represent channel sands of submarine fans (Holder & Leveridge, 1986b). The Gramscatho Group (Porthtowan Formation) comprises a 2500m thick sequence of dark grey-green slates with subordinate interbedded greywacke sandstones and thin siltstones. The

succeeding Mylor Slate Formation comprises grey/green slates interbedded with thin bands and laminae of sandstone, basic lavas and sedimentary breccias. These were interpreted by Wilson and Taylor (1976) as a quiet marine basinal sequence with distal turbidite incursions and slumping generated by faulting to the north. The Famennian age of palynomorphs from the lower part of the sequence (Turner *et al.* 1979) suggests that the formation extends at least to the top of the Devonian. The lower boundary with the Gramscatho Group is placed near the Givetian/Frasnian boundary on the indirect evidence of the facies change being associated in part with the Frasnian transgression recognised all across Europe (Holder & Leveridge, 1986b).

In the allochthon, the Gramscatho Group is subdivided into the Portscatho, Pendower and Carne Formations (Holder & Leveridge, 1986b; Figs 2.5, 2.6), roughly corresponding to the Lower, Middle and Upper Gramscatho Beds of Hendriks (1937). Although palaeocurrent data are lacking in these allochthonous sequences (Barnes & Andrews, 1986), it is likely that this material had a southerly source (e.g. Wilson & Taylor, 1976), and thus contrasts with the sedimentation in the Devonian basins further north. Present models confirm this southerly source with Holder and Leveridge (1986b) implying that it may have been the Normannian High.

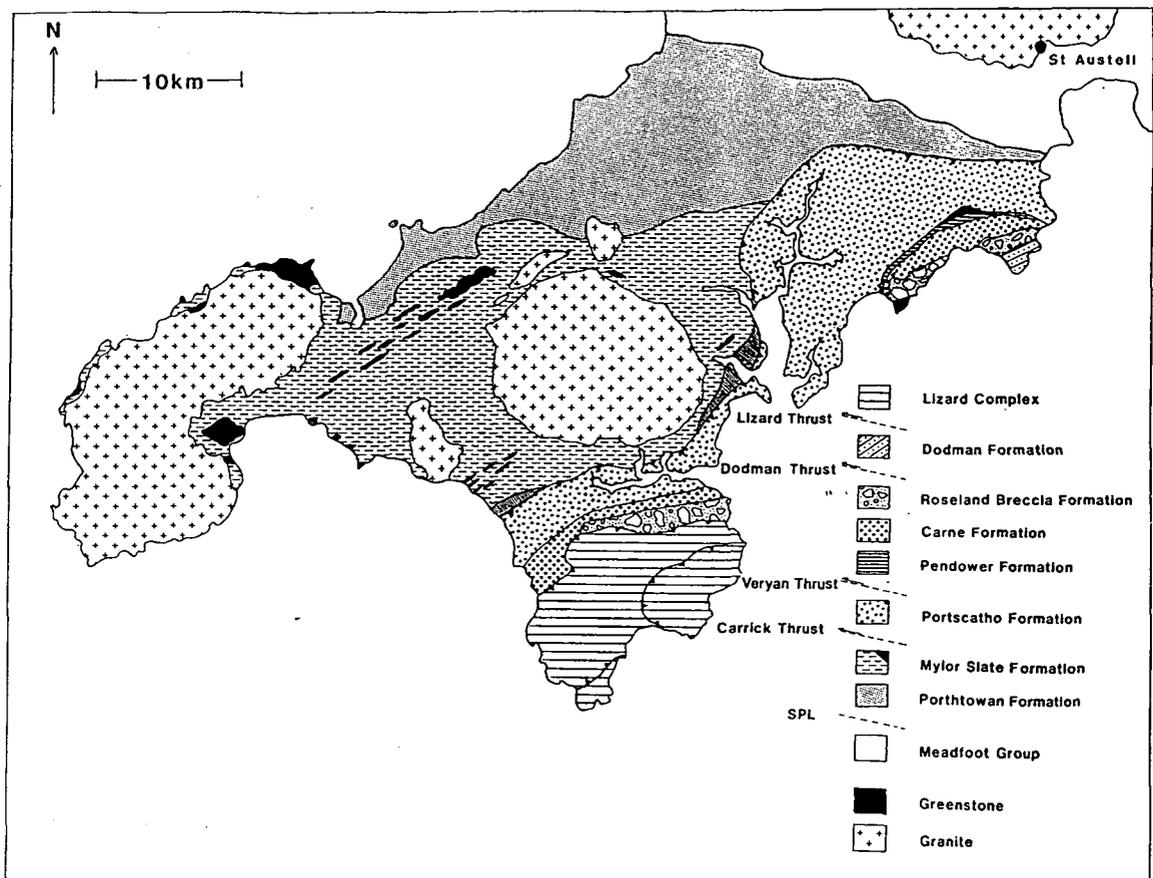


Figure 2.6 A simplified geological map of Cornwall illustrating the position of the Carrick Thrust and Carrick Nappe. After Shail (1992).

ii) Lizard Complex

The structure of the Lizard Complex (Fig 2.6) has been a controversial issue for many years. The tremendous difference between the high-grade metamorphic rocks of the Lizard and the Devonian rocks to the north was recognised by the earliest workers (e.g. De la Beche, 1839). This gave rise to theories that it was either an upfaulted block of basement or part of a large thrust sheet (Flett, 1946). All workers had accepted that the peridotite was essentially a diapir-like intrusion, until the work of Sanders (1955), who suggested that it was a thin sheet-like body. Green (1964) carried out a very detailed study of the peridotite, and concluded that it had formed as a diapiric intrusion of hot mantle into continental crust, with a well developed metamorphic aureole. A borehole by the BGS in the centre of the peridotite showed that it was only 360m thick, proving its sheet-like form and adding further weight to the theories developing during the 1970's that it was a dismembered ophiolite. Subsequently, regional seismic studies revealed that the Lizard Complex is less than 1km thick and, although a volcanic carapace to the ophiolite is not present, tectonically associated metamorphosed lavas (now hornblende schists) showing MORB similarities are consistent with a Lizard ocean-crust model (Floyd *et al.* 1993b). Recent Sm:Nd mineral isochron analyses from the olivine gabbro of the Lizard revealed an age of formation of 375±34Ma (Davies, 1984), whilst Rb:Sr isochron ages of 369±12Ma from the Kennack Gneiss were interpreted as a metamorphic age, recording the time of emplacement (Styles & Rundle, 1984). These isotopic dates were used to refute the earlier idea that the ophiolite had a Late Proterozoic (Cadomian) age.

The Lizard Complex was thrust over sediments and volcanics of a sedimentary *mélange*, termed the Meneage Formation (Barnes, 1983; Fig 2.6), which lies near the top of the Gramscatho Group. The *mélange*, which is in excess of 1km thick, preserves a distinct stratigraphy. The age of the uppermost Meneage Formation is Givetian, possibly extending up to the Frasnian (Barnes, 1984). Barnes and Andrews (1984) suggest that the Lizard Complex was emplaced in a cooled thrust sheet during D₁ deformation, since evidence of an increase in the regional metamorphic gradient across the Meneage is lacking.

iii) Devonian-Lower Carboniferous Intrashelf Basins

These intrashelf basins conform to a general facies model consisting of Lower to Middle Devonian basins with a tectonically active northern margin accumulating volcanoclastics which gives way to a broad shelf, and a siliciclastic starved but carbonate rich southern margin supplying limestone turbidites into the basin. Further carbonate input came from reef complexes developed on topographic highs, whilst an

Upper Devonian influx of flyschoid clastics, eroded from the advancing deformation front, terminated the succession (Bluck *et al.* 1988). The fragmentation and subsidence of the carbonate platform resulting in 'Schwellen and Becken' is a feature common to much of the Rhenohercynian Zone (Franke & Engel, 1982).

a) Trevone Basin (Upper Eifelian-Famennian): The Trevone Basin (Matthews, 1977) includes Middle Devonian to Namurian deposits occurring in an E-W striking zone from the north Cornish coast eastwards to the St. Teath-Portnadler Fault Zone (see Fig. 2.3). Basinal black mudstones indicate that the basin was already developing in the Emsian (Selwood, 1990). The stratigraphy of the basin has recently been revised and two distinct Middle to Upper Devonian successions recognised (the Bounds Cliff Succession and the Padstow Succession; Smith, 1991; Warr, 1991; Fig 2.7), though the distinction between them was lost in the Early Famennian when purple-and-green argillites spread across the whole basin. Beds younger than Famennian age are not recorded (Selwood, 1990).

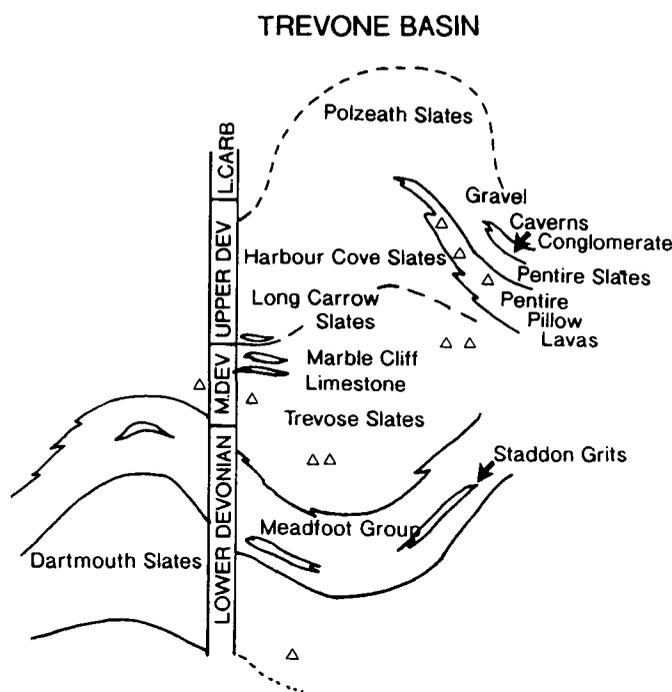


Figure 2.7 The stratigraphy of the Trevone Basin and its northern and southern shelf areas. Boscastle and Buckator formations are considered to have maintained their relative palaeogeographical positions on the northern shelf. Modified after Bluck *et al.* (1988).

In the northern sector, the Upper Devonian Bounds Cliff Formation (the Northern Succession of Selwood & Thomas, 1986a) is dominated by thick mudstone successions which contain intraformational conglomerates, abundant sandy tempestites and slumps, thought to be deposited high on the shelf/basin margin (Bluck *et al.* 1988).

Further south, the Padstow Succession (Smith, 1991), which comprises the Pentire and Trevone Successions of Gauss and House (1972), includes a Middle and Upper Devonian, mudstone dominated, deep marine sequence (partially described by Beese, 1984), plus important intrusive and extrusive basic volcanics (Pentire Volcanic Member) exhibiting an intraplate geochemistry typical of rifting (Floyd, 1982). Intraformational conglomerates containing micritic limestone clasts may have formed as a consequence of fault scarp activation or volcanic activity (Warr, 1991). The Marble Cliff Beds of earliest Frasnian age are limestone turbidites, composed mainly of crinoidal debris, which exhibit evidence of a south-westerly provenance (Tucker, 1969). Apart from these turbidites, there is no evidence for the presence of a carbonate platform on the shelf flanking the basin to the south (Bluck *et al.*, 1988).

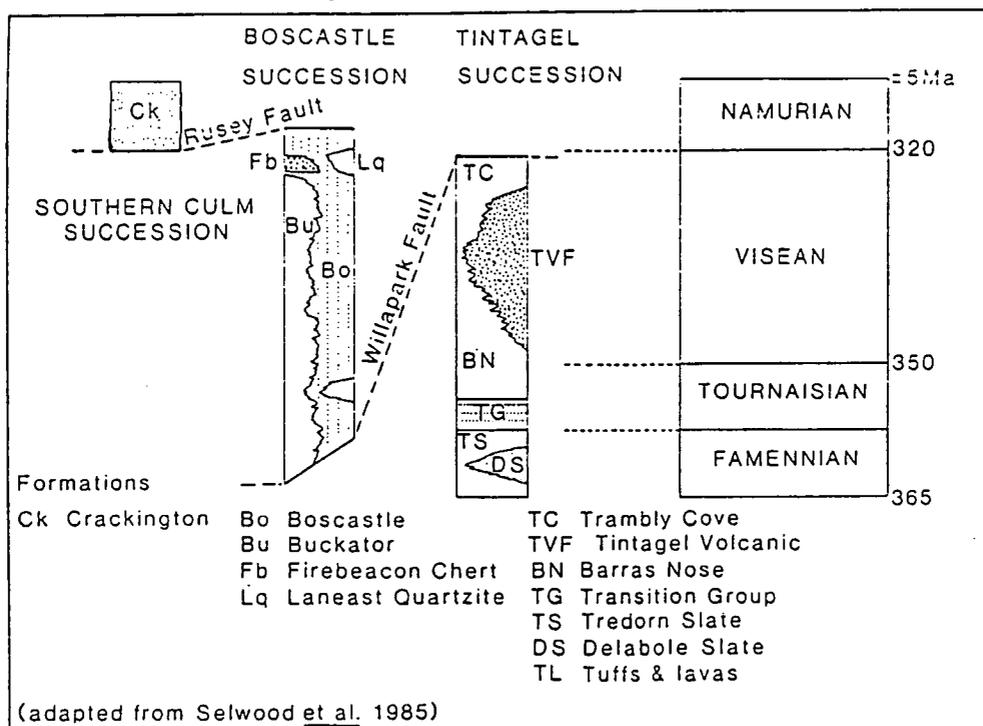


Figure 2.8 Detailed stratigraphic successions of the Boscastle area. After Warr (1991).

There is continued debate as to the status of the 10-15km wide zone of Famennian to earliest Namurian lithologies (the Boscastle and Tintagel Successions; Fig 2.8) which occur between the Trevone and Culm Basins. The successions include the Upper Devonian Delabole Slate Group, Viséan sandstones of shallow marine origin (Boscastle Formation) and the Tintagel Volcanic Group (also of possible Viséan age). Full details of the stratigraphy are provided in Selwood *et al.* (1985) and Selwood and Thomas (1986a,b). Controversy exists as to the pre-deformation position of these successions. Hartley and Warr (1990), suggest that, although complexly folded and thrust, the recognised successions have maintained their relative palaeogeographical positions, whilst Isaac *et al.* (1982), argue that they are contained in southerly derived

nappes (the Boscastle Nappe and associated Tredhorn Nappe). However, a suitable source area, from which to derive such nappe sequences, has yet to be identified (Andrews *et al.* 1988; Warr, 1991).

b) Liskeard High (Emsian-Frasnian): South of Bodmin Moor, Burton and Tanner (1986) demonstrated the existence of a persistent shelf region (Fig. 2.3), limited to the west by the St. Teath-Portnadler Fault Zone, and eastwards by the Otterham Fault Zone. This area accumulated a sequence of Emsian to Frasnian open shelf deposits with interbedded volcanoclastics (Selwood, 1990). The 'Liskeard High' represents a westward extension of the axial ridge proposed by Selwood and Durrance (1982) in South Devon and interrupts the central SW England basinal continuity, separating the Trevone basin in the west from its along strike equivalent, the South Devon Basin, in the east (Burton & Tanner, 1986). There is no evidence in this region for the establishment of the reefal facies that characterises the shelf south of the South Devon Basin (Selwood, 1990).

c) South Devon Basin (Eifelian-Tournasian): The along strike, eastern equivalent of the Trevone Basin, the South Devon Basin (see Fig. 2.3) shows a marked contrast since it is dominated by shallow water carbonates, which developed on an intrabasinal rise (the Chudleigh Rise; Fig. 2.9) generated by a thick volcanic pile (Kingsteighton Volcanic Group), related to basin foundering. Overlying this thick succession of rise carbonates is a complete, but condensed, Late Devonian to Early Carboniferous mudstone sequence. The Chudleigh Rise carbonates are distinct from the Eifelian to late Frasnian limestones of the carbonate platform to the south of the basin. These platform limestones include the Newton Abbott Limestone Group and the Torquay Limestone Group (Scrutton, 1977), which, although occurring as both parautochthonous and highly disrupted allochthonous sequences, can be interpreted as reef complexes and are comparable to less deformed reef complexes elsewhere (Scrutton, 1977).

Middle to Upper Devonian basinal mudstones occur to the north of the platform; these contain southerly-derived limestone turbidites and slump deposits (Van Stratten & Tucker, 1972) that indicate remobilization of platform material. Small, local bioherms developed across the platform prior to complete submergence in late Frasnian times, when condensed nodular limestones and thin shales were deposited (Bluck *et al.* 1988).

The parautochthon north of the basin appears as windows in the allochthon, revealing a sequence of sparsely fossiliferous Upper Devonian outer shelf shales. These contain occasional thin, probably storm generated, sandstones of northern derivation (Bluck *et al.* 1988).

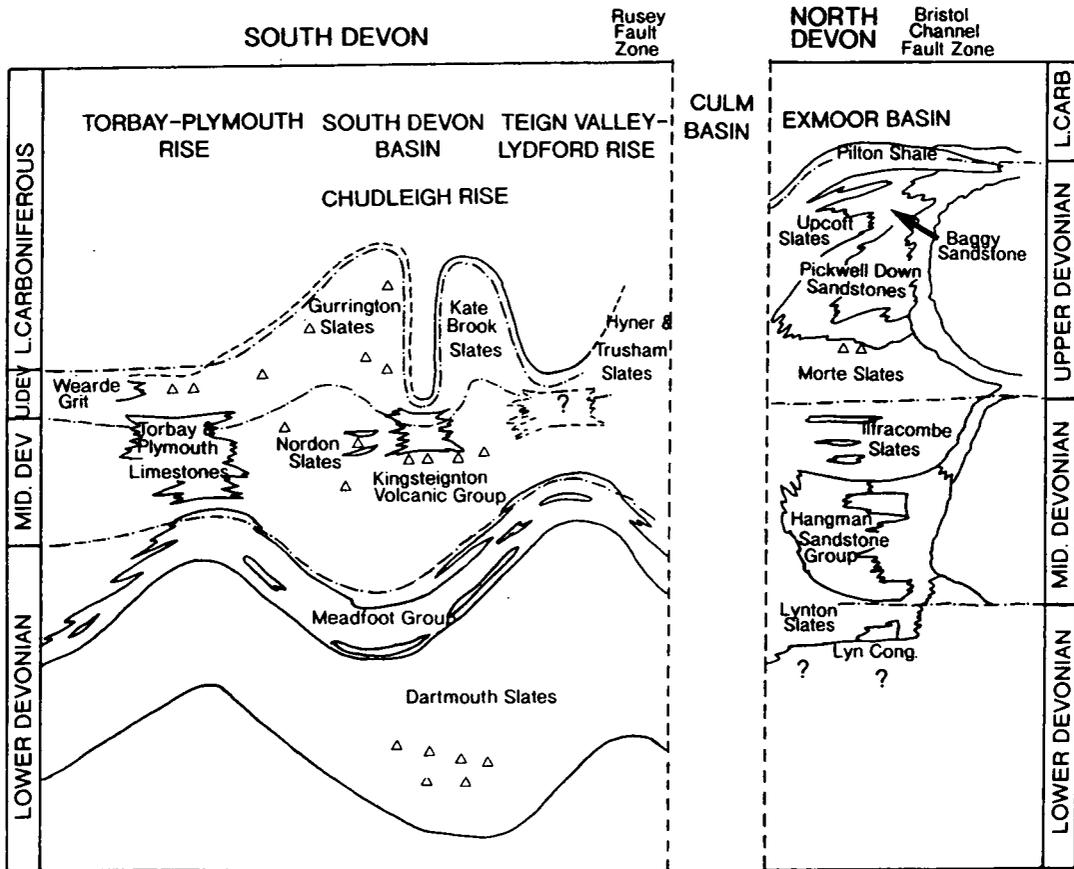


Figure 2.9 Possible stratigraphic relationships between the Devonian rocks of North Wales and North Devon, with those of South Devon. Modified after Bluck *et al.* (1988)

d) North Devon Basin (Emsian-Tournasian): This basin is characterised by thick alluvial sequences, derived from the adjoining continental landmass, which are intercalated with near-shore marine deposits in complete sequences, and thus differ from the wholly marine successions further south (Bluck *et al.* 1988; see Fig. 2.3).

The oldest rocks in this basin are the Emsian Lynton Beds (Fig 2.9). These are the northern equivalent to the upper Meadfoot Group and comprise mainly lenticular bedded, bioturbated heterolithic sandstone and mudstone successions, together with wave rippled and hummocky cross-bedded fine sandstones. They are interpreted as wave dominated shelf deposits (Bluck *et al.* 1988). Occasional matrix supported conglomerates are interpreted as fault scarp deposits and Tunbridge (1986) has

suggested that these syn-sedimentary faults are splays from the Bristol Channel Fault system.

The succeeding Eifelian Hangman Sandstone Group is dominated by two coarse alluvial successions separated by a short mudstone interval (Tunbridge, 1980). These successions indicate the influx of substantial quantities of siliciclastic material due to Late Caledonian (Mid Devonian) uplift in South Wales and possible late uplift along the line of the Bristol Channel Fault (Leeder, 1982; Tunbridge, 1986). This contrasts with the deepening of marine conditions further to the south and attests to the importance of source area tectonics. The uppermost formations of the Hangman Sandstone Group (Sherrycombe and Little Hangman Formations) indicate marine transgressive conditions, as Middle Devonian rising sea levels stemmed the flood of clastics and terminated terrestrial deposition (Bluck *et al.*, 1988).

A return to shallow marine conditions is reflected in the Givetian to Early Frasnian Ilfracombe Beds, which comprise heterolithic mudstones, fine calcareous sandstones and bioclastic limestones, and represent deposition below storm wave base (Bluck *et al.* 1988).

Deepening is marked by the Frasnian delta front deposits of the Morte Slates and the overlying Famennian proximal delta facies of the Pickwell Down Sandstones. Unlike the terrestrial sandstones of the Hangman Sandstone Group, the Pickwell Down Sandstone cannot be demonstrated to be the product of sourceland tectonics and is instead attributed to Early Famennian regression (House *et al.* 1977).

The remainder of the Upper Devonian and early Carboniferous records a transition to deeper marine conditions (Pilton Shale) as the shoreline migrated northwards from its Middle Devonian position in N.Devon towards S.Wales (Bluck *et al.* 1988).

iv) Carboniferous Foreland Basins

a) Culm Basin (lower Namurian-early Westphalian C): The Culm basin developed conformably on deep marine mudstones and cherts of Viséan age (Codden Hill Chert) which succeeded the Pilton Shales of North Devon. A 2-3km sequence is preserved in an E-W trending basin, which crosses north Cornwall and Devon before passing under the Permo-Triassic cover in the Exeter region (Fig. 2.3). Sedgwick and Murchison (1839) established the Culm Measures as the lithostratigraphic name for the basinal Carboniferous rocks, and subdivided them into Lower Culm (~Dinantian) and Upper Culm (Silesian). Subsequently the Upper Culm was subdivided into various smaller units, but Edmonds (1974) revised and refined the lithostratigraphy recognising three formations (Fig 2.10); the Crackington Formation (lower Namurian to early

Westphalian A), the Bideford Formation (?late Namurian to Westphalian A) and the Bude Formation (early Westphalian A to early Westphalian C). These formations, at least in part, probably represent different parts of a linked depositional system comprising basinal, shelf to slope and fluvial dominated deltaic environments respectively (Elliot, 1976; Melvin, 1986; Higgs, 1991).

		SOUTH WALES		CULM&N.DEVON			
CARBONIFEROUS	SILESIAIAN	C-D	1600	Pennant Measures			
		Westphalian	B	920m	Mid. Coal Measures	Bude Fm.	
			A		Lr. Coal Measures	Bideford Fm.	
			Namurian	750m	Shale Group B. Grit	Crackington Fm.	
	DINANITIAN	Visean		1400m	Oy. Beds Ox. Hd. Lst. Hunts Bay Group High Tor Lst. Caswell Bay Mst. Gully Oolite	Codden Hill Chert	
					Black Rock Group		
					Lower Limestone Shale Group		
			Tournasian				
DEVONIAN	UPPER		137m	Quartz Conglomerate Group Cwrt-yr-ala Fm.	Pilton Shales Baggy Ssts. Upcott Slates Pickwell Down Sst. Morte Slates		
	MIDDLE				Ilfracombe Beds		
					Hangman Sst. Group		

Plate 2.10 Summary lithostratigraphy of the Middle Devonian to Upper Carboniferous successions in North Devon and South Wales. Ox Hd Lst= Oxwich Head Limestone; Oy= Oystermouth; B. Grit= Basal Grit; Fm= Formation; *=approximate structural thickness. Thicknesses for the Dinantian of South Wales are maximums. Modified after Hartley & Warr (1990).

The Crackington Formation is envisaged as the turbidite fill of a wide marine basin. The exact location of the source area is unknown, but linear rather than point source derivation is indicated, possibly by shelf margin slumping (Thomas, 1988). The Bideford Formation accords with the model of a delta-complex, which extended southwards into a moderately shallow nearshore basin (Elliot, 1976; Thomas, 1988). Interpreting the conditions for the deposition of the Bude Formation has proved controversial. Melvin (1986) attempted to reconcile this by proposing deposition in a fan which developed in relatively shallow water depths (i.e. shelf as opposed to abyssal), although Higgs (1984, 1991) argues for non-fan shelf deposition.

Palaeocurrent data for the Crackington Formation display a bimodal E-W trend probably reflecting an axial drainage system (Melvin, 1986). Data for the Bude

Formation (Melvin, 1986; Higgs, 1991) show a polymodal palaeocurrent pattern with flow from all quadrants except the south. In general, it appears that detritus in the Culm Basin was derived from its northern margin because fine-grained orthoquartzitic debris from a mature hinterland predominates (Thomas, 1988).

The succession in the Culm Basin shows a transition from the dominantly marine Crackington Formation to the freshwater/brackish Bude Formation, indicating a gradual infilling of the basin in an environment of northward migrating tectonism (Melvin, 1986; Thomas, 1988). Subsidence in the Culm Basin is thought to have been generated by lithosphere loading associated with this Upper Carboniferous northward migration (Thomas, 1988; Higgs, 1991). The role of previous basin history and inversion in controlling the site of this foreland basin has been discussed by Hartley and Warr (1990).

b) South Wales Basin (Namurian-Westphalian): The Late Devonian transgression associated with the deposition of the Pilton Shales in the North Devon Basin led to the development of carbonate shelf conditions in South Wales which persisted through the Dinantian. Gayer and Jones (1989) suggest that the Bristol Channel fault zone was a major Variscan thrust which floored an extensive thrust sheet, the emplacement of which, during Early Namurian to Westphalian times, downflexed the lithosphere to produce the South Wales coalfield foreland basin (see Fig. 2.3). They also suggest that the Culm Basin may have continued to evolve as a thrust sheet top basin as it lay in the hanging wall to this major thrust. The Early Dinantian carbonate platform subsequently broke up and the basin infilled with Namurian and Westphalian clastic sediments (Kelling, 1988; Gayer & Jones, 1989; Hartley & Warr, 1990). Namurian sediments fine upwards and record a change from the high energy, shallow marine Basal Grit, to the mudstone dominated lagoonal sediments of the Shale Group (Hartley, 1993; Fig. 2.10). Sediment was sourced principally from the north, with some input from the south and east. Westphalian A to Early Westphalian C saw the deposition of mudstones and extensive coals, probably in a swampy, coastal plain setting, with clastic input again sourced from the north, south and east (Hartley, 1993).

A marked change in sedimentation occurred during early Westphalian C to Late Westphalian D with the deposition of the sandstone dominated Pennant Measures. These were derived from a tectonic sourceland to the south, possibly an uplifted Variscan thrust sheet in the Bristol Channel Fault area (Gayer & Jones, 1989).

v) *Permian Basins*

During the latest Carboniferous, there was a rapid change from NNW-SSE directed Variscan compression to a protracted period of extension related subsidence, which affected the NW European continental shelf for much of the Permian to Triassic times (Hawkes, 1981). This brought about widespread normal faulting and the reactivation of Variscan thrusts, and was associated with the development of substantial Permian sedimentary basins to the north and south of the present peninsula, and in the East Devon area (Evans, 1990). This change in tectonic regime has been attributed to orogenic collapse in mainland Europe (Ménard & Molnar, 1988) and, in conjunction with mantle activity, led to the widespread eruption of Stephanian lamprophyric rocks and later to the Lower Permian elvan volcanism and attendant mineralization across SW England (Hawkes, 1981).

vi) *Cornubian Granite Batholith*

In Early Permian times, during the closing stages of the Variscan Orogeny, the deformed late Palaeozoic rocks of SW England were intruded by a major granite batholith. This Cornubian Batholith is exposed in a series of ENE-WSW trending inliers from Dartmoor to the Isles of Scilly (Edmonds *et al.* 1975; see Fig. 2.3).

The early workers in SW England suspected that the major granite outcrops were linked at depth. De la Beche (1839) was quite clear that a single granite mass underlay the individual 'protrusions'. Although Ussher (1892) took a similar view, he also concluded that the granite was due to metamorphism of pre-existing rocks of Pre-Devonian age. It was not until the gravity survey of Bott *et al.* (1958) that these early suspicions were confirmed. A large negative gravity anomaly over the Cornubian peninsula revealed the extent of the 250km long batholith (Fig 2.11). Recent gravity data (Tombs, 1977) suggests that the batholith widens from 10km at the surface to between 30-50km at its base, which is considered to lie at a depth of 10-12km.

Onshore, the batholith consists of coarse adamellitic granite with tourmaline, zircon and apatite as the commonest accessory minerals. All the granites contain felsic veins and pods, and quartz/tourmaline segregations are common (Edmonds *et al.* 1975). Late stage alteration effects are well displayed and include extensive tourmalinization, greisenization and kaolinization. This granite was also the source of the hydrothermal Sn-W mineralization, as well as the heat source for associated Cu-Pb-Zn-Fe-As deposits within the margins of the granites and their aureoles (Floyd *et al.* 1993b). According to the Rb:Sr isotopic evidence of Darbyshire and Shepherd (1985), the batholith was emplaced between 290-280Ma, with the main stage of

mineralization at about 270Ma. The K:Ar results of 303-254Ma (Edmonds *et al.* 1975) reflect cooling rather than emplacement ages (Darbyshire & Shepherd, 1985).

A second, smaller, sub-parallel batholith lies some 150km west of Land's End and rises 50m above the sea bed. This submarine shoal, termed Haig Fras, measures 45km by 15 km (Evans, 1990; Fig 2.11). The commonest rock recovered from Haig Fras is a fine-grained, foliated, leucocratic, tourmaline-bearing granite, which yields a K:Ar date of 277Ma (Exley, 1966).

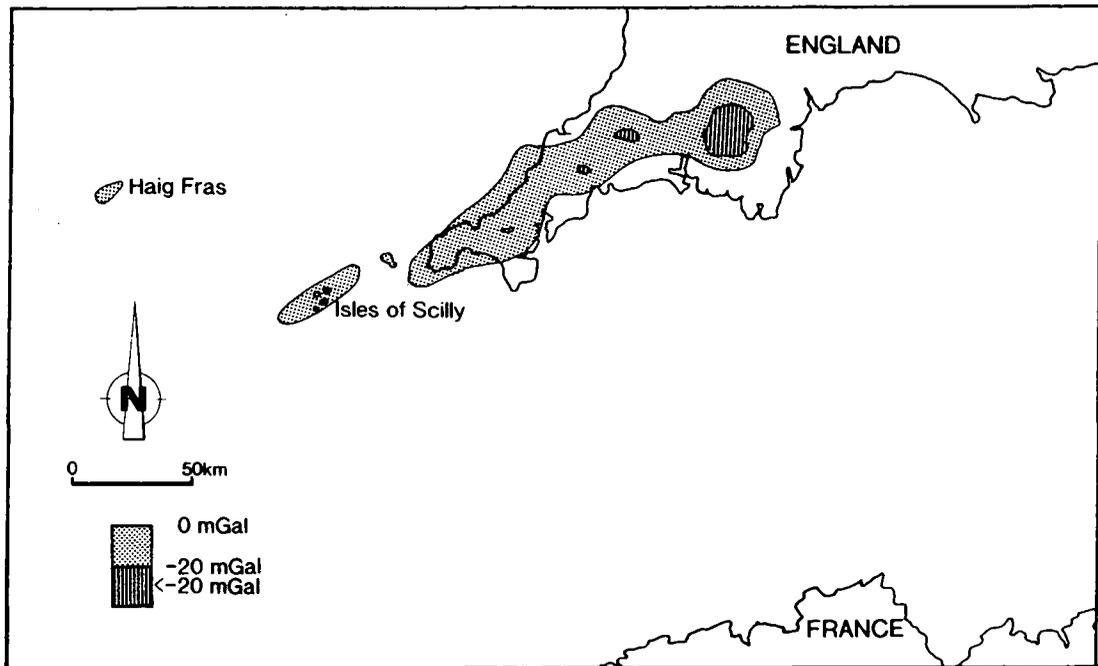


Figure 2.11 Bouguer anomaly map of the SW Approaches area, giving the approximate extent of the SW England granite batholith. Modified after Evans (1990).

2.3 The Tectonic Evolution of Southwest England

2.3.1 Introduction

Poor inland outcrop, incomplete stratigraphic control and the effects of numerous, post-Variscan faults has hindered a greater understanding of the highly variable polyphase sequence of ductile structures across SW England and has inhibited the synthesis of regional models. However, in the last decade, a broader and more complete picture has begun to emerge. Of importance in this respect are the collections of papers edited by Hancock (1983) and Hutton and Sanderson (1984). Contributions in the *Journal of the Geological Society*, and in particular the thematic sets of papers on the Lizard Complex (1984) and the Variscan evolution of SW England (1986), have also been useful. Dineley (1986) provides a brief review of the main advances in this field over the last quarter century.

2.3.2 Early Work

Hendriks (1937, 1939) provided the first major synthesis of Cornubian tectonics and drew analogies between flysch sedimentation observed in south Cornwall and similar flysch deposits seen in the Alps. Using these analogies, and her observations that the 'Start-Lizard schists' were comparable to the Alpine 'Schistes lustrés', she postulated the existence of a single Lizard-Dodman-Start thrust which carried a northward directed nappe over the flysch sequences below. Although this idea is now discredited (Sadler, 1974), the idea that thrust tectonics is fundamental to understanding the evolution of SW England is universally agreed (e.g. Coward & McClay, 1983; Leveridge *et al.* 1984).

During the 1950's to 1970's, much of the work revolved around detailed descriptions of the small-scale structures (e.g. Marshall, 1962), using the approach favoured by Wilson (1951). Dearman and co-workers, together with groups from Exeter University, continued this approach. The recognition that many of the boundaries between different rock units are faults of various types was considered crucial and much of the credit for this belongs to Dearman (1963), who first described a series of NW-SE trending dextral strike-slip faults throughout SW England. These cut both the deformed Upper Palaeozoic sequences and the Permian granites, and some may record Palaeogene movement (Dearman, 1963). In a similar manner, Freshney (1965) commented on the widespread occurrence of low angle, north dipping extensional faults in North Cornwall and Central Devon and suggested that these were Variscan thrusts that were reactivated during granite emplacement. Early studies of the geometry of polyphase deformation were provided by Smith (1965), Dearman (1966) and Dearman and Freshney (1966).

In the early 1970's, the wealth of information that had accumulated through two decades of detailed study of small-scale structures was allied to an improved stratigraphic understanding to produce the first tentative syntheses of Cornubian structure (e.g. Hendriks, 1971; Dearman *et al.* 1971). Subsequent work revolved around the identification of tectonic zones (Sanderson & Dearman, 1973; Fig 2.12) on the basis of early fold axial trends, fold facing directions, associated cleavage fabrics and the number of locally developed fold phases. Whilst the delineation of these tectonic zones provided a much needed regional structural framework, it also artificially emphasised the boundaries between the zones (Hobson & Sanderson, 1983; Hancock *et al.* 1983). The work of Matthews (1977) was an exception in that it stressed the continuity of structural elements and attempted to integrate the pre-deformation history of the depositional basins.

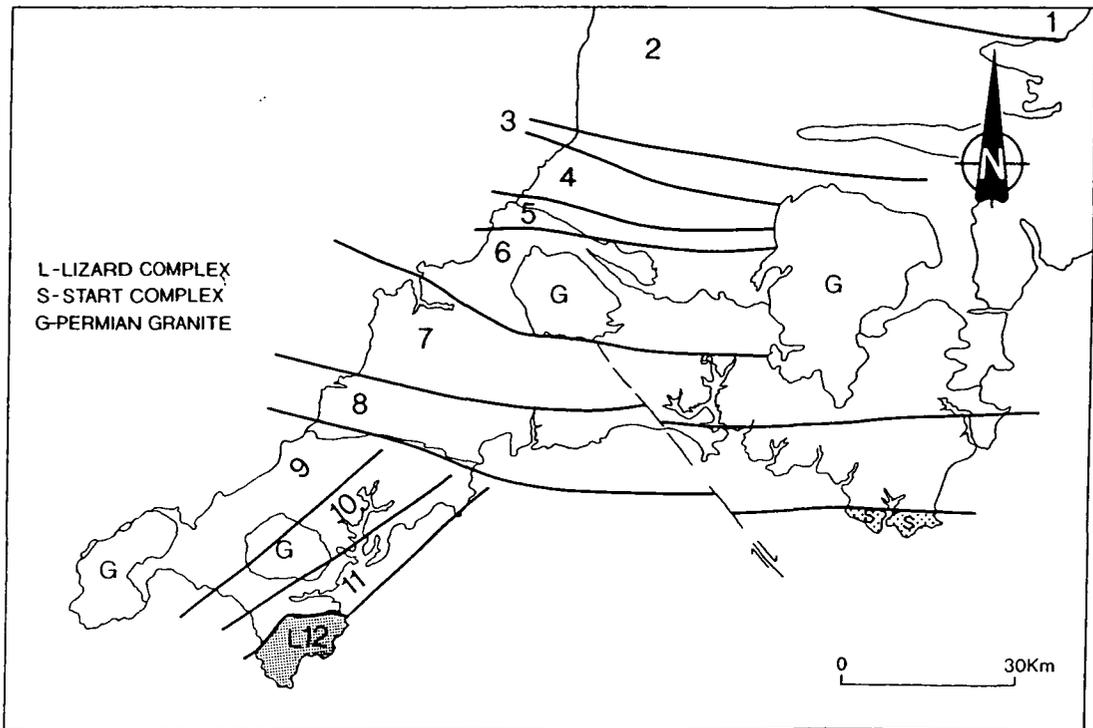


Figure 2.12 Map of the twelve tectonic zones of SW England, defined by fold attitude, facing and axial trend. See Table 1 for detail. After Sanderson & Dearman (1973).

Zone	Fold Attitude	Facing	Axis Trend
1	Inclined S	N	E-W
2	Upright	up	E-W
3	Inclined N	S	E-W
4	Recumbent	S	E-W
5	Recumbent	S*	Oblique
6	Recumbent	SSE	ENE-WSW
7	Recumbent	NNW	ENE-WSW
8	Inclined S	NNW	ENE-WSW
9	Recumbent	NNW	ENE-WSW
10	Inclined SE	NW	NE-SW
11	Inclined SE	NW*	Oblique
12	Metamorphics of Lizard and Dodman Point		

*Original facing direction

During the same period, Dodson and Rex (1971) detailed K-Ar cooling ages for slates across the peninsula. These dates indicated that mica recrystallization began in the Late Devonian in Cornwall and showed progressively younger ages moving northwards (Fig. 2.13). This was inferred to reflect a northward migrating Variscan deformation front. However Warr *et al.* (1991) have recently revised the K-Ar analyses, recognising only three populations of ages; M₂ cooling ages (270-310Ma) related to underthrusting of the Culm Basin, M₁ cooling ages (340-320Ma) attributed to D₁ deformation and cooling ages related to the emplacement of the Lizard Complex (380-360Ma).

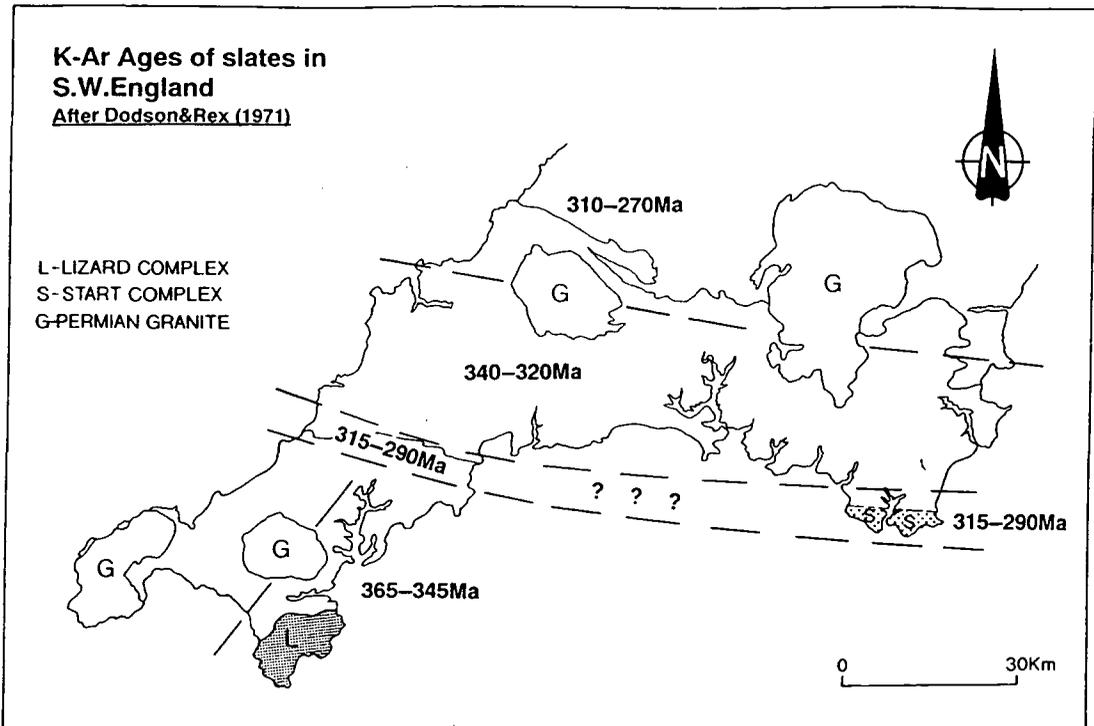


Figure 2.13 K-Ar metamorphic cooling ages for the slates of SW England, showing a progressive northwards younging. After Dodson & Rex (1971)

With the evolution of the tectonic models came a greater understanding of the setting and emplacement of the Lizard Complex. Early work proposed that the Lizard was a diapiric intrusion of hot peridotite into continental crust (Green, 1964). Studies throughout the 1970's, however, argued that the Lizard Complex was a dismembered ophiolite which had been emplaced, as a thrust slice, over the Devonian sequence to the north.

2.3.3 Recent Tectonic Models

During the last ten years, a number of tectonic models have been presented for the Variscan Orogen of SW Britain. Whilst it is clear that the fold and thrust belt of SW England resulted from a NNW directed compression deforming a series of E-W trending sedimentary basins, the models offered to account for this are significantly different.

Thin-Skinned Fold & Thrust Models

During the early 1980's, several authors noted that the Variscan deformation of SW England exhibited many of the characteristics of a foreland fold and thrust belt. This idea was partially inspired by the results of deep seismic profiles in which gently south dipping reflectors were interpreted as Variscan thrusts (Meissner *et al.* 1981; Kenolty

et al. 1981; Chadwick *et al.* 1983). Major Variscan thrusts were also recognised at this time following detailed surface mapping onshore (Coward & McClay, 1983). The gentle southerly dips of cleavages, thrusts and recumbent folds along the NW coast of Devon and Cornwall and the SE coast of Devon were interpreted as evidence of southerly dipping shear zones, one of which was thought to form a regional basal décollement, running from near surface close to the Variscan Front in S. Wales, to a lower part of the crust in S. Cornwall (Shackleton *et al.* 1982). Such thin-skinned models were thought to be supported by the deep seismic data of Meissner *et al.* (1981). Strain data from across the peninsula indicated a total shortening of at least 150km (~50%), but Shackleton *et al.* (1982) suggested that the total thickness of the crust was insufficient to allow generation of the Cornubian batholith by crustal anatexis. This view may be premature since their degree of late orogenic extension was clearly underestimated (<5%).

During the remapping of central SW England (Exeter University mapping contract 1971-1985), Isaac *et al.* (1982) suggested that large-scale northerly moving thrusts were also of fundamental importance in this region. Isaac *et al.* (1982) proposed a very thin skinned (<1km) tectonic regime to explain the distribution of units in the poorly exposed area between the Bodmin and Dartmoor granites. The thin nappes were considered to have undergone gravity-driven northwards sliding from an advancing deformation front. Selwood and Thomas (1986a,b) developed the work of Isaac *et al.* (1982) and Selwood *et al.* (1985) by applying a thin-skinned nappe model to all of North Cornwall and Central SW England. The inland terrane was thought to comprise northward transported nappes, some emplaced by gravity sliding, and much of the interpretation depended on revised, and palaeontologically supported, stratigraphy. However, Warr (1991) noted that the restoration of the supposed nappe sequences was achieved by employing facies analysis and restoring structural units back to an assumed simple basin model, an unsafe assumption given the complexity of deformation in central S.W. England (see below). The new data led Selwood and Thomas (1986a,b) to reject the inland projection of a facing confrontation in the Trevone Basin (detailed later).

Coward & McClay (1983) applied the idea of thin-skinned foreland folding and thrusting, floored by a regional décollement, to the exposures of south Devon and redefined the Dartmouth Antiform (Hobson, 1976a) as a thrust structure on the basis of revised younging and bedding/cleavage data. Although Chapman *et al.* (1984), concurred with this interpretation, Smythe (1984) questioned some of the field observations made at critical localities. The thrusting models proposed that many of the folds and cleavages observed could be related to the propagation of thrusts, a

suggestion that might cast some doubt on the precise regional correlation of cleavages and folds. Coward and McClay (1983) also recognised a zone of large-scale secondary backfolding immediately to the north of the Start Complex and suggested that this structure, which may be traced westwards into Cornwall (Shackleton *et al.* 1982), had imposed an E-W grain on the tectonics of South Devon. Using seismic data, Leveridge *et al.* (1984) reinterpreted the structure of South Cornwall, and in particular the Gramscatho successions, in terms of major system of thrust nappes (section 2.2.2), floored by a décollement at a depth of ~13km.

Coward and Smallwood (1984) summarised the structures of SW England, including a description of backthrusts in the Tintagel High Strain Zone which were thought to allow detachment and backfolding along the southern margin of the Culm Basin. Zones of anomalous E-W lineations were noted, including the area immediately to the north of the Start Complex, and models invoking differential thrust sheet displacement and rotation were proposed to account for such oblique lineations. Whilst these models, allied to palaeomagnetic data, reasonably account for the lineations seen in Pembrokeshire, their application in South Devon and Cornwall seems somewhat more tenuous (see 4.6.1). Coward and Smallwood (1984) also noted that the orientation of the NW-SE dextral wrench faults (Dearman, 1963) approximately parallels the Variscan transport direction and, barring coincidence, it was suggested that the faults could have originally developed as lateral ramps during Variscan thrusting. Finally, the occurrence of late orogenic extensional faults was noted and analogies drawn to similar structures in the Basin and Range Province in the USA.

A set of thematic papers on the Variscan Evolution of SW England was published by the Geological Society in 1986, with many papers continuing to advocate thin-skinned models for SW England (e.g. Holder & Leveridge, 1986b; Selwood & Thomas, 1986a,b; Whalley & Lloyd, 1986; Williams & Chapman, 1986).

Amongst these contributions, those of Selwood and Thomas (1986a,b) fuelled a controversy surrounding the existence of a facing confrontation, where the transport directions of the same deformation event (D_1) oppose each other, in the Padstow area (Roberts & Sanderson, 1971). Termed the 'Padstow Facing Confrontation' by Beese (1982, 1984), this zone has been much studied, both in terms of structural zonation (Sanderson & Dearman, 1973) and in tectonic modelling (Shackleton *et al.* 1982; Sanderson, 1984; Coward & Smallwood, 1984; Selwood & Thomas, 1986a,b; Andrews *et al.* 1988; Pamplin, 1990). Selwood and Thomas (1986a,b) suggested that north facing and northwards transport was dominant and downgrade the significance of the Padstow Facing Confrontation on the grounds that D_{1n} (southerly directed thrusting) was a minor event of only local significance. However, Andrews *et al.*

(1988), Pamplin and Andrews (1988) and Seago and Chapman (1988) confirmed the existence of a 2km wide confrontation between Polzeath and Daymer Bay, by identifying large areas of inverted and southerly facing strata. Andrews *et al.* (1988) suggested that the facing confrontation developed in response to southward directed backthrusting, as movement on the basal décollement was temporarily halted. Renewed northwards transport was thought to have occurred during the Upper Carboniferous along a ductile shear zone, the surface expression of which is termed the Tintagel High Strain Zone (Sanderson & Dearman, 1973; Andrews *et al.* 1988). Ultimately thrust stacking above a lateral thrust ramp formed the Davidstow antiform, with simultaneous emplacement of the Culm sediments over the Trevone Basin along the Rusey Fault Zone, which was interpreted as the roof thrust to a large-scale pop-up structure. Pamplin (1990) argued that backthrusting could not generate a laterally extensive yet narrow zone of confrontation and instead suggested that the Padstow Facing Confrontation developed as a D_1 structure by basin inversion of the Trevone (D_{1s}) and Culm (D_{1n}) Basins. Warr (1991) also attributed the facing confrontation to the fundamental control exerted by pre-existing extensional structures during inversion of the Trevone Basin (see below).

Another somewhat less contentious issue concerns the tectonic evolution and inversion of the Gramscatho Basin. Two contrasting larger scale models for the evolution of South Cornwall were proposed by Holder and Leveridge (1986a) and Barnes and Andrews (1986), although both were essentially based on the same geological data. The former model suggested that the Lizard Complex marks the site of the Main Rhenic Ocean suture and proposed that the Gramscatho flysch developed in a foreland basin sourced by the Normannian High, a large uplifted nappe of crystalline basement. The latter model further developed the ideas of Badham (1982) and Sanderson (1984) in suggesting that the Gramscatho basin was of limited extent and formed due to localised crustal thinning within a major dextral strike-slip fault system. Subsequent continental collision, due to the closure of a major ocean basin south of Armorica (Fig. 2.14), caused the observed foreland thrusting and obduction of the Lizard Complex. Both of these models are an extension of the idea put forward by Leeder (1982), who proposed that the SW England basins owe their location and origin to back-arc continental 'oceanization' processes, caused by the Ligurian Subduction (Fig. 2.14).

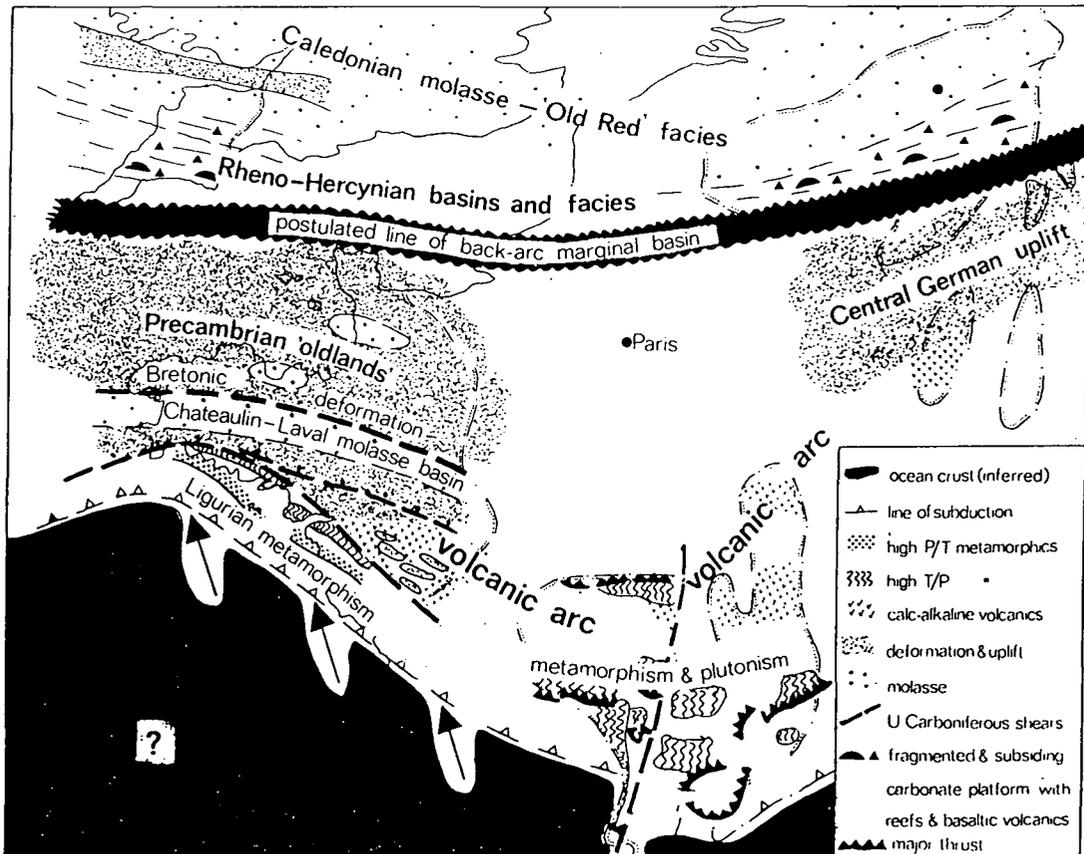


Figure 2.14 A 'radical' palaeogeography for Upper Devonian times in NW Europe, to show Liguro/Bretonic subduction, metamorphic belts, calc-alkaline volcanic arc, core of Precambrian 'oldlands' and Rhenohercynian back-arc basins. After Leeder (1982).

Thick-Skinned/Strike-Slip Models

Sanderson (1984), in a significant review paper, envisaged a more thick-skinned style of deformation involving most of the crust. He presented a dextral transpressive model for the tectonic development of SW England and Ireland in which basins arose as local pull-aparts within a strike-slip regime. The basins, which initiated at different times across the area, were inverted by a northward migrating deformation front. Sanderson (1984) suggested that deformation was characteristically heterogeneous and concentrated within basins where the crust was thinnest and the heat flow was the highest. This model, with revised estimates of late orogenic extension, showed that the Cornubian granites could be generated by lower crustal melting below a thickened crust (*cf.* the model of Shackleton *et al.* 1982 in which a sheet-like granite body is injected northwards during thrusting). The estimated 40% of shortening could be accounted for by the thrusting and underplating of oceanic and thinned continental crust, combined with dextral shear. The Sanderson (1984) model is compatible with the dextral strike-slip tectonic models of Badham and Halls (1975) and Badham (1982), and is consistent with the oblique collision of the European and African plates (Scotese *et al.* 1979). A strike-slip model appears satisfactory for the Gramscatho

Basin, where rift basins and oceanic crust may have arisen as local pull-aparts associated with dextral strike-slip faulting (Badham, 1982). While this model would also explain the localised nature of the Lizard rocks and the NW-SE trending basalt dykes, it is not so applicable to the rest of SW England, where the overall basin structure implies that rifting was controlled predominantly by N-S extension, although a component of strike-slip may have been implicit in the rifting phase (Warr, 1991). Attempts by Sanderson (1984) to model the Culm Basin as a rift basin, in which sedimentation initiated in the Upper Devonian and continued into the Westphalian, are refuted by Warr (1991) who, using stratigraphic interpretations, showed that the Culm Basin developed during the Namurian as a foreland basin whose position was strongly influenced by previous basin inversion (see below). A similar model for foreland basin evolution and inversion is presented by Gayer and Jones (1989), and Hartley and Warr (1990), for the South Wales Basin.

Turner (1986), using the field data of the Exeter University mapping contracts (1971-1985), presented a thick-skinned model for central SW England, in which he proposed that significant strike-parallel variations in stratigraphy and structure were indicative of transpression/transension of a compartmentalised basement. He postulated that such variations occurred across deep-seated NNW-SSE Hercynian fault zones, along which differential vertical movements affected both sedimentation (basin formation) and structural evolution. In his model, Turner (1986) rejected the concepts of a regional décollement, extensive translation and regional backthrusting. Warr (1991), however, noted that there is no relationship between sedimentation and structure and the NNW-SSE fault zones, and goes on to suggest that, while the concept of reactivation of basement fractures is plausible, there is no evidence to suggest that the NNW-SSE fault zones were reactivated basin-controlling faults. The deformation style of north Cornwall is characteristic of a foreland fold and thrust belt with shortening of at least 50%, locally even approaching 70% (Warr, 1991).

Basin Inversion Models

Many of the previous regional structural models for the Rhenohercynian of SW England highlighted thin-skinned thrust processes (see above) without considering the influence of a pre-existing extensional basin development. In recent years, the concept of basin inversion has been increasingly applied to the restacking of highly stretched continental margins within orogenic zones, with examples from the Alps (Butler, 1989; Hayward & Graham, 1989), the Subvariscan foldbelt of South Wales (Gayer & Jones, 1989; Powell, 1989) and the Rhenohercynian of SW England (Pamplin, 1988; Hartley & Warr, 1990; Warr, 1991). Warr (1991) proposed that the complications in the

regional pattern of SW England, such as early phases of backthrusting, confrontation zones and lines of dextral transpression, resulted from controls exerted by the original basin architecture, especially the location and geometry of pre-existing extensional faults. The development of intense slaty cleavage, isoclinal folding and thrusting indicate that significant strains and high degrees of crustal shortening were achieved during Variscan orogenesis. Many workers adopted a thin-skinned approach to accommodate such strains at depth, showing that deformation occurred above a shallow regional decollement. However, Warr (1991) noted that basin inversion can accommodate considerable shortening (by more internal strain in the syn-rift strata), without necessarily requiring major displacements between individual nappes (*cf.* the models of Selwood & Thomas, 1986a,b). Whether a thin-skinned approach is consistent with the basin inversion mechanism remains unclear, because of the lack of structural constraint at depth (Warr, 1991). It appears, however, that basin inversion is more compatible with the thicker-skinned model proposed by Sanderson (1984).

Pamplin (1988) modelled a progressive inversion of rift basins from south to north, and attempted to detail the observed patterns of folding and thrusting by the reactivation of a series of listric basin-forming extensional faults. However, evidence for basin inversion by fault reactivation is lacking as no thrust fault shows signs of having previously been a reactivated extensional fault (Warr, 1991). Warr (1991), studying the inversion of the Trevone and Culm Basins, concurred with Pamplins (1988) observation that pre-existing structural architecture exerted a fundamental control on the tectonic evolution of SW England basins, and went on to suggest that steep, basin-controlling faults at the northern margin of the Trevone Basin acted as a buttress to the northward advancing deformation. This buttressing is used to explain the D₁ backthrusting event and the formation of the 'intra-basinal' confrontation zone at Padstow. The assertion by Warr (1991) that the relative obliquity of the pre-existing extensional framework (E-W) to the NNW regional compression adds further complications to the pattern of deformation, is an extension of the dextral transpression basin inversion model proposed by Holdsworth (1989a) to account for a linear zone of anomalous structure which runs E-W through south Devon and north Cornwall (the Start-Perranporth Zone; see 5.6.1).

More recently the theme was continued by Fleitlitz (1992) who described a shear zone within the northwest central Rhenohercynian slate belts (the Monschau shear zone) in which the pre-existing basin architecture exerted an influence on the cover sequence during Variscan transpressive inversion. Similarly, Chadwick (1993) was able to demonstrate the important role of previous extensional faults during the dextral transpressive inversion of the Wessex Basin in mid-Tertiary times.

2.4 Geophysical evidence and its bearing on tectonic models

Much geophysical evidence has been acquired around SW England and is comprehensively reviewed by Brooks *et al.* (1984). Early magnetic investigations showed that the Lizard Complex was of limited areal extent, probably not much greater than its onshore outcrop. This lent tenuous support to the Lizard-Dodman-Start Thrust concept since it implied that the Lizard Complex could be a small thrust slice rather than a regionally extensive igneous body. Gravity surveys (e.g. Bott *et al.* 1958) confirmed that the individual granite plutons of SW England were linked at depth, with subsequent work showing that this batholith extended to the Scilly Isles.

However, it is offshore deep seismic refraction/reflection data which has contributed the most to the understanding of Variscan tectonics. The early onshore work of Kenolty *et al.* (1981) and Chadwick *et al.* (1983) indicated major detachments beneath Southern Britain. Subsequent work has also imaged these reflectors (Brooks *et al.* 1984; Donato, 1988; Jones, 1991). A large amount of seismic data, from both commercial sources and BIRPS profiles, is now available for the offshore regions of SW England and has recently been synthesised by Evans (1990). Offshore seismic results from the Plymouth Bay and Western Approaches area have been presented by Leveridge *et al.* (1984), BIRPS and ECORS (1986) and Edmonds *et al.* (1989). All these authors emphasised the importance of southerly dipping reflectors, which were interpreted as Variscan thrusts and apparently had the same orientation as structures proposed from field mapping. Leveridge *et al.* (1984) used the offshore data to infer the presence of the Carrick Thrust (see 2.2.2), which was shown as the roof thrust in a major thrust stack involving the Carrick, Dodman, Lizard Boundary and Normannian nappes. Recent onshore stratigraphic revisions (Le Gall *et al.* 1985) suggested that the boundary between the Frasnian Portscatho Formation and the Eifelian Pendower Formation might also be a thrust (section 2.2.2). Offshore seismic evidence used to support this proposal was presented by Edmonds *et al.* (1989) and the Veryan Thrust was invoked. Whilst the position of these detachments in offshore seismic sections is generally agreed, their onshore expression is debatable (5.4.3). Holdsworth (1989a) noted that much of the offshore seismic data is of limited use as it does not image steeply dipping or sub-vertical structures, this clearly compromising some tectonic models. Le Gall (1990) extrapolated SWAT profiles inland and was able to correlate reflectors with known geological features, e.g. the Lizard ophiolitic suture and the Tintagel backthrust zone, although some of the interpretations were subsequently questioned (Brooks & Le Gall, 1992). Variscan thrusting in the south appeared to have involved most of the continental basement, flooring into a deep infra-crustal decoupling zone not imaged previously on SWAT profiles due to its depth (~25km). Northward dipping reflectors below the Culm Basin were tentatively considered to

represent older south verging Caledonian thrusts (*cf.* Brooks & Le Gall, 1992), whilst the pre-existing extensional basin pattern, inferred from deep seismic data, was shown to have highly influenced the geometry of thrusting (Le Gall, 1990, 1991).

The development of Permian and Mesozoic basins in the offshore regions surrounding SW England has a complex history, involving successive episodes of rifting and inversion. Many studies describe how earlier Variscan thrusts and strike-slip faults have been reactivated during basin development (e.g. Zeigler, 1987; Chapman, 1989; Hillis & Chapman, 1992).

2.5 Summary

Upper Palaeozoic sedimentation in SW Britain initiated on continental crust that formed the southern margin of eastern Avalonia. The oldest sediments observed in SW England are the alluvial deposits of the Gedinnian-Siegenian Dartmouth Group. In the absence of palaeocurrent data the succession is presumed to be northerly derived on the basis of its similarity with the Old Red Sandstone sequences of South Wales. During the Siegenian-Emsian there was a marine transgression across much of SW England, and the shallow marine siliciclastic and carbonate sequences of the Meadfoot Group were deposited. In the upper part of the succession, the Staddon Formation provides evidence of deltaic incursions, sourced from periodic local intrashelf uplifts. By the Eifelian, differential subsidence, possibly related to extension on deep seated E-W basement faults, had initiated and allowed the development of separate depositional basins that were separated by rises. The Gramscatho Basin developed seaward of this fragmenting shelf and had a limited clastic input from the tectonically emergent Normannian High in the south. Sediment input to the Trevone and South Devon Basins was reduced and mudstone dominated. Carbonate platforms developed on intrabasinal volcanic highs and contributed limestone turbidites and debris flows to the basinal areas. Widespread basic volcanicity occurred during the Middle Devonian and had intraplate characteristics. Continued extension in the Gramscatho Basin led to the formation of oceanic crust. In North Devon, by contrast, the Middle Devonian is marked by a major clastic influx from Wales and, possibly, the Bristol Channel Area.

During the Upper Devonian, further uplift of the Normannian High allowed enhanced clastic input to the Gramscatho Basin. Reef complexes in the basins to the north were submerged and there was renewed volcanicity. The youngest sediments recorded in the southerly basins are of Late Fammenian age, and are coeval with deformation associated with the overthrusting of the Normannian High and obduction of the Lizard Complex. A 'deformation front' migrated northwards through the

peninsula, with the downflexing of the lithosphere ahead of this margin producing both partial melting at depth, and the formation of the Culm and South Wales Basins. Inversion of the older basins was associated with a thick/thin-skinned tectonic regime involving northward directed thrusting and nappe emplacement. Significant local structural complexities arose due to the interaction between tectonic shortening and the pre-existing basinal architecture. Continued shortening, with both northward and southward directed shear, during the Upper Carboniferous led to the inversion of the Culm sediments. By latest Carboniferous there is evidence of lamprophyric volcanism; granite intrusion occurred and Permian basins began to develop, in part controlled by earlier Variscan structure.

CHAPTER 3

THE LITHOLOGY AND GEOCHEMISTRY OF THE START COMPLEX AND ADJACENT REGIONS, SOUTH DEVON

3.1 Introduction

The Start Complex forms a small (*c.*45 sq. km), isolated outcrop of distinctive lithologies at the southernmost end of the South Hams peninsula of south Devon. The northern limit of the Start Complex extends from Outer Hope village in the west, across the Kingsbridge estuary, to Greenstraight in the east (see Fig. 4.1*). Exposure is largely confined to the rugged coastal cliffs, whilst inland exposures are mostly limited to occasional crags, stream sections, badly weathered quarries and roadcuts. Access to the coastal exposures is generally excellent, although some cliff sections were inaccessible, even by boat.

Mapping was carried out, on scales of 1:5,000 and 1:2,500, during 14 weeks in the summer of 1991. Approximately 3 weeks of this was assigned to the mapping of the poorly exposed inland areas of the Start Complex, principally from the Kingsbridge estuary sections and occasional crags and tors. The Devonian succession to the north of the Start Complex was mapped northwards until the primary mineral stretching lineation switched from its along strike orientation to the more usual regional downdip direction (see 4.4.3, 4.4.4). This switch generally coincided with the first outcrops of the Dartmouth Beds, and thus these lithologies received little attention.

Lithologically, the Start Complex comprises an interlayered sequence of highly deformed schistose pelites ('greyschists') and extensive horizons of basic metavolcanics ('greenschists'), whose outcrop pattern broadly defines a large west plunging and closing antiform (see 4.2.6; Fig. 4.1). The origin and age of the Complex is uncertain, although it is often assumed to be Devonian. Early workers considered the Start Complex to be an exposed on-shore remnant of the pre-Devonian basement. De la Beche (1839) suggested that the schists, whilst sharing some similarities with the Devonian slates, were "...more ancient.." (p36). Subsequent work attempted to resolve the question of whether the Start Boundary Fault was a distinct lineament, separating rocks of different ages, or whether it was a zone of progressive metamorphism, thus

* Locality maps are all contained in Chapter 4

implying that the Start Complex was the metamorphic equivalent to the Devonian sequence. Bonney (1884) rejected the concept of progressive metamorphism, and advocated an Archean age, relating the Start Complex to the Lizard and Eddystone outcrops. Somervail (1887), however, suggested a link between the greenschists and the 'Greenstones' of Dartmoor, implying a Devonian age for the former. Hunt (1892), noting petrographic similarities between the schists and the Devonian slates, concurred with Somervail's (1887) observations. The Survey memoir (Ussher, 1904) objectively reviewed the previous literature, but did little to resolve the question of the age of the Start Complex. J.J.H. Teall concluded, in the Survey memoir (Ussher, 1904), that the Start Complex represented "...a series of arenaceous and argillaceous deposits with which a considerable amount of basic material was associated, and...can be matched by rocks occurring in the Devonian area though the igneous and sedimentary types do not occur in the same relative proportions." (p.45).

Tilley (1923) provided a comprehensive petrographic study of the schists and, based on their higher degree of metamorphism, suggested a Precambrian age for the Complex. Hendriks (1939) linked the Start Complex to the Lizard and Dodman peninsulas along a single thrust, flooring a Variscan nappe. She suggested a Palaeozoic age for this nappe sequence, but noted that the Start Complex lacked Silurian rocks. However, she later (Hendriks, 1959) further constrained her stratigraphic interpretation of the Start Complex and advocated a Devonian age for these schists. Marshall (1965) tentatively suggested that the schists were Devonian in age, citing evidence from heavy mineral analyses, and was certain that "...the schists and slates were derived from similar metamorphic terranes and deposited in similar sedimentary environments..." (p.86).

Much of the subsequent work (e.g. Dodson & Rex, 1971; Hobson, 1977; Coward & McClay, 1983) has served to confirm the Devonian age of the Start Complex, although Doody and Brooks (1986), in interpreting recent seismic refraction data, linked the Start Complex to the Eddystone reef along a shallow refractor, implying a pre-Devonian age for the former (see 2.2.1). This interpretation has subsequently been questioned (Leveridge *et al.* 1984; Holder & Leveridge, 1986a) and is no longer considered valid. Clearly the age of the Start Complex has yet to be unequivocally established, although the recent geochemical analyses of the greenschists, which show close chemical comparisons to the basalts of the Lizard ophiolite, suggest that they may be of similar age, i.e. ~375Ma (Floyd *et al.* 1993a; see 3.2.1).

Petrographic similarities between the Start schists and the Devonian shales have long been understood (e.g. De la Beche, 1839; Hunt, 1892; Ussher, 1904), with the differences in the field appearance of the two groups being attributed to differing

degrees of metamorphism. The degree of mineralogical segregation, and the local development of garnet in some of the schists, suggests that the Start rocks are of a higher metamorphic grade compared to the Devonian slates (Meadfoot Group) which outcrop immediately to the north (see 4.4.5). They share a common deformation history with the Devonian strata, except that the metamorphic segregation fabric in the Start Complex predates the first phase of folds common to both areas (see 4.2). Separating these two units is the Start Boundary Fault (SBF), a brittle normal fault, which coincides with an earlier steeply dipping zone of high ductile strain, characterised by a series of phyllonitic and mylonitic strata (see 4.3).

The aim of this chapter is to offer a study of the sedimentology, geochemistry and general tectonic setting of the Start Complex schists, allied to a brief stratigraphic overview of the adjacent Devonian strata.

3.2 Geochemistry and Petrology of the Start Complex

3.2.1 The Greenschists

Introduction

The first detailed descriptions of the greenschists came from Bonney (1884), who compared them to 'greenstone tuffs' or even highly altered basalts. He concluded that these chloritic schists were essentially fine-grained tuffs and that the banded appearance was indicative of bedding. Bonney (1884) also offered the first petrographic accounts of the greenschists, giving the mineralogy of these as: chlorite (which he considered as diagnostic of this lithology), quartz, epidote, some feldspar, possible kyanite, and subordinate amounts of calcite and iron oxides. Raisin (1887), enlarging on Bonney's (1884) work, noted the presence of hornblende in some examples and the occurrence of an "exceptional" intermediate lithology, containing both mica (considered as diagnostic of the greyschists by Bonney (1884)), and chlorite.

Ussher (1904), in describing the two strikingly different lithologies in the Start Complex, suggested that the greenschists "...may unhesitatingly be regarded as an altered series of basic igneous rocks allied to the diabases in composition and possibly consisting in part of altered tuffs" (p37). He distinguished these from the greyschists on the basis of their "...almost universal green colour" and also noted that, when highly oxidised, the greenschists varied in colour from pale grey/green, through pale yellow to red and brown.

Tilley (1923) subdivided the greenschists into two distinct types based on petrography: a) chlorite-epidote-albite schists, and b) hornblende-epidote-albite schists. He suggested that these subdivisions represented different grades of metamorphism of

the same rock type. Changes in the plagioclase composition, as well as decreases in the amount of epidote, calcite and chlorite, are offered by Tilley (1923) as supporting evidence. He noted that no zonal distribution of the two types occurred, and that varieties intermediate between them were also common.

Tilley (1923) also noted the presence in some greenschist sequences of large unzoned nodular masses, ranging in size from 2.5-15cm and composed principally of epidote, with lesser amounts of chlorite, calcite and albite. These he interpreted as metamorphosed amygdales from the parent rock. He indicated that the greenschists were derived from basic igneous protoliths, suggesting that they might have been a series of lava flows and sills (see below).

Field Appearance and Relationships

Several well exposed coastal sections of the greenschist were examined in an attempt to elucidate their primary geological relationships and origins. The exposures at Rickham Common [SX 2737, 0375], Prawle Point [SX 2772, 0350] and Gammon Head [SX 2766, 0355] provide both the best exposed and apparently least deformed sequences of greenschist. In these 'low strain' areas, the greenschists are finely banded with pale and deep olive green layers, varying from centimetre down to millimetre in scale (Plate 3.1). The darker layers are mica-rich, and can impart a platy phyllitic appearance to freshly broken surfaces of the greenschist.

The original thickness of the greenschists is difficult to assess due to subsequent tectonic strains, although they are likely to have been several tens or hundreds of metres thick. Their present thicknesses are given in 4.2.1. On the basis of their apparent thickness, lateral persistence and uniform mineralogical composition, it is assumed that the greenschists originally represented a series of basaltic lava flows and/or volcanoclastics (rather than intrusives). Although these features are not diagnostic of a metavolcanic sequence, comparisons with texturally similar rocks from the Lizard Complex (hornblende schists, foliated metagabbros) led Floyd *et al.* (1993a) to suggest that coarse-grained intrusive rocks, or finer grained dykes, were not the main initial component of the greenschists. Floyd *et al.* (1993b) note that the delicate nature of the fine laminations and the rapid changes in lithology, especially in low strain areas, strongly suggests that much of the sequence was composed of basaltic volcanoclastics (i.e. tuffs) rather than lavas. They go on, however, to make the tenuous observation that, although curved pillow lava surfaces and other typical lava features are lacking, the variability in terms of the gross banding in some sequences could reflect original differences between lava flows or even sills, all now heavily sheared to a degree of schistose uniformity.

The best area for examining the primary relationships and fabrics is the coastal exposure along Rickham Common [SX 2737, 0375], where early folds are picked out by a fine interlayering of the tuffaceous greenschist. Although this layering in places looks primary (i.e. like an original volcanoclastic layering), close examination reveals fine quartz segregations parallel to fold limbs and quartz rodding in the fold hinges. It appears, therefore, that field evidence for the primary texture of the precursor to greenschists is lacking, due to the extensive recrystallization during later deformation. Boundaries between the greenschists and greyschists are concordant and in places appear to be quite gradational, especially along the coast at Rickham. Despite the clear evidence of primary interbanding and interfolding along here, the original relationships between the two lithologies are impossible to assess given the extensive later recrystallization.

In 'high strain' areas, the greenschists are massive, structureless, have a very uniform, pale green colour, and are typically devoid of folds, probably as a consequence of the increased deformation. Folds, where seen, are defined by thick, crenulated quartz segregation within a fine granular matrix.

The greenschist layering in places is disrupted by epidote pods and segregations, some up to 15cm across, which may be associated with fine quartz rods and segregations (Plate 3.1). This segregation fabric is clearly seen to wrap around the pods, some of which are flattened parallel to the banding. The origin of these structures is still uncertain. Amygdales up to 30 centimetres across have been recorded in basaltic lava flows in Mull (Kerr, 1993), their exceptional size being the result of in situ growth during metamorphism. These unusual amygdales are characterised by coarse epidotic rims and cores of zeolite. Thus the large size of the epidote pods in the greenschists does not preclude them from having originally been amygdales, although Floyd *et al.* (1993b) suggest that it is unlikely that the infill of lava vesicles would retain its shape, given the degree of deformation suffered by these greenschists. Alternatively, it is possible that these pods are the remnants of volcanic bombs within a volcanoclastic sequence, which disrupted the essentially homogeneous tuffs to produce a concentration of strain and the subsequent preferential growth of epidote. Thus these epidote pods would appear to be evidence of element migration during metamorphism, prior to the superposition of the main tectonic fabric.

Some of the greenschist exposures are rich in pyrite (now probably limonite pseudomorphed after pyrite), which may be up to one centimetre across and are weathered out to stand in relief on the rock surface (Plate 3.2). Typically, they are aligned parallel to the mineral stretching lineation and many show well developed quartz pressure shadows (see Plate 4.14; 4.2.4), although the individual pyrite crystals occasionally grow along the foliation planes as flat sheet-like bodies and show no

alignment whatsoever. A few more massive exposures of greenschist exhibit large (up to 5mm) albite porphyroblasts, which once again are emphasised by weathering, and seem to be aligned parallel to the stretching lineation.

Many of the greenschists contain significant amounts of calcite due to the breakdown of calcic plagioclase (this is especially common adjacent to fault zones, see 4.3.2, 4.3.4), and this results in a characteristic 'carious' weathered surface.

Petrology

The greenschists are fine to medium-grained and are characteristically highly schistose, ranging from finely laminated to irregularly banded, the banding being defined by variations in mineralogy and grain size. Petrographically, the greenschists are relatively simple with two main mineral assemblages present: a) an initial metamorphic assemblage of amphibole-epidote-albite, and b) a less common retrograde assemblage with abundant chlorite, muscovite and sphene, together with lesser proportions of the initial minerals. In both cases, minor accessory quartz, carbonate, pyrite and iron oxides may also be present. As the amphibole is predominantly actinolite and no pumpellyite has been observed, the initial assemblage may be considered to be low grade greenschist facies.

The initial assemblage typically comprises prismatic and fibrous actinolite, which is colourless, or a very pale yellow/green with a very weak pleochroism. Associated with the actinolite are small, yellow, anhedral epidotes. The fine lamination seen in hand specimen is due to layers of the actinolite/epidote alternating with bands of anhedral poikiloblastic albite (Plate 3.3). With increased deformation related recrystallization, the segregation banding is accentuated. In the more irregularly foliated greenschists, both amphibole and epidote may be present as large prismatic subhedral crystals. Some of these large pleochroic epidotes show a simple zonation with concentric cores and subhedral rims.

The retrograde assemblage displays variable replacement of the original colourless actinolite by green, strongly pleochroic chlorite, so that some of the remnant actinolites exhibit a patchy bluish tinge due to incipient chloritisation. Associated with the chlorite-rich laminae are sphene granules, minor quartz, carbonate and some muscovite. In many of the laminated greenschists, rapid changes in lithology are common, with intimately interbanded chlorite-rich, epidote-rich and muscovite-rich assemblages. Some layers are especially abundant in pyrite, with adjacent bands being extensively stained by Fe oxide (Plate 3.4).



Plate 3.1 Segregation layering in the metavolcanics defined by epidote-rich and hornblende-rich horizons. Large, slightly flattened, epidote pods and fine quartz rods disrupt the layering. Greenschists cliffs at the Hipples [SX 2738, 0377].

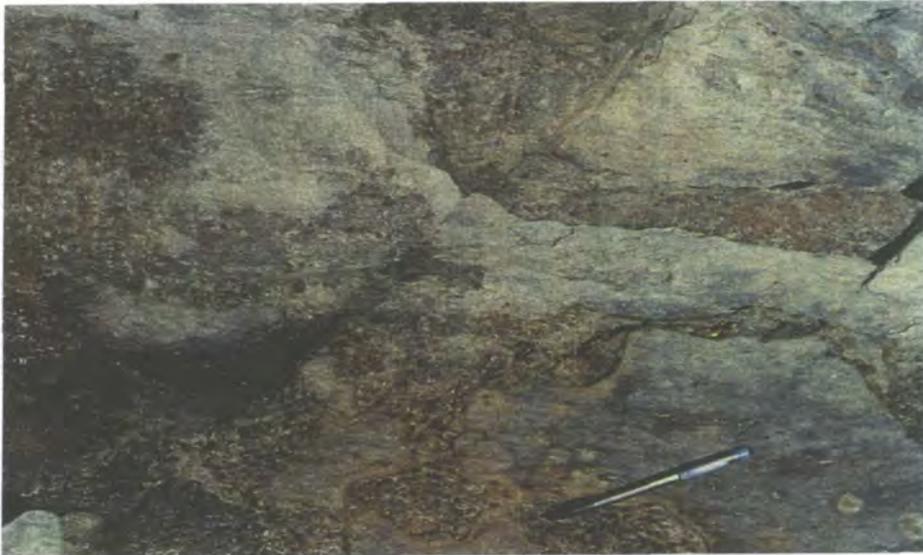


Plate 3.2 Foliation surface in greenschist containing aggregates of highly weathered pyrite (or limonite), some of which are elongated parallel to the mineral stretching lineation. Greenschists south of Black Cove [SX 2770, 0353]



Plate 3.3 Thin section of greenschist, showing the initial assemblage of fibrous actinolite dotted with small, anhedral epidotes. The schistose foliation is defined by aligned actinolites giving a wavy lamination and the alternation between these layers and quartz/albite layers. This foliation is cut by weak crenulating fabric orientated ~E-W across the section. Incipient chloritization is evident as green pleochroic patches in PP in many of the actinolites. Greenschist from Portlemouth, [SX 2741, 0371], under X4 magnification and crossed polars. FoV= 6x3.7mm.

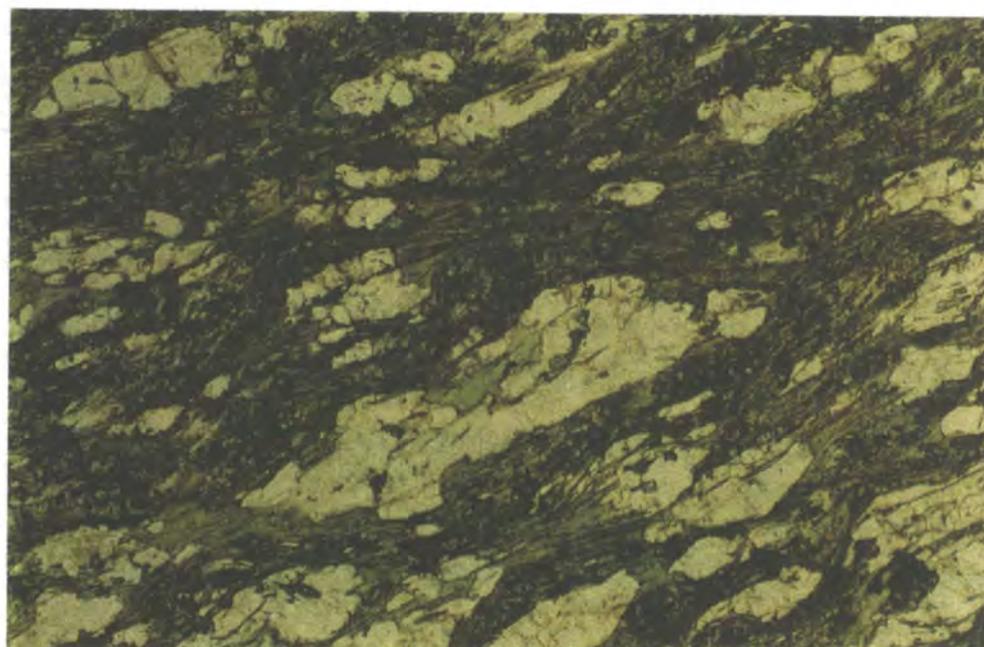


Plate 3.4 Thin section of greenschist, showing the retrograde assemblage of green, strongly pleochroic chlorite, muscovite and remnant primary mineralogy (see Plate 3.3). Foliation is enhanced by segregation of quartz and albite into distinct horizons. Greenschist from Portlemouth, [SX 2741, 0371] under X4 magnification and plane polars. FoV= 6x3.7mm.

Geochemistry

Until recently, the only geochemical studies of the Start greenschists were the major element analysis of Tilley (1923). Samples of each of the chlorite-epidote-albite schists and the hornblende-epidote-albite schists were analysed. The results of both groups revealed an essentially basaltic composition, with the latter group showing affinities to the Hornblende Schists of the Lizard peninsula.

A more comprehensive study of the Start greenschists was carried out by Floyd *et al.* (1993a), who analysed representative samples of the main greenschist outcrops using standard X-ray fluorescence spectrometry (XRF: major oxides and most trace elements), inductively coupled plasma source spectrometry (ICP: rare earth elements) and instrumental neutron activation analysis (INAA: Hf, Sc, Th, Ta & U).

Samples for the study were collected from greenschist localities around the Complex by P.A.Floyd and R.E.Holdsworth. These were then divided into four broad geographical/stratigraphical groups, although the basis for this is now open to question in light of the recent reinterpretation of the structural setting of the greenschists within the Start Complex (Fig. 3.1b; 4.2.6).

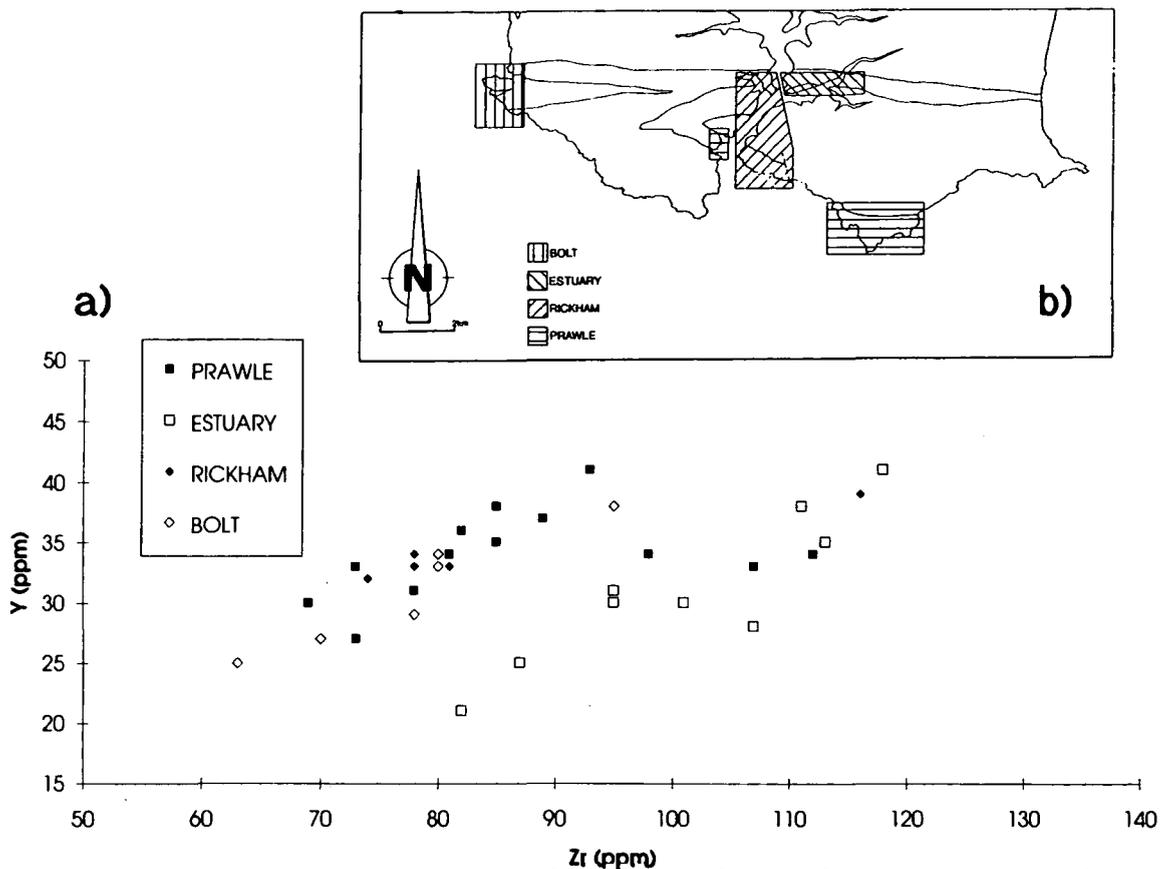


Figure 3.1 (a) Y-Zr plot of the Start greenschists. Samples from the Salcombe estuary area plot as a distinct group with lower Y-Zr ratios. (b) Four geographical/stratigraphical groups defined by P.A. Floyd and R.E. Holdsworth during sampling.

If the greenschists were originally lavas, the initial chemical alteration was probably due to submarine weathering. Indeed, the earliest fabric may be a consequence of submarine deformation (see below). Some of the adjacent metasediments do seem to indicate a deep marine environment. However, the main cause of element mobility is most likely to have been the pervasive regional metamorphism and locally intense high strain.

Floyd *et al.* (1993a) note that, although the absolute abundances of many of the major oxides cannot be relied on to determine the initial composition, the overall bulk chemistry indicates that the greenschists were a restricted series of basic magmatic rocks. Because of the highly metamorphosed nature of the greenschists, the interrelationships of the immobile elements and their ratios are a more reliable means of determining their primary features. The greenschists are characterised by: a) low (depleted) contents and a limited range of incompatible element abundances (TiO_2 1.0-1.8 wt %; Zr 60-110 ppm; Nb 3-5 ppm), coupled with very high Zr/Nb ratios (*c.* 30) and low Th/Hf ratios (*c.* 0.03); b) a low FeO/MgO ratio (1-2) and a small range of transitional element contents (Cr 200-500 ppm); and c) a range of chondrite normalised rare earth element patterns from light rare-earth-element-depleted to flat ($[\text{La}/\text{Yb}]_n = 0.4-1.0$). The geochemical data indicate that the greenschists are a relatively coherent suite of moderately differentiated tholeiites, with an overall depleted and primitive composition. Initial results, especially the variability of the Y/Zr and Zr/ TiO_2 ratios, indicated the possibility of two different chemical groups (P.A. Floyd, *pers. comm.* Fig. 3.1a), although there is no field evidence to suggest that the different trends on the Zr/Y plot represent two different igneous units (Fig. 3.1b). However, further modelling in the course of this study, using immobile element ratios, has shown that two distinct groups do appear to occur, with the greenschists from the Salcombe estuary area forming a distinct group. Since these samples are not spatially distinct from the other greenschist outcrops (see Fig. 3.1b) and appear to lie at the same structural level, the immobile element ratio variations can best be accounted for in terms of mantle heterogeneity, rather than an evolved mantle source. The proximity of these more distinct specimens to the Start Boundary Fault, which is thought to be the surface expression of a significant crustal fracture (see 4.5), is not likely to have influenced the immobile element ratios, since immobile element abundances are unaffected by metamorphism or hydrothermal alteration.

The most revealing data arises from the chemical discrimination diagrams for basaltic rocks e.g. the Hf-Ta-Th diagram (Wood, 1980). When the greenschists are plotted onto this diagram they are seen to exhibit the characteristics of spreading axes with a depleted or normal-type mid-ocean ridge basalt (N-MORB) chemistry (Fig. 3.2).

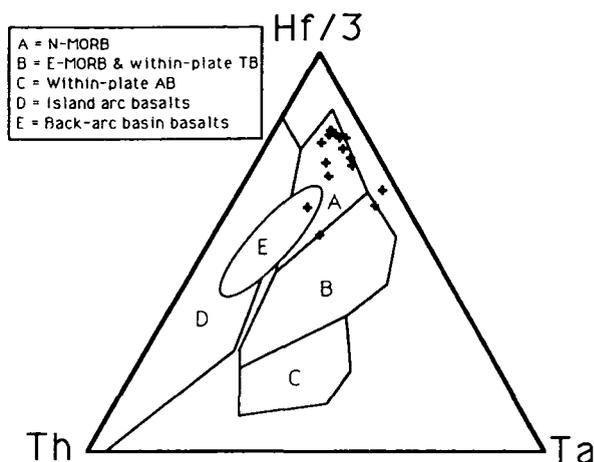


Figure 3.2 Hf-Ta-Th tectonomagmatic discrimination diagram (Wood, 1980), showing the concentration of the Start greenschists in the N-MORB field. After Floyd *et al.* (1993a).

Similarly a multi-element N-MORB normalised plot (Sun & McDonough, 1989) reveals a flat pattern (Fig. 3.3), with the exception of the large-ion-lithophile elements whose deviation from the unity line reflects their greater mobility during metamorphism. Floyd *et al.* (1993b) note that the depleted N-MORB character of the Start greenschists is significant in that they are much more primitive than any ocean-floor basalts recognised in S. Cornwall, and in this respect have some affinities with some of the Lizard dykes and the Landewednack hornblende schist.

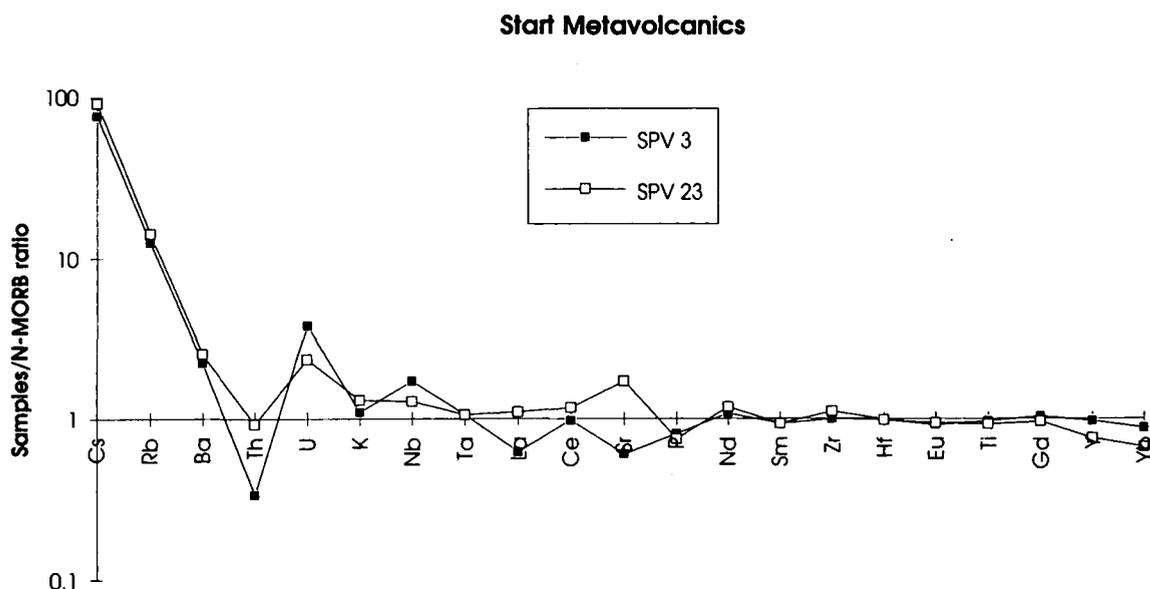


Figure 3.3 Representative N-MORB normalised (Sun & McDonough, 1989) multi-element pattern for the Start greenschists. Deviation from the unity line for some of the LIL elements reflects their greater mobility during metamorphism. After Floyd *et al.* (1993a).

Implications of the geochemical results

The basaltic geochemistry of S.W. England defines two major magmatic provinces, which are geographically distinct and separated by a broad E-W zone, which is approximately coincident with the northern margin of the Gramscatho Basin and the Start Complex. These provinces are: a) S. Cornwall, typified by tholeiitic submarine basalts of predominantly Lower Devonian age, and b) N. Cornwall and Devon, which is dominated by alkali basalts, ranging in age from Middle Devonian to Lower Carboniferous. The S. Cornish tholeiitic basaltic lavas display many primary chemical features akin to various types of MORB (Floyd, 1982, 1984) and it is the presence of these lavas, along with the Lizard ophiolite, which has been used to infer the presence of ocean crust flooring the Gramscatho Basin (Barnes, 1984; Holder & Leveridge, 1986a).

As shown above, the Start greenschists have predominantly N-MORB type geochemical characteristics, which link them with the Lizard dykes and the S. Cornish Landewednack hornblende schists. These characteristics suggest that both the Start greenschists and the Lizard ophiolite may represent oceanic crust of similar age and derivation, the former having a more restricted composition (Floyd *et al.* 1993a/b). Whether the greenschist evolved in the same basin as the S. Cornish basalts (Start-type schists may be traced towards Plymouth in offshore cores; see 3.2.4), or in two separate basins, cannot be decided using geochemical data alone (Floyd *et al.* 1993a). The highly deformed nature of these schists and the presence of a unique early foliation may be evidence of sub-oceanic ridge-related deformation, similar to that noted by Vearnecombe (1979) in the Landewednack Schists of the Lizard (Floyd *et al.* 1993a). The markedly oceanic character of the Start greenschists further emphasises the presence of two distinct magmatic provinces in S.W.England, and strongly suggests that the E-W divide between these two areas represents a significant, pre-Variscan tectonic feature.

3.2.2 The Greyschists

Introduction

Bonney (1884), who provided the first detailed description of the greyschists, noted that they dominate across the Start Complex, are mineralogically very uniform and are often highly deformed. He described the principal constituents as quartz, white mica, possibly chlorite and feldspar and an unnamed 'black mineral', whose colour was the result of abundant iron oxides and organic matter. He concluded that the greyschists formed a distinct stratigraphic unit overlying the greenschist. Raisin (1887) suggested that this black mineral was kyanite which grew in situ to enclose fine laminations of iron oxides, this imparting the dark colour.

In the Geological Memoir (Ussher, 1904), a number of petrographic descriptions of the greyschists confirm their compositional uniformity and show that the dominant mineralogy is quartz, muscovite, chlorite, and accessories. Ussher (1904) disagreed with Bonney's (1884) stratigraphic interpretation and suggested that the greyschists lay beneath the greenschists, accounting for the outcrop pattern in terms of a westward plunging antiform in the east closing towards a synform in the west.

Tilley (1923) divided the greyschists into two distinct stratigraphic/geographic groups, the Start schists (the oldest and structurally lowest) and the Bolt schists (structurally highest), separated by a single horizon of folded greenschist. These two groups are, however, lithologically and petrographically indistinguishable, and whilst they may conveniently subdivide the Start schists, they have no obvious stratigraphic basis. Tilley (1923) noted the presence of albite, and showed that the black mineral, described by earlier workers, was albite containing banded inclusions of dark organic matter. The organic matter is seen to clearly cut across albite twins and then pass laterally into adjacent chlorite/muscovite layers (see below). Tilley (1923) was also the first to note the presence of garnet, developed in the schist in association with chlorite and muscovite. He suggested that whilst garnet is relatively rare in the 'pure' greyschists, it is well developed in the intermediate or composite lithologies.

Field Appearance

The greyschists comprise a series of highly deformed pelites and semi-pelites and represent the dominant lithology within the Start Complex. The schists have a characteristic striped appearance due to alternate layers of white or pale grey quartz/albite and dark grey to black mica/chlorite, the individual layers ranging from 0.5 millimetres up to a few centimetres in thickness. In some areas, especially the east coast sections north of Start Point [SX 2818, 0386], the dark pelitic schists are interbanded with thin pale grey/brown horizons of psammite. However, no sedimentary structures or primary bedding structures are preserved.

Typically the greyschists contain numerous quartz segregation veins and lenticles, many of which have been extensively sheared to leave thick crenulated quartz rods and mullions, which weather markedly in relief of the surface (see Plates 4.1 & 4.6). The quartz veins, often containing significant amounts of albite, can reach 1-1.2m in thickness, and may be associated with extensive veins of chlorite (Plate 3.5).

Where the greyschists are devoid of quartz/albite veins, they take on a characteristic platy, phyllitic appearance, bearing a strong resemblance to examples of the more deformed Meadfoot Beds north of the Start Boundary Fault (see 4.4)



Plate 3.5 Large quartz/albite vein with layers and pods of chlorite. Greyschists at Bolt Head [SX 2725, 0362].



Plate 3.6 Greyschists covered in large (up to 2 cm across) euhedral crystals of pyrite, many of which have well developed asymmetric quartz pressure shadows. Hugh's Hole [SX 2678, 0388].

These greyschists are especially common in the Kingsbridge Estuary sections (Fig. 4.1), and may be one reason why the early workers (e.g. Somervail, 1887) thought that the junction between the Devonian sequences and the Start Complex was gradational. Thin horizons of this black phyllitic material occur throughout the main outcrops of greyschist in both the Start and Bolt areas, but are generally rare.

As with the greenschists, there are numerous horizons within the greyschists which are rich in pyrite. These range from thin, banding parallel, lenticular aggregates

of small (1-2mm) bright gold^{coloured} pyrite cubes, up to extensive areas studded with large (1-1.3cm) euhedral crystals of either pyrite or limonite, which weather in relief of the surface (Plate 3.6; see Plate 4.3). In areas where the strain is slightly higher, the individual pyrite cubes develop elongate quartz/chlorite pressure shadows, which preserve a perfect cubic shape when the original pyrite weathers out. More typically however, the greyschist has an irregular pitted appearance, associated with much iron oxide staining, where the pyrite/limonite has been removed. Thin, cross-cutting pyrite rich veins also occur, with the vein material being seen to permeate laterally along the foliation planes. These pyrite-rich horizons appear to reflect local anaerobic conditions within an accumulating basinal sequence. It would be reasonable to expect an increase in the amount of such sulphides at increasing depth within the sequence. The West Cliff [SX 2961, 0382] to Bolt Head [SX 2727, 0360] section provides an across strike traverse through steep greyschist cliffs, but did not reveal any such trend. Associated with the pyrite horizons are thin black 'cindery' layers, no more than 4-5 centimetres thick and not very laterally extensive.

Petrology

Petrographically the greyschists are relatively simple, the principal minerals being quartz, albite and muscovite with lesser amounts of biotite, chlorite, garnet, epidote, zircon and iron oxides. Although quartz is generally predominant, some sections are composed entirely of aligned and crenulated micas, usually muscovite associated with biotite, sericite and knots of chlorite. Other sections are dominated by large, porphyroblastic albites in a fine mosaic of recrystallised quartz grains and elongate laths of mica. It is probable that some of what appears to be biotite is, in fact, iron-rich chlorite.

Compositional banding occurs on a millimetre scale, with thin leucocratic segregations of quartz and albite separated by lenticular melanocratic masses of poorly aligned phyllosilicates. With increasing deformation, the metamorphic segregation is enhanced and the phyllosilicates take on a pronounced alignment defining a strong tectonic fabric. Thin, sub-parallel, anastomosing seams of iron oxides coat the phyllosilicates, further emphasising this schistosity. In some schists, large anhedral poikiloblastic and subhedral porphyroblastic albites overprint all of the deformation fabrics, although there is some suggestion of the tectonic fabric wrapping the smaller porphyroblasts, implying syn- to post-tectonic growth. The fine iron oxide seams which help to define the schistosity pass through the poikiloblastic albites as inclusion trails, but are largely absent from the often simply twinned porphyroblasts (Plate 3.7). The unusual albite poikiloblasts are presumably the 'black mineral' referred to by early workers, although the enclosed trails are clearly iron oxides and not organic matter

(e.g. Tilley, 1923). The greyschists exposed in the interbanded green and grey schists between Prawle Point and Start Point contain numerous small, euhedral almandine-rich garnets (Plate 3.8). The garnets are well developed in the thin micaceous seams in these quartz-rich rocks of intermediate affinities. In these examples, the tectonic fabric, as defined by an alignment of the phyllosilicates, is only moderately developed and can be seen to clearly wrap around the garnets. Quartz pressure shadows are variably developed and are further evidence of growth, prior to the formation of the main tectonic fabric.

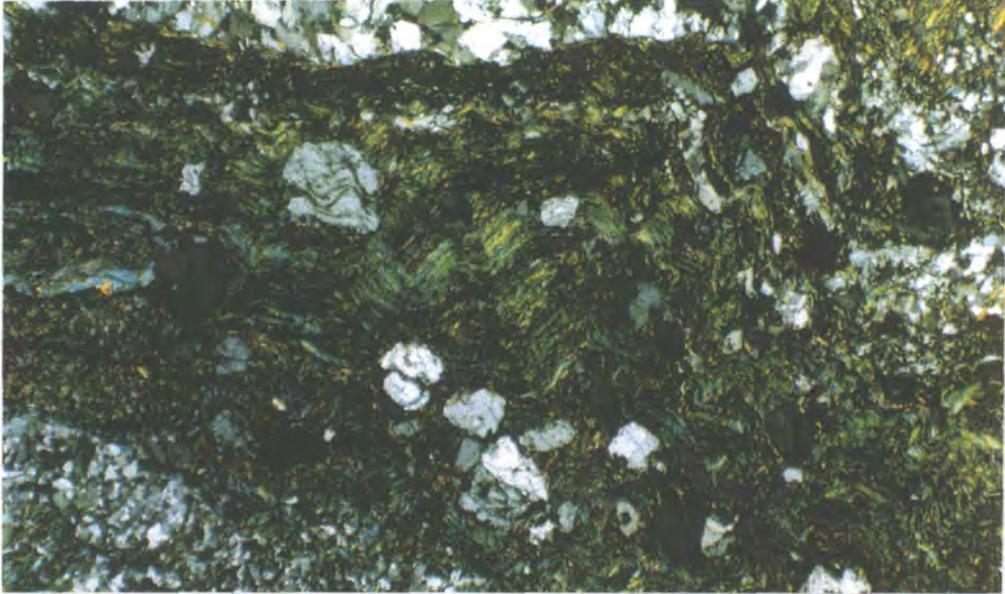


Plate 3.7 Thin section of greyschist showing the inclusion of the fine schistose fabric, defined by aligned muscovites and seams of finely disseminated oxides, being enclosed by large poikiloblastic albites. Such inclusion trails are largely absent from the simply twinned albites. Greyschist from Deckler's Cliff, [SX 2756, 0367] under X4 magnification and crossed polars. FoV= 6x3.7mm.

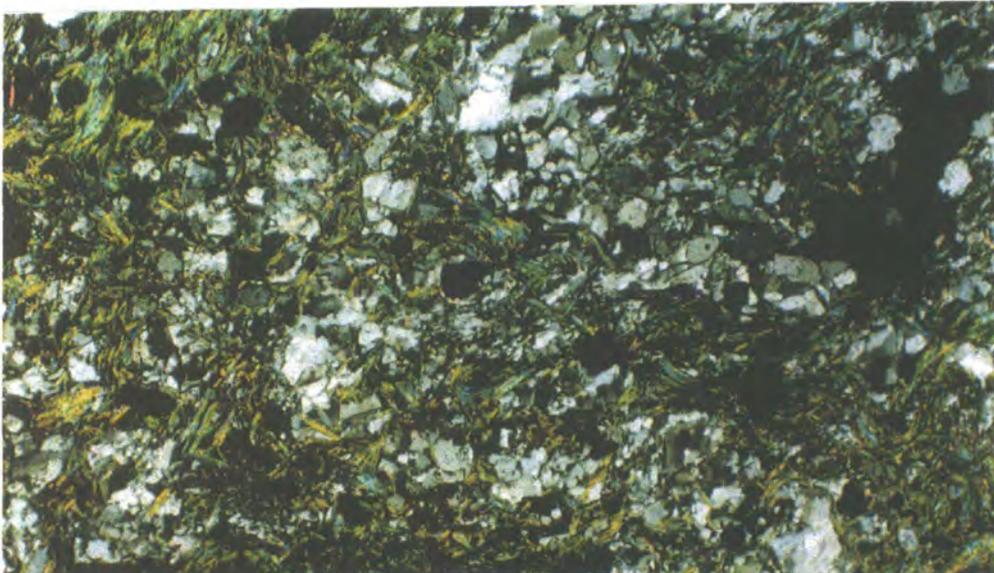


Plate 3.8 Thin section of greyschist showing large, euhedral almandine-rich garnets growing preferentially in the thin micaceous layers, which define a crude foliation within the quartzose schists. Greyschist from Lannacombe beach, [SX 2803, 0371] under X4 magnification and crossed polars. FoV= 6x3.7mm.

Geochemistry

Whilst collecting specimens for the geochemical analysis of the Start Complex greenschists (see above), P.A.Floyd and R.E.Holdsworth also collected representative samples of the greyschists (samples SPS 1-SPS 40). These rocks, along with some specimens of intermediate lithologies and fault rocks collected by the author during the summer of 1991, were analysed using standard X-ray fluorescence spectrometry (XRF; major oxides and most trace elements) at the University of Durham (see Appendix I for methodology and results).

Recent geochemical analysis carried out on other basin sequences in SW. England (e.g. the Gramscatho Basin; Floyd & Leveridge, 1987) have revealed usable geochemical trends throughout the stratigraphic sequence, thought to be related to changes in the source composition. Also, some of the oxide ratios (e.g. $\text{Al}_2\text{O}_3/\text{SiO}_2$) exhibited trends related to grain size variation, with coarse-grained quartz-rich turbiditic sandstones having the lowest ratios. However, Floyd *et al.* (1991) note that many factors may affect sandstone chemistry, including weathering, diagenesis and metamorphism. These all tend to alter the relative abundances of LIL elements and thus it is more accurate to use immobile elements to characterise provenance and tectonic setting. This is especially likely to be the case with the highly deformed and recrystallised Start Complex greyschists. A consequence of this extreme alteration is a complete lack of both detailed stratigraphic evidence and evidence of primary sedimentary features e.g. bedding, grain size variation, etc. and thus a geochemical analysis of these metasediments has limited application. Further problems arise due to element migration, resulting in changes in the absolute abundances of various oxides. For example, the SiO_2 content, which varies markedly in these schists (46%-92%), probably reflects extensive recrystallisation and quartz segregation.

The use of stable trace element ratios in determining the tectonic environment of sediments is questioned by Floyd *et al.* (1991), who note that sorting, heavy mineral content and proportion of mafic input tends to spread the chemical data across a number of geologically unrelated fields. Again it is impossible to assess the above factors in the Start Complex given their complete lack of detailed sedimentary structure and extensive alteration. Floyd *et al.* (1991) suggest the use of upper continental crust-normalised multi-element patterns to provide an approximate indicator of the tectonic environment, but do note that supplementary petrographic data is necessary to further constrain this. They cite data from two turbiditic basins in the Rhenohercynian zone (Giessen Greywacke Unit and Gramscatho Group) and show that these plot as continental arc+active margin (CAAM) sequences. Although primary petrographic data was not available for the Start Complex, the greyschist geochemistry was normalised against average upper continental crust values (from Taylor &

McClennan, 1985), resulting in a pattern broadly similar to those from the Giessen and Gramscatho sequences (i.e. indicative of a CAAM environment, Figs. 3.4a&b). Subsequent normalisation against CAAM values, from Floyd *et al.* (1991), resulted in a flat trend with some Zr enrichment and marked Sr depletion (Fig. 3.5). The high Zr values may be attributable to zircon input, whilst the depleted Sr values may be a consequence of this elements greater mobility during metamorphism.

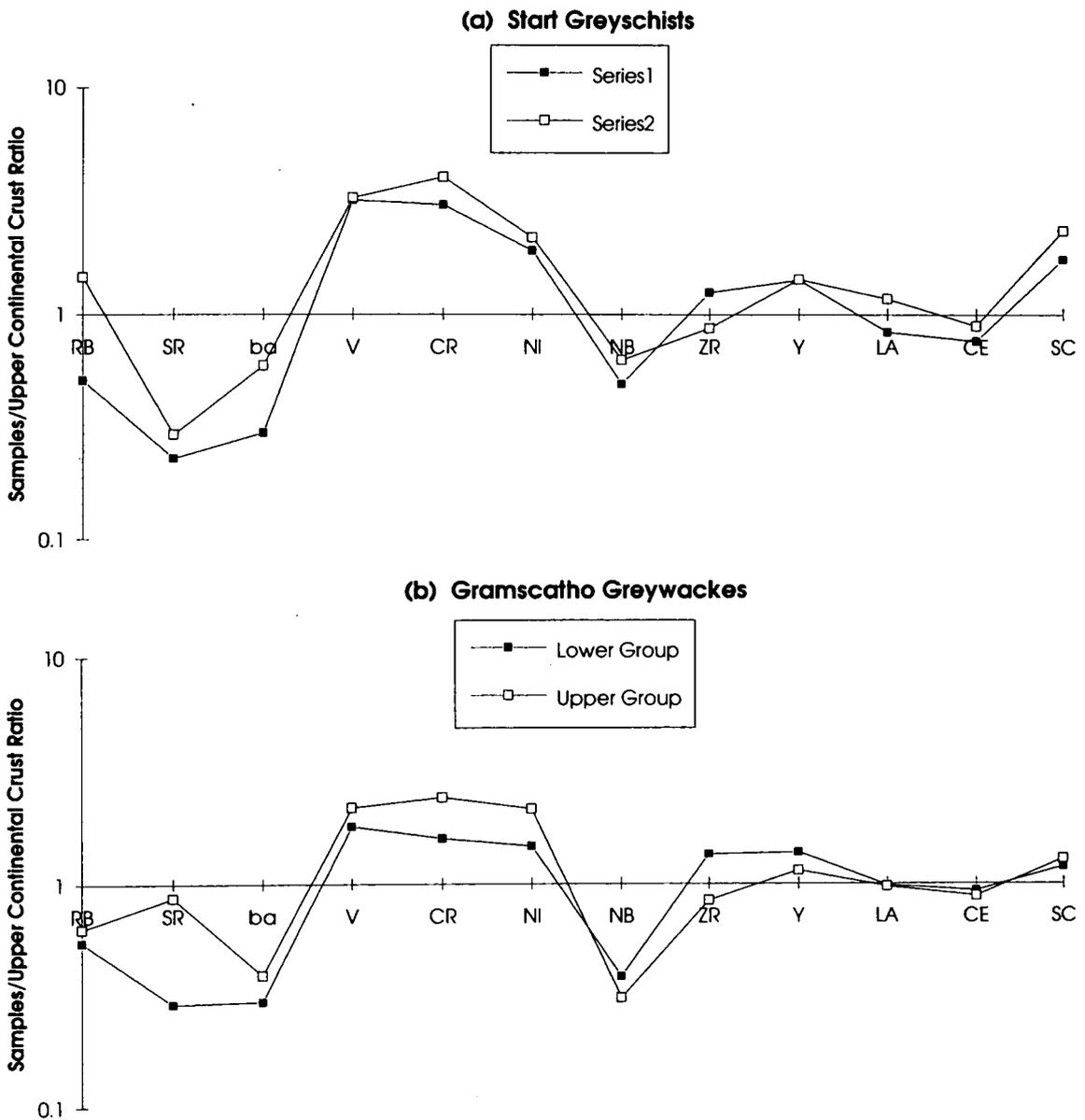


Figure 3.4 (a) Upper continental crust normalised multi-element pattern for two averaged groups of the Start greyschists, showing general correspondence to the continental arc+active margin tectonic environment. (b) Similarly normalised data for the Gramscatho Basin (data from Floyd *et al.* 1991). Upper continental crust normalisation values from Taylor & McLennan (1985).

The greyschists are often intimately interbanded with greenschist horizons, both as a primary interbanding and as secondary tectonic interleaving, and thus there is likely to be a degree of metavolcanic contamination even in the 'pure' greyschist units. This contamination will further hinder any geochemical study. It is notable, however, that these highly metamorphosed sediments do show geochemical similarities to the Gramscatho and other Rhenohercynian basinal sequences.

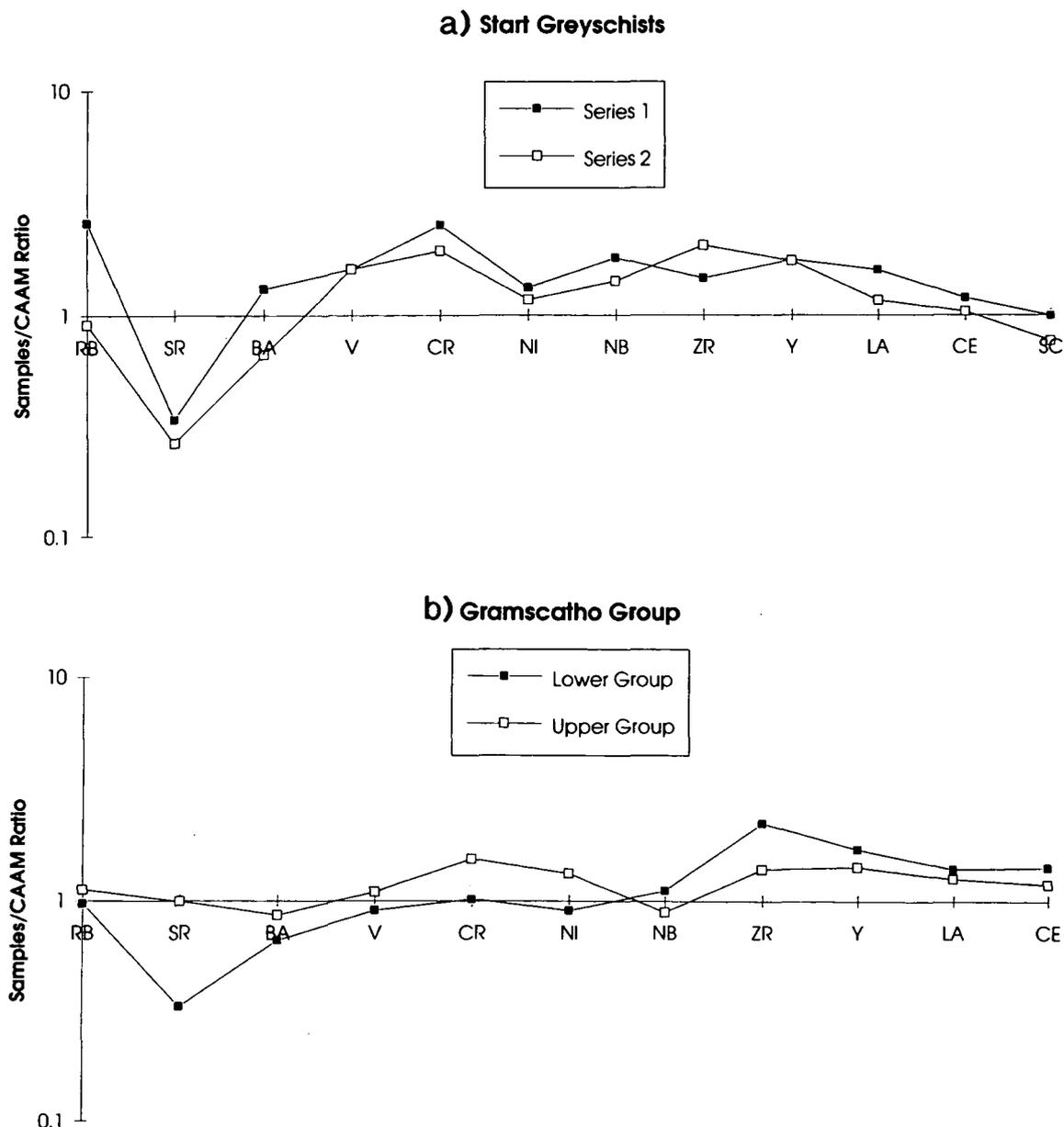


Figure 3.5 (a) Continental arc+active margin normalised multi-element pattern for the Start greyschists showing approximate trend parallel to unity (Sr depletion may be due to element mobility during metamorphism). **(b)** Similarly normalised data for the Gramscatho Group, again with Sr depletion (data from Floyd *et al.* 1991).

3.2.3 Intermediate Lithologies

Raisin (1887) coined the term 'micaceo-chloritic' to describe schists with similar amounts of mica and chlorite, the minerals she considered to be diagnostic of the greyschists and greenschists respectively. She acknowledged that chlorite was present in typical greyschists, but only in small quantities and that white mica was similarly subordinate in the greenschists. Her micaceo-chloritic schists contained sufficient quantities of these minerals to warrant classification into a distinct group. These specimens came from areas where greenschists and greyschists occurred in alternate layers.

Tilley (1923) noted that the intermediate schists were best developed, not surprisingly, at the upper and lower boundaries of the main greenschist outcrop. He considered these schists to be more akin to the greenschists, suggesting that they were tuffaceous in origin with an intermixture of sediment.

Petrographically, these intermediate schists contain most of the mineral phases common to both the greenschists and the greyschists, i.e. chlorite, quartz, calcite, epidote, albite, muscovite, actinolite and garnet, and a more or less continuous spectrum can be observed between the two end members. They may be distinguished from the true greenschists by both the greater amounts of quartz and by the presence of almandine-rich garnet. They differ from the true greyschists in that actinolite is present, and in general the amounts of epidote and chlorite are greater. Whilst Tilley (1923) chose to subdivide these schists into two further groups, based on the occurrence of garnet, this seems unnecessary in what are relatively unimportant lithologies. Field evidence has shown that the junction between the greenschists and the greyschists varies from sharp, usually tectonic, boundaries to a gradational interbanding of the two units, with the interbanding having clearly been enhanced, in places, by tectonic interleaving. Whilst intermediate lithologies of primary origin are important in understanding the original nature of the greenschists, the highly deformed nature of these outcrops precludes them from offering unambiguous evidence.

3.2.4 Offshore extent of the Start Complex

Phillips (1964) recorded Start-type schists extending offshore westwards into the Plymouth Bay area, to form part of the east Rutts Reef. Marshall (1965) provided a comprehensive comparison between these schists and those of the Start Complex. He noted that some of the cored quartz-mica schists bear a striking resemblance to the greyschists (especially those at Bolt Head), although these core specimens have significantly larger and more abundant albite porphyroblasts. Samples of chlorite schist, recovered adjacent to the above mica schists, were indistinguishable from the onshore

greenschists. The approximate areal extent of these lithologies is given in Figure 3.6, although this has not been accurately constrained and may be considerably greater.

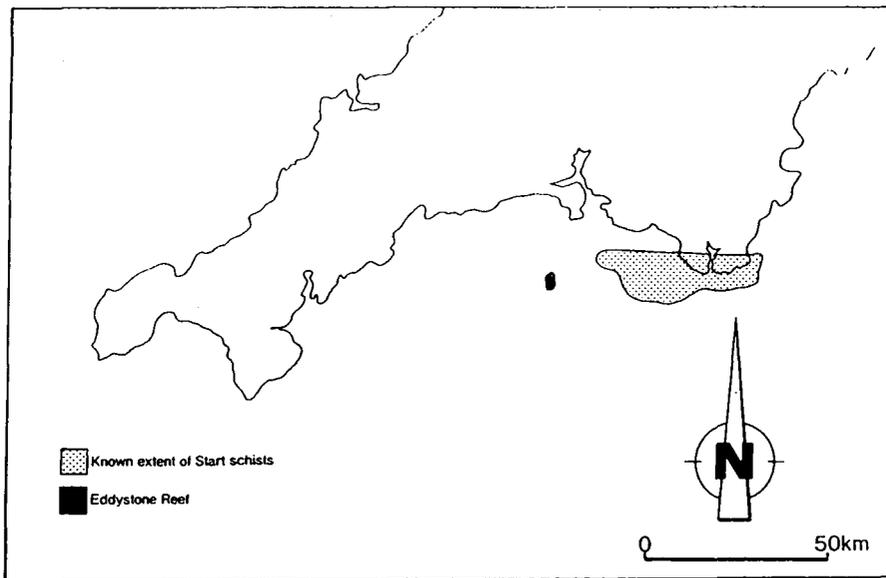


Figure 3.6 Approximate offshore extent of Start-type schists within the Plymouth Bay area from borehole data. Modified after Evans (1990).

3.2.5 Summary

The Start Complex consists of an interlayered sequence of highly altered and deformed schistose pelites and variously deformed basic metavolcanics, of probable Devonian age. The metavolcanics resemble finely layered tuffs, although they may be highly altered basaltic lavas. Their relationship to the schistose pelites is uncertain, as all primary fabrics have been overprinted by extensive later recrystallization and tectonism. The schistose pelites are quartz- and mica-rich and are occasionally interleaved with thin psammitic horizons. Petrographically they are very uniform, the strong schistose fabric being defined by an alignment of micas.

Geochemical analyses of both lithologies were revealing. Not surprisingly, the data yielded by the pelites was quite inconclusive in the absence of petrographic and stratigraphic constraints, but did show similarities to other Rhenohercynian basinal sequences in terms of the broad tectonic environment. The basic metavolcanics have the geochemical signature of depleted N-type MORB and are amongst some of the most depleted in this region. This links them to the Lizard complex and related S. Cornish basalts, and sets them apart from the alkali basalts of N. Devon and Cornwall, which have distinctly intraplate characteristics. The metavolcanics may thus represent incipient ocean crust flooring a small basin of limited areal extent, in which a thick sedimentary sequence was accumulating (the pelites). Structural controls on the basin development and its tectonic setting within the Rhenohercynian zone are discussed further in 3.5.

3.3 The Lower Devonian Succession

3.3.1 Introduction

The general stratigraphy and sedimentology of the Lower Devonian succession in SW. England is outlined in Chapter 2 (see 2.2.2 part *i*). The Lower Devonian (Dartmouth Group and Meadfoot Group) forms a broad E-W belt running between Dartmouth in S. Devon and Watergate Bay in W. Cornwall (Fig. 4.1*) The scope for further detailed stratigraphic study of the Meadfoot (Seigenian to Upper Emsian) and Dartmouth Groups (?Gedenian to mid Seigenian) in South Devon is limited due to a) generally poor exposure, b) the high strain state of the shales, c) the great uniformity of the lithologies across the area, and d) limited coverage of the succession during fieldwork.

Study of the Dartmouth Beds was very limited as these lithologies had very little exposure within the field area (3.3.2; see also 3.1). On the east coast, a short stretch (c. 0.5 km) of monotonous red shales was studied at Pilchard Cove [SX 2843, 0466], whilst on the west coast, outcrop was restricted to a single, poorly exposed outlier in the Avon estuary [SX 2665, 0444].

The Meadfoot Beds form most of the Devonian outcrops within the field area (3.3.3). In general, many east coast sections are poorly exposed, with long stretches of sandy shoreline. Well exposed cliff sections are confined to the short stretches of coastline at Torcross [SX 2823, 0419], and from Beesands [SX 2819, 0402] to Tinsey Head [SX 2819, 0396]. The west coast, north of the high strain zone at Hope Cove [SX 2673, 0397], displays the most accessible exposures of Devonian slates, with a more or less continuous section. Occasional carbonate-rich horizons within the Meadfoot Beds yielded some almost complete fossils, and an abundance of fossil fragments, typically orthocones, crinoids, brachiopods and corals, many of which were flattened and distorted within the cleavage.

The Devonian shales are cut in places by large, foliated, pre-orogenic intrusive igneous bodies. These intrusives were originally of basic composition, but are now largely composed of replacive carbonate (see 3.3.4). The best examples occur at Torcross on the east coast, with thinner, less extensive, but compositionally similar rocks occurring sporadically along the west coast. A single outcrop of an unfoliated and relatively unaltered intrusive occurs at Long Stone [SX 2661, 0433], to the north of Thurlestone, and appears to be the only example of a post-orogenic intrusive within this area (see 3.3.5).

* See also the enlarged maps of the east and west coasts

3.3.2 The Dartmouth Beds

Dineley (1966) provided the first detailed description of the Dartmouth Beds and chose to divide them into four stratigraphic subdivisions, based on his mapping south of Plymouth, although Seago and Chapman (1988) and Seago (1991) subsequently modified these (see 2.2.3). Richter (1967) gave a comprehensive account of the Dartmouth Beds along the east Devon coast, from Torquay to Beesands. Smith and Humphreys (1991) noted that the lack of detailed work on the Dartmouth Group probably reflects its intensely deformed nature, i.e. the biostratigraphic control is very poor, and most sedimentary structures are obscured by cleavage. In this part of Devon, the Dartmouth Beds belong to the Warren Formation (Seago & Chapman, 1988; Smith and Humphreys, 1991; the Warren Sandstones of Dineley, 1966). They comprise red/green or lilac mudstones, with a significant proportion of thin interbedded sandstones and pass conformably and gradationally upwards into the basal unit of the overlying Meadfoot Group, the Bovisand Formation (Seago & Chapman, 1988).

On the east coast, the conformable boundary is obscured by the shingle spit of Slapton Sands [SX 2827, 0444], although it has been mapped inland (BGS sheet 355) and is most probably a stratigraphic contact. The first outcrops of the Dartmouth Beds to the north of this region occur at Pilchard Cove [SX 2843, 0466], and are pale buff to maroon/lilac shales, with a noticeably lustrous or glossy appearance. Many of the hard, shiny surfaces reveal large micaceous minerals, which impart the slaty cleavage to these rocks. Rare coarse grey sandstone horizons, up to 25 centimetres thick, occur throughout this section, whilst thin (5-6cm) pale greenish/grey silty lenticles are quite abundant. Where these sandstones occur as distinct units they may be confused with pale pre-orogenic intrusives (see below), especially when the sandstones are fine-grained.

On the west coast at Bigbury [SX 2655, 0433], the boundary between the Dartmouth and Meadfoot Beds is exposed and is clearly gradational, the purple mudrocks of the Dartmouth Group being interbedded with the dark pelites of the Meadfoot Group (see Fig. 4.11b). Traversing north-eastwards from Sharpland Point [SX 2657, 0442], the pale grey/black Meadfoot shales are seen to contain thin (2-3m) interbands of purple shales, which increase in proportion eastwards to become the dominant lithology around Cockleridge Point [SX 2664, 0444]. The boundary with the small inlier of Warren Formation is complicated by the presence of numerous thin pre-orogenic intrusives, which markedly affect the adjacent rocks, resulting in contact zones of hard purple/grey shales, which are generally difficult to distinguish from the true Dartmouth Beds in weathered surfaces. Seago (1991) suggests that this northern contact may be thrust controlled, though this is probably not the case (see 4.4.3). The Dartmouth Slates vary from a deep maroon on fresh faces through to a pale lilac/buff on the more weathered examples. Outcrop of these shales deteriorates rapidly east of

Cockleridge Point and the southern stratigraphic contact with the overlying black Meadfoot pelites, as mapped by Ussher (1904), is not exposed. A thin horizon of hard reddish-purple shales at Hams End [SX 2662, 0442] may represent the westward extension of the southern junction.

Dineley (1966) provided a detailed petrographic description of the Dartmouth Beds in the Plymouth area, noting muscovite, quartz and chlorite, with very minor amounts of pyrite, haematite, epidote, limonite, rutile and tourmaline. A single section of the Dartmouth slate from Pilchard Cove [SX 2843, 0466] revealed the presence of aligned phyllosilicates, principally muscovite, with very occasional patches of chlorite and a variety of finely disseminated oxides.

3.3.3 The Meadfoot Beds

Ussher (1904), who provided the first detailed work on this lithology, chose to divide the Meadfoot Beds in this region into four E-W trending units. The oldest of these was the Ringmore type, characterised by reddish, silty slates, which passed upwards into the Torcross type, typified by dark grey slates. The Torcross type was separated from the uppermost Tinsey Head Slate Series by a thin band of coarse grey grits (which are, in fact, sandstones), the Beeson Grits. However, these divisions, apart from the Beeson Grits, were not differentiated on the survey map, with the Meadfoot Beds being mapped as a single, homogeneous unit. Richter (1967) studied these sediments at their type locality (Meadfoot Bay, Torquay), but did little stratigraphic work in south Devon. Most recently, Seago and Chapman (1988) and Seago (1991) reassessed the Meadfoot Group stratigraphy and divided it into the Bovisand Formation overlain by the Staddon Grit Formation. Seago and Chapman (1988), note that, due to the complex facies relationships within the Meadfoot Group, it is often difficult to distinguish one formation from the other, especially in areas of poor exposure. They suggest that the coarse grit horizon (Beeson Grit) noted by Ussher (1904), was probably a lateral equivalent of the Staddon Grit Formation.

The east coast exposures of the Meadfoot Beds comprise short cliff outcrops separated by lengthy beaches. At Torcross [SX 2823, 0419], the shales are typically light to mid-grey slates with pale buff, sandy partings and thin pale green silty lenticles. Fine-grained sandstones, ranging in thickness from 5-30 centimetres, are quite abundant within the dark shales. The original bedding is obvious at this locality, and defines an upright antiform/synform pair. The shales remain unaltered adjacent to the large 'Torcross' intrusives, apart from a marked increase in the density of quartz veining. Typically the intensity of deformation is higher in these quartz veined zones and many such slates bear a strong resemblance to the quartz-rich greyschists of the Start complex. At Dun Point [SX 2823, 0416], a second group of large pre-orogenic

intrusives occurs, and the shales here are similarly unaffected. It appears that the alteration adjacent to the intrusives is independent of the scale of these bodies. Many of the paler shale horizons are noticeably calcareous, bearing distorted solitary corals and occasional crinoid ossicles. These layers weather to result in a pitted appearance.

The cliffs at Beesands [SX 2819, 0402] comprise dark shales with thin (2-3cm) pale sandy horizons. The strain here is higher than in the southernmost outcrops of Dun Point, with the sandy horizons and silty lenticles being transposed into the dominant cleavage and clearly enclosed by an early tectonic fabric (see Plate 4.28). Unfortunately the point at which the strain increases is obscured in the shingle of Beesands beach (See 4.4.4). Moving south towards the SBF, bedding generally becomes less obvious, with only occasional pale silty partings visible in the dark shales, although the frequency of the thick sandstone layers remains unchanged. Contrary to Ussher's (1904) observations, intrusive rocks do occur along this section, but are thin (5-70cm) and not very laterally extensive. Many of the thinner intrusives may be mistaken for sandstone bodies, as they display deformation features very similar to soft sediment deformation structures, and are finely interbanded with the shales. In places, this interleaving is tectonically enhanced, making it difficult to distinguish these intrusives from the true buff coloured sandstone partings. The structure and appearance of these pre-orogenic intrusives is dealt with in more detail in 5.3.2.

The interbedded shales and sandstones at Tinsey Head [SX 2819, 0396] are highly faulted, with some of the fault zones being marked by deep red oxide staining. Presumably these sequences of thin sandstone bodies within the dark mudrocks are the Beeson Grits of Ussher (1904), although whether they may be classified as a separate lithology is questionable. South of Tinsey Head, the slates appear to be more calcareous, with numerous carbonate-rich horizons bearing distorted solitary corals (identified as *Zaphrentis* by Ussher, 1904), and crinoid ossicles. Moving south towards the boundary with the Start Complex, the strain increases markedly and as a consequence bedding is reduced to attenuated intrafolial lenticles. Immediately to the north of the SBF, the Meadfoot shales are very soft and highly weathered, coated in a film of silt and extensively overgrown with vegetation.

There is little evidence to support Ussher's (1904) original subdivisions of these shales and he did admit that the basis for some of this zonation was tenuous, depending on such unreliable variables as the coloration of the slates, the presence of fossils, the presence of intrusives, the occurrence of grits, etc. Clearly in deriving his zonation, Ussher (1904) has overemphasised the importance of some of the above variables.

The west coast exposures of the Meadfoot Beds are more or less continuous, and again lend little support to Ussher's (1904) zonation. North of Challaborough [SX 2647, 0449] the shales are highly variable in colour, ranging from pale green/grey

through to deep black, the coloration being largely unaffected by the presence of many thin intrusives. Moving south towards Bigbury Bay [SX 2655, 0433], reddish-purple shales are also seen within these outcrops. Around Sharpland Point, as mentioned above, many thin intrusives have affected the appearance of the dark shales, resulting in extensive red oxide surface staining. The reasons for this being so acute here are unclear, as the intrusives across this area belong to a single suite, and the slates are lithologically similar to those to the north

A different type of intrusive, which appears to be post-orogenic, occurs at Long Stone [SX 2661, 0433] in a zone of what is mapped as Beeson Grit. Whilst numerous thin sandy horizons do occur intimately interbanded with the large intrusives, the extent of outcrop is far less than indicated on BGS sheet 355. Thin zones of red/purple shales, similar to the Dartmouth Beds, are interspersed with the black/grey shales along this section. At Broad Sands [SX 2665, 0427] the purple shales dominate and here there is evidence of this purple coloration permeating along faults and fractures in the originally grey pelites.

At Warren point [SX 2670, 0421], the pale grey Meadfoot Beds are richly fossiliferous, bearing numerous flattened and distorted solitary corals, stretched orthocones and a few crinoid ossicles. Although brachiopods have been identified in the Meadfoot shales with a degree of accuracy, Evans (1985) notes that the rest of the fauna is not determinable even to the generic level and probably comprises fragments of bryozoans, tentaculids and trilobites. Samples collected in the field were too incomplete and deformed to allow useful identification.

Again moving south towards the Start Boundary Fault, the strain increases markedly, with the last true examples of bedding occurring at the Leas Foot Sands promontory [SX 2675, 0420] where 3-5 centimetre pale sandy partings define a series of upright, open folds within the black shales (Plate 3.9), just to the north of the faulted outlier of red Permian conglomerates. This single outcrop of New Red Sandstone is relatively thin (10's of metres thick) and, although its inland projection was not assessed, Seago (1991) suggests that it has 2-3km² of areal extent.

The short stretch from Thurlestone Rock [SX 2675, 0420] to Outer Hope [SX 2675, 0402] comprises high strain shales in which the bedding is reduced to fine silt partings which are attenuated and transposed into the dominant cleavage. Ussher (1904) mapped several Beeson Grit horizons terminating against the cliffs along this section and, whilst thin (10-15 cm) sandstone bodies are quite common here, they do not appear to constitute a lithology in their own right.

In general, the Kingsbridge estuary sections of the Meadfoot Beds are poorly exposed and very badly weathered. Shales are a very uniform pale grey to buff colour, with bedding only occasionally visible as thin brown sandy horizons. Much of the

outcrop is friable and has suffered from varying degrees of recent creep. 'Torcross' type intrusives are relatively common in these sections (e.g. at Halwell Point [SX 2749, 0404] and west of Charleton Point [SX 2745, 0421]), ranging in size from a few centimetres up to 1.5m in thickness and again they are highly weathered and in places virtually indistinguishable from the shales.

Petrographically very little work has been done on the Meadfoot Beds in this area bar the detailed accounts of Marshall (1965), who noted that the fabric of these shales was defined by aligned micas, principally muscovite. He also suggested that garnet may possibly be present. Several representative thin sections were studied and the presence of chlorite, quartz, very occasional plagioclase, finely disseminated oxides and possible organic matter was noted, although many of the accessory minerals recorded by Marshall (1965) were impossible to identify. No garnet was seen and its presence in these low grade pelites must be questioned.

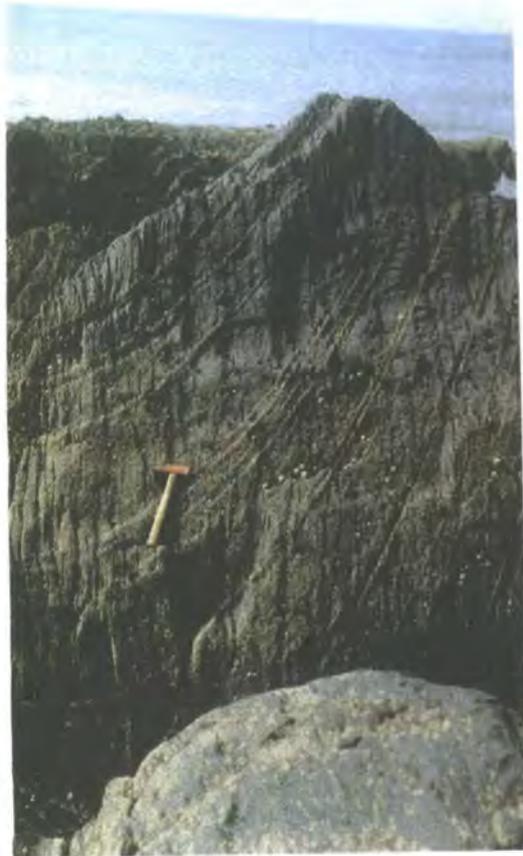


Plate 3.9 Upright north verging primary folds in the Meadfoot Beds defined by pale sandstone layers in the black pelites. Leas Foot Sands promontory [SX 2674, 0420].

3.4 The Intrusives

3.4.1 Pre-orogenic Intrusives

Field Appearance

The Lower Devonian strata of south Devon are cut by numerous foliated and altered intrusive igneous bodies (Fig. 4.1*; see also Plate 5.12). The type locality for these rocks is at Torcross, where several large ribs of pale greenish/buff weakly foliated rock have weathered out of the dark Meadfoot shale cliffs. The surface of the intrusive is pitted and marked by diffuse rusty brown patches, 3-15 millimetres across and elongated in a consistent direction, which were presumably originally pyrite and are now probably limonite or leucoxene. Parallel to the above are numerous diffuse elongate, pale grey xenoliths, which range in size from a few millimetres up to 6-7 centimetres, and whose original composition is impossible to determine as they are now composed mostly of carbonate (Plate 3.10). The larger intrusives exhibit narrow relict chilled margins, typically only a few centimetres in width, with a distinctive mid-brown colour. The adjacent shales remain typically unaltered, apart from the presence of numerous fine anastomosing quartz veins. There is limited evidence of a 'soft sediment' type of emplacement of these intrusives, e.g. pseudo-flame structures, etc., which lends support to the premise that they were emplaced into wet sediments. The marked lateral thickness variations combined with the interbanding of shales and silts provides additional evidence. Richter (1967) noted that these intrusives cut the bedding, but are themselves weakly foliated and in places even appear to have ^{been} emplaced along this foliation. He suggested that they ^{were} emplaced during an early stage of cleavage development prior to complete development of the tectonic fabric.

'Torcross' type intrusives occur throughout the Meadfoot sequence in this area, with variations in the field appearance attributable to differing degrees of alteration, deformation and the effects of surface weathering. This is clearly demonstrated by the variety of intrusives around Warren Point [SX 2670, 0421]. The large intrusive forming the Leas Foot Sands promontory is a deep rust colour on the surface, although a freshly cut sample shows that this coloration only penetrates 1-1.5 centimetres, beneath which the rock is a pale grey/buff. The presence of many large normal faults, which also cut the Permian outlier (see above), are probably responsible for much of the surface staining. Significant quantities of Torcross type intrusive are associated with the Start Boundary Fault, although they have been very highly altered (see 4.3)

Just to the north of Warren Point is a thin, pale pink intrusive with a curious mottled appearance and a very intense foliation. Superficially this appears different to the typical 'Torcross' intrusives, but a section reveals that it has been subjected to a higher degree of deformation and is now an almost pure carbonate mylonite. The

* See also the enlarged maps of the east and west coasts for exact locations

mylonitic foliation is coated in anastomosing films and seams of opaque oxides which impart the unusual pink colour, whilst weathering along the foliation results in the mottled appearance.



Plate 3.10 Weathered foliation surface of large pre-orogenic intrusive at Torcross [SX 2823, 0419], viewed looking approximately north. Pyrite crystals are flattened within the foliation and elongated in an E-W direction. The elongate, diffuse pale patches, paralleling the pyrites, were xenoliths, and are now composed entirely of replacive carbonate.

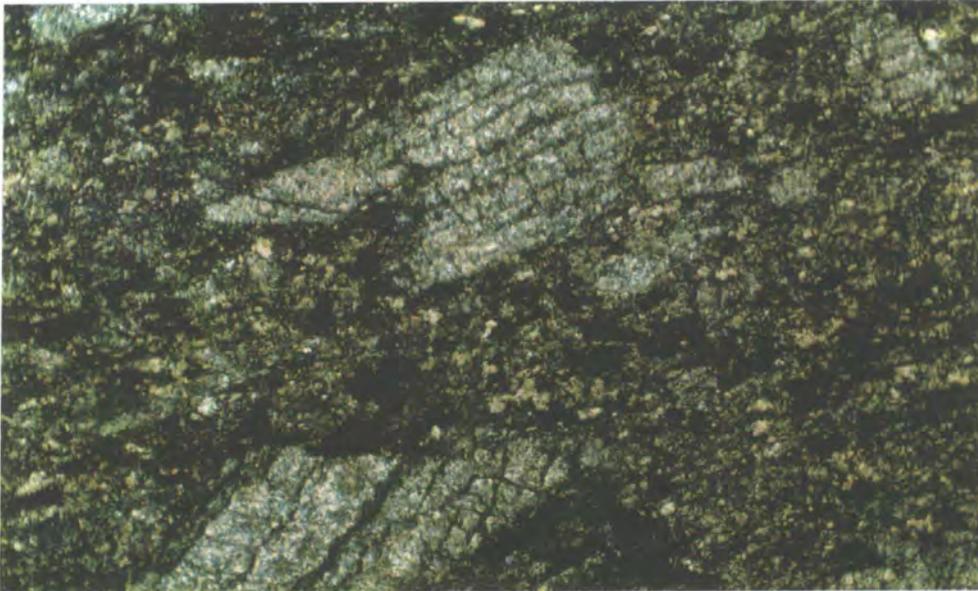


Plate 3.11 Thin section of pre-orogenic intrusive with large diffuse oblong patches, now composed of calcite, serpentine and talc and which were probably originally olivines in medium-grained basic rock. Groundmass is now very fine-grained calcite, with occasional euhedral crystals and finely disseminated oxides. Pre-orogenic intrusive from Torcross, [SX 2823, 0419] under plan view (X1) and crossed polars. FoV= 12x6.8mm.

Petrology

Due to the high degree of alteration these rocks yield little petrographic information and are now principally composed of a fine-grained mass of carbonate, with lesser amounts of quartz, albite, white micas and chlorite. In some examples, large rectangular patches, now composed of chlorite with seams of calcite, serpentine and talc, probably represent the replacement of original olivine phenocrysts (Plate 3.11). Large isolated porphyroclasts of pyrite are typically fractured and distorted within the foliation, which is defined by anastomosing seams of white mica. In the more mylonitic examples, the fragments of pyrite porphyroclast are linked by quartz/chlorite pressure shadows and tails (Plate 3.12).



Plate 3.12 Thin section of pre-orogenic intrusive showing large pyrite crystals fringed with intergrown quartz and chlorite pressure shadows. Fragments of broken crystals are similarly linked. Groundmass is fine mylonitic carbonate and chlorite, the mylonitic foliation defined by calcite augen wrapped by fine anastomosing oxide seams. Pre-orogenic intrusive from Torcross, [SX 2823, 0419] under X4 magnification and crossed polars. FoV= 6x3.7mm.

3.4.2 Post-orogenic Intrusives

Field Appearance

Long Stone [SX 2661, 0433], north of Thurlestone, is a large rocky promontory composed of sheets of hard greenish/grey intrusive, bands of coarse grit and quartz-rich shales. The northern face of this promontory is a significant normal fault. This sub-vertical intrusive has been emplaced along the dominant foliation, and trends ENE-WSW. It is completely unfoliated, implying that it is post-orogenic. The sedimentary/igneous junction is irregular, with random splays of intrusive trapping slivers and wedges of shale. The igneous body is medium-grained with an overall pale greenish/grey surface spotted with pink euhedral phenocrysts and cut by diffuse veins

of this pinkish material. The intrusion is over 2m thick, although it thins markedly offshore, and, not surprisingly, the adjacent shales and grits are noticeably affected with a 1-2m reddened contact zone. There are, however, no other visible signs of contact metamorphism, e.g. spotting, within the immediate area. The promontory as a whole is shredded with dense networks of quartz veins, some of which are up to 1.5m in thickness. The intrusive would appear to be the only example of post-orogenic igneous activity seen within this region.

Petrology

Despite evidence of hydrous alteration, this specimen is the freshest of the intrusives examined. The predominant mineralogy is simply twinned plagioclase, pyroxene (probably augite) and a curious form of epidote, which in hand specimen appears as the pale pink 'phenocrysts' and the vein infill (Plate 3.13). The epidote is unusual, not only in colour, in that it has an inclined extinction and an anomalous 'dusty' appearance. The effects of alteration are still apparent and there is a considerable amount of carbonate within the groundmass. This intrusive is not unlike the greenstones recorded in the Devonian argillites of north Cornwall (e.g. Floyd *et al.* 1993b; see 5.3.2).



Plate 3.13 Thin section of post-orogenic intrusive showing groundmass of simply twinned plagioclase, laths of pyroxene (in places fringed with retrogressive chlorite), diffuse patches of calcite and finely disseminated oxides. A large vein, which cuts the section from left to right, is composed entirely of unusual, dusty pink epidote, fine veinlets splaying off this throughout the section. Post-orogenic intrusive from Long Stone, [SX 2661, 0433] under X4 magnification and crossed polars. FoV= 6x3.7mm.

Geochemistry of the Intrusives

Samples of several of these intrusives were analysed using standard X-ray fluorescence spectrometry (XRF: major oxides and most trace elements) at the University of Durham (see Appendix I for methodology and results). Unfortunately the absolute abundances of the major oxides were very unreliable due to the extensive alteration of the specimens. This was especially so in the cases of i) K_2O , which has been leached out, ii) SiO_2 , which, due to extensive quartz veining and segregation, recorded anomalously high values (e.g. one specimen was 92% SiO_2), and iii) CaO , which was high due to the breakdown of feldspars resulting in replacive carbonate. Trace element plots showed quite high levels of enrichment although no trend emerged. Unfortunately, they also showed variable values for the more immobile elements (Zr, Y). It is felt that the degree of alteration of these intrusives, some of which are now reduced to carbonate mylonites, precludes any reasonable geochemical modelling and that further REE analyses (using ICPMS) are not warranted.

3.5 Discussion

Geochemical evidence suggests that the Start Complex may be linked to the basaltic rocks of S. Cornwall and is quite distinct from the basalts in the Devonian sequence to the north. Similarly, illite crystallinity and phyllosilicate cell parameters connect the Start Complex schists to the pelites of the S. Cornish Gramscatho Basin, rather than to the lower grade shales to the north (Primmer, 1983)

Badham (1982) suggested that the Gramscatho basin may have evolved as a dextral pull-apart along the southern margin of a major intracontinental transform fault. Evidence for such E-W dextral extension was offered by Barnes and Andrews (1986), using the orientation of the Lizard dykes. Holdsworth (1989a) extrapolated this idea to the Start 'basin', and further suggested, on structural grounds, that this E-W basin bounding fault possibly represented a Devonian terrane boundary, separating the ORS continent to the north from the Armorican microplate to the south. However, the orientation of basic volcanics within the Mylor Slates suggests that dextral transtension was unimportant during the later history of the evolving Gramscatho basin. Also stratigraphic continuity across the northern margin of the Gramscatho basin indicates that the terrane boundary was more probably pre-Devonian in age (Shail, 1992). Alternative orthogonal models have been offered for the evolution of the Gramscatho basin (Leveridge *et al.* 1984; Holder & Leveridge, 1986a) and, as no unequivocal evidence of basin formation has yet to emerge from south Cornwall and none is likely to be forthcoming from the highly deformed Start Complex, it remains impossible to produce a definitive regional model for the kinematic evolution of these basins.

Nevertheless, it does appear that the Start Complex represents an (?) Early Devonian basin in which accumulated a sequence that is in some ways similar to that of the Gramscatho Basin. With continued extension, basic volcanics of N-type MORB characteristics were erupted onto the basin floor, and may have been deformed in the process. These lavas/volcaniclastics were finely interbedded with the basinal sediments. The present day metavolcanics probably do not represent true ocean crust (*cf.* the Lizard), as they are underlain by pelites, but are evidence of incipient crustal formation.

The extent of the Start 'basin' is difficult to assess. Despite the strong similarities, it is impossible to say geochemically whether the Start Complex forms part of the Gramscatho Basin. The limited extent of the Start-type schists in offshore cores (see Fig. 3.6) does tend to suggest that there is was no link between the two areas. Sadler (1974) noted that Devonian palaeogeography indicates the presence of a sedimentary basin separating the Lizard and Start areas. Holdsworth (1989a) suggested that the NW-SE dextral faults which cross SW England, acted as oblique linking fault segments, defining a series of small pull-apart basins. He also suggested that the Eddystone Gneisses, bound to the east by the Plymouth Bay Fault, represents the Normannian High (Fig. 3.7) further implying that the Start 'basin' is spatially distinct from the Gramscatho.

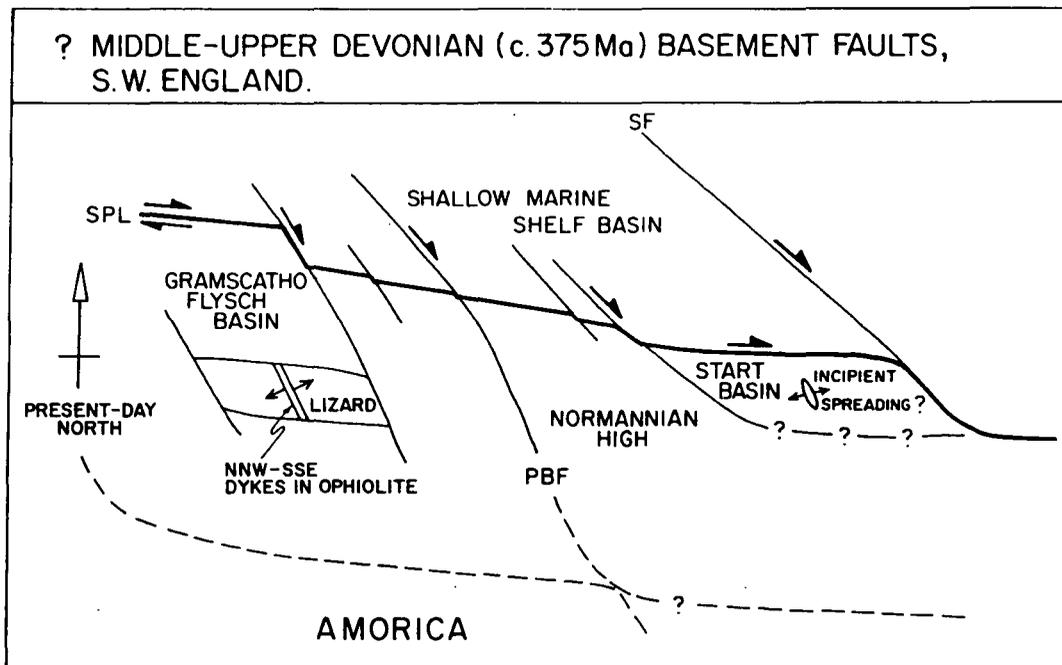


Figure 3.7 Speculative sketch map showing suggested fault and basin configuration in pre-Devonian basement c. 375Ma. SPL, Start Perranporth Line; PBF, Plymouth Bay Fault; SF, Sticklepath Fault. After Holdsworth (1989a).

However, the NW-SE dextral faults cannot be proven to be early Devonian in age, and are more probably related to later Variscan orogenesis. For example, Coward and Smallwood (1984) argue that these faults represent reactivated lateral ramps to NNW transporting Variscan thrusts, whilst Rattey and Sanderson (1980) note that differential thrust movement could have generated transport parallel (i.e. NNW trending) strike-slip faults across this region. Thus these faults may not have been present during basin evolution. Franke *et al.* (1989) suggest that, because of similarities between the Start schists and the metavolcanics of south Cornwall, these two units were deposited in a single 'Cornwall Basin' which formed during Late Silurian to Early Devonian times. It appears that the size of the 'Start Basin' and its relationship to the Gramscatho basin and/or Cornwall Basin is still unclear, although the available evidence does show that even if these sequences evolved in spatially distinct basins/sub-basins, they do show a remarkable number of similarities.

The subsequent closure and inversion of these basins, and the tectonic significance of the proposed major E-W basin bounding fault along their northern margins, is discussed in 4.5, 4.6 and 5.6.

CHAPTER 4

THE STRUCTURAL EVOLUTION OF THE START COMPLEX AND ADJACENT REGIONS

4.1 Introduction

This chapter presents the results of the detailed structural mapping and observations that were carried out across the Start Complex and adjacent Devonian shales (Fig. 4.1). The small-scale structures of the greyschists and greenschists are described in chronological order from representative localities (sections 4.2.3, 4.2.4). Following this, the narrow zone of intense deformation straddling the Start Boundary Fault is dealt with (section 4.3). Then the structure of the Devonian sequences to the north is detailed, allowing a comparison between the Start Complex and the typical deformation style of SW. England (section 4.4). A review of the structural correlation between the Start Complex and the Devonian sequence (section 4.5) is followed by a final discussion on the regional setting of the Start Complex, and the evidence for dextral transpression therein (section 4.6).

4.2 The Start Complex

4.2.1 Introduction

Distribution of Lithologies

The large-scale structure of the Start Complex has been variously interpreted. Bonney (1884) suggested that the greyschists were a distinct unit lying stratigraphically above the greenschists, whilst Ussher (1904) concluded that the greyschists lay beneath the greenschists, forming the core of a greenschist anticline on the east side of the Salcombe estuary. Tilley (1923) provided a more accurate interpretation of the regional structure of the Start Complex, recognising two identical units of greyschist repeated by a single greenschist formation, all folded into a west plunging antiform (Fig. 4.1). Marshall (1962) suggested that this structure was D2 in age, an observation confirmed by Hobson (1977), who mapped out several large scale parasitic F2 folds. Hobson (1977) suggested that the greyschists were a single formation repeated across an early (F1) isocline, cored by greenschists. He proposed that major F1 isoclines were originally present in the greenschist unit, but that their inverted limbs had been cut out by northerly directed (D1) tectonic slides. These slides, which lay close to the greenschist/greyschist boundary, would have inverted the sequence, and led Hobson



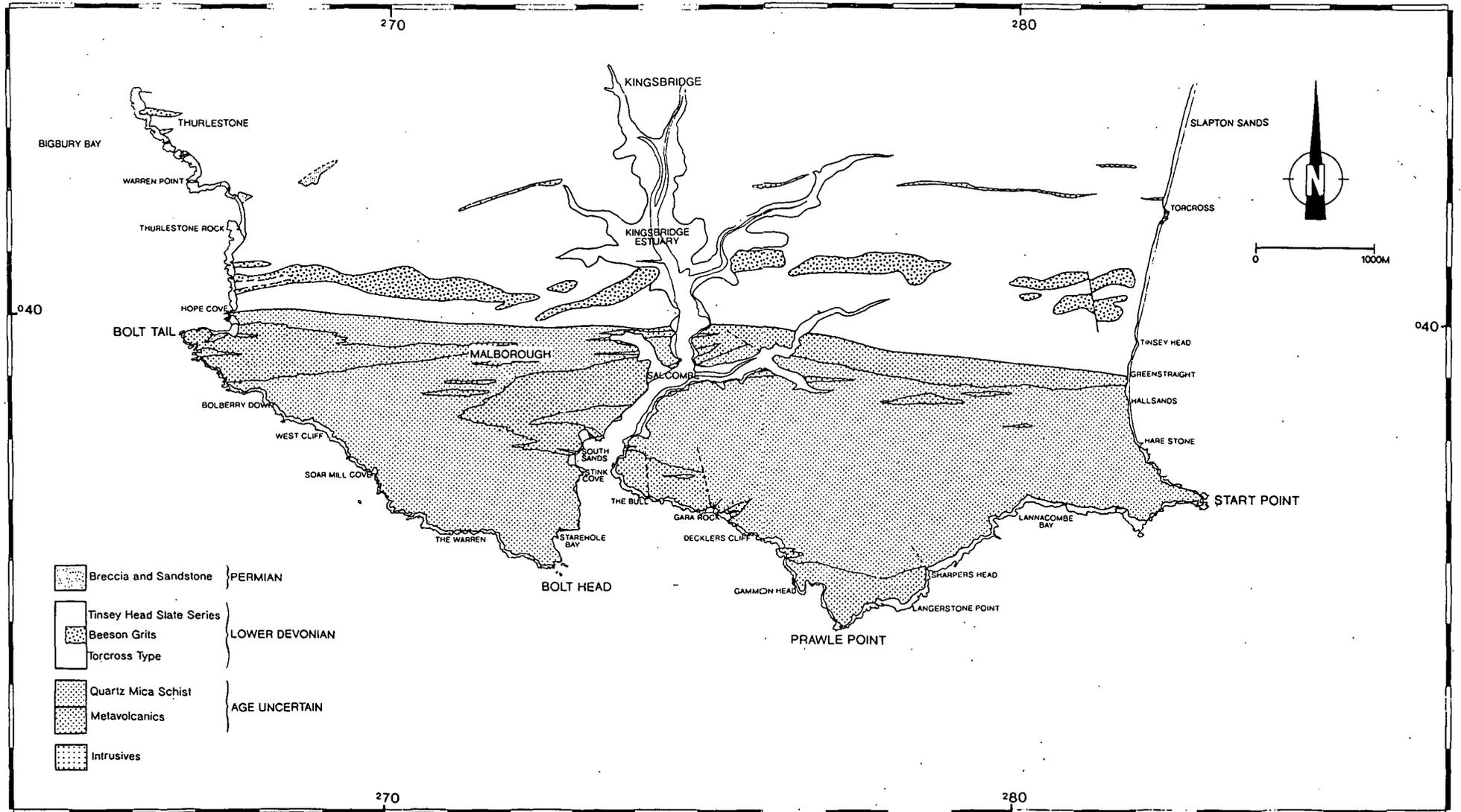


Figure 4.1 Geological map of the South Hams district showing the key localities referred to in the text. Geology of the Start Complex based on the authors own mapping. Geology of the Devonian sequence based on BGS sheets 355/356.

(1977) to conclude that the greenschists were two separate formations (see 4.2.6). However, the remapping of the Start Complex, carried out in the course of this study, has somewhat altered the regional interpretation of the lithologies.

The present study suggests that the greenschists east of Salcombe do comprise a single folded horizon, the antiformal fold hinge plunging gently SW through Rew [SX 2715, 0383] and West Cliff [SX 2691, 0382], rather than westwards through Bolberry [SX 2691, 0393] and Bolt Tail [SX 2669, 0395; Fig. 4.1]. Hence, the greenschist units at Bolt Tail must form a separate and structurally higher horizon. Also, in light of the observations of Seago and Chapman (1988), and the mapping carried out during this study, the age of this antiform is now interpreted as D3, with F3 folds being the predominant structure on all scales across the Start Complex (see 4.2.6). Careful study of the primary folds in the greenschist units suggests that Hobson's (1977) theory of D1 slides is incorrect (see 4.2.6).

Assessing the thicknesses of the individual units comprising this regional fold is complicated due to the degree of early folding and tectonic interleaving. This is especially the case in the more massive greenschist units, where all remnants of the primary structure is usually lost. The present thickness of the greenschists is quite variable, the largest outcrop occurring at Prawle Point [SX 2772, 0350], where the successions of greenschists are at least 600m thick (Fig. 4.1; 4.4). The Portsmouth outcrop [SX 2743, 0373] on the southern limb of the fold is ~400m thick, whilst its continuation into South Sands [SX 2730, 0381] preserves a maximum thickness of 300m (Fig. 4.5; 4.3). The northern fold limb is cut out against the Start Boundary Fault and, as a result, the maximum recorded thickness is reduced to about 250m. The upper greenschist horizon, exposed at Bolt Tail [SX 2669, 0395] contains considerable amounts of interfolded greyschist and is 500-600m thick, thinning dramatically eastwards towards Malborough [SX 2710, 0397]. The original thickness of all these units was clearly considerably less than at present and, at some outcrops, e.g. Shapers Head [SX 0785, 0358], the degree of tectonic thickening due to isoclinal folding is clearly preserved.

4.2.2 Previous deformation chronology

The re-interpretation of south Devon in terms of a thrust belt model (e.g. Coward & McClay, 1983) brings into question the validity of using traditional chronological notations of structural elements, e.g. D1, D2, etc. Tobisch and Paterson (1988) also point out that deformation partitioning, essentially the concentration of deformation into discrete domains in a rock mass, can affect the regional deformation chronology. For example, different structures can form simultaneously in contiguous domains,

whilst individual thrusting events may produce a local structural sequence with little or no regional significance. Thus, in a regional correlation, strict adherence to a traditional chronological notation could produce a very misleading picture, and it would be more valid to use regionally consistent terms such as primary deformation, secondary deformation, etc. However, given the small size and consistent deformation style of the Start Complex, it is felt that the traditional deformation chronology may be used locally for simplicity, whilst remaining aware of the regional limitations of using such a system.

Ussher (1904) suggested that although the lithologies of the Start Complex appeared to exhibit a higher metamorphic grade than the Devonian slates to the north, the two areas had a common deformation history. He reviewed the previous literature and showed that there were two different theories concerning the regional setting of the Start Complex. The first, proposed by Bonney (1884) and Raisin (1887), suggested that the Start Complex was considerably older than the Devonian sequence to the north, and separated from it by a fault. The second theory favoured progressive metamorphism (e.g. Holl, 1868; Somervail, 1887; Hunt, 1892), possibly within the aureole of a granite further to the south. The conclusions drawn by J.J.H. Teall in the Memoir (Ussher, 1904) were quite vague, suggesting that the survey work tended to confirm the latter theory, as the two areas had a number of strong similarities. However, Teall also noted that these similarities did not rule out the possibility of there being a faulted junction.

Marshall (1962) was the first to describe the deformation chronology of the Start Complex in detail. He initially suggested that all phases of deformation were common to both the lower Devonian shales and the adjacent Start schists. His chronology is as follows:

D1 Deformation: Small, isoclinal recumbent folds which plunge moderately ($10-20^\circ$) to the west ($250-280^\circ$). The axial planar schistosity dips slightly to the south, but is generally strongly overprinted by D2.

D2 Deformation: These folds plunge gently westwards, have vertical axial planes and exhibit no axial planar schistosity. The S1/S2 intersection is parallel to the F2 fold axes and results in a strong mica crenulation and quartz rodding.

D3 Deformation & D4 Deformation: Two sets of sub-vertical, D3 kink bands, trending NE-SW and NW-SE, with dextral and sinistral vergence respectively were noted. A second, later set of gently inclined D4 kinks was also noted, but were thought to be very weakly developed in the schists.

Subsequently, Marshall (1965) suggested that F2 folds in the Start Complex were coeval with F1 in the Devonian, implying that there is an additional early deformation phase in the Start schists which has no counterpart in the Devonian shales.

Although his modified correlation now appears to be correct (see 4.2.3, 4.2.5), it was at the time based solely on the upright attitude of F1/F2 fold axial planes, rather than on any study of associated fabrics. Hobson (1976b) refuted this correlation, suggesting that the upright attitude of the F1 folds in the Devonian sequence was the result of later refolding by an open, late antiform (the Dartmouth Antiform). Marshall (1965) also examined the Start boundary in detail, concluding that it was a steep F2 fault, and disagreed with Hendriks' (1939) hypothesis that it represented a major Variscan thrust.

Hobson (1977) also recognised four deformation phases in the Start Complex and advocated direct correlation of these deformation phases across the Start Boundary Fault. He also described what he considered to be a distinct set of "post-D2" recumbent folds which had an unclear relationship with the kink bands (D3 & D4 *sensu* Marshall, 1962). Hobson (1977) noted that these late recumbent folds had no counterpart in the Devonian sequence to the north, although they were similar to those recorded in south Cornwall, leading him to suggest that the structure of the Start Complex was similar to that of south Cornwall, and that the Start Boundary Fault was therefore an extension of the Mevagissey-Perranporth tectonic line. The regional significance of this 'phase' of recumbent folds is discussed in 4.6.3. In considering the regional setting of the Start Complex, Hobson (1976b) made the important observation that early (F1) south-dipping, north-facing folds steepen southwards along the western coast of south Devon, ultimately passing through the vertical to dip steeply north and face south adjacent to the Start Boundary Fault. Richter (1967) had previously shown that this also occurs along the east coast. Hobson (1976b) attributed this phenomenon, at least in part, to refolding (see 4.5).

Seago and Chapman (1988) reviewed the relationships of deformation phases in this area, and reconfirmed the presence of an early deformation phase in the Start Complex, which was not preserved in the Devonian sequence to the north.

4.2.3 The Minor structures of the Greyschists

Pre-F3 Structure

The greyschists are the predominant lithology at outcrop level across the Start Complex and are remarkably consistent in their deformation style.

i) Early folds and fabrics: There is no trace of original bedding in any of these schists, the earliest visible fabric always being metamorphic in nature and associated with a penetrative early cleavage. They are strikingly lineated, with many specimens fracturing readily into mullions, suggesting that this is an L>S tectonite. Small centimetre scale, intrafolial isoclinal folds, which plunge gently both ENE and WSW (Plate

4.1) and lie at low angles to the lineation. These are generally poorly preserved and, together with the lineation, are refolded by later structures (Plates 4.1, 4.2). On close examination, these intrafolial folds are seen to deform thin quartz veins or metamorphic segregation fabrics, which lie parallel to an early, sub-parallel penetrative cleavage. This is confirmed in thin sections of the early fold hinge regions (see 4.2.5). Thus, these intrafolial folds must be at least D2 in age. The mineral stretching lineation is best seen on the surface of folded quartz veins (Plate 4.2), where it is defined by elongate quartz grains. Many of the greyschists exposed along the coast at West Cliff [SX 2691, 0382] are rich in iron pyrites. The larger (up to 2cm) euhedral pyrites are wrapped by quartz pressure shadows, which lie parallel to the stretching direction (Plate 4.3).

F3 Folds

The structures in the deformed greyschists at Great Mattiscombe Sands [SX 2817, 0369] are typical of this lithology across the entire Start Complex. Here centimetre to 10's centimetre scale, open to tight folds, which vary in attitude from upright to recumbent, are dominant (Plate 4.4). F3 folds deform small F2 intrafolial isoclinal folds (Plate 4.1) and whilst this refolding appears at first sight to be coaxial, many of the later folds clearly refold the mineral stretching lineation, which can be shown to be F2 axis-parallel. Although in some cases this refolding is strictly coaxial (L2 parallel to F3 axes), in most examples there is a degree of obliquity, the lineation lying several degrees clockwise of F3 axes, or even wrapping around the later fold core by up to 30° (Plate 4.2). Rarely, anticlockwise lineations are also observed, and are statistically significant enough to warrant attention. The above relationships suggest that the F2 folds are rotated clockwise prior to refolding by F3.

In general, F3 folds are gently whaleback, plunging from E-W through to ENE-WSW, although some display a marked variability of plunge, from steeply E, through the horizontal, to moderately WSW, resulting in pronounced whaleback geometries. The variation of F3 plunge may be primary, as evidenced by variations in the angle of refolding of the early mineral lineation, or may be secondary, possibly due to later modification by shears/faults, e.g. at Mill Bay [SX 2741, 0382], where there is a consistent relationship between the mineral lineation and the variably plunging axes.

Locally, F3 folds are variable in vergence, reflecting the frequency of metre scale structures, although on a larger scale, vergence is very consistent throughout the Start Complex (See 4.2.6)

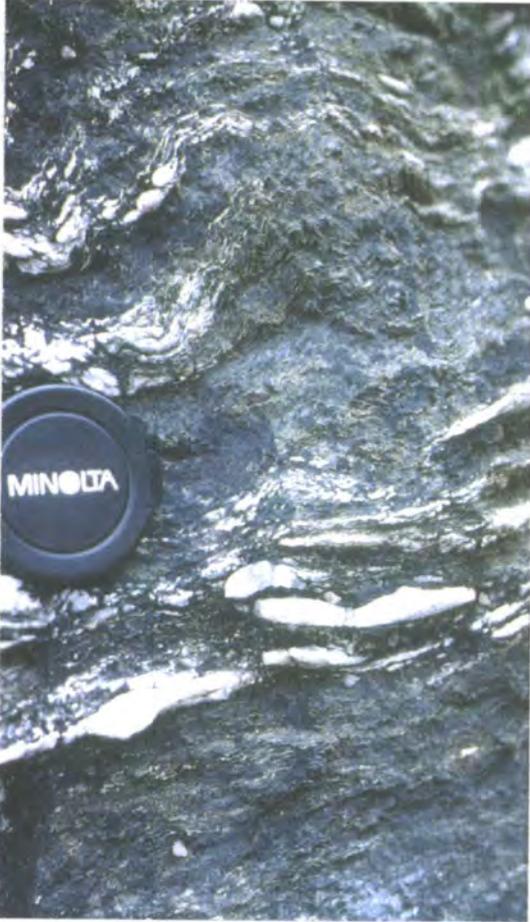


Plate 4.1 Flat lying intrafolial F2 isoclinal folds in dark pelitic greyschist cut by sub-vertical weak S3 cleavage. Note that the fine quartz veins are strongly crenulated by S3, whilst the thicker veins are largely unaffected. Greyschists west of Bolt Head [SX 2720, 0361].

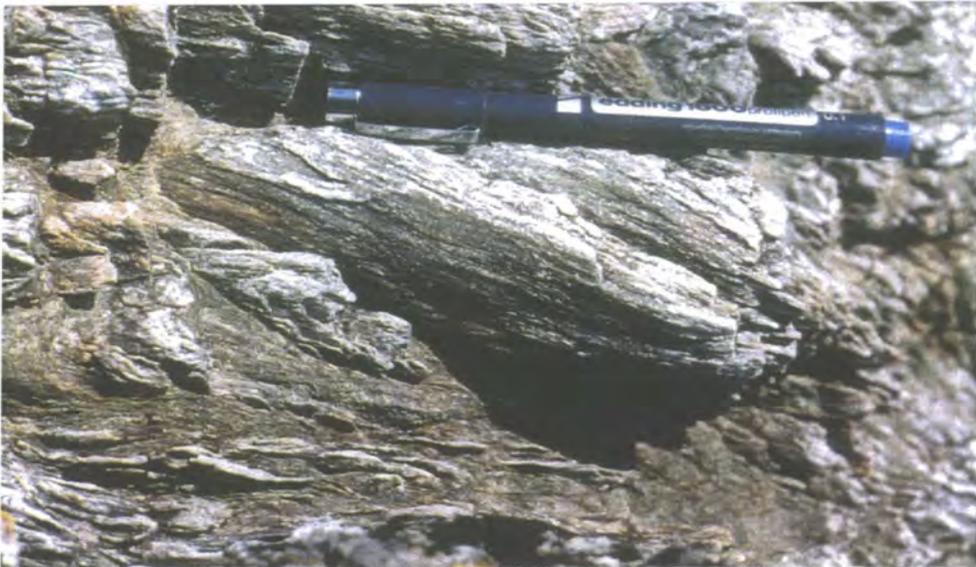


Plate 4.2 Refolded L2 lineation wrapping clockwise around the core of a gently east plunging F3 fold (pen approximately parallels fold plunge). The lineation is defined by a quartz grain elongation. Note also the planar sub-vertical fractures, trending N-S, in the top left of the field of view. Greyschists in the cliffs north of Salcombe Castle, [SX 2736, 0384].



Plate 4.3 Cubes of iron pyrites wrapped by quartz pressure shadows. These quartz tails parallel the stretching lineation and show variable shear senses. Greyschist cliffs at West Cliff [SX 2691, 0382].



Plate 4.4 Typical open, upright F3 folds with gentle whaleback hinges. Fold profile surface is a planar N-S joint. Greyschists forming the wave cut platform at Great Mattiscombe Sands, viewed looking west, [SX 2817, 0369].

S3 Cleavages

The axial planar cleavage associated with F3 folds is variably developed. In part this is lithologically controlled, in that more phyllitic/micaceous greyschists display a fine, well developed, closely spaced cleavage, whilst the coarser, quartz-rich lithologies exhibit a weak and quite sinuous, widely spaced cleavage. In places, this is responsible for the appearance of what appear initially to be two synchronous fabrics, cutting the F3 folds, e.g. in strongly deformed greyschists exposed along the wave cut platform between Great Mattiscombe Beach and Lannacombe Beach [SX 2812, 0372; see Fig.

4.2a]. Here, the variable cleavage orientations appear to result from marked strain discontinuities between deforming layers, so that some layers are intensely folded with centimetre scale folds cut by a strong fanning cleavage, whilst others display 10's centimetre scale open folds with moderate axial planar cleavage development. Close examination shows that this is a single cleavage, variably developed from layer to layer, and that there is no demonstrable overprinting relationship. Further examples occur at Hugh's Hole [SX 2679, 0387] where the quartz-rich greyschists are highly deformed, with much disharmonic folding and variable cleavage development. Here, there are two very clear, discrete and apparently synchronous S3 cleavages associated with a series of steep south verging F3 folds. The steep S3 cleavage (S3a), axial planar to these upright F3 folds, is strongly developed, and results in thin oxide coated pressure solution seams. In the long fold limbs of the F3 folds, a weak sub-horizontal fabric (S3b), fans around small open recumbent folds. However, the weak flat lying fabric is never developed in the upright F3 hinge zones, there are no fold interference patterns, and there is no clearly demonstrable crenulation of either fabric. Thus it would appear that these cleavages, although in places almost mutually orthogonal, are synchronous and part of a single, continuous deformation event. This is further confirmed by several examples (e.g. at Soar Mill Cove [SX 2697, 0375]) of upright F3 folds flattening laterally into sub-recumbent structures, the S3a cleavage weakening in the process to ultimately align with S3b. These patterns inevitably lead to a local variability of F3 vergence, as a consequence of variable fabric/cleavage relationship. The interpretation and possible regional significance of these recumbent fabrics is discussed in 4.6.3.

The excellent three-dimensional exposure of the small-scale F3 folds in the greyschists demonstrates that the cleavage is usually axial planar. However, at Splat Cove [SX 2731, 0375; Fig. 4.3], the strongly developed S3 cleavage, which appears as a 2-3 centimetre spaced, iron oxide coated fracture cleavage, dips moderately to the north and crenulates the S2 fabric. In three dimensions, this cleavage can be seen to be non-axial planar, lying several degrees both clockwise and anticlockwise of the F3 axes. Such non-axial planar fabrics are only a localised phenomenon, and no consistent regional sense of transection is evident (4.6.3).

The lineated appearance of the greyschists may locally be accentuated by a strong intersection lineation, formed between the S3 cleavage and the early metamorphic fabric, which, where sufficiently intense, results locally in a crude pencil schist. The intersection fabric is best developed in the fine micaceous layers of the greyschist, whilst the mineral lineation is often only seen on folded quartz surfaces. The intersection fabric is quite sinuous and usually F3 parallel (Plate 4.5), whilst the mineral lineation has variable relationships with these later fold axes.

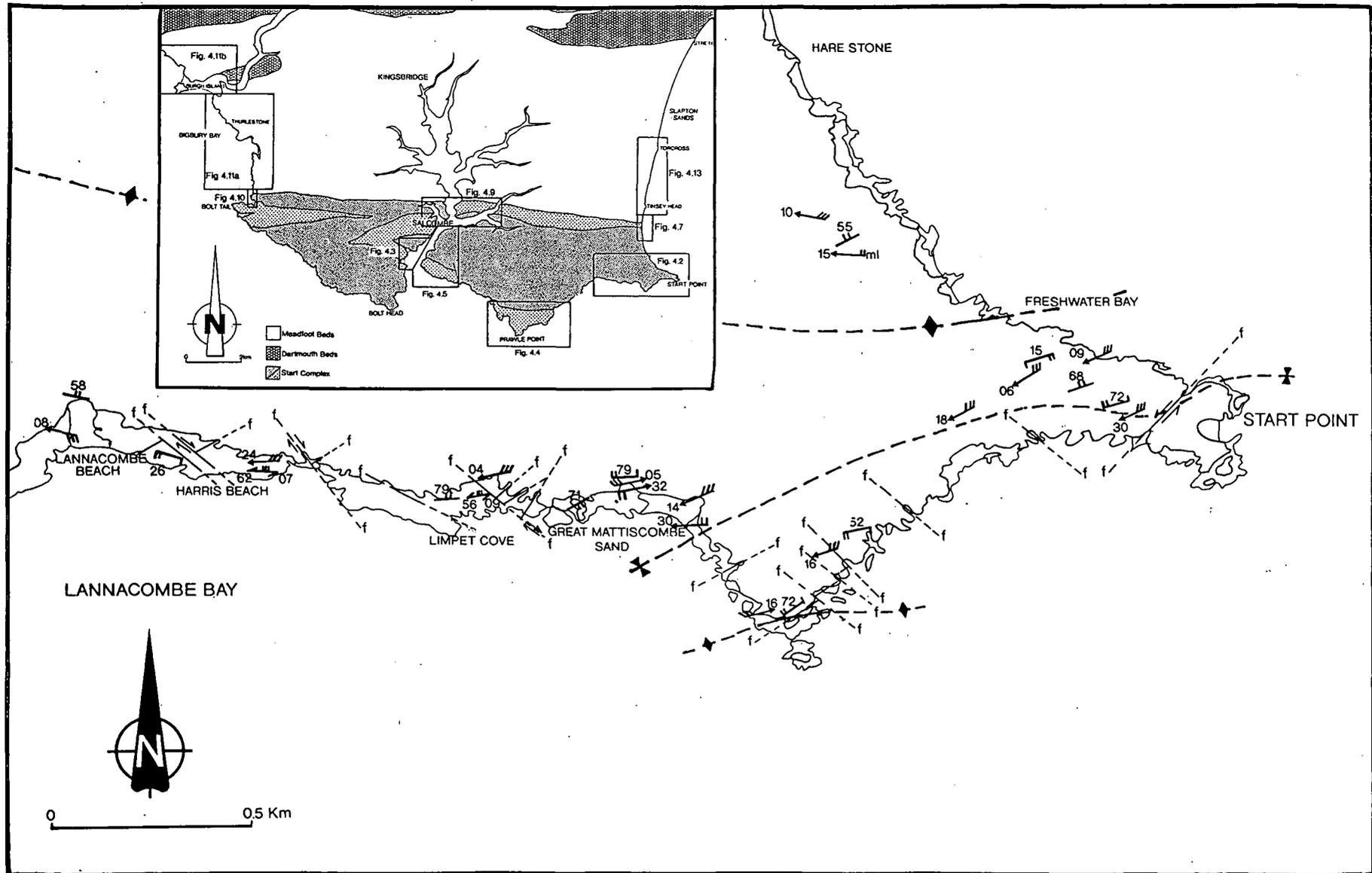


Figure 4.2 a) Geological map of the Start Point area. See Fig. 4.3 for key to lithologies and structure, NB. entire area comprises greyschist. b) Map of the South Hams area to show the location of the enlarged map views referred to in this chapter.



Plate 4.5 Sinuous S3 intersection lineation on the surface of folded quartz veins. Note: i) that the folds tend to fracture along the S3 surfaces, and ii) the approximate parallelism with the F3 axis (pen). Greyschist reefs west of Salcombe Castle [SX 2733, 0381].

Quartz Rodding

The lineated nature of the greyschist is further accentuated by the presence of quartz rods, which weather in relief of the rock surface. The best examples come from the greyschist cliffs north of Salcombe Castle [SX 2734, 0381; Fig. 4.3], where they range from a few millimetres up to 8-10 centimetres in thickness. In cross section, these rods vary from bouquinaged quartz layers, or fold limbs (Plate 4.6), through to rootless fold hinges with a more or less cylindrical cross sections. The quartz rods are essentially the hinges of small parasitic folds which have been thickened, detached and isolated from the limbs during vein attenuation. They parallel fold axes and are useful in that they allow a close examination of the relationship between the fold axes and the refolded lineations.



Plate 4.6 Incipient quartz rod formation Greyschist cliffs at Salcombe Castle [SX 2734, 0381]

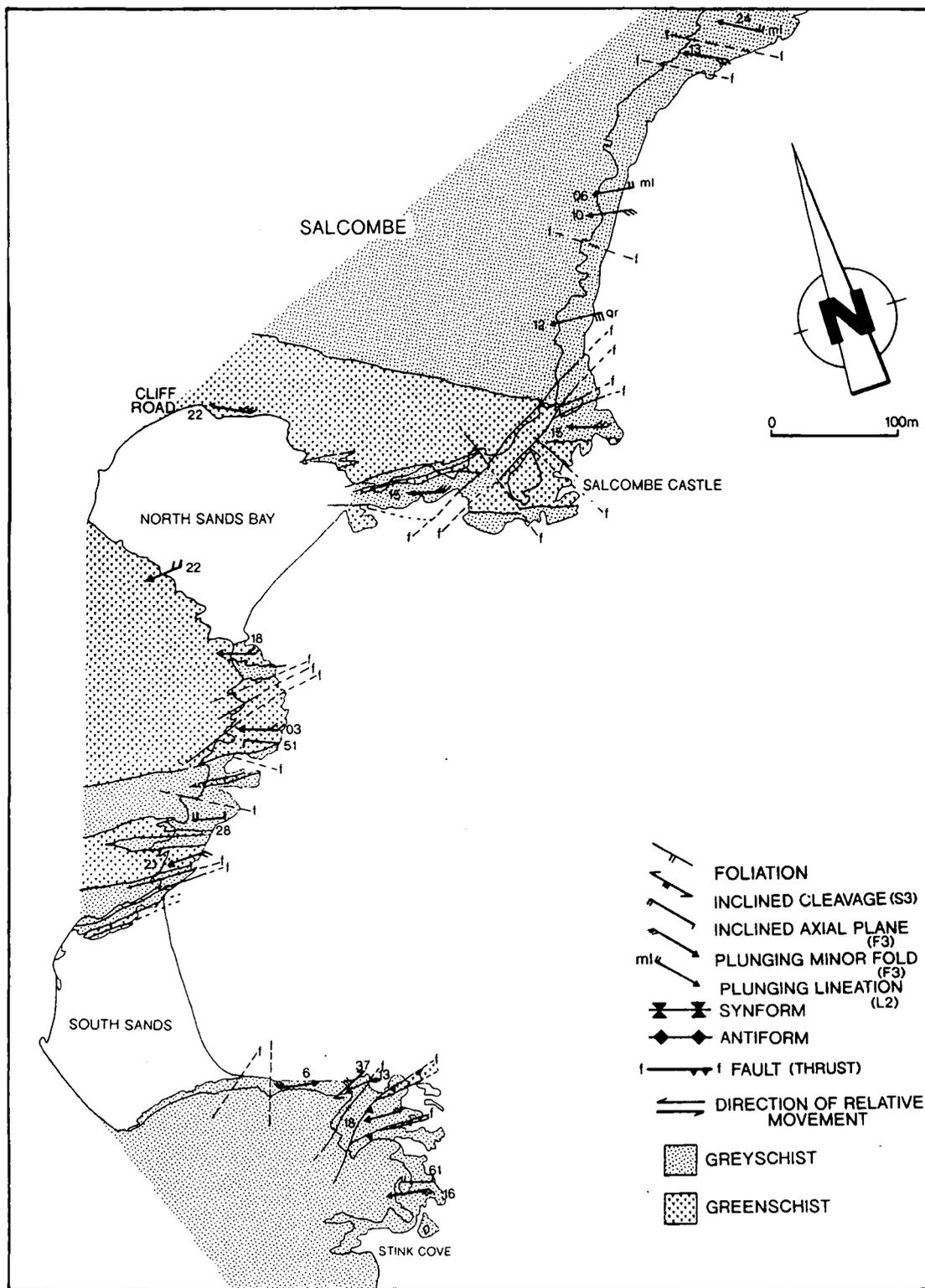


Figure 4.3 Geological map of North and South Sands. Fig. 4.2b shows the location of this map within S. Devon.

Shear bands

Postdating the ductile fold phases in the Start Complex are several sets of brittle/ductile later structures, the most conspicuous and significant of which are variably developed, generally steeply dipping, shear bands. They occur in two distinct orientations; dextral shear bands trending ESE/WNW and sinistral shear bands trending NE/SW.

In areas of high syn-D3 strain, where the greyschists are characterised by very tight, disharmonic F3 folds and pronounced S3 cleavages, the shear bands are poorly developed, occurring as localised discrete curvilinear surfaces with limited (1-10cm) rather brittle offsets. More typically, where syn-D3 strains are lower, the shear bands occur as anastomosing networks of ductile shears defining elongate domains of schist in which the fabric has suffered varying degrees of rotation, due to the displacement on the adjacent shears (Plate 4.7). Dextral shears are generally larger in scale, varying from centimetre spaced up to several metre spaced, and are dominant over sinistral structures. Displacements on the shears appear to be scale-dependant, so that the more extensive dextral shears have the larger offsets. In places, especially in the greyschists at the base of Salcombe Castle [SX 2734, 0381; Fig. 4.3], the shear bands are so intense and closely spaced as to shred the rocks into a 'button schist'.

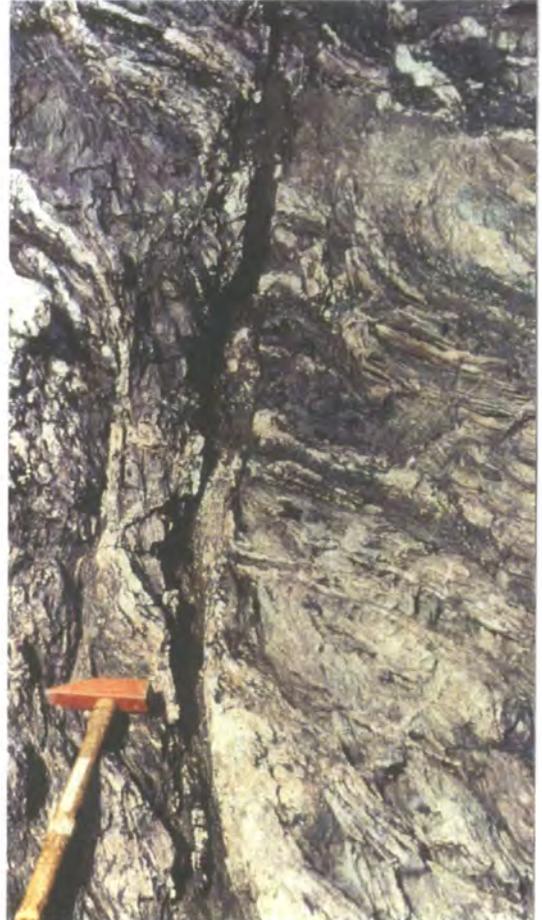
The well exposed shear bands in the greyschist cliffs along the Warren [SX 2709, 0368] are representative of greyschist shears in general and demonstrate a number of points:

- a) Dextral shear bands are clearly predominant, and may occur in two, or even sometimes three, distinctly orientated sets. Offsets are similar in all of the dextral shears and are always greater than the displacements on the discrete, poorly developed sinistral shears.
- b) The relative timing between the dextral and sinistral shears is hard to assess. They would appear to be more or less synchronous, resulting in a variety of cross cutting relationships, although the larger size and greater displacements on the dextral shears gives the impression of them being later.
- c) In many cases, the larger and more brittle shears show varying degrees of dilation, with fibrous infillings of quartz or albite, that may be sheared or offset. (Plate 4.8).
- d) Sinistral shears range from pervasive anastomosing networks of ductile shears, on which there is usually a clearly demonstrable offset and rarely any associated quartz mineralisation, through to the larger and apparently more brittle structures, which occasionally show rare infills of quartz/albite or calcite fibres ^{and} are variably sheared or sinistrally offset. Difficulty arises where the displacements on the shears are negligible, and these shears lie close to N/S orientations, in which case they may be hard to distinguish from extensional fractures.



Plate 4.7 (left) Anastomosing arrays of centimetre scale ductile shear bands in which the schistose fabric has been variably rotated within elongate domains. Greyschist at Mill Bay [SX 2741, 0382].

Plate 4.8 (right) Large dextral shear bands showing dilation and infill with quartz/albite. Note that sub-vertical shear surfaces are coated in quartz/albite, within which growth fibres indicate dextral shear. Greyschist foreshore at the Warren [SX 2709, 0368].



e) Block rotation is very difficult to prove on a larger scale, and is only obvious on a centimetre scale where the shears form fine anastomosing networks (e.g. Plate 4.7). Although shears with anomalous orientations (e.g. sinistral shears parallel to dextral shear orientations) have been recorded and may relate to large scale block rotation, there is no direct evidence of this phenomenon. However, in those domains where the strike of the schistose foliation is locally anomalous, this may usually be accounted for by drag imposed along contiguous shears/faults and may often be traced laterally, without break, in to areas of more usual strike.

f) Many of the larger, well exposed dextral shears have a marked component of locally consistent dip-slip displacement, evidenced by fabric rotation and offset quartz veins in vertical sections. Unfortunately fibres are rarely preserved on the shear surfaces, although considerable evidence along this section, of oblique displacement on shear bands, corroborates the observations made on the larger faults at Hope Cove (see below). Oblique displacements on shears is discussed in 5.6.

g) In places, small brittle kink bands could be traced a few 10's of centimetres along strike into ductile shears, suggesting a continuum of structures between ductile shears, brittle shears/fractures and the late brittle kink bands. The offset folia between the kink planes are replaced by displaced quartz fibres within the shears, with both giving a consistent sense of shear.

Extensional Fractures

Sub-vertical, ~N/S trending extensional fractures, are consistently developed across the Start Complex, irrespective of lithology or deformation state. They vary from laterally extensive planar joints, e.g. along Great Mattiscombe Sands (Fig. 4.2a), where large regular planar joints dissect the wave cut platform (see Plate 4.4), down to short (10-15cm) fractures, occasionally infilled with quartz or albite (see Plate 4.2). Where mineralised, the fibres are always normal to the fracture wall and show no evidence of horizontal displacement, thus distinguishing these fractures from anomalously orientated dilatent shears. Occasionally, in three-dimensional exposures, the larger fractures do show a small degree of vertical displacement, causing minor offsetting of the infill fibres. This vertical displacement, which is only ever in the range of a few centimetres, is probably the result of later brittle reactivation of these fractures, as no fracture infill appears to have grown during vertical movement and the original orthogonal fibres are often fractured and displaced.

Late Faults

The greyschists across the Start Complex are cut by a variety of faults, ranging from syn-D3 thrusts, through to sub-vertical strike-slip faults. The steep faults are discrete

planar surfaces, devoid of slickensides/slickenlines, but often with much associated brecciation and narrow (30-80cm) zones of friable fault gouge. A characteristic feature of the faulting in this area is the extensive iron oxide staining which occurs in the adjacent schists extending for several metres away from the fault zone (Plate 4.9). This, allied to the deformation associated with the faulting, results in a lithified, deep maroon, fault rock in which individual lithologies may be indistinguishable. In general, the kinematics of individual faults are hard to assess, especially in the absence of slickenside indicators. This is exacerbated by extensive erosion, such that most steep faults have smooth weathered surfaces, and may even be completely eroded to leave deep gulleys which dissect the foreshore. However, in many places, it was noted that these steep faults are approximately parallel to the small ductile shear bands, e.g. at Limpet Cove [SX 2812, 0369]. Here, the sub-vertical faults appear to have two distinct trends; an early NE-SW trending set, cut by an apparently later NW-SE trending set, some of which appear to have a dextral offset (Fig. 4.2a). Also, many of the shear band arrays tend to intensify into fault zones, e.g. at Deckler's Cliff [SX 2757, 0366], with dextral shears intensifying into the NW/SE faults and the sinistral sets intensifying into the NE/SW faults. Thus offsets on these faults may be broadly assessed, in the absence of more direct kinematic indicators, by their relationships to the small ductile shears. In some areas, shears antithetic to the larger faults occur, a phenomenon which is seen on a variety of scales. For example, along the southern edge of Hope Cove [SX 2673, 0397] a series of steep NNE/SSW faults display fibres with a sinistral oblique offset sense (see Fig. 4.10). These faults appear to be antithetic to a major dextral WNE/ESE fault, which cuts through the cove, and forms a deep valley running inland at least as far as Bolberry [SX 2691, 0393].



Plate 4.9 Thrust fault within interleaved greenschist and greyschist. Note the extensive iron oxide staining, especially within the footwall. Salcombe Castle, [SX 2734, 0381].

Occasionally, rotation of the schistose fabric has occurred about a steeply plunging axis adjacent to the faults. Rotation is consistently clockwise adjacent to the NW/SE faults, and anticlockwise adjacent to the NE/SW faults (Plate 4.10). This rotation of fabric can also be used to determine offset sense. For example, at the eastern end of Start Point promontory [SX 2829, 0372], a large NE/SW fault is responsible for much of the recent cliff collapse in the area (see Fig. 4.2a). Anticlockwise rotation of the schistose fabric strike, F3 fold axes, etc. adjacent to this fault indicates that it is sinistral. Similarly, at Soar Mill Cove [SX 2697, 0375], a 40° clockwise strike rotation adjacent to WNW/ESE faults indicates dextral offsets.

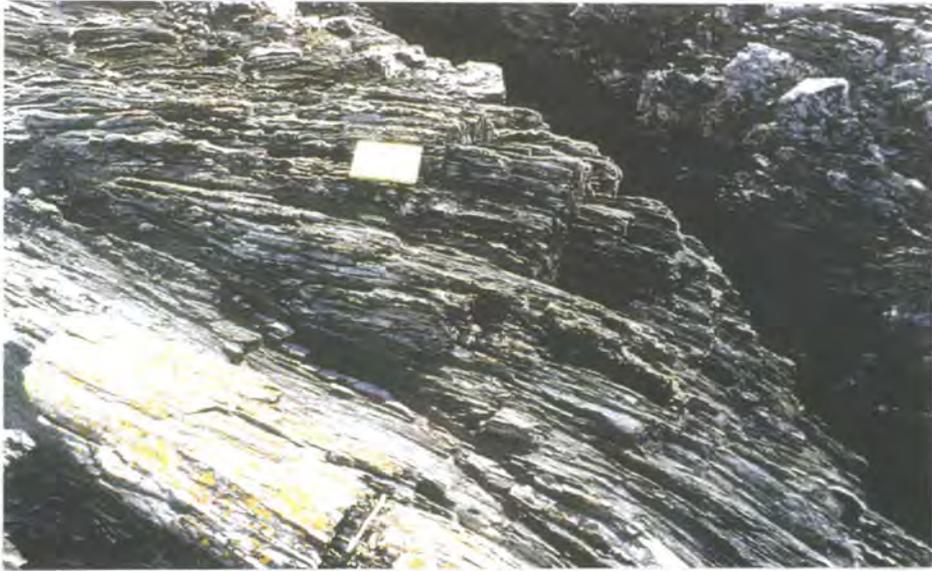


Plate 4.10 Clockwise rotation of the schistose strike adjacent to a large brittle dextral fault. Note that the planar N-S joints have a consistent orientation, and thus post-date the fault movement. Greyschist, Starehole Bay, [SX 2726, 0364].

At Salcombe Castle [SX 2733, 0381; Fig. 4.3], thin offset horizons adjacent to large scale faults indicate complex fault histories, although offset horizons are not always accurate indicators of displacement sense (*cf.* Hobbs *et al.* 1976, Fig 7.15). For example, along the NW edge of Salcombe Castle, a NW dipping *thru* fault runs parallel to the cliffs and is associated with extensive mineralisation (see Plate 4.9). This fault, and a second sub-parallel structure, appear to dextrally offset the greenschist/greyschist boundary, and are cut by several steep NW-SE orientated faults with apparent sinistral offsets (Fig. 4.3). Thus it would appear that the large brittle faults have the opposite sense of displacement to the small scale ductile shears. However, fibres on the prominent normal fault indicate transport towards 283° and this, allied to the swing in strike of the adjacent fabric, shear bands, etc, and the orientation of extensional vein arrays, clearly indicates a significant sinistral oblique component of movement. Thus, despite these faults appearing to have up to 50m of

dextral displacement, close examination of the fault planes proves a significant component of sinistral oblique reactivation.

Low angle thrusting is generally poorly developed in the greyschists, shortening being accommodated by folding on a variety of scales rather than by layer parallel slip. However, in the greyschists around Splatcove [SX 2731, 0375; Fig. 4.3] several northwards transporting thrusts have interleaved ~1m slices of greenschists into the greyschists.

4.2.4 The minor structures of the Greenschists

Pre-F3 Structure

The greenschists range from massive exposures in which the only visible structures are the fine, crenulated quartz veins and occasional shear bands, through to intensely folded outcrops. In general, the syn-D3 strain seems to be lower in the greenschists relative to the greyschists, and in places a dramatic strain discontinuity may be seen where the two lithologies are interbanded. The pre-F3 structure is thus generally better preserved in the greenschists.

a) Early Folds and fabrics

The earliest folds in the greenschists are intrafolial isoclines with variable attitudes, which deform a strong metamorphic fabric and sub-parallel quartz segregations. They range from small, centimetre scale structures, through to large (10's centimetre up to metre) scale sub-recumbent folds. These larger folds have the effect of markedly thickening the greenschist sequence (Plate 4.11). There is no evidence of a primary layering, even in the very low syn-D2 strain outcrops, and it appears that the earliest visible fabric in all the greenschist outcrops is always metamorphic, and associated with a sub-parallel early cleavage. The early folds would thus appear to be F2 structures.

Along the south side of North sands beach, are several upright 10's centimetre scale F2 folds are refolded by gently inclined to recumbent F3 folds (Plate 4.12), forming Type III interference patterns (Ramsay, 1967). Between North and South Sands [SX 2730, 0378], many of the greenschist units display well developed arrays of tight, north verging F2 folds, with somewhat angular geometries. Moving southwards to South Sands [SX 2728, 0376], there are numerous upright, moderate to tight F2 folds, occasionally cut by a flat-lying S3 fabric, which must have considerably thickened the sequence, although the degree of thickening is impossible to assess given the essentially homogeneous nature of the greenschists. Where the two lithologies are intimately interbanded, the degree of F2 thickening becomes immediately apparent

(Plate 4.13), and it would appear that this effect has been largely neglected by earlier workers (e.g. Hobson, 1977a), who related the changes in centimetre scale F2 vergence to regional structures (e.g. his flat lying isoclines) rather than to these metre and 10's of metre scale folds. Where the vergence of the F2 interfolds can be assessed, it is generally to the north, as is the vergence of the centimetre scale F2 folds in the massive greenschist units. This vergence is very consistent in the greenschist units throughout the Start Complex.



Plate 4.11 Flat-lying metre scale F2 folds in greenschist, which considerably thicken the sequence and may produce local vergence changes. Note the small localised 'hook' refolds produced by later F3 refolding (top right). Greenschists at Sharpers' Head [SX 2785, 0358].



Plate 4.12 Type III (Ramsay, 1967) interference pattern resulting from the refolding of a flat-lying F2 isocline by an upright F3 fold. Greenschists along southern edge of North Sands beach, [SX 2731, 0380].

At the Hipples [SX 2737, 0377; see Fig. 4.5], a sequence of low strain greenschists allows a detailed examination of F2 structures. In some of the very low strain areas, it is easy to see why previous workers described the early folds as D1, as the deformed layers look original, and there is even a suggestion of volcanoclastic grading. In detail, however, the primary folds always deform a previous tectonic fabric, defined by aligned quartz and epidote segregations. In general, F2 folds in the greenschists across the Start Complex have a characteristic geometry, with hinges and short common limbs thickened relative to the attenuated southerly dipping long limbs. Hobson (1977) attributes the attenuation of the normal limbs to small localised tectonic slides, but in the absence of any ductile displacements, even on a centimetre scale, it seems more likely that these F2 geometries relate to syn-F2 vertical shortening. Although Hobson's (1977) small-scale slides are not seen, there are many large (metre scale), approximately fabric parallel, brittle detachments (see Plate 4.16), which clearly thin the overall sequence. These are probably due to later extension and may be related to the many gently south dipping, normal faults, which occur in arrays all along this section. Displacements along these extensional detachments may also account for the development of localised sub-horizontal cleavages which Hobson (1977) described as cutting the F2 folds in this area, and which he related to D4 refolding. Hobson (1977) also attributes the local steepening of the F2 axial planes in this section to later F4 refolding. This area is cut by gently NE dipping thrust faults and, as will be discussed in section 4.5, backthrusting on a variety of scales is well developed across the South Hams region, and it is quite likely that the steepening here is due to backfolding, and not to late D4 refolding.



Plate 4.13 Metre scale F2 interfolds of greenschists and greyschist. Note the generally massive appearance of the core, with the S3 cleavage visible in the greyschist not crenulating the greenschists fabric here. The marked strain discontinuity between the two lithologies would appear to result from recrystallisation of the greenschists. Rickham Common, [SX 2736, 0375].

b) Early Lineations

A well developed mineral stretching lineation is seen throughout the greenschists, and is usually defined by the parallel alignment of prismatic amphiboles. Occasionally, pyrite or feldspar aggregates occur, apparently elongated parallel to this lineation. In some cases, these record a genuine stretch, as evidenced by stretched and fractured pyrites and feldspars with asymmetric quartz pressure shadows (Plate 4.14). The fractured and displaced pyrites here indicate dextral displacements (e.g. a in Plate 4.14), as do the displacement controlled quartz fibres around individual pyrite crystals (e.g. b in Plate 4.14), although examples of sinistral shear or symmetrical pressure shadows also occur. However, in many examples, the elongation is an apparent effect, the result of later mimetic growth of tabular minerals along the foliation planes. This strong mineral lineation typically lies parallel to the F2 axes and, whilst in places it is gently refolded by F3 axes, in general it also lies parallel to these later fold axes.

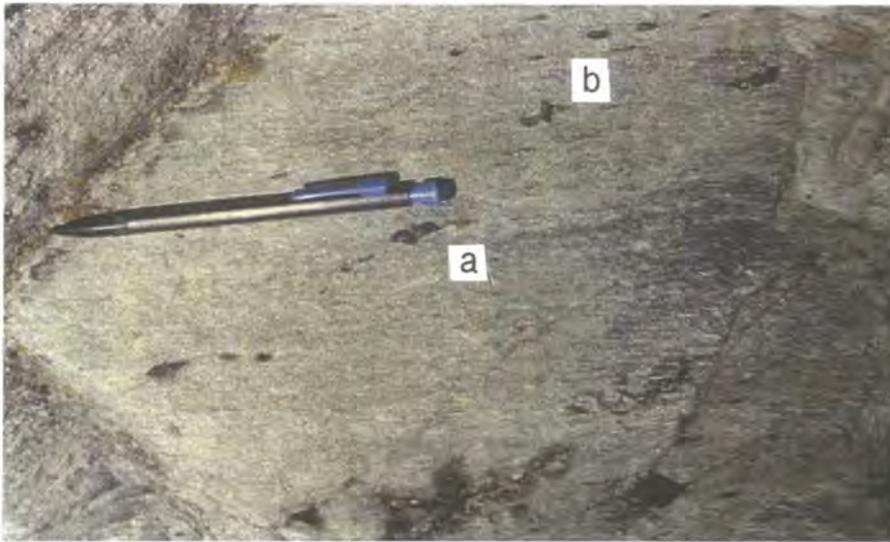


Plate 4.14 Asymmetric quartz pressure shadows around pyrite (?limonite) crystals. Fractured and displaced crystals in the lower half of the field (a) indicate dextral shear, as do the displacement controlled quartz fibres in the upper half (b). All fibres parallel the well developed mineral stretching lineation. Greenschist, Portlemouth, [SX 2738, 0372].

F3 Folds

In many greenschist localities, F3 folds are poorly developed or absent, occurring as rather open refolds of the isoclinal F2 structures. The attitude of the F3 fold axial planes is locally quite consistent, although throughout the greenschists it varies from upright to recumbent. The plunge of the F3 fold axes is likewise locally very consistent, in general being gently to the west. However, where the syn-D2 strain is high, the later folds become more variable in both attitude and plunge, as a consequence of the marked pre-existing anisotropy. A variety of 'hook' interference

patterns are seen in the greenschist cliffs at the Hipples [SX 2737, 0377], generally with steeply south dipping F2 folds refolded by sub-recumbent F3 folds, although gently inclined F2 folds cut by upright F3 folds also occur. Similar interference patterns are seen in the steep greenschist cliffs at The Bull [SX 2743, 0371], where metre scale, flat lying, north verging F2's are locally refolded by south verging, upright F3's. At the Hipples, both the degree and sense of refolding is variable, and is clearly dependant on the variability of the F3 plunge implying that much of this F3 plunge variation is primary in nature.

Where the syn-D3 strain is higher, the F3 folds tighten and the F2 folds are reduced to small, intrafolial isoclines, e.g. in the greenschist exposures beneath Cliff Road [SX 2731, 0382; Fig. 4.3]. The tightening of the folds is generally accompanied by the development of a strong S3 cleavage, although, in many cases, the strong penetrative fabric observed in profile sections which appears to be axial planar to the metre scale folds, is in fact a weak dextral shear band fabric. In good exposure it is always possible to distinguish these two fabrics.

Due to the layered, competent nature of the greenschists, low angle thrusting is an important mechanism in accommodating much of the shortening, whilst associated folds are open, large scale (metre up to 10's metre) features. Much of this thrusting appears to be syn-D3 in age, as the S3 cleavage intensifies locally into the fault planes. In many cases, these thrust related F3 folds strongly refold the mineral lineation. However, there are cases of shortening structures, e.g. folds, cleavages, etc., developing in the hanging walls of large-scale late extensional detachments, which are quite common in the greenschists. These low angle detachments cut the shear band fabrics and some of the large sub-vertical extensional faults, and are probably related to later regional extension, although they can be difficult to distinguish from earlier thrusts, where poorly exposed.

Quartz Rods

These are rarer in the greenschists compared to the greyschists, generally only occurring as isolated structures. As they are essentially modified folded quartz layers, they parallel the fold plunge, and in many of the greenschist profile sections are useful for assessing both the F2 and F3 fold plunge.

Shear Bands

Shear bands in the greenschists are generally smaller in scale and offset compared to the greyschists. These shears are fine, millimetre spaced, somewhat brittle in nature, and with limited offsets in the range of a few millimetres up to 1-2 centimetres. The shears themselves tend to be short and sub-parallel, rarely anastomosing into networks.

Displacements on individual shears are highly variable over short distances, these diminishing rapidly from layer to layer (Plate 4.15). Some of the larger shears are slightly dilatent, with infills of quartz/albite, similar, but on a smaller scale, to those seen in the greyschists. On passing across a lithological boundary into the greyschists, these shears display at least an order of magnitude increase in both spacing and offset, a phenomenon well displayed in the interbanded green and greyschists along the northern edge of North Sands beach [SX 2732, 0382].

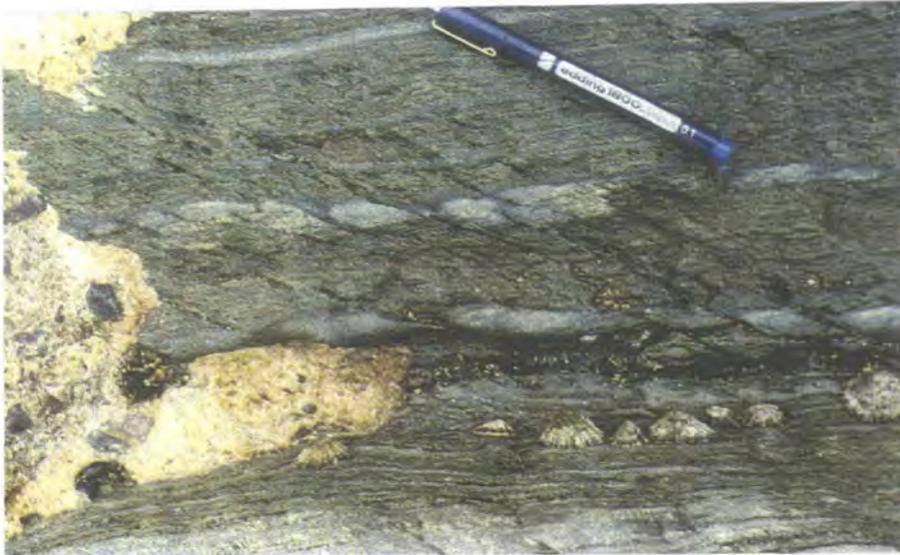


Plate 4.15 Centimetre scale dextral shear bands offsetting thin tuffaceous horizon in massive greenschists. Note that the displacements die out laterally (Pen is parallel to shears, pen top marks layer of zero displacement). Greenschists, northern edge of North Sands beach, [SX 2732, 0382].

At Dutch End [SX 2785, 0356; Fig. 4.4], the greenschists are cut by a series of NE/SW trending sub-vertical extensional shears, one of which displays a morphologically unusual quartz vein infill (Plate 4.16). Here, thin strips of the host greenschist have been detached and folded into the dilating vein to produce a type of bridge structure. Exactly how these structures have evolved and what they represent is discussed in 4.6.5

Extensional Fractures

As in the greyschists, these late structures are remarkably consistent in orientation across the greenschist outcrops in the Start Complex, and range from extensive sub-vertical N/S joints, through to short lenticular fractures with quartz/albite or carbonate infills. Their significance is discussed in 4.6.5.



Plate 4.16 Dilational bridge structures with an apparently sinistral shear sense, within a thick quartz vein. Strips of host greenschist are deformed into a series of 'parallel' folds, some of which are cut by limb parallel detachments, giving the appearance of an along-strike duplex. View is downwards onto the sub-vertical vein. Dutch End, [SX 2785, 0356].

Late Faults

Faulting in the greenschist is more subtle than that of the greyschists, with gently dipping fabric-parallel detachments, which locally both thicken and extend the sequence, being the more common. Extensive iron mineralisation is again associated with the larger faults, producing a hard, lithified maroon fault rock in the more faulted areas, e.g. the steep cliffs along the southern edge of Inner Hope [SX 2675, 0399]. Unfortunately, slickenfibres are rarely preserved on these fault surfaces, with occasional coarse ridges being the only remnant. At Yeovil Rock [SX 2671, 0398], several quite extensive carbonate veins are associated with the steeper faults, material splaying off these veins and infilling fractures, etc. in the adjacent flaggy greenschists. This rather coarse-grained carbonate material is evidence of CaCO_3 rich fluids being active during late faulting in this area (see 4.3.5)

As in the greyschists, there are two broad trends of steeply dipping faults. In this case, a dominant NNW-SSE group and an approximately E-W group. Unfortunately there is little to constrain fault displacements and, whilst the NNW-SSE faults probably accommodate dextral displacements, as evidenced by shear band intensification, movements on the E-W faults are uncertain and possibly variable.

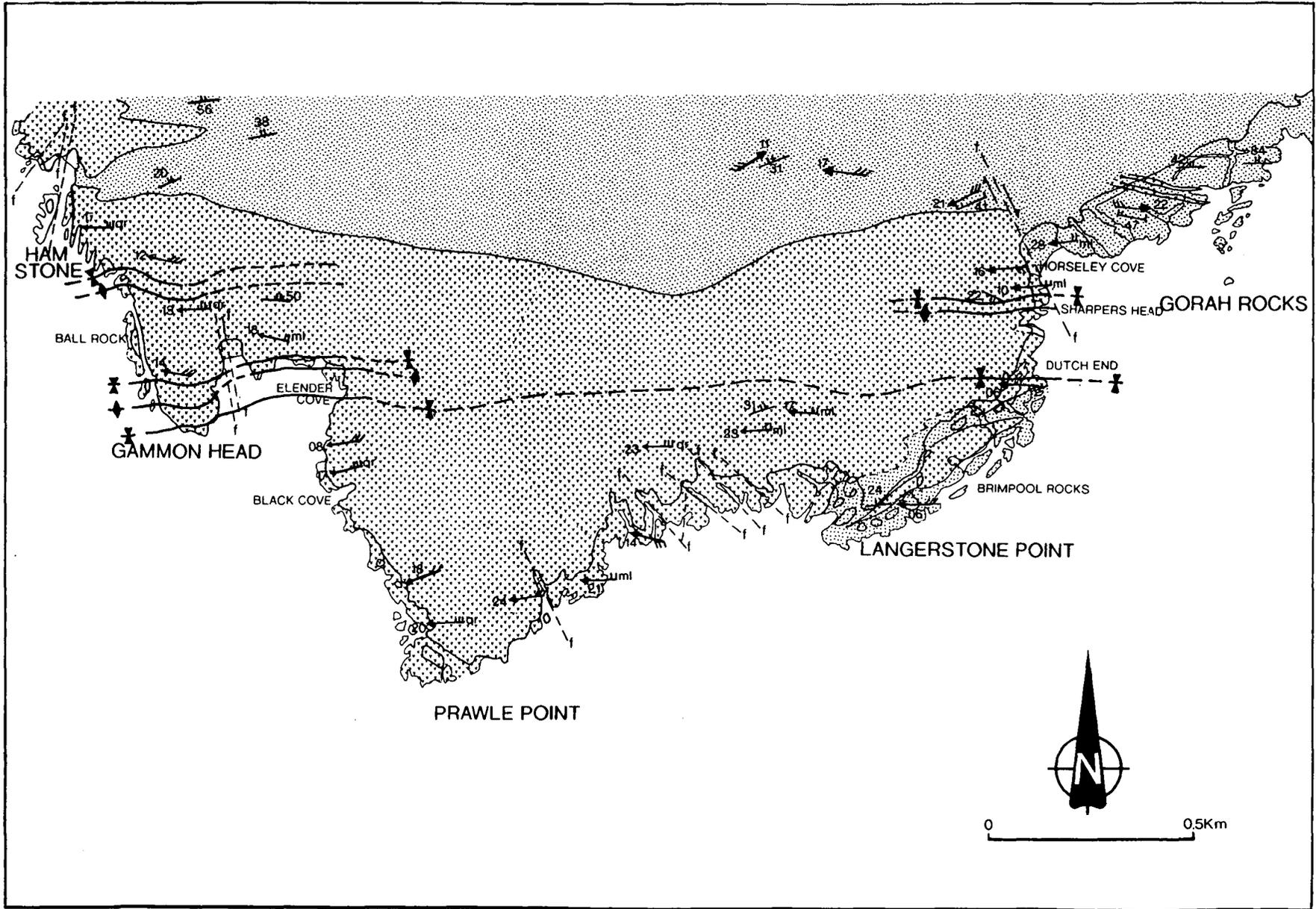


Figure 4.4 Geological map of Prawle Point. See Fig. 4.3 for key and Fig. 4.2b for location in the Start Complex.

4.2.5 The Microstructure

Of the two schistose lithologies in the Start Complex, the greyschists typically appear to be more deformed, displaying a wide spectrum of deformation microstructures, whilst the greenschists are generally massive to poorly foliated.

Greyschists

There is evidence of deformation under a wide range of temperatures, from brittle cataclasis, through to ductile recrystallisation processes. The microstructure detailed is from 'typical' greyschist, with high D2 and moderate D3 strain. Low temperature ($\sim 300^\circ$) plasticity in quartz results in flattened, elongate grains and occasional ribbons, all with a marked undulose extinction and discrete deformation lamellae. Some of the larger relic quartz grains are fractured, although there is no evidence of displacement along the fractures (Plate 4.17a). Evidence of recovery, either due to an increase in temperature ($>300^\circ$), or a lowered strain rate, results in subgraining of the relic quartz. Subsequent higher temperature dynamic recrystallisation is abundant, with finely recrystallised polygonal aggregates of quartz, intimately intergrown with chlorite in places, mantling the relic cores and coating the intragranular microfractures (Plate 4.17b). Sub-grain rotation, indicative of greenschist facies deformation, appears to be the dominant process in these greyschists, although there are examples of what appears to be grain boundary migration, with small lobate protrusions migrating into adjacent highly strained crystals. However, this cannot be proven, and many of the examples are simply the result of sub-graining, the lobate protrusions not being in optical continuity with the host crystal. It is possible that some higher temperature deformation did occur (upper greenschists-lower amphibolite) to produce the clear examples of grain boundary migration, these fabrics being subsequently overprinted by retrograde deformation dominated by sub-grain rotation, although this is difficult to prove in these highly deformed schists.

The marked mineralogical segregation of mica and quartz into separate domains is suggestive of diffusive mass transfer (DMT) processes, although this is difficult to prove due to recrystallisation overprints. Within quartzose domains, a crude foliation is defined by short discontinuous films of mica and chlorite coating the lensoid quartz aggregates. Sub-parallel to this are anastomosing pressure solution seams coated in opaque oxides. Within mica domains, the foliation is intense and the muscovites show a high degree of parallelism. Deformation is accommodated by open kink banding, these kink folds (F3) passing laterally into the quartz-rich areas. Fine calcite veins, which gently cross-cut the main mica foliation and are offset by the kinks, and are evidence of early solution precipitation processes. Evidence of late DMT comes from the aligned segregations of subhedral post-tectonic albite porphyroblasts,

which overprint all fabrics, enclosing the kinks as folded inclusion trails, and further enhancing the crude alignment in the main foliation.

An examination of the primary fabric relationships around early (?F2) fold cores reveals a strong early pervasive phyllosilicate fabric, coated in sub-parallel films of opaque oxides. This fabric is crenulated to result in a new spaced cleavage, through localised reorientation of the micas, into sub-parallelism with the axial planes, and a segregation of quartz into the crenulation hinges. The crenulation cleavage is further enhanced by anastomosing, sub-parallel solution seams, and further evidence of DMT is widespread. This early fabric-parallel cleavage confirms field observations (see 4.2.3) and indicates a D2 age (at least) for these folds. It appears that there are no obvious differences between D2 and D3 associated microstructure, and thus PT conditions cannot have changed significantly between the two deformation events.

In the field the formation of quartz rods appears to be due to solution processes, with loss of quartz from the attenuated limbs and concentration of material in the thickened hinge zones. Although thin sections of these structures were not made, the evidence above is consistent with DMT being the dominant deformation process.

Greenschists

In hand specimen the greenschists appear to exhibit considerably lower strain than the greyschists, being generally massive or poorly foliated. This is borne out in thin section where it appears that low temperature DMT, rather than crystal plastic processes, has been the dominant deformation process, as evidenced by anastomosing solution seams, mineral segregation and several generations of cross-cutting veins. A crude foliation, where developed, is defined by segregations of aligned, elongate albites and a weak alignment of the intergrown mica/chlorite and fibrous amphibole. The weak fabric is typically cut by discontinuous veins of either quartz or coarse calcite, and the strain seems to have localised along these, such that the quartz grains here are highly undulose, subgrained and fringed with dynamically recrystallised new grains. These features are virtually absent from the quartz and albite grains within the matrix. Many of the greenschists contain post-tectonic albite porphyroblasts and small subhedral epidotes, probably the result of late stage hydrothermal alteration. These minerals overprint all earlier fabrics, and result in the massive unfoliated appearance.

Although primary (F2) folds are very common in the greenschists, thin sectioning of these structures revealed little, as microfabrics are so poorly developed. They do show that the hinge zones of these folds are defined by a weak, but clearly tectonic fabric, locally crenulated by a spaced (S2) cleavage.

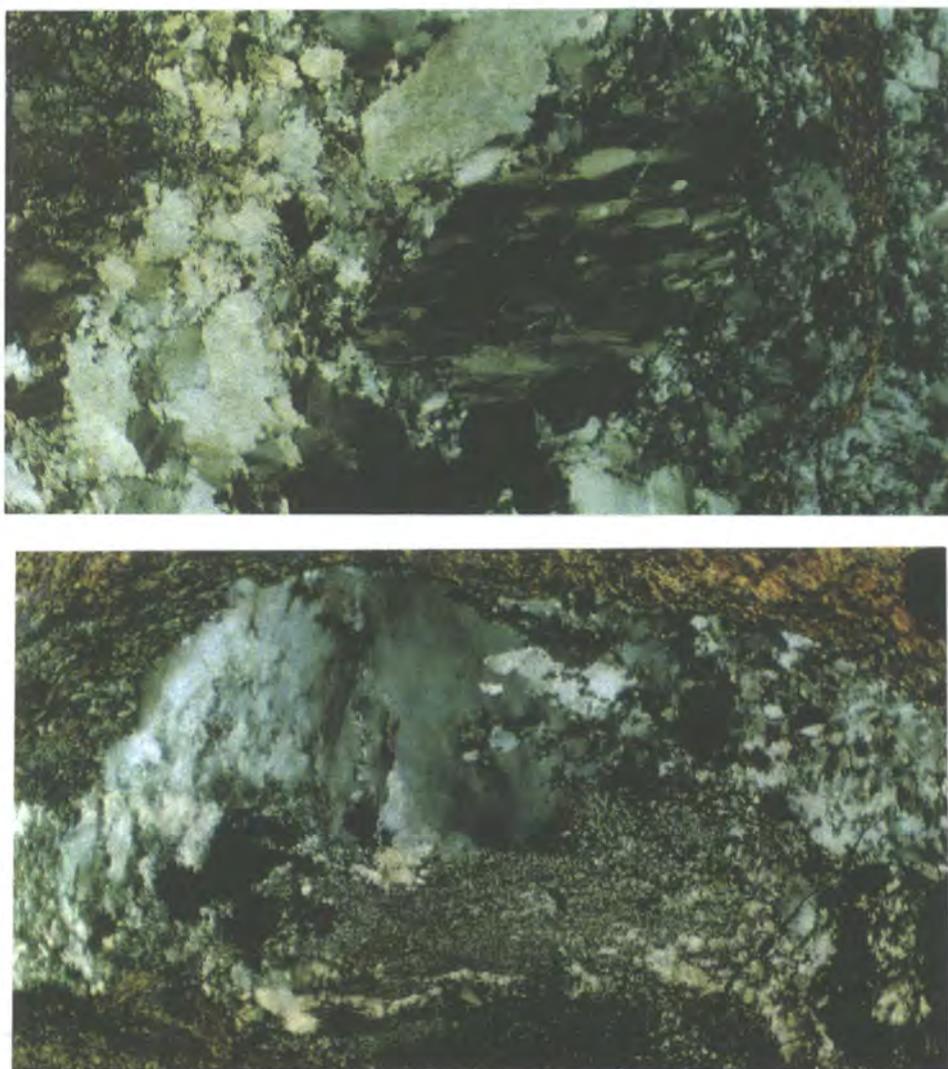


Plate 4.17 a) (top) Thin section of greyschist showing a highly fractured and strained quartz grain. Minor recovery results in well defined sub-grains, and some dynamically recrystallised new grains along the grain margins and intragranular microfractures. b) (bottom) Further dynamic recrystallisation of a) resulting in an elongate quartz ribbon comprising finely recrystallised quartz aggregates, intergrown with patches of chlorite. Greyschists from Start Point [SX 283, 037], X4 magnification and crossed polars. FoV= 6x3.7mm.

4.2.6 Large scale structure of the Start Complex

F2 Folds

The marked uniformity of the two major greyschist horizons in the Start Complex led to the idea that they are the same unit repeated by a recumbent early isocline, cored by greenschist (e.g. Ussher, 1904; Tilley, 1923). Hobson (1977) attempted to establish the presence of this isocline by assessing the primary (F2) fold vergence in the greenschist horizons around the Salcombe estuary. He produced an elaborate model based on the F2 vergence at Portlemouth, at South Sands, and in the main Salcombe estuary. He suggested that the hinge of the isocline lay close to the lower greenschist boundary at Portlemouth, and that the hinge of the complementary synform lay close to the upper greenschist boundary at South Sands. The proximity of their axial surfaces to the lithological boundaries led Hobson (1977) to suggest that

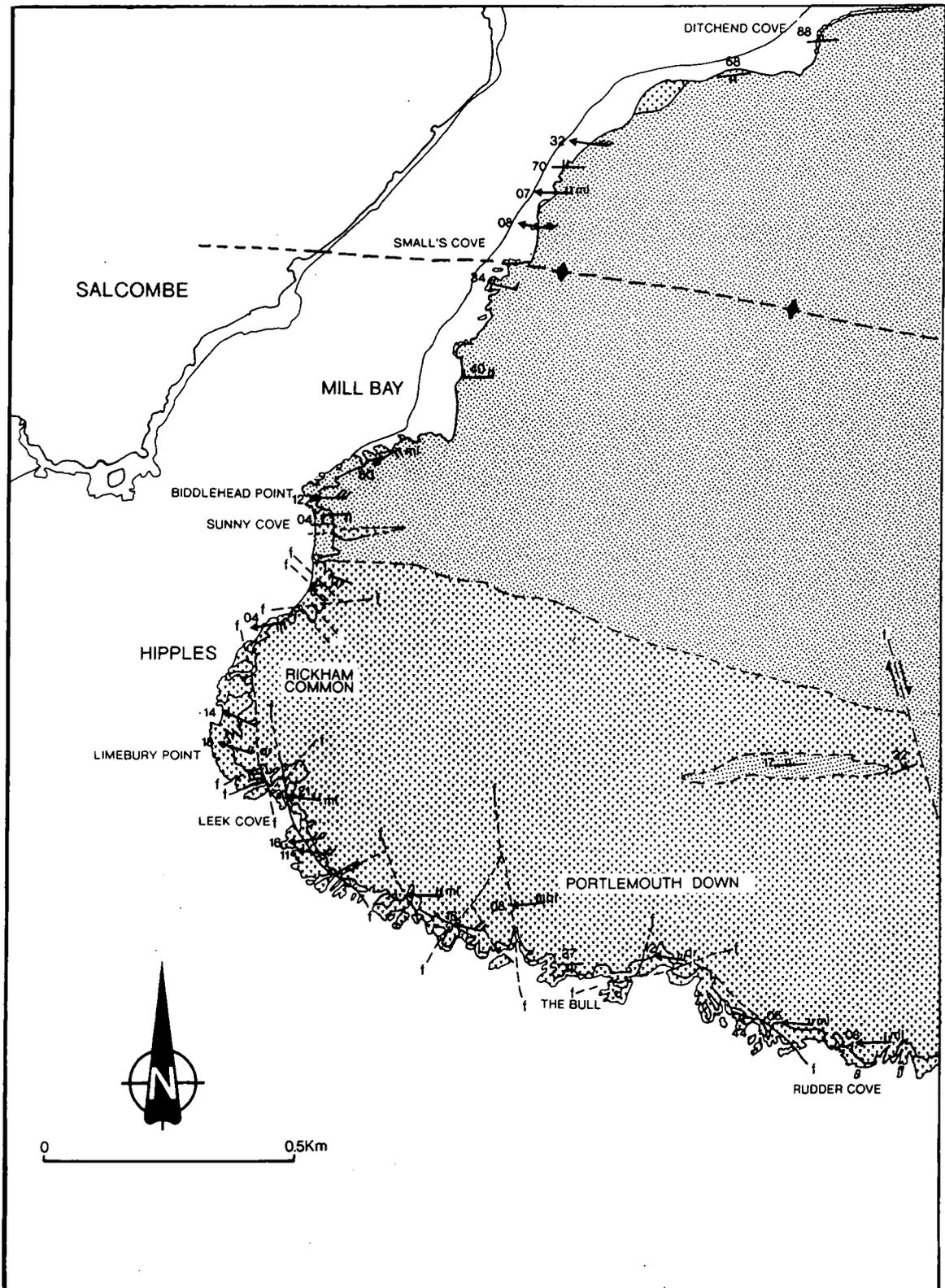


Figure 4.5 Geological of Portsmouth. See Fig. 4.2b for location. See Fig. 4.3 for key.

these represented northerly-directed slides, which had inverted the sequence, placing the structurally lowest greyschist over the structurally highest. He went on to conclude that the greyschists represented two separate mappable formations, effectively interleaved with a single greenschist unit by large scale displacements. However, a detailed examination of the greenschists across the area has shown that the predominant sense of vergence is to the north, e.g. of the 31 F2 folds recorded at Portlemouth [SX 2740, 0372; Fig. 4.5], some 28 verge to the north, whilst at South Sands all ten are north verging (Fig. 4.3). This tends to refute the presence of a major F2 isoclinal fold axis, and it appears that the few south verging F2 folds are simply on the inverted limbs of tens of metre scale, north verging F2 structures (e.g. see Plates 4.11 & 4.13) and are of only localised significance. The primary vergence (F2) is remarkably consistent across the Start Complex, although assessment of the F2 vergence in greyschist is virtually impossible. Assuming that the vergence dataset is representative, it appears that F2 folds within the Start Complex verge towards an antiformal closure within the uppermost Variscan nappe exposed in south Devon.

F3 folds

Minor F3 folds are the dominant structures at outcrop level across the Start Complex, and can be used to define a series of major folds, which in turn define a kilometre scale west plunging antiform (Fig. 4.1). This antiform is the only major structure which may be traced across the entire peninsula.

At Freshwater Bay [SX 2825, 0374], in the hinge zone of the antiform, minor F3 folds have steeply S and SSE inclined axial planes (Fig. 4.2a). Projected along strike to Small's Cove [SX 2742, 0386], in the Salcombe estuary, minor fold axial planes pass through the vertical to dip moderately N and NNW inclined axial planes (Fig. 4.5). Across the estuary, the antiformal axis is obscured by the sea wall, although the minor F3 folds in the greyschist cliffs immediately to the south, all verge north and have steep northerly inclined axial planes (Fig. 4.3). The west coast outcrop of this structure occurs as a broad zone of upright M folds in the greyschist cliffs of West Cliff [SX 2691, 0382]. No sense of overturning is obvious, the major antiform here appearing to be more or less upright. Other major folds are variably developed in the limbs of the antiform, the majority being overturned towards the south, with moderately to steeply north dipping axial planes. North of West Cliff, at Hugh's Hole [SX 2679, 0387], a major synform/antiform pair is defined by moderately to gently northerly inclined F3 folds. It is reasonable to assume that these folds also occur north of the antiform in the Salcombe estuary. Unfortunately the sea wall obscures a significant area of schists here, with the rest of the estuary outcrops being poorly preserved and inconclusive (see Fig. 4.9). At Deckler's Cliff [SX 2755, 0366], a major

northerly inclined antiformal axis runs E-W through the greyschists, but is not seen in the east coast exposures. The presence of major NW-SE dextral strike-slip faults throughout this area, e.g. at Horseley Cove [SX 2786, 0359], may, in part, be responsible.

Approaching Prawle Point [SX 2772, 0350], a sequence of tight, moderately to gently north dipping antiform/synform pairs run E-W through the massive greenschist outcrops, and may be traced inland to reappear on the east coast of Prawle at Sharpers' Head and Dutch End (Fig. 4.4). A sequence of major folds running through Great Mattiscombe Sands and Start Point promontory are similarly moderately northerly inclined (Fig. 4.2).

Thus it appears that the majority of major E-W trending F3 folds in the Start Complex are moderately inclined to the north, implying a degree of regional backsteepening. This is further emphasised by a contoured stereonet of minor F3 axial planes (Fig. 4.6), which shows an ENE-WSW distribution of poles and a predominance of moderate to steep NNW dips. Evidence of backsteepening has been described from the Devonian sequence to the north (e.g. Richter, 1967; Hobson, 1976a; Coward & McClay, 1983), most authors attributing this to backfolding (e.g. Coward and McClay, 1983; see 4.5). The backfolding has been related to southerly directed shear (e.g. Shackleton *et al.* 1982), although Coward and McClay (1983) suggest that no major southward verging backthrust has been recognised in S.Devon, implying that if it does occur, it must lie beneath the present erosion level.

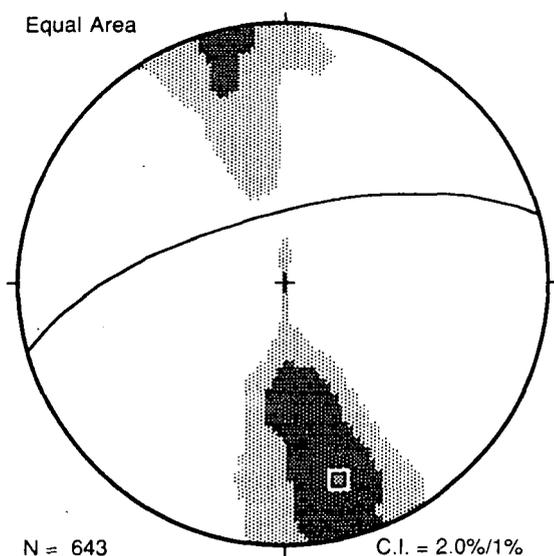


Figure 4.6 Contoured equal area stereogram of 643 poles to F3 fold axial planes in the Start Complex. The predominant dip is moderate to steeply NNW (mean pole is 078/73N), evidence of the regional secondary backsteepening in this region.

4.3 The Start Boundary Fault Zone

4.3.1 Introduction

The Start Boundary Fault (SBF) has been much discussed by Variscan workers, as it is of fundamental importance in understanding the relationship of the Start Complex to the Devonian sequence. More recently, a greater understanding of the tectonic evolutionary models of SW England has further emphasised its regional importance.

Early studies of the Start Boundary Fault concluded that the SBF is not a zone of progressive metamorphism, but represents a distinct tectonic break. Raisin (1887) described the fault as a 'great thrust', a theory subsequently extended by Hendriks (1939), who postulated the 'Start-Dodman-Lizard' thrust hypothesis, suggesting that these three areas belonged to a single Variscan nappe. Although Scrivenor (1949) questioned this interpretation, arguing that there is little evidence of thrusting at Start, Hendriks (1951) reaffirmed the hypothesis and compared the 'brecciation' along the SBF to that of the Meneage. Marshall (1965), however, noted that "...such a description could scarcely be further removed from the field relationships as seen at the Start boundary." (p. 18). The 'Lizard-Dodman-Start Thrust' hypothesis was finally disproved by Sadler (1974), who showed that there is no evidence of thrusting at the Start boundary and that a sedimentary sequence may be recognised in the 'crush-breccias' beneath the Lizard. He described the SBF as a steep north dipping normal fault, one of a series of narrow shear zones. Hobson (1976b) also described the SBF as a sub-vertical fault and, following Dearman (1971), subsequently suggested that the boundary may be an extension of the Perranporth-Mevagissey line (Hobson, 1977). Although evidence of thrusting is lacking at Start, Coward and McClay (1983) considered this fault to be an originally low angle thrust, which had emplaced the Start Complex over the Devonian rocks to the north, implying that the complex is allochthonous. They argued that the thrust was subsequently backsteepened to its sub-vertical position by later southerly directed shear. An alternative model was offered by Holdsworth (1989a), who proposed that the SBF is the surface manifestation of a major basement fault, possibly a Devonian terrane boundary forming the northern margin to the 'Start basin' (see 3.5).

Geophysical evidence across the SBF is ambiguous. Modelling of gravity gradients (e.g. Bott *et al.* 1958; Bott & Scott, 1964) indicates a wedge shaped northern extension of the complex, downfaulted to a shallow depth by the SBF. However, Brooks *et al.* (1984) note that, in the absence of other geophysical constraints, many alternative interpretations of the gravity data may be proposed. Overall, there seems to be consistent evidence for a steep northern margin to the Start Complex. Seismic refraction data may also be variously interpreted (e.g. Doody & Brooks, 1986; see 3.1) but does again indicate that the SBF may be a steep fault.

The SBF is clearly exposed on both the east and west coasts, which allowed detailed examination of the structure itself and the adjacent high strain zones. These zones encompass ~0.5 kilometres of the Start Complex, and ~200 metres of the Devonian to the north, the rest of the Devonian structure being described in 4.4. Although the fault is not exposed in the estuary sections, the high strain zones to the north and south were readily mapped out.

4.3.2 Greenstraight

The SBF zone is exposed on the east coast at Greenstraight [SX 2818, 0391], where it crops out as a series of moderately north dipping faults (Fig. 4.7). A 200m shingle beach separates the fault from the most northerly of the Start schists at Hallsands [SX 2818, 0388] and thus again critical exposure is obscured.

The Start greyschists exposed from Long Rock to Hallsands are offset by steep north dipping, E-W to ENE-WSW trending faults, which in places (e.g. Wilson's Rock [SX 2818, 0385]) form a series of deep caves. Slickensides are not preserved on the fault surfaces due to the extensive erosion, although many of the E-W orientated faults apparently downthrow to the north, based on offsets of pale porcellanous mylonites. At Long Rock [SX 2819, 0382], the greyschists are phyllitic, bearing a strong resemblance to the Meadfoot shales further north. The dominant folds (F3) are open, upright structures which plunge moderately west and clearly deform small intrafolial isoclinal. Well developed mineral lineations typically lie several degrees anticlockwise of F3 axes. The F3 folds verge towards a major north dipping antiform further to the south. Approaching Wilson's Rock, the folds tighten noticeably and display a marked variability of plunge and vergence. Close examination reveals small, double-vergent structures, occasionally cored by 'eyed' sheath folds (Plate 4.18). These curvilinear structures display Z geometries on plunging through the vertical. The increased frequency of such sheath folds moving north along the section explains the observed plunge and vergence variations. In zones of more gentle F3 plunge the vergence variations also define a series of metre scale, north dipping antiform/synform pairs, and the large-scale vergence is consistently southwards. At Hallsands, the greyschists are highly faulted and there are thin (<2m) interbands of greenschists, which are folded into a series of tight, possibly F2 folds. In the footwall of one of the steeply south dipping faults, thin (40-50cm) bands of a hard porcellanous material occur. This appears to be a pure calc mylonite and is interlayered with ~20 centimetre strips of dark slate/phyllite (Plate 4.19).



Plate 4.18 (left) Centimetre scale 'eyed' sheath fold closure within a doubly verging F3 fold. Viewed looking west, greyschist cliffs south of Wilson's Rock, [SX 2818, 0384].

Plate 4.19 (right) Porcellanous calc mylonite, probably originally 'Torcross' type intrusive, interleaved with dark 'shale', within the footwall of a steeply south-dipping late fault. Such lithologies are typical of the boundary zone. Greyschists, Hallsands, [SX 2818, 0387].



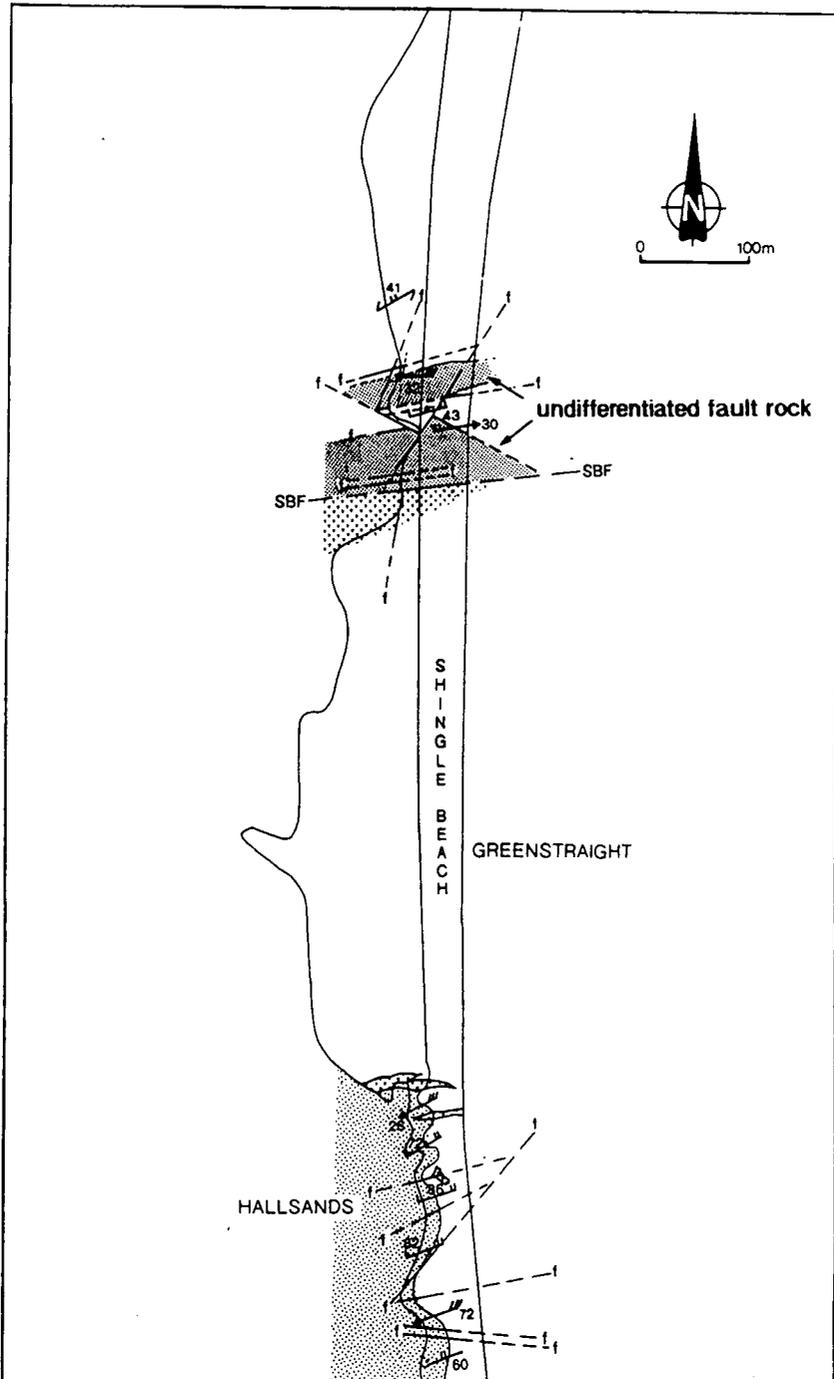


Figure 4.7 Enlarged map view showing the structure of the Start Boundary Fault zone exposed on the east coast at Greenstraight. Figure 4.2b shows the position of this enlarged area in S. Devon

The first exposures immediately to the north of Greenstraight beach are altered and weathered almost beyond recognition and are cut by numerous faults (Fig. 4.8a; see also 4.7). The most southerly of these outcrops forms an E-W trending promontory, and appears to be highly deformed and iron stained greenschists interbanded with considerable amounts of interfolded pale calc mylonite, possibly defining an ENE-WSW trending synform (Plate 4.20). The low cliffs immediately to the north are cut by a series of moderately north dipping faults of uncertain displacement sense. The first of

these (082/40N) is thought to mark the boundary between the Start Complex and the Devonian (Plate 4.21, fault is arrowed; see Fig. 4.8b). These faults define an approximately 3m wide zone of brecciated fault rock (Plate 4.22; Fig. 4.8c) in which large angular to sub-rounded clasts of shale, greyschist and calc mylonite 'float' within a hard lithified matrix composed predominantly of carbonate (see 4.3.5). The larger clasts contain fragments of an earlier dark breccia, which, in turn, appears to bear the remnants of a previous fault rock. Thus at least three phases of brittle movement have occurred within this fault zone. Unfortunately, fault kinematics are impossible to assess as fault surfaces are highly weathered and fibres are largely absent, although some fine quartz fibres do indicate dip-slip displacements, normal to the fault strike (see Plate 4.22).

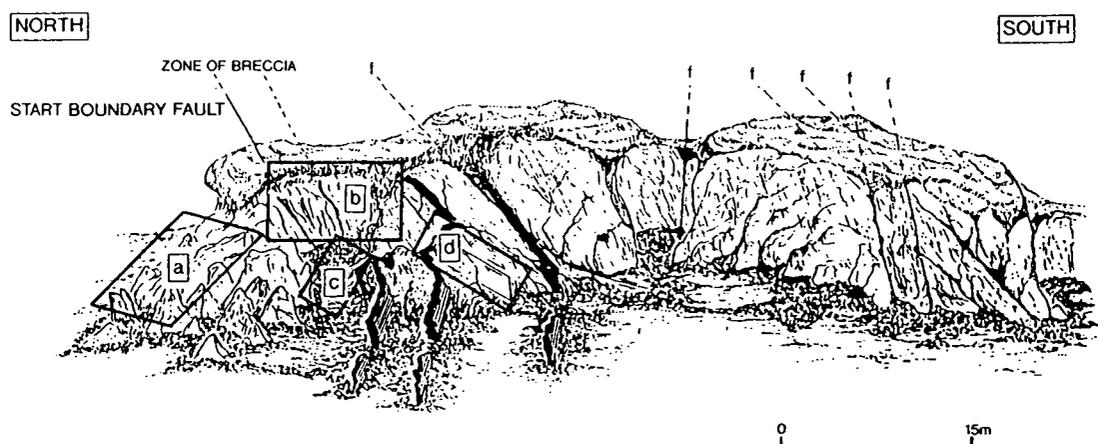


Figure 4.8 Sketch of the Start Boundary Fault, as exposed at the northern end of Greenstraight beach. Boxes a-d indicate the views shown in plates 4.20-4.23 respectively. Refer to Fig 4.2b for exact location in S. Devon. Modified after Marshall (1965).

To the north of the brecciated zone, approximately 50m of ambiguous, highly faulted lithologies occurs, comprising Devonian shales, calc mylonites and Torcross-type intrusive material (see 4.3.5). Evidence from further north (see 3.4.1) has shown that many of the Torcross-type intrusives are now composed largely of replacive carbonate, and thus the calc mylonites could be derived from the Torcross intrusives. Numerous samples were collected along the zone so that thin sectioning could be used to distinguish the unusual lithologies (see 4.3.5).

The area is cut by three separate fault sets: NE-SW, NW-SE and E-W. The NE-SW faults dips steeply NW and are extensively mineralised (Plate 4.23; Fig. 4.8d), predominantly by pyrite and limonite with lesser amounts of carbonate.



Plate 4.20 Highly altered and iron oxide stained greenschists with interleaved 'Torcross' type intrusive forming a small promontory south of the boundary fault. Tight F3 folds here appear to define an ENE-WSW trending synform. Viewed looking north, fault zone north of Greenstraight, [SX 2818, 0390]. NB. Fig. 4.8a-d shows the location and field of view of Plates 4.20-4.23.



Plate 4.21 Series of moderately north dipping faults within the low cliffs. Hammer rests on fault zone breccia and gouge. Planar fault surface to the left of the hammer (arrowed) represents the Start Boundary Fault. Fibres on this indicate dip-slip displacement, possibly downthrowing to the north. Viewed looking due west, fault zone north of Greenstraight, [SX 2818, 0390].

Marshall and Falla (1962) suggested that these bodies bear evidence of D1 folding and are thus pre-tectonic and sedimentary in origin. Examination of these outcrops shows that they are extensively offset by smaller fractures and are not crenulated by S3, and are thus post-D3 in age (see Plate 4.23). It is more likely that these mineral veins are precipitated from fault related fluids. The three fault sets exhibit a complex movement

history, although determination of displacements from offset horizons is probably inaccurate (see 4.2.3).



Plate 4.22 Zone of pale grey brecciated fault rock. Several generations of angular clasts, typically greenschist, shale, and various mylonites, float within a fine calcareous matrix. Fine elongate quartz fibres on some surfaces, e.g. to right of the hammer, indicate dip-slip displacements, although the displacement sense cannot be assessed. Viewed looking southwest, fault zone north of Greenstraight, [SX 2818, 0390].



Plate 4.23 Extensive iron mineralisation (pyrite, limonite and carbonate) along fault surfaces. Mineral veins are offset by later fractures and are not cut by the S3 cleavage which crenulates the adjacent fabric (e.g. centre of field of view). Viewed looking north, fault zone north of Greenstraight, [SX 2818, 0391].

The shales are cut by an intense, moderately north dipping crenulation fabric, which defines tight, moderately west plunging and north verging folds. This crenulating fabric would appear to be S3, due to its similarity with, and parallelism to, the crenulating fabrics in the shales to the north. Anastomosing arrays of quartz veins permeate the

outcrop and have variable age relationships. Pre-D3 veins are tightly folded and cut by S3 parallel veins, which are then offset by a later vein array. Intense quartz veining is characteristic of the badly weathered Meadfoot shales north of the fault zone, where the veins are crenulated (F3) and sheared out into asymmetric boudins by discrete, gently north dipping shears. The displacement sense on these, and the occasional low angle north dipping thrusts, is top to the south.

4.3.3 Salcombe Estuary

This section encompasses the inland outcrops along Gullet [SX 2762, 0395], Batson [SX 2740, 0396] and the Kingsbridge Estuary [SX 2747, 0395]. Exposure is generally poor, the river bank outcrops being highly weathered and frequently slumped. Although the fault itself is only exposed on the eastern side of the Kingsbridge Estuary, its presence could be inferred in other sections and the junction placed to within tens of metres (Fig. 4.9).

Gullet

The outcrop along this part of the estuary (Fig. 4.9) is generally the poorest. As a result of the extensive faulting in this area, the two schistose lithologies are interleaved and variously oxide stained. Thus much of the available exposure is a deep maroon intermediate lithology with little discernible structure. The dominant folds (F3) are moderate to tight, tightening northwards towards the SBF. Axial planes are moderately north dipping in the south, but steepen northwards to dip steeply south adjacent to the SBF. Vergence is consistently southwards, although approaching the SBF, the minor fold vergence defines a series of steep north dipping antiform/synform pairs (e.g. along Westerncombe [SX 2759, 0395]). The F3 folds plunge moderately west and locally, especially in the greenschists around Gullet Farm [SX 2765, 0396], small 'eyed' sheath folds are well developed, implying a high shear strain adjacent to the SBF. This is further evidenced by marked pressure solution striping of the schists, resulting in attenuated fold limbs and transposition of the early fabrics into the upright cleavage. Mineral lineations are difficult to see in these highly weathered shales.

The most obvious feature of this area is the intensity of late brittle faulting, with a number of distinct fault sets occurring: an E-W or ENE-WSW set of sub-vertical faults of uncertain displacement and two later sets of faults in NW-SE and NE-SW orientations, which appear to be predominantly strike-slip. These consistently displace the E-W faults and are more variable in orientation. The strike slip faults appear to have dextral and sinistral offsets respectively and are an extension of the

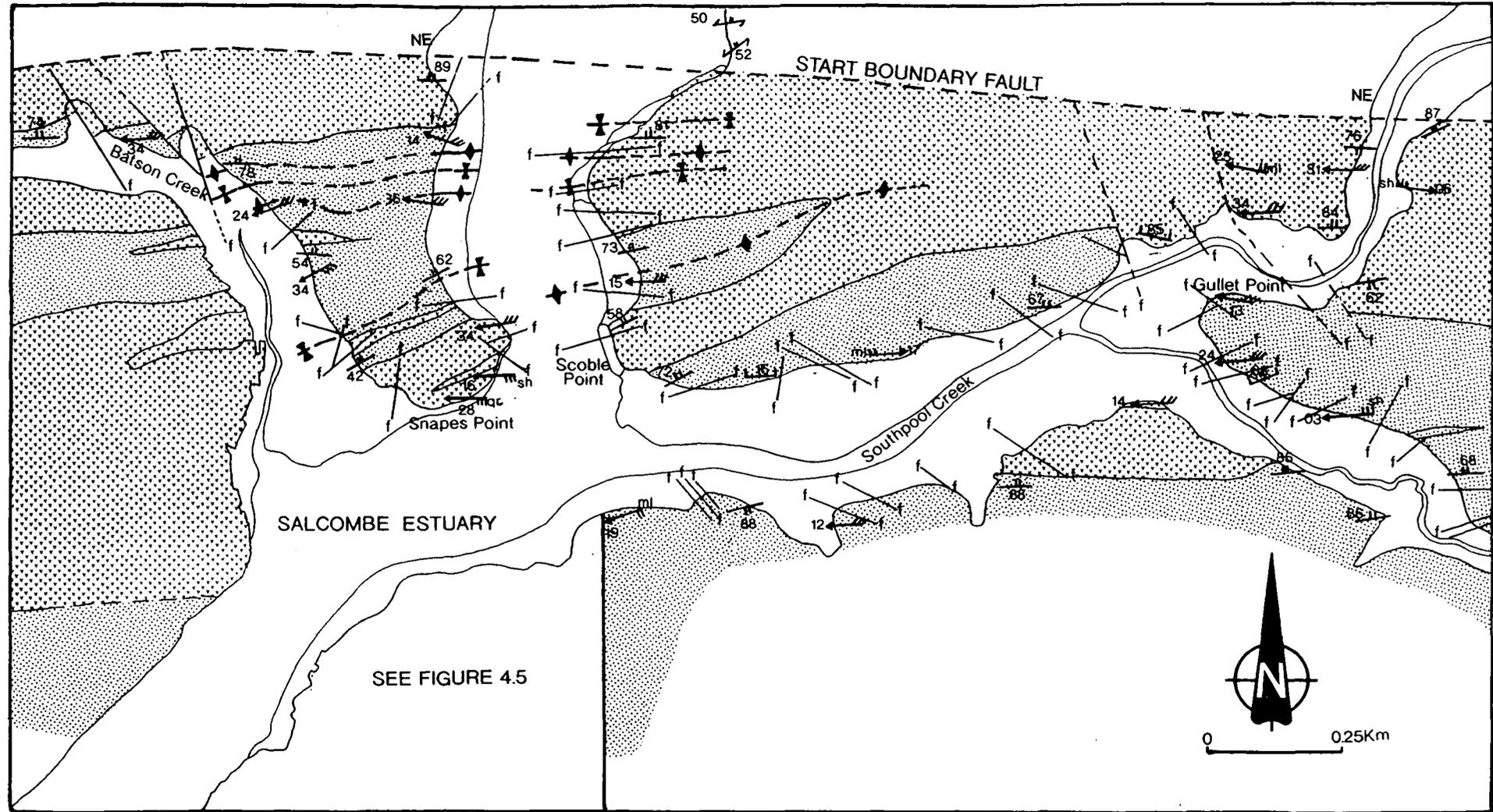


Figure 4.9 Geological map of the Salcombe estuary exposures of the Start Boundary Fault. See Fig. 4.3 for key and Fig. 4.2b for location in the Start Complex.

smaller scale pervasive shear bands common to these outcrops. Unfortunately kinematic indicators are very poorly preserved, or are absent on these fault surfaces and thus a detailed fault analysis is difficult (see 4.6.6).

Batson

The exposures along the southern side of Batson Creek (Fig. 4.9) are obscured by a sea wall. Moving NW from Snapes Point [SX 2746, 0392], the moderately north dipping fabric steepens to sub-vertical, whilst the centimetre scale fold vergence (F3) defines a series of upright synform/antiform pairs, locally disrupted by later brittle faults. There is no appreciable tightening of these folds moving northwards, although there is a marked increase in fold plunge, so that east of Lower Batson, the folds plunge through the vertical to result in locally complex vergence patterns. There is no direct evidence here, however, of sheath folding.

The inlets between Snapes Manor and Lower Batson are cut by steep NW-SE trending strike-slip faults which appear to have dextral offsets, displacing a series of sub-vertical E-W trending faults. Once again it appears that three distinct fault sets occur: variably inclined NW-SE (dextral) and NE-SW (sinistral) trending strike-slip faults and E-W orientated sub-vertical faults which appear to have oblique-slip displacements, as evidenced by offset markers. Also the intensity of this faulting increases northwards, implying a major controlling structure immediately to the north of these exposures. Unfortunately the SBF is not exposed, although must lie very close to the small creek inlets.

Kingsbridge Estuary

This encompasses the estuary exposures from Snapes Point and Scoble Point northwards to Halwell Point [SX 2749, 0405], and provides the best inland cross section through the high strain boundary zone. Within the Start Complex, the schistose fabric dips moderately to steeply northwards and the dominant folds (F3) define a series of antiform/synform pairs which again increase in frequency moving towards the SBF. These folds plunge moderately westwards, locally steepening to plunge through the vertical. At Snapes Point [SX 2746, 0392], a single 10's centimetre scale sheath fold occurs (Plate 4.24), again indicating high shear strain in these outcrops. Along the western edge of the estuary, the SBF is not exposed and the transition from greenschist to soft shales occurs across a 40 metre wide highly silted section. On the eastern side [SX 2571, 0339], however, the fault is marked by a narrow zone of dark phyllitic shales, interleaved with thin bands (20-30cm) of pale calcareous fault rock. This passes south into a 20 metre zone of a dark green/black lustrous lithology, which appears to be highly strained greenschist. The narrow zone of calcareous fault rock is

considered to mark the junction, and in thin section is seen to be a very finely recrystallised calc mylonite. There is, however, no obvious fault and nowhere are the Start schists and Devonian shales juxtaposed. Samples were again collected in order to distinguish the boundary lithologies in thin section (4.3.5).

North of the fault, the Devonian sequence is highly deformed, with much pressure solution striping evident in the pale Meadfoot shales. Thin (<15 cm) sandstones are abundant and are associated with anastomosing arrays of fine quartz veins of various generations. A gently north dipping fabric crenulates the main cleavage, and is axial planar to tight centimetre scale north verging folds (locally F3), similar to those seen north of Greenstraight. Due east of Tosnos Point [SX 2747, 0401], a narrow zone of highly variable fold plunge occurs. Fine continuous mudstone layers are folded into a series of tight F3 folds, which plunge gently west, steepen through the vertical to plunge moderately east before returning to gentle westerly plunges (Plate 4.25). This may be seen in a number of folded layers and results in very complex vergence patterns. These patterns may reflect sheath folding, the individual fold cores occurring on different parts of a larger sheath structure. North of this zone, the monotonous shales are cut by frequent, thin Torcross type intrusives and thin pale sandstone bodies, the two often being confused in the badly weathered shales. Small, gently north dipping ductile shears offset the shales, resulting in localised north verging crenulations. In Frogmore Creek, the exposure is generally poor and heavily silted with little discernible structure.

Inland Exposures

The SBF zone is very poorly exposed inland, with the only occurrence being along the roadside at Kellaton [SX 2802, 0393], where weathered shales and pale calcareous bands are interleaved. However, such calc mylonites have been observed much further north in the Meadfoot Beds (3.4.1), and their presence here is not necessarily indicative of the boundary, although they do occur more or less along strike of the SBF at Greenstraight. There is little to be noted from these poorly exposed outcrops, bar the generally steep northerly dips of the slaty cleavage and the rather high strain nature of the shales.

Summary

The estuarine exposures of the SBF and its concomitant high strain zone reveal a consistent pattern. Within the Start Complex the dominant folds (F3) consistently verge to the south and, approaching the SBF, define a series of tight, upright ENE-WSW trending antiform/synform pairs. The F3 folds tighten and steepen towards the fault and the strain state of the schists increases, with pressure solution striping and



Plate 4.24 Tens of centimetre scale sheath fold in interleaved greenschist and greyschist. Note the curvature of the axial surface and the presence of smaller 'parasitic' sheaths above and below the main sheath closure. Viewed looking due west, Snapes Point, [SX 2746, 0392].

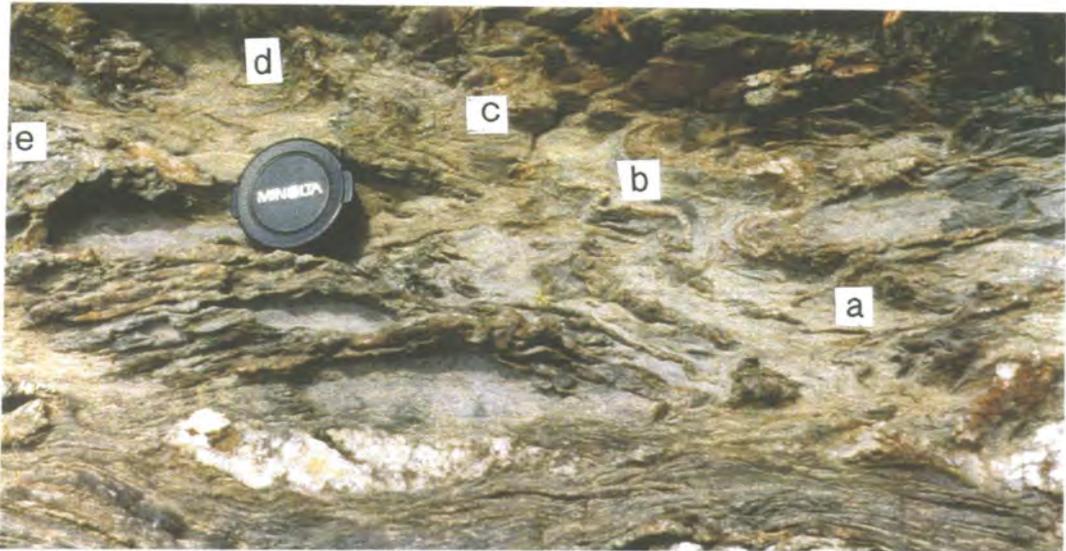


Plate 4.25 Single continuous folded mudstone layers within the pale shales deformed into a series of variably plunging folds, which may reflect sheath fold hinge line variations. Folds at a plunge gently west, steepening up to plunge moderately to steeply west at b. At c the folds are vertical, displaying Z geometries, before switching to plunge steeply east at d. At e the folds are moderately east plunging before returning through the vertical to again plunge west (left of field of view). Viewed looking north, Kingsbridge estuary, [SX 2747, 0401].

quartz veining being very apparent. The F3 folds consistently plunge moderately west, although narrow zones of steeply plunging folds occur and locally well developed intrafolial sheath folds are seen. Late brittle faults are especially common in these areas, occurring in three distinct orientations: Large sub-vertical E-W faults with apparent oblique-slip displacements are offset by NW-SE (dextral) and NE-SW (sinistral) steeply dipping strike slip faults of various scales. The strike-slip faults parallel the small ductile shears, these shears ranging from millimetre to metre scale (4.6.5).

4.3.4 Inner and Outer Hope

This section provides the clearest across strike traverse through the SBF, and allows a detailed examination of the relationships between the Devonian shales and the Start schists.

The steep cliffs which front Inner Hope village [SX 2676, 0399] comprise variably deformed and oxidised greenschist and greyschist (Fig. 4.10). In places, these schists are virtually unaltered and possess open to tight south verging F3 folds, which can be seen to deform small (4-5cm) north verging intrafolial isoclinal folds (F2). Overall the fabric dips moderately to the south although it is locally highly disrupted by later brittle faults. The effect of this faulting increases northwards, with narrow lenses of fractured, discoloured and rotated schists, separated by seams of deep maroon fault rock. Again fault kinematics are difficult to assess, and locally fault chronologies are quite complex. It does appear that the steep faults all post-date the F3 folds, and also the occasional post-F3 crenulations associated with gently south dipping thrusts. The sub-vertical E-W faults are consistently offset by steep NE-SW and NW-SE strike-slip faults. Again these large late brittle structures parallel smaller ductile shears. The strike-slip faults result in significant local rotations of strike, this being especially apparent on Barney Bank [SX 2675, 0400]. Here, tight F3 folds define a series of NE-SW trending antiform/synform pairs, which are cut and offset by arrays of variably inclined metre scale faults with oblique displacements. The elongate discontinuous lenses of greenschists within the greyschists here may be either a primary feature or may result from later F2 interfolding, although this is impossible to conclude due to the extensive later disruption by pervasive shear bands. Between Barney Bank and Hope Headland [SX 2675, 0401], the fabric rotates through $\sim 70^\circ$ and thus a significant ESE-WNW sinistral fault must run through here, although there is no direct evidence for a fault.

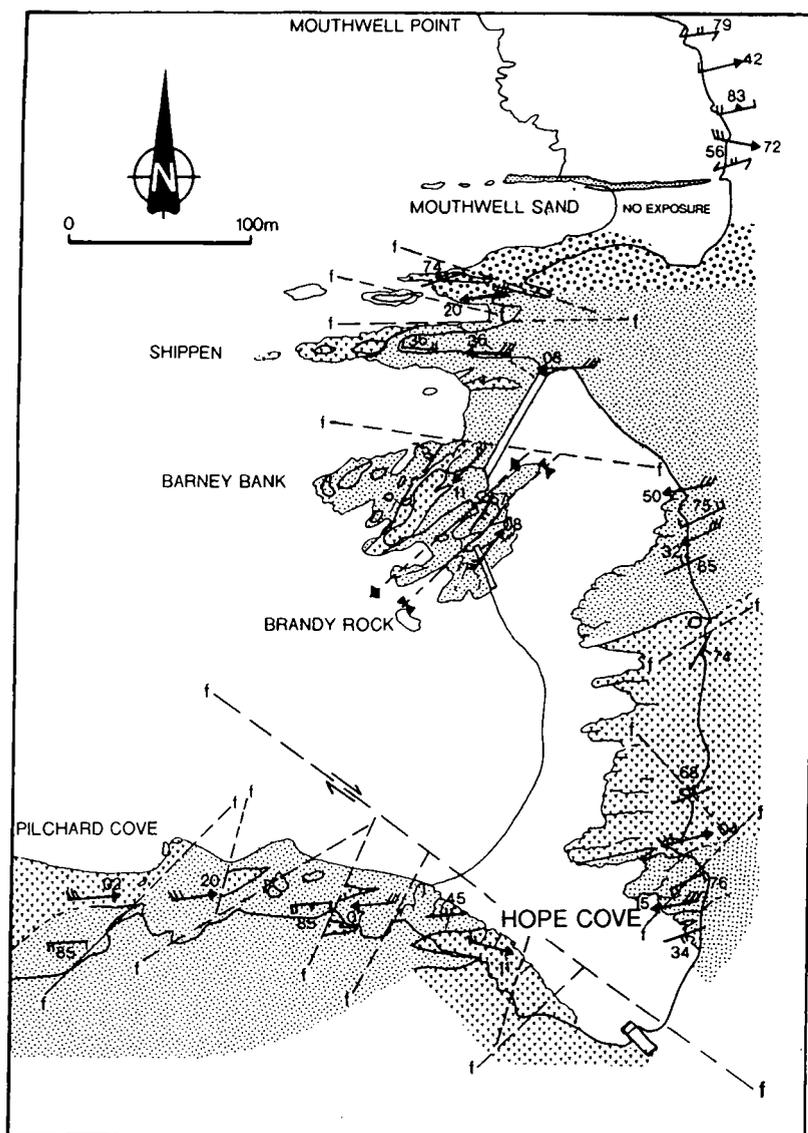


Figure 4.10 Detailed map of the Start Boundary Fault, as exposed at Inner and Outer Hope.

At Hope Headland, the strain increases markedly, the schists being sliced into centimetre to metre scale discontinuous lenses by arrays of anastomosing ductile shear bands. This is associated with the development of highly variable F3 fold plunges, folds plunging through the vertical giving locally complex vergence patterns.

The north face of Hope headland is marked by a deep E-W trending cave, cut by two steeply dipping brittle faults, which again lack displacement fibres. The greyschists here are very dark and phyllonitic, and are interleaved with thick seams of pale grey/green porcellanous material, which is interleaved ^{with} calc mylonite and thick quartz/albite veins (4.3.5). It is quite probable that the pale hard fault rock in this area is altered Torcross type intrusive, now reduced to a calc mylonite. It is also likely that some of the 'greyschist' here is, in fact, faulted-in highly deformed Devonian shale. Distinguishing the Devonian shales from the schists in the field was difficult. The shales are generally very 'platy', have a well developed slaty cleavage, and contain much evidence of pressure solution. The greyschists are extensively quartz veined, have less obvious evidence of pressure solution and are phyllonitic and rather irregular in

appearance. Numerous samples were collected across the boundary zone so that thin sections could be used to decipher the complex lithologies of this area (see 4.3.5).

Moving northwards around the headland, the greyschists are cut by several large, steeply north dipping faults. Narrow 'greyschists' units preserved in the hanging walls of the faults preserve tight to isoclinal south verging F3 folds, which re-fold fine intrafolial isoclines of uncertain vergence [e.g. at SX 2675, 0401]. Much of the exposure here is dominated by finely interbanded dark phyllonitic 'greyschist' and thick seams of pale calc mylonite. Greenschist occurs as subordinate, highly altered lenses, sheared in to the main rock mass.

The steep cliffs along the southern edge of Outer Hope beach [SX 2676, 0401] reveal a complex interleaving of lithologies, all extensively disrupted by later faults and pervasive shear bands. Greyschists and calc mylonites are shredded into centimetre scale 'augen', wrapped by fine grey mylonitic seams. Isoclinal F3 folds within the greyschist 'pods' are almost intrafolial and can be seen to deform the high strain platy fabric. These folds plunge steeply, often through the vertical, where they predominantly show Z-geometries. As a result of the intensity of shearing here, the schistose fabric is progressively obliterated and transposed to form an intense E-W trending high strain zone. Shear criteria, such as vertically plunging 'Z' folds, σ -porphyroclasts, offset horizons and shear bands, all indicate significant dextral shear along this part of the fault zone.

A 20 metre wide E-W gully through the rocky foreshore, presumably marking a late brittle fault, separates the sheared cliff section from the obvious Devonian shales to the north (Plate 4.26). North of this, several metres of calc mylonite, intercut with the shales by arrays of shear bands, marks the northern limit of the boundary zone (Plate 4.27). Here the Devonian shales are tightly interfolded with mylonitic greenschist and phyllonitic greyschist in 10's centimetre scale, steeply plunging folds. It is possible that some of these folds are D2 (*sensu* Start) in age, although the majority appear to be D3, as they deform the high strain fabric. There is much evidence of pressure solution within the shales, with fine iron oxide coated solution seams pervading the outcrops.

The Meadfoot shales north of the fault are highly deformed and characterised by tight north verging crenulations, similar to those seen north of Greenstraight (4.4.3). Just south of Mouthwell Point [SX 2676, 0402], a narrow zone of anomalously plunging F2 folds displays plunge patterns similar to those seen at Greenstraight (4.3.2) and the Kingsbridge Estuary section (4.3.3). Tight 'Z' folds steepen from gentle west plunges through the vertical to plunge steeply eastwards and then return to moderate west plunges (Plate 4.28). Some of the steep fold orientations



Plate 4.26 The Start Boundary Fault exposed in the foreshore at Outer Hope beach. Shingle gully at left marks a large brittle fault which may be traced inland for several hundred metres and which is considered to represent the junction between the Start Complex and the Devonian shales. The buff coloured fault rock is calc mylonite, interleaved with mylonitic greenschist. Outer Hope, [SX 2676, 0401].

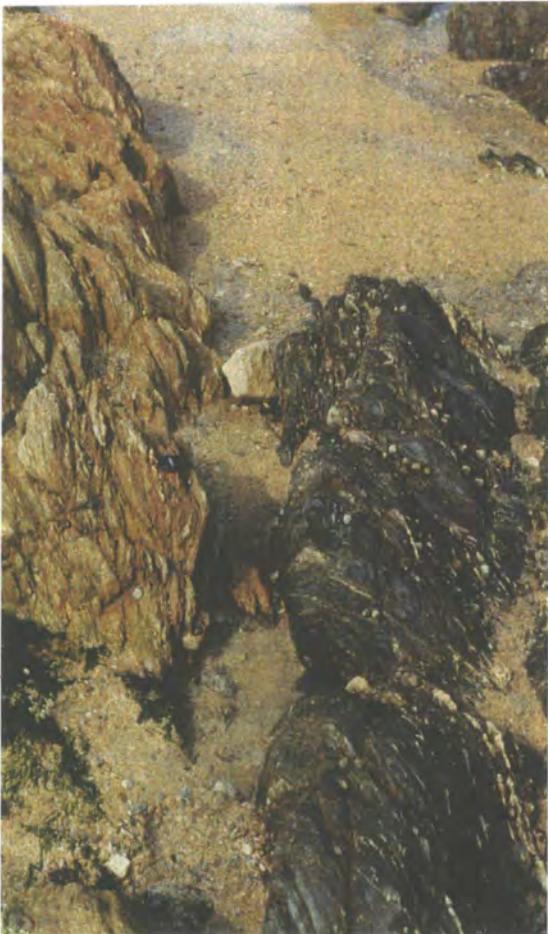


Plate 4.27 Close-up of Plate 4.26, showing dark Meadfoot shales interleaved with boundary zone mylonites by late dextral shears. Locally the two lithologies are tightly folded into steeply plunging folds, which deform the strong platy fabric in the shales and are probably D3 (*sensu* Start) in age (e.g. extreme top left of field of view). Early D2 interfolds also occur and attest to the complex, long-lived deformation along this zone. Note the iron oxide coated solution seams and quartz veining in the shales, indicating extensive pressure solution during deformation. Outer Hope, [SX 2676, 0401].



Plate 4.28 Steeply plunging folds which vary from plunging moderately west through to plunging steeply east, displaying Z-geometries on passing through the vertical. Tracing these structures out of the field of view reveals marked plunge variation, which, as in Plate 4.25, may reflect sheath fold hinge geometries.

may be related to adjacent shear bands, although this mechanism cannot account for all the observed variation and would require each of the folds to be isolated into narrow shear-bound zones. The possible relationships of the above fold patterns to sheath fold processes, within a dextral shear environment, is discussed in 4.6.4. Beyond Mouthwell Point the strain state of the shales gradually diminishes, and is documented in 4.4.3.

4.3.5 The petrology and microstructure of the fault rocks

As noted above it is generally impossible to distinguish the highly altered and interbanded lithologies along the SBF. A number of thin sections were cut to attempt to resolve this and identify the fault rock precursors and deformation mechanisms.

At Greenstraight, the fault rocks exposed immediately to the south of the SBF are highly altered and iron mineralised greenschists. In some sections, all of the primary mineralogy has been replaced by the very fine-grained, carbonate matrix, whilst in the others the original greenschist has been reduced to a mylonite, comprising alternate bands of aligned micas, imparting an intense foliation, and finely recrystallised quartz. Opaque iron minerals coat the mylonitic foliation and thin chlorite films and calcite seams further enhance the fabric. Euhedral pyrite crystals are undeformed and lack pressure shadows suggesting that the iron mineralisation is a late stage phenomenon. In some examples, there is evidence of later cataclasis, with cross-cutting transgranular

calcite-filled microfractures. Several generations of veining are visible, attesting to the prolonged influence of elevated pore fluid pressures.

The fault breccia noted along the SBF itself comprises angular to sub-rounded quartz clasts (probably originally greyschist), and angular fragments of shale, in a fine grained muddy matrix of calcite and chlorite (Plate 4.29). In places ghost outlines indicate where original clasts have been replaced by the calcite matrix. The quartz clasts are composed of fine, dynamically recrystallised polygonal aggregates, cut by thin quartz and calcite veins. Some of the calcite veins follow along original ductile shears, and are probably the result of replacement rather than true veining. None of the veins transgress into the matrix (i.e. they are all pre-brecciation). Discrete late transgranular calcite veins also occur throughout the matrix, but do not cut the clasts.

North of the SBF the rocks are very highly altered and even thin sectioning reveals little. The majority comprise lensoid knots of white mica wrapped in anastomosing films of opaque minerals, set in a very fine-grained, dynamically recrystallised calcite matrix. In places these are reduced to a calc mylonite, in which obdurate quartz augen (now finely recrystallised polygonal aggregates) are wrapped in films of chlorite and fringed with pressure shadows of quartz/chlorite (Plate 4.30). Some of the augen are markedly asymmetric, although no overall shear sense could be deduced. Seams of opaque oxides wrap the augen and enhance the mylonitic foliation, which is locally cut by thin quartz/pyrite veins. Thus the textures at Greenstraight suggest intense early ductile deformation, with dynamic recrystallisation of quartz and calcite, accompanied by mineral segregation during the mylonitisation. Associated with this is extensive microfracturing and veining, which suggests an elevated pore fluid pressure. Late stage brecciation then occurred along the fault plane and resulted in several generations of protocataclasite, characterised by sub-rounded to angular clasts in a fine calcitic matrix.

The fault rocks exposed at Hope Cove are also dominated by fine-grained calcareous rocks of uncertain origin. Based on the observed carbonate replacement of Torcross type intrusives elsewhere, it is probable that the carbonate fault rocks along the SBF were originally such intrusives, that have been extensively veined and replaced by calcite, both prior to, and after, mylonitisation. The hard porcellanous rocks in this area are interleaved calc mylonites and thick quartz/albite veins, in which the quartz grains are flattened and strained, but show little sign of recovery or dynamic recrystallisation (i.e. late stage veining). The greenschists on the south side of Outer Hope are less iron stained and, in places, are mostly less deformed than those at Greenstraight. They comprise ribboned quartz grains set in a finely recrystallised polygonal quartz mosaic. The presence of randomly orientated lenticular segregations of muscovite (+chlorite), wrapped by opaque seams, imparts a structureless

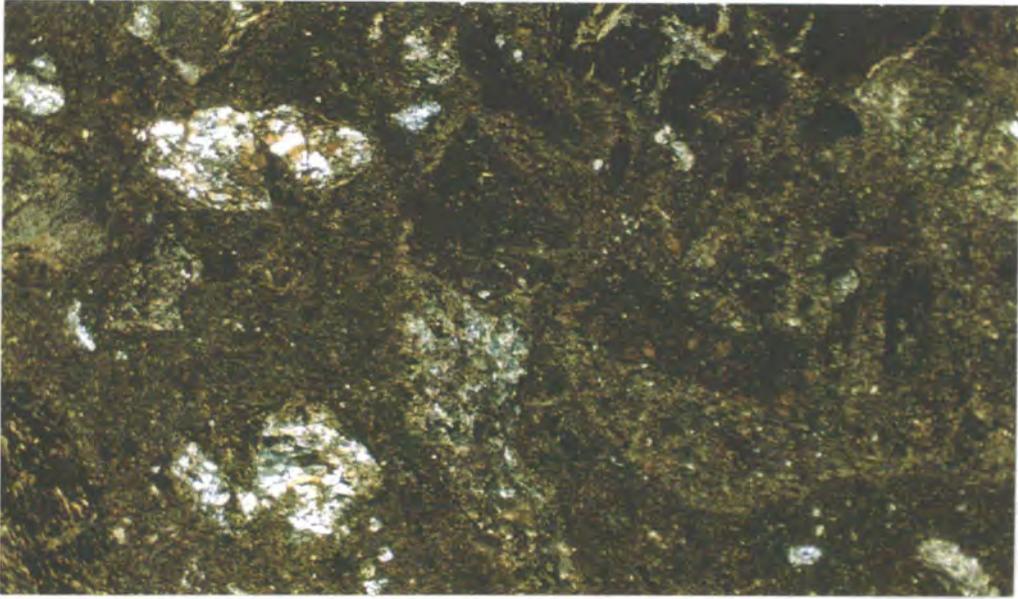


Plate 4.29 Thin section of fault zone protocataclasite. Note the replacement of the finely recrystallised quartz clasts by the calcite matrix and the 'ghost' outlines. Shale fragments and quartz clasts have been extensively veined prior to cataclasis. Start Boundary Fault zone at Greenstraight, [SX 2818, 0390], X1 magnification (plan view) and crossed polars. FoV= 12x6.7mm.

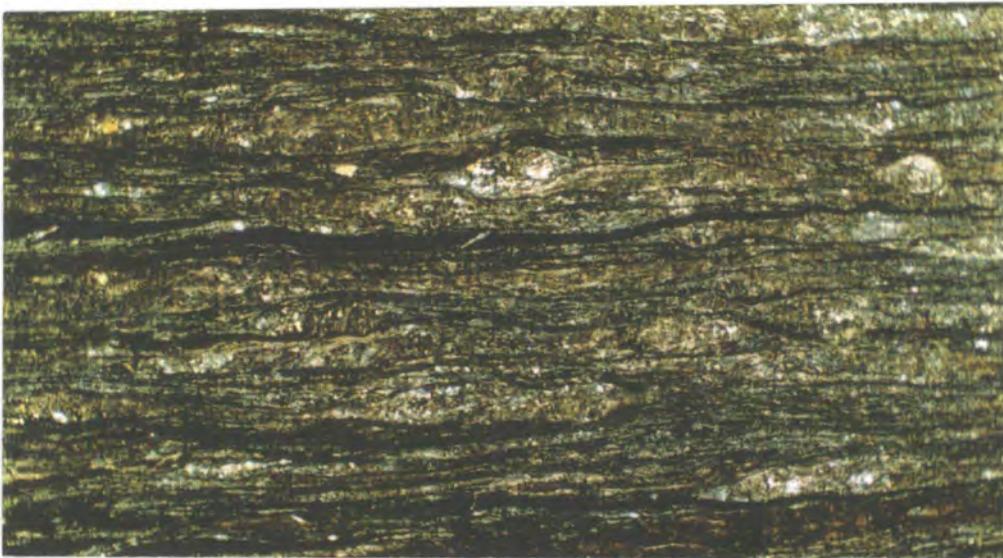


Plate 4.30 Thin section of fault zone calc mylonite from the altered zone north of the SBF. Note the asymmetric augen of dynamically recrystallised quartz. Altered Torcross type intrusive, Greenstraight, [SX 2818, 0391], X4 magnification and crossed polars. FoV= 6x3.7mm.

appearance. In places, the greenschists are reduced to fine mylonites and are difficult to distinguish from altered Torcross intrusives, even in thin section, due to the extensive calcite replacement.

One consistent feature of the SBF lithologies is the presence of veining on all scales and at all stages of deformation, indicating that fluid-assisted DMT was important during deformation along this shear zone. Whilst the calcite vein infills and pressure shadows are evidence of sink processes, there is no obvious indication of the

source and transfer paths. The breakdown of calcic plagioclase (An) to leave albite was observed by the author in thin sections of Start Complex lithologies, and it can be shown that the Devonian shales are carbonate-rich. Whether these are sufficient to source the large volumes of calcite found all along the SBF is uncertain. The apparent replacement of the Torcross-type intrusives by calcite indicates that some external source is needed and the localisation of calcite mineralisation along major faults does suggest that the fluids responsible may be far travelled. Devonian limestones are common across the basinal sequences of SW. England, but there are none in the immediate Start area to provide an obvious source. Thus the volume of calcite associated with the SBF remains problematic, and whilst the fault rocks may act as large sinks, there is no clear evidence of sources or mass transfer paths.

4.4 The Devonian Sequence

4.4.1 Introduction

The Devonian shales lying immediately to the north of the Start Boundary Fault were mapped northwards for several kilometres to allow a comparison of deformation styles and chronology, and to thus interpret the regional setting of the Start Complex within SW. England. Usable exposure is largely confined to the east and west coastlines (Pilchard Cove to Greenstraight and Challaborough to Hope Cove, respectively) and, whilst the Kingsbridge Estuary does provide some additional inland outcrop, these inland areas are generally poorly exposed.

4.4.2 Previous Deformation Chronology

The first detailed work on the deformation chronology of South Devon was carried out by Richter (1967), who, working on the Devonian successions in the Torquay area, recognised four fold phases, a view later substantiated by Hobson (1976b). More recently, Coward and McClay (1983) reinterpreted the structure of south Devon as a series of northward verging thrust nappes. They noted that deformation was locally diachronous, suggesting that local polyphase sequences can be attributed to progressive deformation associated with thrust belt processes (*cf.* Coward & Potts, 1983). On the larger scale, they noted an increase in both the complexity and intensity of strain moving south towards the Start Complex, so that, just to the north of the Start Boundary Fault, there was one set of pre-main cleavage folds, and two sets of cleavage age folds, a significantly more complex deformation history than at Torcross. They suggested that these extra deformation phases were developed in the immediate

footwall region of the Start Boundary Fault (which they tentatively interpreted as a thrust), during thrusting of the Start schists over the lower Devonian rocks.

Much of the study of the structure of SW Devon has centred on the west coast from Plymouth to Bolt Tail (e.g. Hobson, 1976b; *et al.* 1984; Seago & Chapman, 1988; Seago, 1991), and, allied to the work on the east coast (e.g. Richter, 1967; Coward & McClay, 1983), has led to a variety of structural interpretations and a general deformation chronology for the Devonian successions in the South Hams District. This is as follows (after Hobson, 1976b; Seago & Chapman, 1988; Seago, 1991):

D1 Deformation: F1 folds trend ENE-WSW and are sub-horizontal to gently plunging, though some steeply plunging ($\sim 50^\circ$) folds are recorded. Folds are asymmetric, overturning to the NW, indicating NW directed tectonic transport (Seago & Chapman, 1988). The associated axial planar penetrative cleavage is best developed in the mudrocks, where it is defined by a strong alignment of phyllosilicates. The S1 cleavage dips moderately to the south and the younging evidence suggests that these NNW verging folds also face to the NNW. Approaching the Start Boundary Fault, the southerly cleavage dip steepens, ultimately passing through the vertical to dip steeply north. Coward and McClay (1983) suggested that this backsteepening occurred due to the presence of a late Variscan deep level backthrust. The backsteepening effects were thought to produce a local northwards dipping crenulation cleavage and northwards dipping thrusts including many of the F2 and later structures described below (Seago & Chapman, 1988).

D2 Deformation: F2 folds deform the early cleavage and are mainly developed in fine grained lithologies. The F2 fold axes trend NE-SW and refolding of the F1 folds is approximately coaxial, resulting in Type III interference patterns (Ramsay, 1967). Facing and vergence is to the north and north-west, except for the zone immediately to the north of the Start Boundary Fault, where folds are south overturned due to backthrusting. Marshall (1962) noted that there were a number of other D2 structures associated with the F2 folds, including a 'shear set' trending NW-SE (dextral) & NE-SW (sinistral), *a-c* joints at 90° to the S1/S2 intersection lineation, and E-W trending sub-vertical dip-slip faults (e.g. the Start Boundary Fault).

D3 Deformation: Richter (1967) described minor south verging folds which are moderately to steeply E-W plunging. Axial planar fabrics, which locally crenulate both S1 and S2, are poorly developed and dip steeply north. These folds may be due to minor, localised thrusting. Hobson (1976b) described similar folds in the Plymouth area, which he tentatively interpreted as F3 although he was unable to unequivocally establish their relative age. It is highly likely that such localised phenomena are thrust related and are of limited regional significance.

D4 Deformation: Two sets of kink bands have been described from south Devon. The first set is sub-vertical and NNW-SSE trending, perpendicular to earlier fold axes and parallel to local wrench faults. Richter (1967) described a second set of E-W trending, flat lying kink bands although Hobson (1976b) could not distinguish these from geometrically similar F2 folds. Although kink bands occur sporadically across south Devon, their development is usually related to local fault movements.

4.4.3 The West Coast Section

The Meadfoot shales exposed in Hope Cove form part of the Start Boundary Fault zone and are thus considered in 4.3.4. This section is concerned with the structures outcropping between the north side of Mouthwell Point [SX 2674, 0402; Fig. 4.11a] and Black Stone [SX 2641, 0453; Fig. 4.11b], approximately one kilometre north of Challaborough.

Resumé

The Devonian sequence of the west coast defines a series of major upright F1 folds. Some of the larger anticlines may be cored by the upper units of the Dartmouth Group (Warren Formation), although this is hard to establish unequivocally. Primary folds plunge gently E-W in the southern areas, although rotation of the regional strike means that they plunge ENE-WSW or even NE-SW north of Bigbury. Minor fold facing is up to the north and south, depending on their position relative to the larger structures. The primary cleavage fans about the vertical, although is generally steeply north-dipping in the south and moderately to steeply south-dipping in the north. This cleavage is locally crenulated by gently dipping later fabrics, which may be related to local thrusting or the distribution of intrusives. Post-main cleavage crenulations intensify southwards, broadening from narrow discrete zones in the north, to form the dominant fabrics in the south, where they occur as tight north verging folds with markedly variable plunges. In the southern areas, flattened and stretched fossils record a strong strike-parallel extension in the slates, whilst deformed xenoliths and pyrite fragments also indicate foliation parallel extension in the intrusives. Rarely, fine along-strike mineral stretching lineations may also be seen within the slaty cleavage. At Bigbury there is little evidence of extension, whilst at Challaborough deformed pyrites indicate a powerful down-dip stretch within the intrusives, implying an approximately 90° switch in the maximum finite extension direction.

At Mouthwell Point, the outcrops are dominated by moderate to tight, upright, north verging (F3) crenulations of the main cleavage (S2). The crenulation fabric dips moderately to the north and these centimetre scale folds plunge gently E-W (Fig.

4.12). Traces of bedding are always wrapped by an early tectonic fabric, sometimes manifest as fine, ^{iron}oxide coated pressure solution seams, resulting in the folds weathering out in relief of the shales. Approaching Woolman Point [SX 2675, 0204], the shales are intensely deformed and dominated by the somewhat angular crenulations. This 50m wide zone is very similar to that seen to the north of Tinsey Head on the east coast (see 4.4.4). The fine bedding parallel oxide coated seams are evidence of considerable flattening and shearing after fold formation in these high strain shales. Occasional distorted fossils or fossil fragments (solitary corals, crinoid ossicles, etc.) also indicate high oblate strains. Elongate orthocones with fine quartz fibres infilling the displaced fragments, indicate incremental E-W extension within the cleavage (Fig. 4.11a).

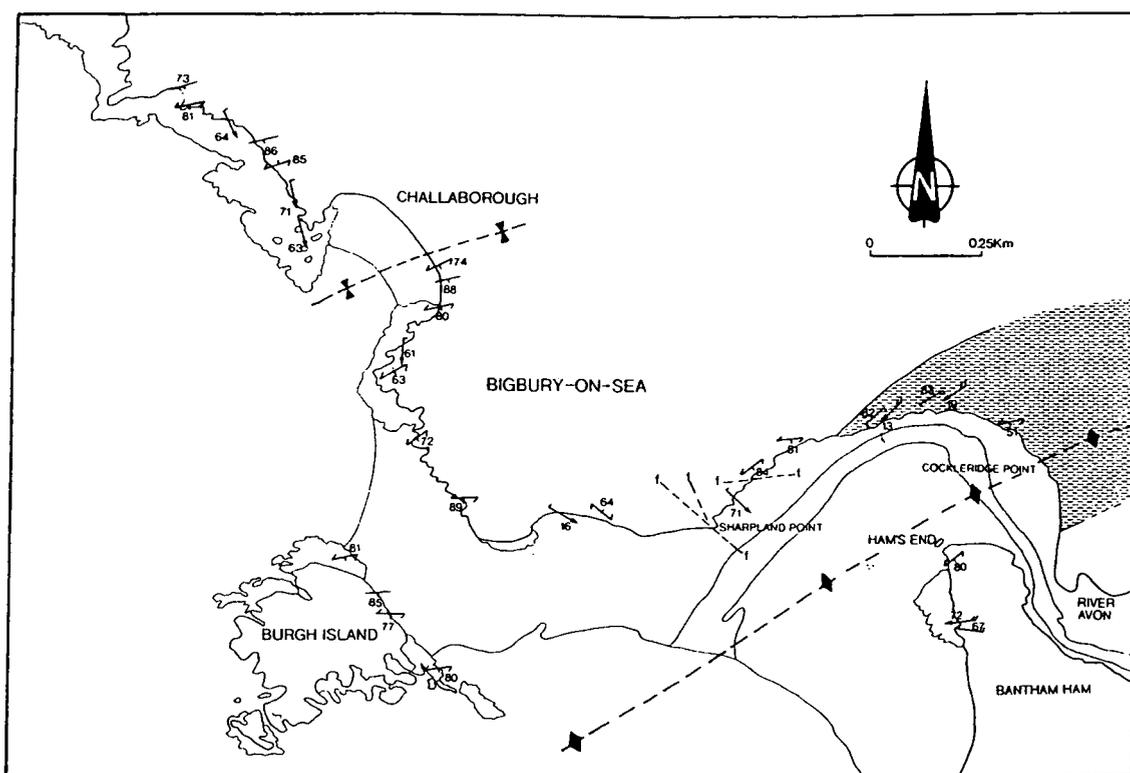


Figure 4.11b Geological map of the western South Hams coastline, from Bigbury Bay to Challaborough. See Fig. 4.2b for location and Fig. 4.3 for key.

At Beacon Point [SX 2675, 0406], the first traces of undeformed bedding appear and indicate a primary north vergence. A sequence of upright, south dipping primary folds occurs from here to Thurlestone Sands, and sporadic way-up data from the sandy horizons indicates that these are upwards north facing folds (Fig. 4.12). North of Great Ledge [SX 2675, 0409], bedding becomes obvious in these lower strain shales, defining rather open and upright folds similar to those seen at Leas Foot Sand (see Plate 3.9).

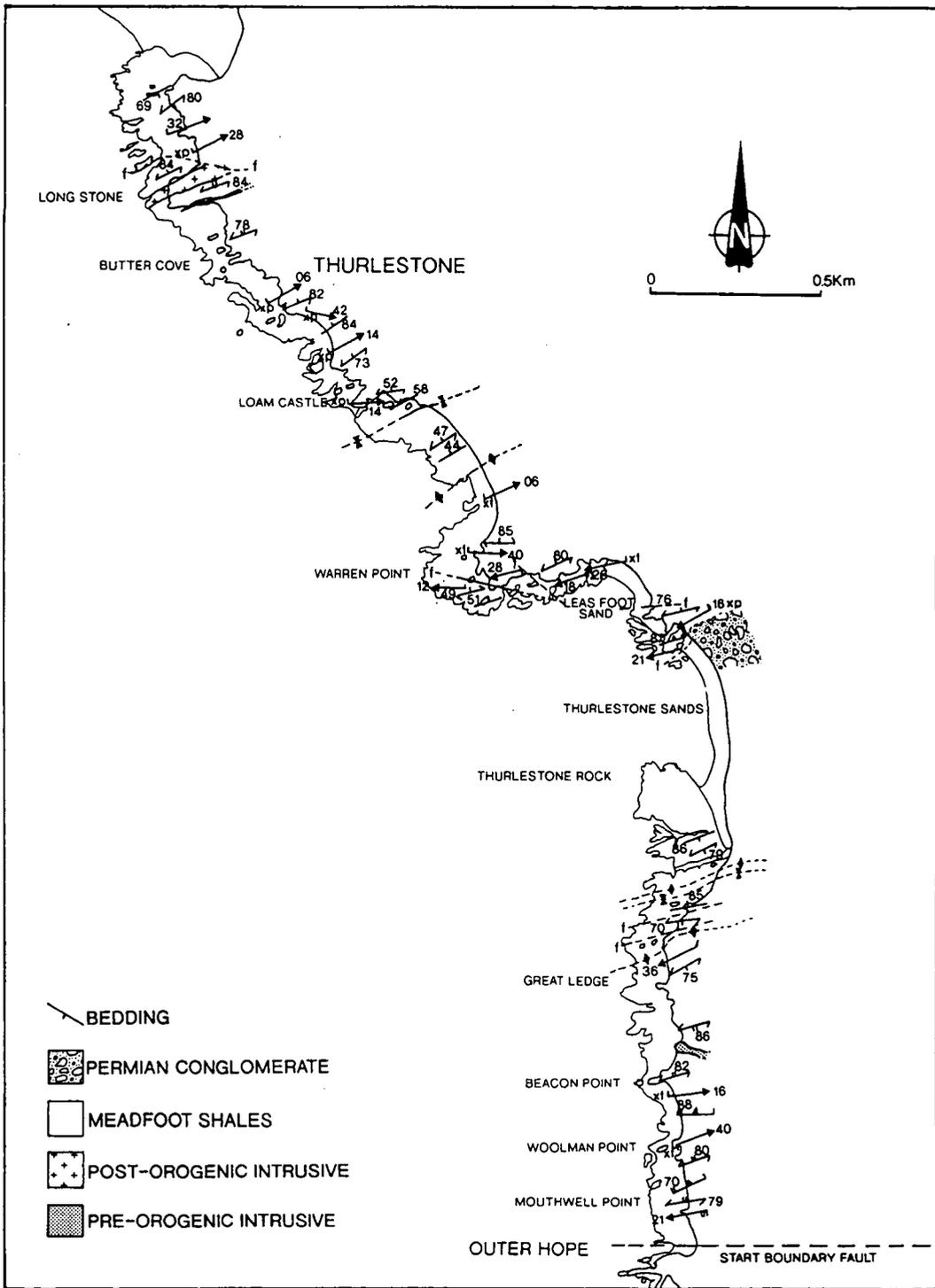


Figure 4.11a Geological map of the western South Hams coastline from Hope Cove to Thurlestone. See Figure 4.2b for location and Fig. 4.3 for key.

The dark shales around Warren Point [SX 2670, 0421] are cut by numerous thin foliated intrusives, carrying fragments of pyrite and stretched xenoliths that record a gently to moderately plunging ENE-WSW extension. This parallels the quartz fibres linking fossil fragments in the numerous narrow zones of calcareous and richly fossiliferous shales. The shales are cut by two sets of predominantly ductile shear bands, dominated by WNW-ESE trending dextral shears (Plate 4.31). The mean

orientation of these shears is similar to those recorded in the Start Complex (see 4.6.5). Primary vergence, where discernible, is to the north and the primary cleavage is overturned to dip steeply northwards, resulting in southward facing. A major F1 anticline, cored by red and purple shales, which may be the upper units of the Dartmouth Group (Warren Sandstones), runs ENE-WSW through Yarmouth Sand [SX 2669, 0424]. A sequence of metre scale open F1 folds may be picked out in the cliffs to the north and the complimentary major syncline is exposed at Loam Castle [SX 2665, 0425]. Northwards, a monotonous sequence of grey and purple shales occurs in which bedding is reduced to pale silty partings and occasional sandy bodies. Primary vergence is consistently northwards, as is the vergence of the occasional zones of late crenulations (Fig. 4.12).

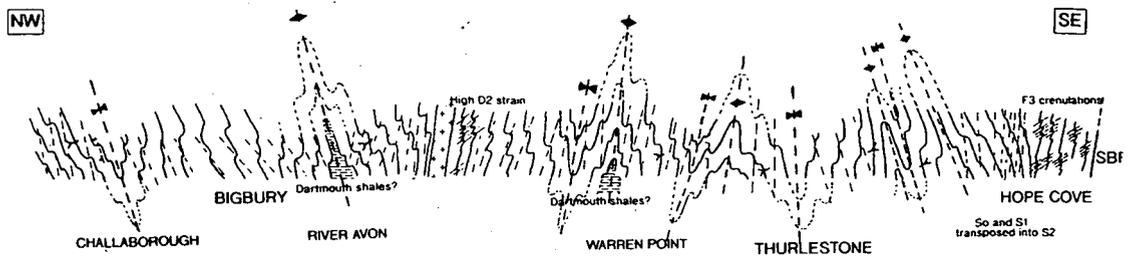


Figure 4.12 Simplified cross section through Fig. 4.11a&b.

The fine intrusives cutting the shales at Butter Cove [SX 2662, 0431] are characterised by a very intense ENE-WSW stretch, evidenced by markedly streaked out xenoliths and displaced pyrite fragments, many of which are wrapped by fine elongate quartz pressure shadows. There does not appear to be a correlation between the narrow zones of post-main cleavage crenulations and the areas of heightened along strike extension. Some of the intrusives contain aligned euhedral bright gold^{coloured} pyrite with no discernible extension. Closer examination shows that they are wrapped by crude down-dip orientated quartz pressure shadows, which appear to be the first evidence of regional down-dip extension, although the majority of strain markers in this area still indicate along-strike extension (Fig. 4.11a). A narrow zone of steeply dipping late folds occurs at Hams End [SX 2662, 0442], and are very similar to structures recorded adjacent to the Start Boundary Fault in Hope Cove (4.3.3, 4.3.4.).

A major F1 anticline appears to run ENE-WSW through the mouth of the River Avon, just south of Bigbury [SX 2665, 0440; Fig. 4.12]. Its axis approximately coincides with the inlier of Dartmouth shales, again suggesting that these represent upper units (Warren Formation) exposed in the fold core. The northern junction of this

inlier has been interpreted as a thrust fault because of the generally high strain here and the presence of tight north overturning F2 folds (Hobson, 1976b; Seago, 1991). Although zones of rather brittle folds do occur along the boundary, there is no evidence of a distinct faulted junction and the purple and grey shales appear interleaved. The orientation of the dominant cleavage in this area is variable, generally steeply north dipping, but occasionally passing through the vertical to dip steeply to the south. Between Bigbury and Challaborough, steep southerly dips begin to predominate until at Challaborough beach, the dominant cleavage assumes its moderate to steep southerly dip. Narrow zones of crenulation occur here and appear to be localised around the numerous thin intrusives. Xenoliths and pyrite in these seem relatively undeformed with very little obvious extension. North of Challaborough beach, however, the intrusives bear markedly elongate pyrites, the majority of which indicate extension down the dip of the foliation. Thus a switch in extension from along strike to down-dip appears to occur in the 100m zone straddling Challaborough. Previous work in S.W. England (e.g. Shackleton *et al.* 1982), and limited reconnaissance to the NW, confirms that NNW trending, down-dip lineations are predominant across the rest of this region. The F1 fold vergence also changes across this zone due to a major synclinal axis running ENE-WSW through the beach (Fig. 4.11b).



Plate 4.31 Brittle dextral shear bands offsetting a 'Torcross' type intrusive into the dark Meadfoot shales. Meadfoot Beds, Warren Point, [SX 2670, 0421].

4.3.4 The East Coast Section

This section is concerned with the variably deformed Meadfoot shales outcropping from immediately north of the SBF, as far north as the first outcrops of Dartmouth Beds at Pilchard Cove [SX 2837, 0460].

Resumé

The Meadfoot Shales immediately to the north of the Start Boundary Fault are very highly deformed with several generations of crenulation cleavage. Outcrop here is dominated by tight, north verging post-main cleavage age folds (F3), which dip steeply to moderately to the north. Traces of bedding seen in these high strain exposures always exhibit a parallel early tectonic fabric, an observation largely overlooked in previous work. F2 folds, which locally display highly variable axial plunges, verge towards an antiformal closure beneath Beesands. The shales at Torcross Point are cut by numerous thick pre-orogenic intrusives, carrying strain markers recording pronounced along strike extensions. Although narrow zones of late crenulations are present, the later strain is generally lower and several main cleavage age (S1) folds may be mapped out, including a major upward facing syncline at Limpet Rocks. A considerable area of critical structure is not exposed due to Slapton Ley. The outcrops north of this are monotonous purple Dartmouth shales, with a moderately southerly inclined main cleavage (S1), within which the mineral stretching lineation plunges down-dip. The open F1 folds verge to the north.

North of the Start Boundary Fault zone at Greenstraight, is ~150m of highly eroded and slumped shale cliff, with little discernible structure. Between here and Tinsey Head [SX 2819, 0396] is a succession of high strain shales dominated by moderate to tight north verging F3 crenulations (Figs. 4.13a, 4.14). The main cleavage (S2) dips moderately to steeply north and is locally transposed into the intense north dipping S3 crenulation fabric (Plate 4.32). Coward and McClay (1983) suggest that the crenulation itself is folded by southwards facing moderately inclined folds in the southern outcrops adjacent to the Start Boundary Fault. There is little evidence for these, although poorly developed kink bands are quite common and always re-fold the earlier structures. The vergence of the F3 crenulations is consistently northwards and the facing of these structures is virtually impossible to establish, given the highly strained nature of these shales.

At Tinsey Head the strain is still high, with lenticles of bedding clearly wrapped by an earlier tectonic fabric (evidenced by fine red ^{iron}oxide seams), indicating that the main cleavage is at least S2 in age. This early fabric has not been noted by previous workers (e.g. Chapman *et al.* 1984, Seago, 1991) who ascribe a D1 age to the main cleavage and map these later structures (F2) as primary folds. The F2 vergence appears to be northwards, where measurable, whilst the facing appears to be predominantly upwards to the south (Fig. 4.14). Coward and McClay (1983) observed both upwards and downwards facing folds in the Beesands area, and argued that the main structure here must therefore be at least D2 in age. Whilst there is ample

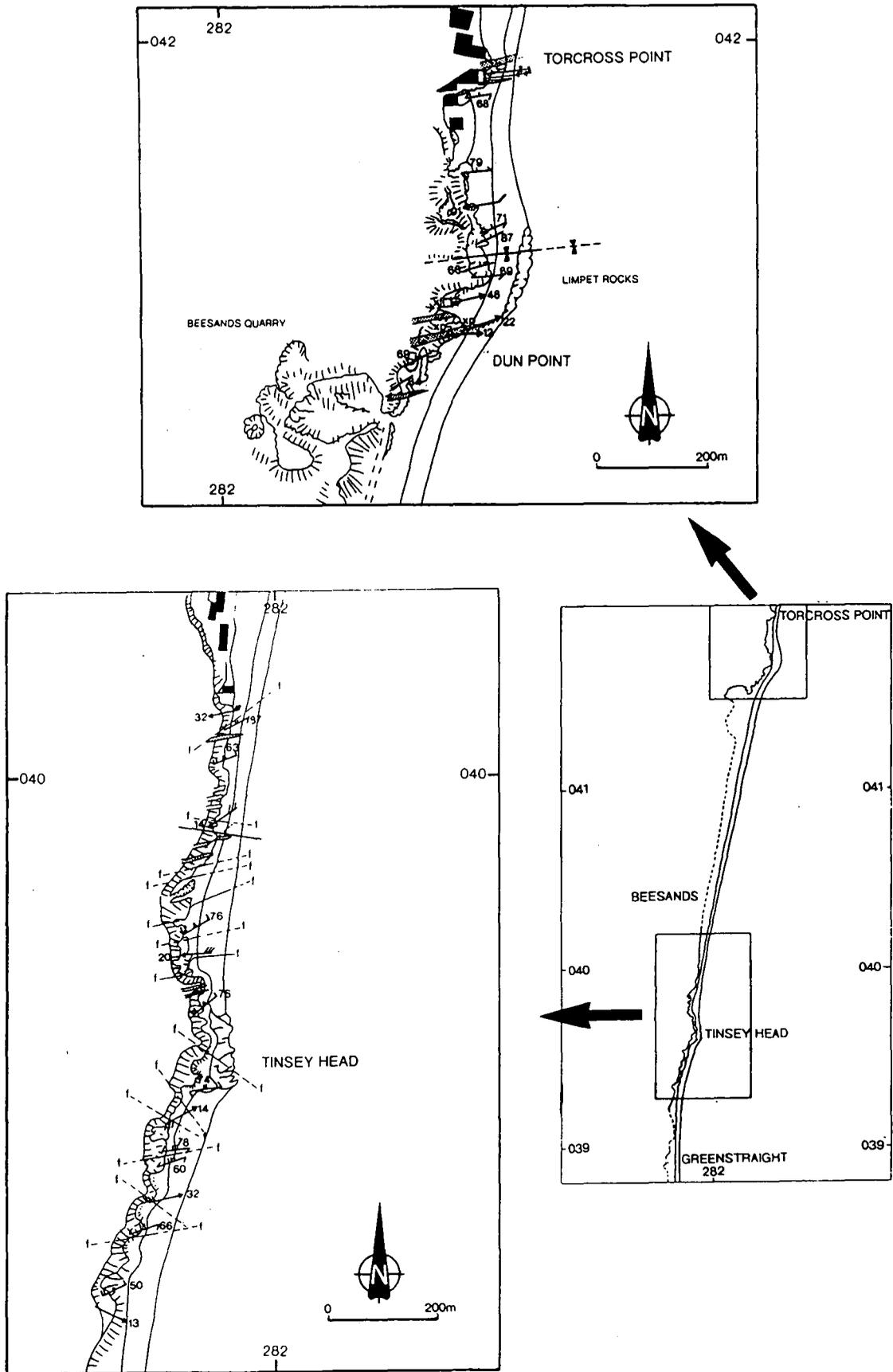


Figure 4.13 Detailed geological map of the Torcross area (**top-b**) and the Tinsey Head area (**bottom left-a**). Relative position of enlarged areas (a & b) along coast is shown bottom right. Figure 4.2b shows the overall location of these maps in the South Hams area. Fig. 4.3 gives the key.

evidence of refolding in the Beesands area, such facing variations can also arise due to primary controls (see below). As the S3 overprint diminishes, small F2 folds become apparent, their vergence being highly variable as a consequence of their steep plunges, some of the folds plunging through the vertical to verge southwards. Also, there are examples of these centimetre scale folds displaying a double vergence, which is strongly suggestive of sheath folds. If the double verging fold is projected towards its closure, the fold limbs have Z-geometries where they pass through the vertical. If sheath folds do occur, then facing variations may be primary, as a sheath fold has two diametrically opposed facing orientations (see 1.6). The significance of such structures and their relationship to the Start Boundary Fault is discussed in 4.6.6.

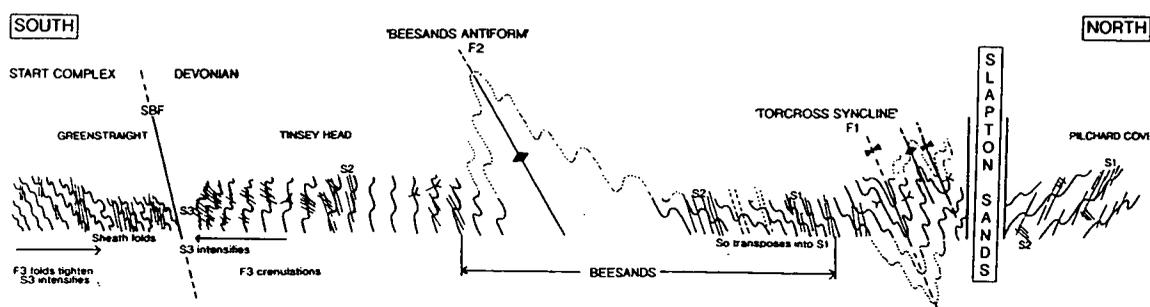


Figure 4.14 Simplified cross section through Fig. 4.13

Between Tinsey Head and Beesands, a monotonous succession of shales occurs in which the main cleavage (S2) is crenulated in localised narrow zones by tight, somewhat angular north verging F3 structures. The main cleavage is locally offset by both dextral and sinistral shear bands, with again the dextral shears being predominant. These structures are discussed along with those of the west South Hams coast and the Start Complex in section 4.6.5. Traces of apparent bedding appear in the outcrops at Beesands village [SX 2819, 0402], but on close examination display a bedding parallel tectonic fabric. Cleavage/banding relationships again show that F2 folds verge towards an antiformal closure somewhere within the shingle of Bee Sands.

At Dun Point [SX 2823, 0416; Fig. 4.13b] the strain is lower, although narrow zones of tight post-main cleavage (S1) crenulations do occur, and may be related to the many intrusives here, i.e. there is some evidence of strain intensification adjacent to these bodies, with zones of enhanced pressure solution and transposition of bedding into locally intense cleavages. The first evidence of original bedding occurs in the exposures south of Dun Point, and thus the transition from S2 to S1 dominated fabrics must lie obscured by the shingle along Bee Sands. Bedding/cleavage relationships define an open, northerly-inclined F1 syncline within the shales at Limpet Rock [SX 2824, 0418; Fig. 4.14]. Northwards, an upward facing and moderately WSW plunging

anticline/syncline pair with a steeply north dipping axial planar cleavage (S1) is developed. Mineral stretching lineations are hard to discern within the shales, but appear to be ENE-WSW. This is confirmed by the deformed strain markers within the large intrusives, which all display pronounced strike parallel extension (see Plate 3.10; Fig. 4.13b).

There is a complete lack of exposure between Torcross Point and the cliffs south of Pilchard Cove [SX 2837, 0460] because of Slapton Ley. The first exposures to the north of this area are very monotonous Dartmouth shales, with numerous silty and sandy horizons. The main cleavage dips moderately south and a fine mineral stretching lineation plunges down the cleavage dip, so once again an important structural transition is obscured by beach, this time within Slapton Sands. Upright, north facing primary folds may be mapped as far as Pilchard Cove, beyond which access is limited. The main cleavage (S1) is locally cut by moderately north dipping crenulations (F2), although these zones are generally rare.

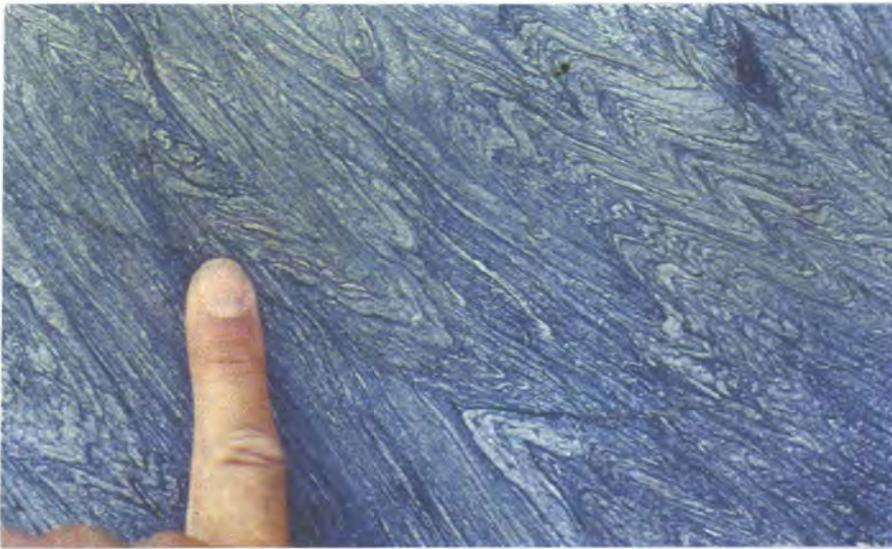


Plate 4.32 Crenulation and transposition of early fabric [bedding+main cleavage (S2)] into localised zones of intense north-dipping S3 cleavages. The S3 crenulations consistently verge northwards and are related to the backsteepening of the fabric in this region. View looking due west, Meadfoot shales, Tinsey Head, [SX 2819, 0396].

Inland Exposures

Due to the lack of critical exposure on the east coast, the inland exposures around the Kingsbridge Estuary were briefly examined in order to provide an along strike projection of the important structure. Unfortunately, exposure is poor and, whilst the orientation of the primary folds could be generally assessed, their facing and overall vergence was impossible to determine.

The exposures around Newbridge [SX 2745, 0431] display south verging primary folds of Meadfoot shales, cut by an axial planar sub-vertical slaty cleavage. This generally dips north, with minor bedding/cleavage changes defining a series of parasitic folds on the steep limb of a large south verging structure. Stretching lineations are impossible to assess in these exposures, and the occasional intrusives lack any obvious deformed markers. Narrow zones of higher strain, characterised by intense quartz veining and tight, steeply plunging later folds, may be related to local thrusting and could not be mapped out regionally. Considerable evidence of thrusting occurs along these sections, ranging from metre scale, gently north dipping fault zones with variable offsets, down to discrete steeply dipping ductile shears associated with arrays of boudinaged quartz veins. Shear sense on these north dipping structures is consistently top to the south. North of Newbridge, the stream section exposure was very poor and the structural transitions lacking on the east coast could not be mapped with any certainty. The estuary exposures immediately to the north of the Start Boundary fault are described in 4.3.3.

4.4.5 The Microstructure

The Devonian shales are very fine-grained and their mineralogy indicates deformation at low temperatures and pressures (e.g. Warr *et al.* 1991; Warr, 1993), and thus it is reasonable to expect that DMT will have occurred. Several sections were studied, ranging from relatively undeformed Dartmouth Beds in the north of the area, through to intensely crenulated Meadfoot Beds adjacent to the Start Boundary Fault.

In thin section, the slates predominantly comprise quartz and white mica, which are segregated into diffuse domains to produce an intense penetrative spaced cleavage. Sub-parallel to the aligned micas are continuous anastomosing solution seams, coated in films of opaque oxides/graphite. In the Dartmouth shales, thin folded grit horizons are flanked by zones of micas. In the fold limbs, the larger grit grains are relatively undeformed, exhibiting a mild undulose extinction and there is no evidence of recovery, sub-graining or dynamic recrystallisation. Grains are markedly flattened, wrapped by thick cleavage seams and fringed with elongate mica beards, this imparting a trapezoidal appearance to the fabric (Plate 4.33a). The mica domains are crenulated into open asymmetric kink folds cut by a weak spaced cleavage (S2), which lies at low angles to layering. The mica crenulations do not transgress the thick planar truncated boundaries with the grit horizons (Plate 4.33a). In the fold (F2) hinge regions, the primary mineralogical segregation breaks down due to the intense, spaced crenulating cleavage, which cuts the grit horizons and reorientates the elongate quartz grains into parallelism. The new segregation of quartz into the hinges and micas into the limbs

further serves to destroy the original segregation. Again the flattened, weakly strained quartz grains show no evidence of recovery or dynamic recrystallization (Plate 4.33b). Even in the multiply cleaved Meadfoot shales, the aligned detrital quartz grains are relatively undeformed and show no signs of extensive crystal plasticity. The presence of intense tectonic fabrics defined by mineralogical segregation and sub-parallel solution seams, allied to the extensive precipitation of material into pressure shadows, beards and veins, is convincing evidence that fluid assisted DMT is the dominant deformation process in these shales. This is consistent with the suggestion that they deformed at considerably lower temperatures and/or strain rates, compared to the Start schists.

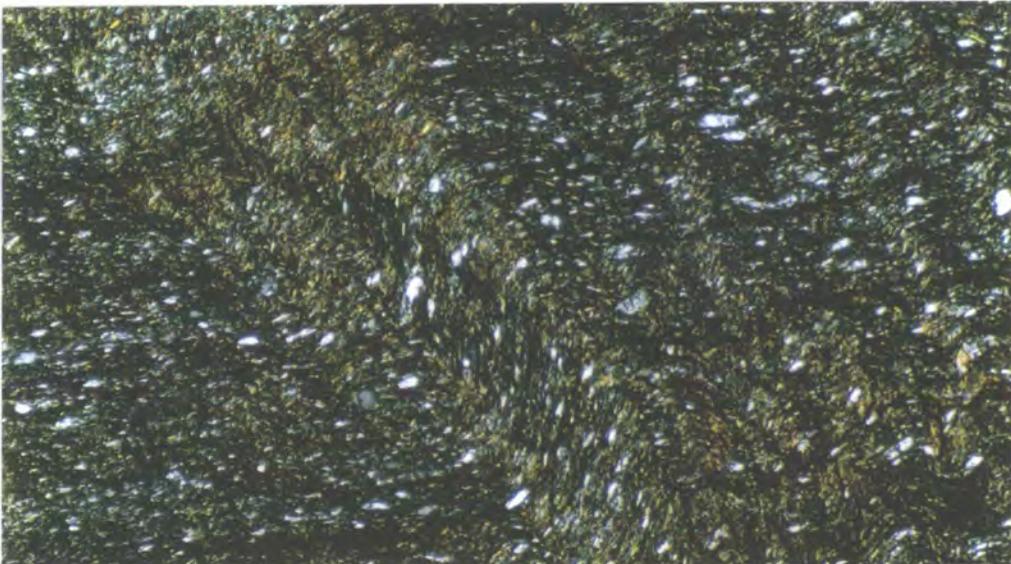
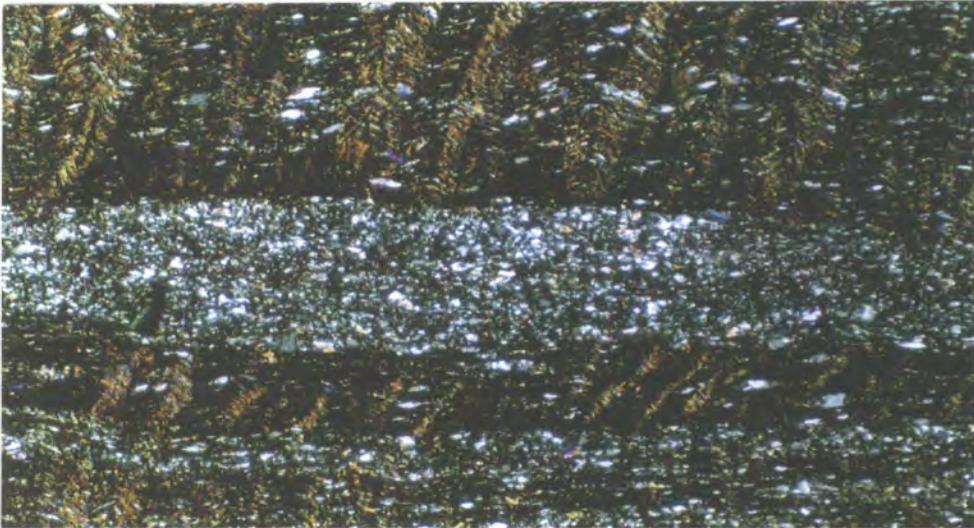


Plate 4.33 a)-top Thin section of Dartmouth shales showing thin folded grit horizon. Flattened detrital quartz grains are relatively undeformed and wrapped by thick anastomosing cleavage seams. Micas segregations are crenulated and separated from the quartz domains by thick planar truncations. **b)-bottom** in the hinge region the crenulating cleavage is at a higher angle, breaking down the distinct primary segregation and transposing the elongate quartz clasts into a new spaced cleavage. Dartmouth Beds, Pilchard Cove, [SX 2844, 0466] under X10 magnification and crossed polars. FoV=3.2x1.7mm.

4.4.6 Summary

The structural pattern of SW Devon is one of north verging and facing primary folds inclined to the south, which steepen progressively southwards to dip north and face south at the Start Boundary Fault. The primary folds also tighten, from open, upright structures in the north, to transposed intrafolial lenticles on approaching the Start Boundary Fault (*cf.* Hobson, 1976b). Late north verging and dipping crenulations intensify southwards into the SBF, whilst 'eyed' sheath folds and doubly verging structures attest to the increased strain approaching this fault zone. Accompanying these changes is a switch in the finite stretching direction from the usual down-dip orientation to an anomalous along-strike position.

4.5 Structural correlations across the SBF zone and its regional significance

Within the Start Complex schists, south of the SBF, the dominant folds (F3) verge south, dip north and have moderate west plunges. These centimetre scale folds define a series of upright metre scale antiform/ synform pairs, which tighten moving northwards towards the SBF, whilst the schistose fabric steepens towards the sub-vertical. Locally, zones of anomalously steeply plunging F3 folds, and apparently complex vergence patterns, are evidence of sheath folding. Occasionally, centimetre scale 'eyed' sheath closures occur within the schists, and attest to the increasing shear strain approaching the SBF. Further evidence of increased strain comes from the intense pressure solution striping, and transposition of early fabrics into the dominant cleavage.

The structural pattern of the Devonian sequence is summarised above (4.4.6), and again shows an intensification of strain into the SBF. Whilst the F3 folds in the Start schists tighten and steepen into the high strain zone, the structures in the Devonian shales are transposed into several generations of cleavage and the outcrops appear to be more highly deformed. This may be a consequence of either rheological differences, in that the shales are essentially 'softer', or a consequence of the stronger pre-existing anisotropy in the schists controlling the distribution of strain. The tight, locally F3, crenulations, which are well developed north of the SBF, would appear to have no counterpart in the Start Complex, and are probably related to the regional backsteepening (see below). The late, flat-lying fabrics which occur within some greyschist sequences are thought to be a consequence of local strain heterogeneity (4.6.3), and are not regionally significant. Field relationships (see Figs. 4.12 & 4.14) and microstructural evidence (see 4.4.5) strongly suggests that the F2 folds in the Devonian may be correlated with the main phase (F3) folds in the Start Complex, and thus the F1 Devonian folds are analagous to the early isoclines in the Start schists.

The backsteepening of the main fabrics across the South Hams region (Fig. 4.15) has been variously interpreted. Richter (1967) considered it to be primary, related to the strain regime in which the folds formed, this imposed by deep-seated basement structures. Unfortunately, he did not expand on his mechanism of basement control. Hobson (1976b) attributed the backsteepening, in part, to backfolding, an idea enlarged upon by Shackleton *et al.* (1982) and Coward and McClay (1983), who related this to southerly directed backthrusting. Evidence for southerly directed shear comes from the small north dipping crenulations (F3) of the main cleavage, which, along with southerly directed shears and thrusts, intensify southwards. Chapman *et al.* (1984) concur with the above but attribute a local D2 age to the crenulations. Whilst the backsteepening is undoubtedly a regional D2 modification of the primary structure, the post-main cleavage backfolds may have variable local chronologies.

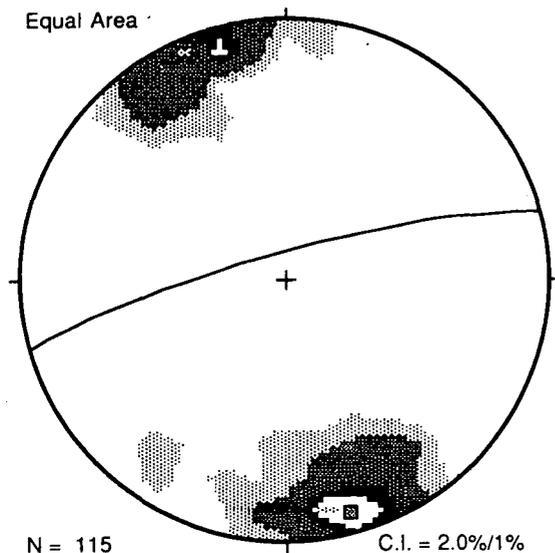


Figure 4.15 Contoured equal area stereogram of 115 poles to the main cleavage in the Devonian shales immediately north of the Start Complex. Predominant dip of slaty cleavage is steeply to the north (mean pole is 076/84N), evidence of secondary backsteepening in this area.

Coward and McClay (1983) reinterpreted SW Devon in terms of thin-skinned thrust tectonics and suggested that the backthrusting here is comparable to the 'retrocharriage' zones of the Alps. However, Holdsworth (1989a) argued that the Start Boundary Fault coincides with a major basement fault at depth (see 3.5), this structure exerting a local control on the overlying Variscan thrust nappes. This fault margin buttress could have caused the secondary backfolding seen in both the Devonian sequence and the Start schists (4.2.6). Shackleton *et al.* (1982) note that the backfold may be traced across S. Devon, through Plymouth, to Perranporth in Cornwall, although its wavelength and amplitude decreases westwards. They suggest that the structure imposed an E-W grain on this part of SW Devon, as further to the north (Torbay-Newton Abbott) the strike trends NE to ENE. However, thrusts generally propagate in an arcuate pattern and pass laterally into ramps. Differential thrust movement is well documented in foreland propagating thrust sheets in SW England (e.g. Rattey & Sanderson, 1982), and it is reasonable to expect backthrusts to behave

similarly. Also, where backthrust zones are recorded in SW England, they are not very laterally extensive (e.g. Tintagel, Shackleton *et al.* 1982). It is thus difficult to explain the great lateral continuity and linear outcrop pattern of the backfolded zone across this region in terms of simple backthrust geometries. It is unlikely that a backsteepened ductile thrust could have exerted such a profound influence on the orientation of the overlying structure. It also cannot account for the considerable evidence of along-strike dextral shear in this region. A significant steep basement fault would be more likely to exert such a control in the immediate Start area, and could readily explain the great lateral continuity of the backfold zone. Gravity data from across the Start area indicates a sudden anomaly several kilometres north of the Start Boundary Fault (Bott & Scott, 1964), implying that this is one of several major steep faults in the South Hams area, these quite probably related to a deep basement structure. Sadler (1974) suggests that the E-W dykes of the South Hams region are a further surface expression of these basement controlled faults.

Thus the SBF appears to mark the site of an E-W trending, deep-seated basement fault, which acted as a buttress to the to the NNW transporting Variscan nappes. The high angle obliquity of this buttressing would induce a degree of dextral shear in the shortening cover sequence, and would result in a strain localisation in the immediate SBF area. The presence of sheath folds along the SBF zone attests to this intense along-strike shear, whilst the abundant shear sense indicators are evidence that dextral simple shear, rather than pure shear, was the predominant deformation process (see 4.6.4). The westwards continuation of the backsteepened zone from South Hams into Cornwall (Shackleton *et al.* 1982), and the correlation of the SBF with the Perranporth Pentewan Line (e.g. Dearman, 1971; Hobson, 1977), may suggest that the basement controlling structure in S. Devon extends westwards into Cornwall (e.g. Holdsworth, 1989a; see 3.5). Field evidence for such a structure is discussed in 5.6. The evidence for dextral transpression within the S. Devon area, due to oblique Variscan orogenesis, is discussed in 4.6, overleaf.

4.6 Discussion and evidence for Transpression

Evidence presented in the previous sections shows that the small scale structures across South Hams (4.2, 4.3, 4.4), and the regional structural pattern (4.5) are consistent with the cover sequence being deformed by dextral simple shear, superimposed upon regional shortening. A detailed analysis of the evidence for dextral transpression is given below.

4.6.1 Origin of east-west lineations

The majority of the field data used in this discussion comes from the Start Complex, as this represents the greatest area mapped. Also, the data recorded in the schists came largely from excellent three-dimensional exposures, whilst that of the Devonian sequence was restricted to two-dimensional slaty cleavage (XY) surfaces. Although it is likely that the measurement of data within this slaty cleavage resulted in biased azimuth readings, the data from the Devonian sequence largely concurs with that of the Start Complex.

Field observations and contoured stereonet data for the Start Complex and the Devonian sequence 5 km to the north apparently show that the primary lineation (L2 in the Start Complex; Fig. 4.16) lies parallel to the E-W trending and gently east plunging early fold axes (F2 in the Start Complex; Fig. 4.17). This along axis orientation of the lineation is anomalous with respect to most of SW England (except for some localities in N. Cornwall; see 5.6.2), where the mineral stretching lineation plunges down the dip of the cleavage, normal to primary fold axes and parallel to the regional tectonic transport direction (e.g. Sanderson & Dearman, 1973). The regional stretching orientation may be traced south, within the Devonian rocks, until around Challaborough, a location some 4.5 km north of the Start Boundary Fault [SX 2646, 0449], it switches to the anomalous E-W orientation (4.4.3; see Fig. 4.11b).

Orogen-parallel lineations have previously been attributed to arcuation, rolling or superimposed deformation (e.g. Shackleton & Ries, 1984), but such mechanisms cannot account for the localised nature of the anomalous lineations in SW England. A number of models have been proposed to account for the oblique lineations in this area: Coward and Smallwood (1984) suggested that strike-parallel extension occurred during NNW thrusting due to one of three mechanisms: i) local hangingwall strains formed by stacking of thrust sheets in culminations (Fig. 4.18a). ii) Stretching at lateral tips, caused by thrust sheet displacement rate being greater than lateral migration rate of the thrust tip. This gives rise to oblique folds in the zone of differential movement (Fig. 4.17b). iii) Sticking at lateral tips, which then act as poles of rotation, resulting in arcuate extension parallel to the thrust trace and normal to the movement direction (Fig. 4.17c). However, in the case of S. Devon, the lineation switch from distinctly

down-dip to along-strike occurs within ~200m (4.4.3, Fig. 4.11b), and no oblique trends are noted. This would tend to rule out the above models, as they produce arcuate lineations or gradational changes in the lineation orientation.

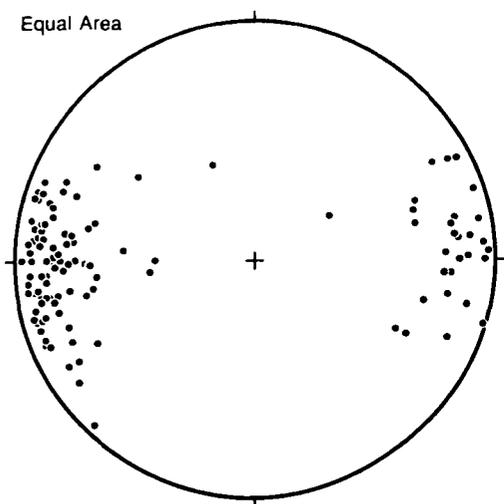
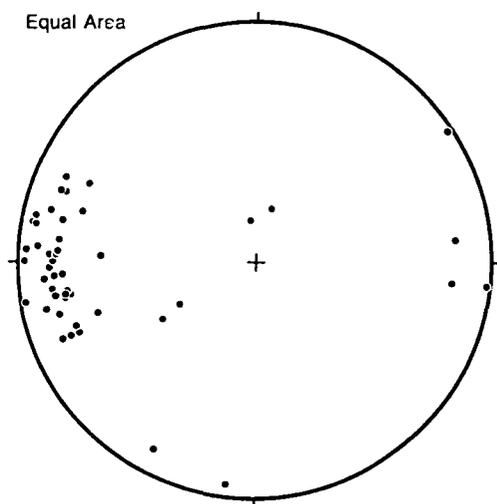


Figure 4.16 (left) Equal area stereogram of 128 mineral lineations in the Start Complex. Mean lineation plunge is 9.4° towards 268° .

Figure 4.17 (right) Equal area stereogram of 58 F2 fold axes in the Start Complex. Mean axial plunge is 3.4° towards 268° .



A further possibility is that the lineation arose through axis parallel extension during folding, normal to the tectonic transport direction. Such axis parallel lineations are, however, both highly unusual and very unlikely, necessitating a sudden change in the mechanism of folding moving from the zone of axis parallel to axis normal lineations.

The lineations are probably best explained in terms of a switch in the maximum finite extension direction due to partitioning of transpressional strain. In S. Devon, there is much evidence of dextral shear (4.6.5), rotation of early fold axes and lineations (4.6.1, 4.6.2), and switching in the lineation direction (4.4), all observations consistent with this area being a linear E-W zone of primary transpression. Sanderson and Marchini (1984) show that, in certain situations, the maximum finite extension direction (X) can switch from vertical, in the zone of net shortening, to horizontal in the zone of net along-strike simple shear (see Chap. 1.2). The orientation of X is dependant on which of the transpressive strains dominates, such that the strain may partition into zones of pure shear compression and strike-slip simple shear, with vertical and horizontal lineations respectively. McCoss (1986) showed that, given the stability of the strain regime, if axis swapping occurs in zones, it is likely to be the result of changes in the displacement direction at the zone boundary, rather than due to

progressive strain. A spatial partitioning of the transpressive strain effectively alters the local displacement direction and would thus precipitate the axis swapping. Local displacement directions calculated from tension gash arrays, using the McCoss (1986) construction (see 4.6.5), confirms this partitioning effect, with transport parallel and transport normal displacement vectors (S) occurring.

The parallelism of the early lineation (L2) with consistently north verging early fold axes (F2) still remains a problem, however, and would seem to indicate that these structures are not strictly coeval. It appears that the F2 folds would have to initiate first during the regional shortening. As these folds tighten, the deformation switches from simple shortening to transpression, with the superimposition of the dextral shear component. Thus the finite stretching direction switches into the along strike orientation, parallel to fold axes, and normal to the regional trend. These structures then undergo clockwise rotation in the dextral shear environment, into parallelism with the lineation and prior to F3 refolding. Given the high angle obliquity of the collision in this area (4.5), it is quite probable that shortening (pure shear) was the predominant mechanism during early deformation. Only with increased deformation was dextral simple shear superimposed on the shortening cover sequence within a broad zone centred either side of the SBF (4.6.5). The presence of calc mylonites and small intrafolial sheath folds in some localities along the SBF indicates that syn-D2 shear strain was high in this area.

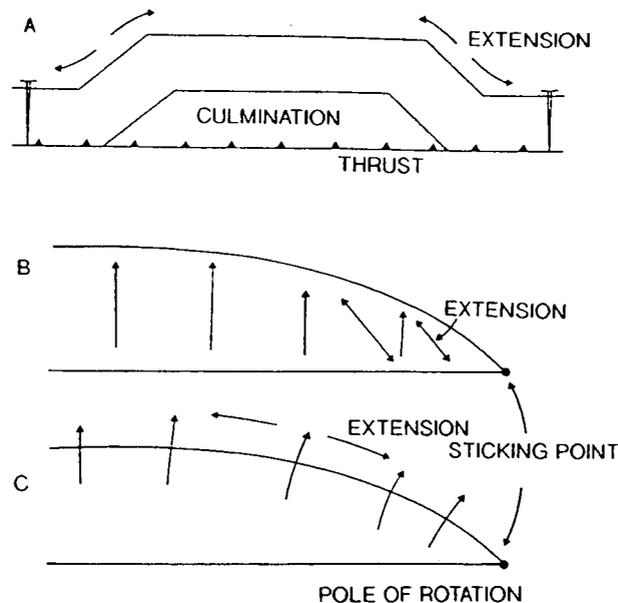


Figure 4.18 Models to produce extensional lineations normal and oblique to the transport direction. a) cross section through a culmination, with transport normal to the section. b) plan view of a thrust tip zone, showing oblique extension due to shear produced by differential movement. c) plan view of a thrust tip zone showing extension along the thrust belt due to rotation round the lateral tip. After Coward & Smallwood (1984).

4.6.2 Relationship between F3 folds and early lineations in the Start Complex

The early penetrative mineral lineation appears to be parallel to F2 axes (see 4.6.1 above) and is clearly refolded by later F3 folds (see Plate 4.2). Field observations suggest that, whilst some cases of strictly coaxial refolding do occur, the majority of examples show a degree of obliquity, with the early lineation typically lying 5-35° clockwise of the later fold axes. This is borne out by stereonet data from across the Start Complex: a stereonet of 658 F3 axes gives a mean fold plunge of 12° towards 260° (Fig. 4.19). For 128 recorded mineral lineations the mean plunge is 9° towards 268° (see Fig. 4.16), an 8° clockwise disparity. Also, contoured plots of F3 axial planes have a mean plane dip of 60° towards 348°, whilst those of F2 folds have a mean plane dip of 45° towards 178°, this again giving a ~10° clockwise disparity (Figs. 4.6, 4.20). Clearly the fold and lineation data are consistent with a degree of clockwise rotation of F2 folds prior to, and during, F3 folding, resulting in a small degree of obliquity between these structures.

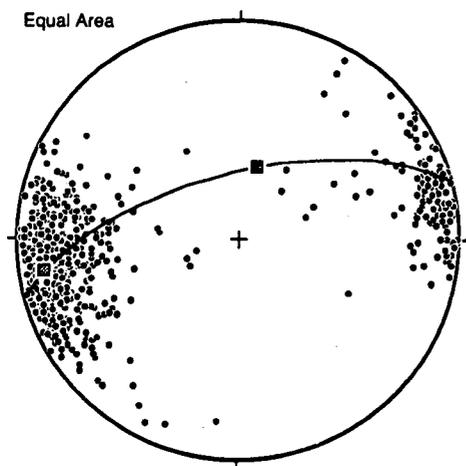
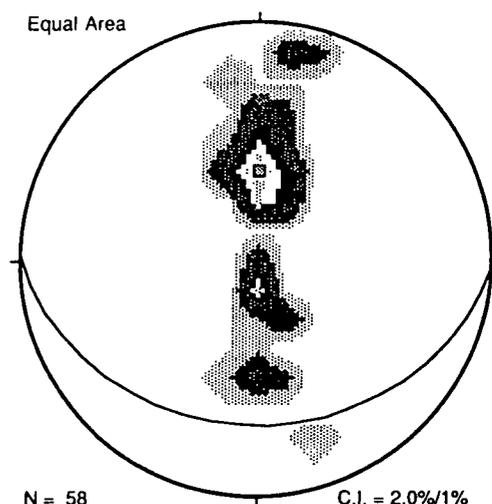


Figure 4.19 (left) Equal area stereogram of 636 F3 fold axes in the Start Complex. Mean fold axial plunge is 12° towards 260°

Figure 4.20 (right) Equal area stereogram of F2 fold axial planes in the Start Complex. Mean pole is 093/54S



Sanderson and Marchini (1984), in applying the transpressive model to folding, show that the incremental minimum stretch is at high angles to the transpressive zone, and thus folds, etc. would be expected to initiate at low angles to the zone (see Fig. 1.04, Chapter 1). These folds subsequently tighten and rotate passively towards the shear plane. Under dextral transpression this rotation would be clockwise and, for constant incremental strain, later folds will form in the initiation position, and thus lie

anticlockwise of the earlier fold axes. Thus the relationship between the F3 fold axes and the mineral lineation is consistent with dextral transpression.

The cases where the mineral lineation lies parallel to the F3 axes could arise due to the earlier lineation being overprinted by either an intersection lineation, or a later F3 related stretching lineation. However, cases of the earlier lineations clearly lying anticlockwise of the later fold axes are harder to explain in this way. Such lineations typically only lie a few degrees anticlockwise off axis, (never $>10^\circ$). The simplest explanation for the anomalous F3-lineation relationships is that the F3 axes had significant initial curvature, this resulting in a locally variable obliquity of lineation with both clockwise, parallel and anticlockwise relationships arising. The predominance of clockwise relationships would, in addition to indicating that the lineations have rotated clockwise during shear, suggest an asymmetry in the initial curvature of the F3 axes relative to the lineation, which is also consistent with a dextral shear sense.

4.6.3 F3 fold development in the Start Complex

Variably orientated and apparently synchronous S3 cleavages were noted in the greyschists across the Start Complex, and in none of these cases could any refold or crenulation relationships be established between fabrics in different orientations. In some cases, e.g. those at Great Mattiscombe Sands [SX 2817, 0369] the effect appears to reflect a marked strain discontinuity between deforming layers and a refraction of the cleavage between them. Treagus (1983) has shown that cleavage development from layer to layer in a deforming multilayer will vary in intensity in inverse proportion to the competence, assuming that competence contrasts remain fairly constant during deformation. This means that competent psammitic layers will develop a weak fabric irrespective of the number of deformation phases, whilst incompetent pelitic layers will develop strong cleavages. Refraction of the cleavage at the layer boundaries, which results from the competence controlled refraction of the principal strain axes, enhances the effect of two apparent cleavages, the intense fanning cleavage in the pelitic layers refracting into the weak spaced cleavage in the psammitic layers. However, F3 structures around Hugh's Hole [SX 2679, 0387], include a steep, strongly developed cleavage (S3a) that fans around tight, upright F3 folds, whilst in the steep limbs of these folds, a sub-recumbent weak spaced cleavage (S3b) occurs. There is no evidence of either fabric being crenulated or refolded. Explaining these anomalous structures is difficult. Treagus and Treagus (1981) modelled the development of fold axes and axial surfaces which are non-orthogonal to the axes of the finite strain ellipsoid. They show that folding on two axes, i.e. the simultaneous propagation of two fold sets, their axial migration and interaction, can occur in areas of constrictional strain, and that the

structures that arise during such single phase constrictional folding are broadly similar to those seen in areas of polyphase deformation. However, the distinguishing feature is that the former are only ever associated with a single cleavage, such that folds of widely varying plunge, attitude and scale may arise, initiating on pre-existing anisotropies, but they will always be cut by a single orientation of cleavage (which forms parallel to the XY plane of the finite strain ellipsoid). Even with the marked anisotropy of the greyschists, this model cannot account for the observed cleavage relationships.

It is possible that these structures are the result of conjugate folding. Ramsay and Sturt (1963), in describing folds from Sørøy, Norway, noted that such folds are often markedly different in size, with one smaller set occurring as minor folds, of opposite vergence and axial planar dip, to the other, larger set. However, they related such geometries to movement on sets of complementary, gently inclined shear planes, with unequal displacements on these shears giving the disparity in fold size. They further noted that the intersecting axial planar fabrics are crenulated, the minor fold set often only manifesting as crenulations of the major cleavage. Ramsay (1962b) also noted such conjugate folds, making the important observations that the fold axial planes tend to represent shear surfaces and that these folds only occur in thinly bedded, finely laminated rocks, and are thus morphologically similar to conjugate kink bands.

It would appear that, although the geometry of folding at Hugh's Hole is suggestive of conjugate folding, the nature of the lithology, the occurrence of almost mutually orthogonal cleavages, and the lack of axial planar shear surfaces or crenulation cleavages, rules out such a mechanism. The development of two synchronous (D1) cleavages during a single deformation phase is discussed in Harris *et al.* (1976), from the Dalradian sandstones of the Tay Nappe. In this case, the first cleavage, considered to be a slaty cleavage, is the result of parallel grain alignment, and is cut by a fanning, spaced pressure solution cleavage. In short, inverted fold limbs these two fabrics are oblique and are thus mappable as distinct, but synchronous cleavages. Unfortunately the cleavages recorded at Hugh's Hole are both clearly pressure solution cleavages of variable intensity, and are developed in highly anisotropic rocks, late in the deformation history. These cleavages remain problematic and, whilst the marked inhomogeneity of the strain in the greyschists may partly be responsible for the observed structures, it cannot solely be responsible.

Hobson (1977) recorded small, poorly developed late recumbent folds of the schistosity throughout the Start Complex, ascribing these to a post D3 deformation event. He correlated these folds with similar structures seen in S. Cornwall. However, it is likely that the structures that he observed are similar to those described above and are not a regional fold phase. Further examples were noted at Start Point, Deckler's

Cliff and especially Portlemouth and in none of these areas does the sub-recumbent (S3b) cleavage result in refolds of the upright folds. Thus these structures are unlikely to be of regional importance, and Hobsons (1977) correlation of these fold phases with folds in S. Cornwall seems untenable.

In the majority of localities, the S3 cleavage appeared to be more or less axial planar to the F3 folds. At Splatcove [SX 2731, 0375], however, the strongly developed cleavage is clearly non-axial planar, transecting the F3 folds in both a clockwise and anticlockwise sense. This transection may be a consequence of the variability of the cleavage surfaces along axis or may be due to the pre-existence of a strong linear anisotropy at various angles to the later fold axes in the schists. This predisposes the later folds to develop with their axes parallel to the earlier linear fabric, whilst the cleavage surfaces are obviously not as sensitive. Probably the simplest explanation for the transection, however, is that of Treagus and Treagus (1981; see above). Their modelling, in its simplest form, assumes homogeneous layers, with slight initial irregularities to allow fold initiation. In the case of the greyschists, the pre-existing anisotropy is very marked, and would have a profound effect on fold axis initiation, migration and interaction. Treagus and Treagus (1981) note that "...a wide field of main fold axes,...would be expected especially in layers with strong initial inhomogeneities." (p. 15). Thus it is possible to generate folds with axes which are markedly oblique to the cleavage surfaces and, given the marked inhomogeneity of the greyschists, the observed transections are quite small.

Given the widespread evidence of dextral transpression associated with F3 folding in the Start Complex, it might be expected that a consistent sense of anticlockwise fold transection would be expected (*cf.* models of Soper & Hutton, 1984; Soper *et al.* 1987; Woodcock *et al.* 1988). However, as noted in 1.2.4, Treagus and Treagus (1992) show that transpression alone does not provide an immediate theoretical explanation for folds transected by synchronous cleavage, and thus transected cleavage relationships should not be universally applied to transpression zones. The transected folds recorded at Splatcove above are isolated and anomalous with respect to the rest of the Start Complex.

4.6.4 Sheath Folds at the Start Boundary Fault: Geometry and Significance

A characteristic feature of the highly strained lithologies along the SBF is the presence of sheath folds. These range from tens of centimetre-scale elliptical 'eyed' closures, through doubly verging folds which close on themselves, to localised zones of steeply plunging folds which typically have a characteristic Z-geometry as their hinges pass through the vertical.

If the SBF is considered to coincide with a major high strain zone, with the dextral simple shear predominating, the geometry of the sheath folds should reflect this strain regime. The vast majority of the steeply plunging folds within the sheath fold zones display Z- or dextral geometries as their axes pass through a vertical plunge. This is also seen in the small doubly verging structures in the schists along Hallsands. If the fold limbs of these structures are projected out of the plane of section towards the sheath closure, the limbs always display a Z-geometry on plunging through the vertical (Fig. 4.21). These steeply plunging Z-folds may be explained in a number of ways: They may simply be shear sense criteria, produced by sub-vertical dextral shear along the SBF. Alternatively they could be reorientated folds, rotated from sub-horizontal E-W plunging to sub-vertical plunging during transpression, according to the rotational models of Fossen and Tikoff (1993; see Fig. 1.10). A third possibility is that they represent the steep hinge zones of sheath folds, although this can probably only account for a small number of the observed sub-vertical Z-folds. The marked bias towards dextral asymmetries in these vertically plunging folds indicates that their geometries are significant. In many cases, these folds simply reflect intense sub-vertical dextral shear, and may be considered as one of the many dextral shear sense criteria. In the demonstrable examples of sheath folds, the marked bias towards dextral asymmetries indicates that dextral shear is an important mechanism in the formation of these structures.

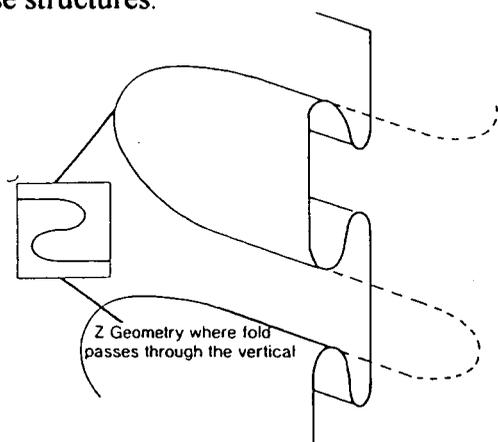


Figure 4.21 Diagrammatic representation of the curvilinearity of sheath fold hinge lines, showing how the fold limbs display Z-geometries as the fold plunges through the vertical.

Further evidence of dextral shear comes from the well exposed sheath structures within the Devonian shales exposed in the Kingsbridge Estuary and the north side of Hope Cove. In these localities, single continuous folded mudstone horizons within the dark shales are seen to vary from plunging moderately east, through to plunging steeply west (displaying Z-geometries where they plunge through the vertical), and back to gentle easterly plunges (see Plates 4.25 & 4.28). If this hinge rotation is viewed within the cleavage surface, it also shows a Z- asymmetry. Such hinge line variations in sheath folds are probably the result of initial perturbations along the hinge of the original whaleback fold, which are accentuated as the hinge is elongated during shear. This

should result in a sheath fold with small 'parasitic' asymmetric sheaths along its' hinge, which verge towards the sheath closure (Fig. 4.22). Thus, the sheath fold hinge vergence noted in the above localities could indicate the position of a larger sheath closure. However, all four observed examples of this hinge variation had the same dextral asymmetry, so either all these localities fortuitously lie on the single long limb of a larger structure, or the consistent dextral vergence reflects a pronounced dextral asymmetry to the larger sheath fold. The latter model would necessitate a degree of obliquity between the original hinge and the shear couple and, whilst it could offer more evidence of dextral simple shear along this high strain zone, it cannot be fully tested here and remains speculative.

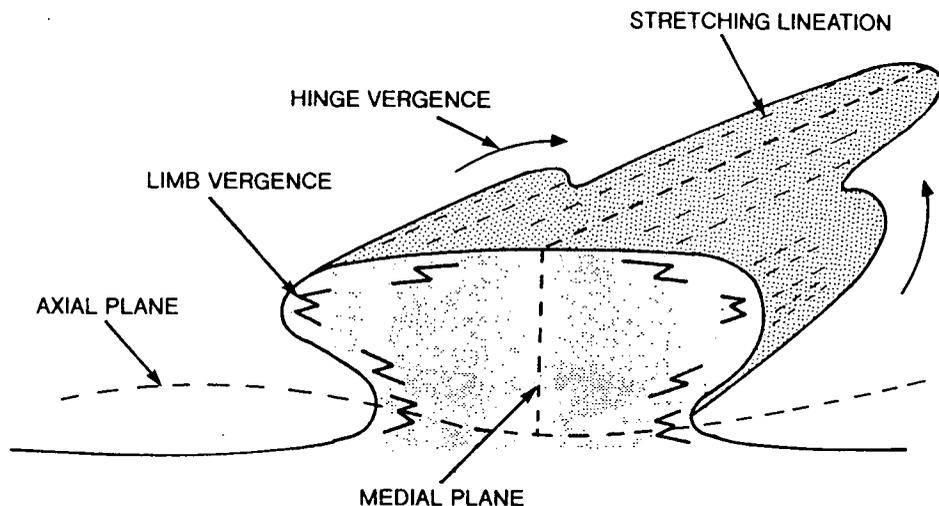


Figure 4.22 Diagrammatic representation of sheath fold hinge and limb vergence. Initial irregularities in the original whaleback hinge are accentuated into 'parasitic' folds on the flanks of the larger sheath. Limb asymmetry verges towards the original antiformal closure, whilst hinge asymmetry verges towards the sheath closure.

In conclusion, sheath fold geometries along the SBF indicate that dextral simple shear is an important component of the overall strain regime, pure shear being insufficient to generate these structures. This is further evidence for localised dextral tranpression along the SBF zone.

4.6.5 Shear Bands and Extensional Fractures

Shear Bands

Shear bands are well developed across this region, especially in the Start schists. In many of the more sheared greyschist localities within the Start Complex, the outcrop appears to develop an S/C type of fabric (e.g. Plate 4.7), with obvious shear planes offsetting and rotating the strong planar schistosity. In all of these cases, however, the shear planes are affecting an earlier foliation, i.e. the shear planes and foliation are

never synchronous, and thus these structures are shear bands rather than an S/C fabric (e.g. Lister & Snoke, 1984).

Within the Start Complex, dextral shear bands (DSB) are clearly predominant, outnumbering the sinistral shear bands (SSB) by a ratio of ~3:1. The mean strike of the dextral shears is 301° , whilst that of the sinistrals is 042° (Figs 4.23 & 4.24). The presence of these shear bands offers considerable evidence of large-scale orogen-parallel dextral shear across this region, and collectively they appear to represent Reidel and Anti-Reidel shears to an approximately E-W dextral shear zone.

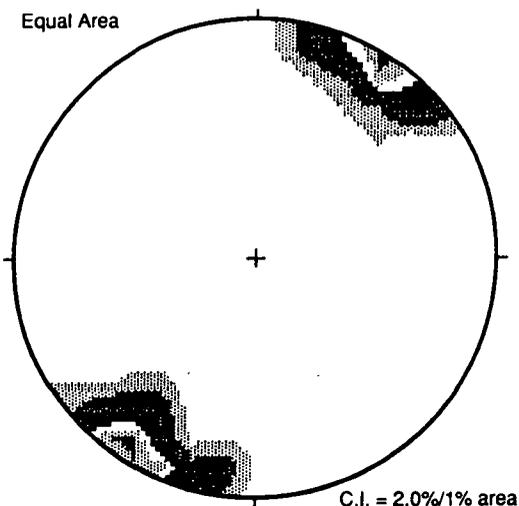
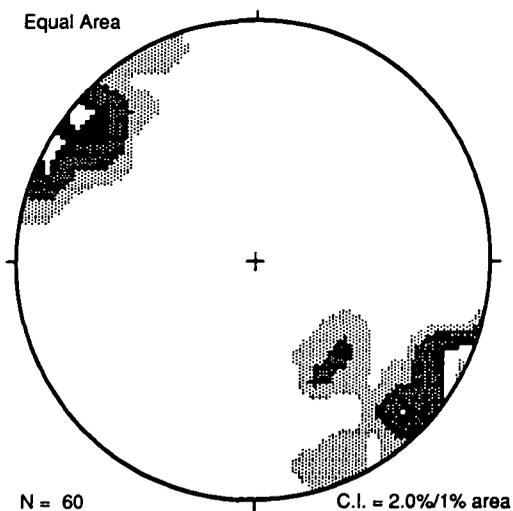


Figure 4.23 (left) Contoured equal area stereogram of 154 poles to dextral shear bands in the Start Complex. The mean pole is $121/84\text{NE}$.

Figure 4.24 (right) Contoured equal area stereogram of 60 poles to sinistral shear bands in the Start Complex. The mean pole is $042/86\text{NW}$.



Assuming the SBF lies sub-parallel to a regional shear plane (see 4.6.4), then the DSB's (Reidels) strike, on average, 31° clockwise from this, whilst the SSB's (Anti-Reidels) strike 48° anticlockwise from it. Naylor *et al.* (1986), show that for a simple shear zone in dry sand (i.e. where σ_1 lies at 45° to the shear zone), the R shears behave as Coulomb-Mohr faults, and thus initiate at $45-\phi/2^\circ$ to the σ_1 direction, where ϕ is the angle of internal friction. For models involving a free surface (i.e. the earth's surface) R will initiate at $17-20^\circ$ to the main shear, and will accommodate much of the synthetic displacement. The R' will initiate at high angles to the main shear zone ($\sim 80^\circ$) and will be generally short lived as they are in a kinematically unfavourable position for taking up large displacements. Naylor *et al.* (1986) also show that under transpressive

regimes the R and R' shears will initiate at higher angles to the main shear zone (see Chapter 1, Fig. 1.4), and thus it would appear that the shear bands observed across the Start Complex are in an orientation consistent with dextral transpression.

If it is assumed that these do represent R and R' shears in an overall dextral transpressive regime, then it is reasonable to expect the shears have undergone both active and passive rotation. Active rotation takes place as a consequence of the relative motion of the shears, this rotation being antithetic in sense to the displacement along the shears (Fig. 4.25a). In passive rotation, the shears act as strictly passive markers within the overall deforming rock mass, and will thus rotate in the direction of the regional shear (Fig. 4.25b). As long as displacement along each shear plane continues, it can be assumed that active rotation would appear to predominate. Whenever the shears lock up to become fixed planar surfaces, passive rotation takes over. In the case of the DSB's, active rotation will tend to flatten the shears towards the E-W shear zone, whilst the passive clockwise rotation, imposed by the overall dextral shear regime, will rotate them away from this position.

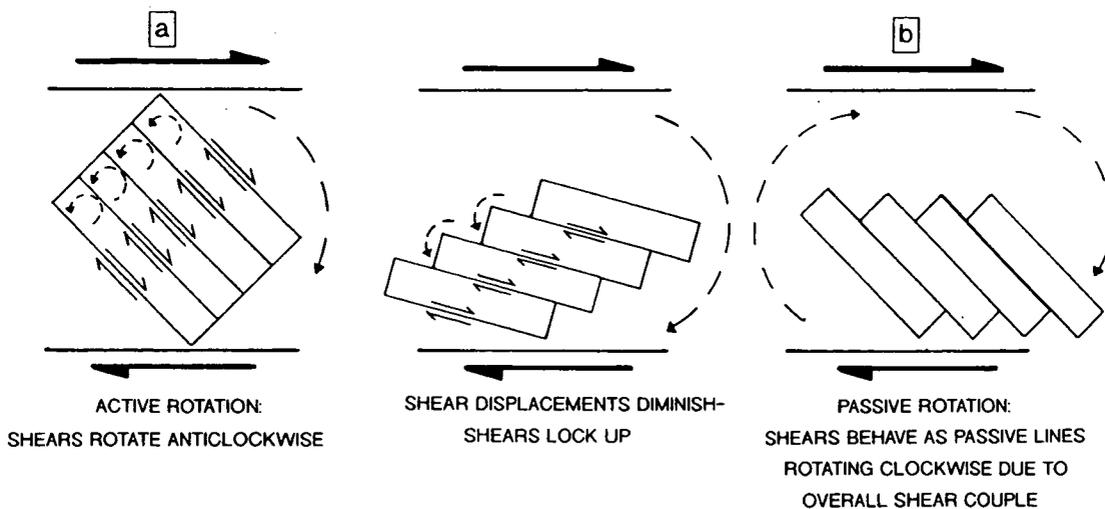


Figure 4.25 Rotation of shear bands during dextral shear: a) active rotation in which the change in orientation of the shears results from relative movement between shears, irrespective of the overall shear couple. i.e. dextral shears will rotate anticlockwise despite the overall clockwise shear. b) passive rotation in which the shears lock-up and rotate as passive line markers within the overall shear regime.

A histogram of the Start DSB's (Fig. 4.26), shows a more or less symmetrical distribution, suggesting that neither of the rotation mechanisms had predominated on a regional scale. Histograms of individual field areas, however, show a variety of distributions from clockwise skew, through symmetrical to anticlockwise skew. One possible explanation for this local variation is lithological control. It was noted, at North Sands for example, that as shears pass from greyschist into the greenschist there is at least an order of magnitude change in both the scale and the offset of the shear

surfaces. It is reasonable to expect that shears in the greenschist localities will have undergone less active rotation than those in the greyschists, if the amount of relative displacement along the shear planes is less in the greenschists relative to the greyschists. A composite histogram of dextral shear bands in greenschist shows a much smaller spread of data compared to one for dextral shear bands in greyschist, whilst the mean shear plane in the latter lies 8° clockwise of that in the former, further suggesting that the shears in the greenschists have undergone less active rotation. In the greyschists, where it is possible to broadly group the shears in brittle or ductile, there appears to be little pattern in terms of shear distribution, i.e. the more brittle shears should lie at higher angles to the shear zone as they have undergone less rotation. Unfortunately, field observations do not bear this out, with brittle and ductile shears intimately occurring in a variety of orientations, probably as a consequence of both the rotation and the preferential reactivation of certain shears.

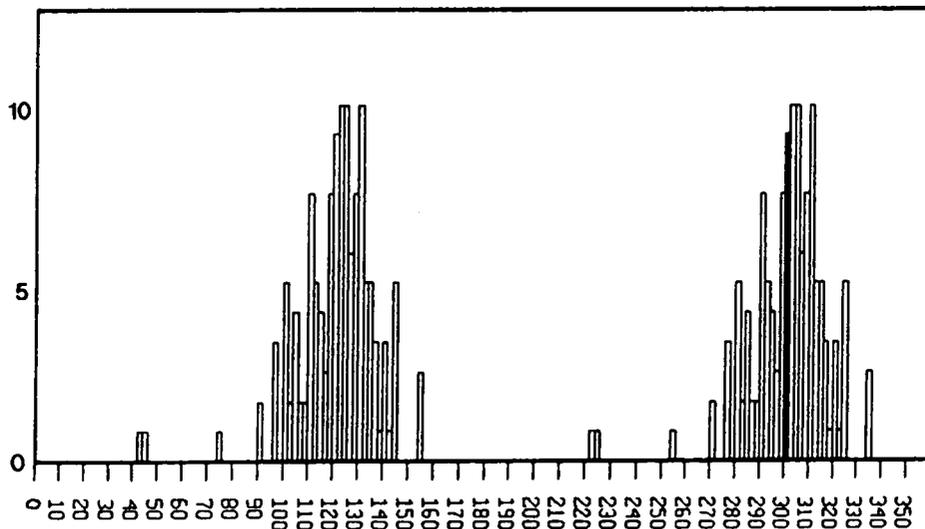


Figure 4.26 Histogram of dextral shear bands within the Start Complex showing a reasonably symmetrical spread of data.

Passive rotation, if it does occur, necessitates further shears/faults to effectively accommodate block rotation on a variety of scales, and whilst direct field evidence for this is not seen, the possibility cannot be ruled out, especially as shears with anomalous orientations do occur in many localities.

In the case of the SSB's, both active and passive rotation will rotate the shears clockwise towards the shear plane. A histogram of Start SSB's (Fig. 4.27) shows a very broad spread of data, implying that R' shears may have rotated considerably away from their initiation position.

The limited data collected in the Devonian shales to the north of the SBF shows a similar pattern to the Start Complex. Here the dextral shear bands strike, on average, 295° , whilst the sinistral shear bands strike 048° . However, little could be

concluded in terms of active/passive rotation or brittle/ductile relationships, as the available data was restricted.

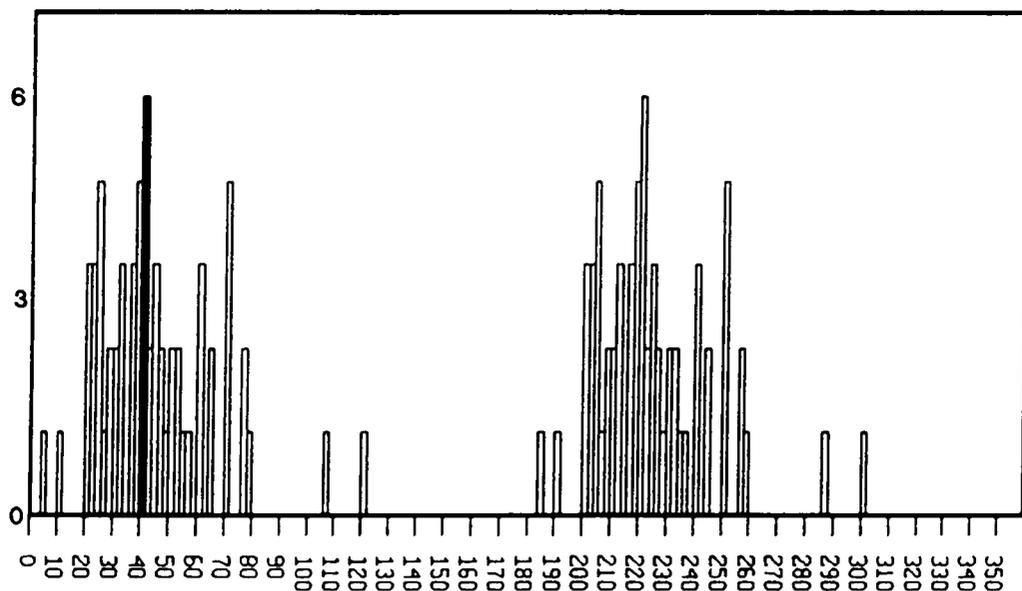


Figure 4.27 Histogram of sinistral shear bands within the Start Complex showing a symmetrical distribution and a very broad spread of data.

The unusual 'bridge' structures noted at Dutch End [SX 2785, 0356; 4.2.4] are included with the shear bands, rather than extensional structures, as their formation clearly involves a significant amount of along strike displacement. In these, the host rock has fragmented off the fracture wall and folded into the dilating vein (see Plate 4.16). The usual mechanism to account for bridge structures (e.g. Nicholson & Pollard, 1985; Nicholson, 1991) assumes fracture propagation followed by dilation, with the resultant bridges being folded due to shortening imposed by the adjacent fracture dilation. Nicholson (1991) suggests that the observed shear displacements are a consequence of dilation alone and not of zone-parallel ductile shear strain (as in the models outlined in Ramsay & Huber, 1987). However, the bridge structures at Dutch End occur as remarkably consistent parallel folds (Plate 4.16), which in places are cut by anastomosing shears to produce what appears to be a sinistral strike-slip duplex. The marked along strike consistency of these bridges, and the lack of cross fractures are difficult to explain using the model of Nicholson (1991). The evidence strongly suggests that zone parallel shear was important during dilation, and indeed these veins parallel fine centimetre scale sinistral shears in the adjacent greenschist. These veins are cut in places by dextral shears, again suggesting that the veins are the result of dilation along pre-existing sinistral shears, rather than an extensional vein set in its own right. Although slightly dilatent, quartz/albite infilled shear bands are quite common across the Start Complex, structures like those described from Dutch End are not seen elsewhere in this region.

Extensional Structures:**i) Joints and Fractures**

These structures are best developed in the Start Complex, possibly because the strong slaty cleavage in the Devonian sequence to the north is too strong an anisotropy to allow such structures to develop. In the schists, the joints are consistently sub-vertical, have a mean dip of 82° towards 082° (Fig. 4.28) and vary from laterally extensive planar joints down to short (5-6cm) lenticular fractures infilled with quartz or albite. The quartz fibres are normal to the fracture wall and are not displaced. A comparison between Figure 4.28, of poles to extensional fractures, and Figure 4.19, of F3 axes, shows the two stereonet to be remarkably similar. The mean pole for the former is 08° towards 262° whilst that of the latter is 12° towards 260° . It would appear that these fractures are the extensional structures which develop parallel to the maximum compressive axis (e.g. Sanderson & Marchini, 1984; see 'V' on Fig. 1.05, Chapter 1) and normal to fold axes.

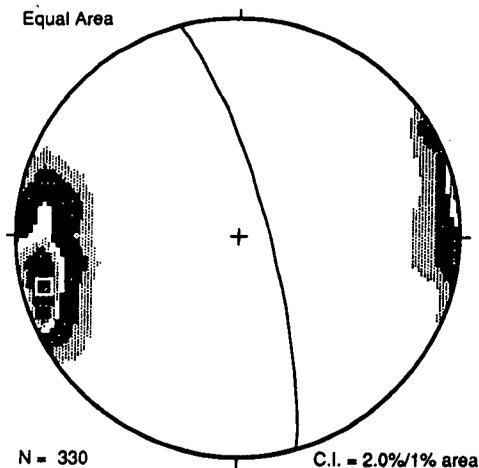


Figure 4.28. Contoured equal area stereogram of 330 poles to extensional fractures across the Start Complex. Mean pole is $172/82E$.

Figure 4.29 summarises the relationships of the shear bands (R & R') to these fractures. The mean strike of the fractures is $172^\circ/352^\circ$, implying that the maximum compression direction was very close to this orientation. Whilst rotation, passive in this case, is important in modifying the orientation of structures in this area, these extensional fractures appear to be very late in the deformation history, cutting all folds, most shear bands and even some late kink bands. Thus it is unlikely that they will have undergone much rotation and will lie close to their initiation position. A histogram of the Start extensional fractures shows a slight clockwise skew and very little spread of data, tending to support the idea that these structures are little rotated. Assuming this to be the case they may now be tied in to the shear bands (Fig. 4.29). The mean dextral shear orientation is 121° whilst that of the sinistral shears is 042° , these being perfectly bisected by the 352° position for the extensional structures ('T' in Fig. 4.29). The fold

axis data showed that F2 axes had undergone $\sim 8\text{-}10^\circ$ of passive clockwise rotation prior to F3 refolding, and this, combined with the active rotation which is especially important in the case of the dextral shears, is more than enough to account for the large obtuse angle ($\sim 100^\circ$) the shears make about T.

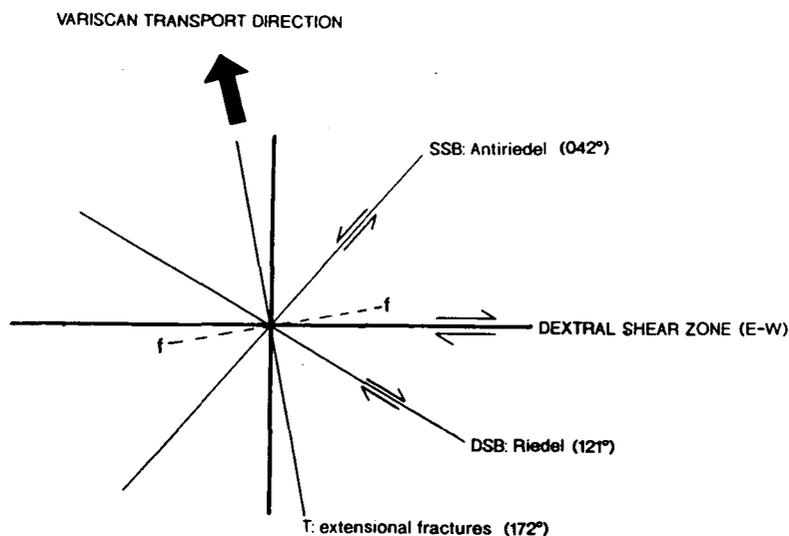


Figure 4.29 Diagram showing the relationship between shear bands (R & R'), extensional fractures (T) and the maximum compressive stress during deformation. Modified after Petit, 1987.

ii) Tension Gashes

In some of the lower strain localities, poorly developed tension gashes occur typically as discrete structures and rarely as en echelon arrays. Such extensional fractures are better developed in the more homogeneous slates to the north of the SBF, although they are common in both schistose lithologies within the Start Complex. These tension gash arrays were of use in applying the geometrical construction of McCoss (1986), in order to establish the zone boundary displacement vectors (\underline{S} ; see Chap. 1.2). The techniques and methodology of this simple construction are included in Appendix 2. In the case of the Start tension gashes, method B (McCoss, 1986) was applied to a number of examples in which the maximum and minimum principal axes of the infinitesimal strain ellipsoid were inferred from the tension gash orientation and recorded along with the orientation of the tension gash zone boundary. Few accurate readings were possible in these schistose lithologies. Constructed values of \underline{S} vary, with two distinct trends emerging: apparently dextral tension gashes recorded a displacement vector towards $\sim 344^\circ$, whilst apparently sinistral tension gashes gave displacement values of $\sim 273^\circ$. These vectors are approximately parallel to the inferred transport direction on a regional scale (Variscan transport) and to the directions of displacement along the zone of \sim E-W shear. Why the tension gash asymmetry so closely correlates with the partitioned displacement vectors is unclear and does not appear to be related to conjugate array geometries. The change in the displacement

vector seems to reflect the inferred strain partitioning and, as the relative ages of these structures cannot be assessed, it is impossible to say whether this is spatial partitioning, e.g. the synchronous distribution of strain into distinct domains. Given the close geographical proximity of some of these structures it is more likely to be a temporal partitioning effect. It is also interesting that these structures record very late increments of strain and even at this late stage of the regional deformation history, the strain appears to be readily partitioning.

4.6.6 Late Faults

The zone of high ductile strains associated with the SBF appears to have profoundly influenced fault orientation, distribution and geometry within the South Hams area. Whilst there is a local control on faulting, e.g. pre-existing anisotropy (see below), an underlying basement influence is very likely, and especially apparent in the zone of deformation straddling the SBF. Here, three distinct fault orientations occur: Sub-vertical E-W trending faults with variable displacements, including oblique-, dip- and strike-slip offsets of both sinistral and dextral sense. Offsetting these are NW-SE (dextral) and NE-SW (sinistral) trending strike-slip faults, which are presumably accommodating considerable dextral shear across the region. The intensification of the faults into the SBF is strong evidence that these structures mark a long-lived zone with a complex kinematic history, including dextral E-W shear intermittent with N-S dip-slip displacements.

Presumably the essentially conjugate sets of strike-slip faults relate to the bulk strain in a similar way to the shear bands, only on a larger scale. For example, along Limpet Cove [SX 2673, 0397; Fig. 4.2a], steep fault gulleys dissect the wave-cut platform and parallel local shear band orientations. The intensification of the sinistral shear bands into the NE/SW trending faults, and of the dextral shear bands into the WNW/ESE trending faults is clear evidence that these large-scale, essentially late brittle features, are accommodating the same displacements as the small ductile shears. Also within the greyschists at Portlemouth, the dextral shear bands are up to ~10's metre spacing with significant offsets. These are both brittle, with associated breccias, and ductile with a localised rotation of the schistose fabric, and there is a continuum of structures from small early ductile shears, through to late brittle faults, all accommodating the strong E-W dextral shear in this area

The faulting across the South Hams region is generally very brittle in nature, with much associated fault gouge, brecciation and mineralisation (e.g. see Plate 4.9). Unfortunately, fault surfaces rarely preserve any slickensides/slickenlines, and a direct assessment of fault kinematics is often difficult. In many cases, the relationships of the small-scale shear bands to the large faults provide the only usable displacement

constraints. Of the faults in this area, the large dextral strike-slip faults, which trend NW-SE, are dominant and, especially in the Start Complex, profoundly influence outcrop patterns (see Figs. 4.4, 4.5). Some of the large sinistral faults in this region may be antithetic to these major NW-SE dextral strike-slip faults, e.g. the faulting observed along the south side of Hope Cove [SX 2673, 0397; Fig. 4.10]. Here, rarely preserved fault fibres confirm that displacements on these antithetic structures are oblique and, as with the oblique slips on the shear bands (see 4.6.5, above), this is necessary to accommodate triaxial transpressive strains. Modelling of the faults to test for transpressive quadrimodal distributions of poles (Underhill & Woodcock, 1987) could not be applied in these cases as displacement fibres were so rarely observed. Also the Underhill and Woodcock (1987) technique assumes strain hardening, with each individual fault strand locking up to record the incremental strain history. Evidence from the Start Complex, e.g. cross-cutting relationships between the steep faults and shallow dipping thrusts and later extensional detachments, suggests that many faults have reactivated and record an intermittent displacement history, probably also involving a degree of later extensional reactivation, thus limiting their use in the above modelling.

It is worth noting that the steep, generally strike-slip faults, only occur where the fabric is upright whilst the shallow dipping thrusts and later extensional detachments only occur in the areas with a gently inclined fabric, i.e. the early fabric orientation has exerted a profound influence on the siting of later structures. The control exerted by the earlier fabric (essentially a pre-existing anisotropy) may be one reason for the partitioning of transpressive strains into domains of strike-slip displacements and domains of dip-slip shortening. In this case, the partitioning is spatial, with narrow zones of steeply inclined fabric preferentially taking up the along strike shear, and separated by broader zones of gently inclined fabric accommodating shortening. With continued shortening, temporal partitioning of the strain may occur, as the early fabric is reorientated into steeper positions, e.g. rotation in fold limbs during shortening, backsteepening by later structures, etc. In such cases, the initial dip-slip displacements give way to progressively more oblique movements, ultimately with strike-slip structures predominating. Whilst there is no direct mesoscale evidence of this phenomenon, it is clear that within the S. Devon area the along-strike simple shear component of the transpressive strain partitions progressively into narrower zones, ultimately localising along the SBF zone. This zone coincides with the site of considerable backsteepening (see 4.5) and tightening of folds, etc. and it is possible that one control on this temporal partitioning is the marked rotation and steepening of the early fabric into a sub-vertical orientation within this area.

4.7 Summary

The Start Complex contains two compositionally different schistose lithologies, which are interleaved and interfolded on a wide variety of scales and which are clearly dominated in outcrop area by the greyschists. These greyschists are typically a variably deformed lineated (L>S) tectonite, record higher strains and are characterised by centimetre scale, open to tight late folds (F3), which are the dominant structure across the Start Complex. The highly lineated nature of the schists is a consequence of a powerful E-W mineral grain alignment, combined with a variably developed intersection fabric and extensive fold axis parallel quartz rodding. The mineral stretching lineation lies in an anomalous strike-parallel orientation, parallel to primary fold axes. The lower syn-D3 strain greenschists better preserve earlier fabrics and morphologically distinctive early folds (F2), which are locally refolded by weakly developed later folds (F3) to result in type III interference patterns (Ramsay, 1967). Even in these outcrops, primary fabrics and fabric relationships have been largely obscured by the extensive later recrystallisation. The large-scale structure of the Start Complex has been variously explained, and is here interpreted as a gently SW plunging, steeply northerly inclined F3 antiform.

The Devonian sequence to the north of the Start Complex is characterised by open north verging and facing primary folds, which vary from gently southerly inclined in the north, steepening to sub-vertical moving southwards, and ultimately dipping steeply north adjacent to the SBF. The gently E-W plunging primary folds also tighten moving southwards and, adjacent to the SBF there are several generations of crenulating cleavage, transposition of the early fabrics into the main cleavage and the reduction of primary folds to intrafolial lenticles.

The anomalous northerly inclination of the main cleavage and the steep northerly dip of the F3 antiform in the Start Complex are accounted for in terms of backfolding against a deep seated basement fault. The SBF is considered to be the surface manifestation of this buttress, and outcrops as a series of north dipping, possibly normal, faults, straddled by a zone of ductilely deformed fault rocks. Large volumes of altered intrusive ^{gne} intimately associated with the fault zone and this, along with the Start schists, has locally been reduced to a mylonite and extensively iron mineralised. Sheath folds are well developed along the SBF zone, their geometries indicating that they are the result of high dextral shear strains.

Evidence for dextral transpression is abundant across the region. The anomalous along-strike orientation of the extensional lineation and its sudden switch to the regional down-dip orientation at Challaborough may be accounted for by a deformation regime dominated initially by pure shear and then by marked dextral

transpression. Further evidence of this change in the regional shear comes from the sheath folds along the SBF. The consistent refold relationships of the early fold parallel lineations and late fold axes implies clockwise rotation of the early structures in a dextral shear environment. Late structures such as arrays of pervasive shear bands, faults and extensional fractures are all consistent with dextral shear across the region, the shear bands being interpreted as Reidel and Antireidel shears respectively to a large E-W dextral shear regime. Stereographic analysis of these indicates Variscan transport towards 350° , across this area, this being high angle oblique to the proposed basement fault (assuming its surface orientation reflects its position at depth). The high angle obliquity could induce shortening, followed by dextral transpression, during Variscan orogenesis, and would account for the observed localisation of strain along the SBF. The proposed along-strike continuation of this backsteepened high strain zone into S. Cornwall, and its large-scale regional significance, are investigated in Chapters 5 & 6.

CHAPTER 5

THE STRUCTURAL EVOLUTION OF THE NORTHERN GRAMSCATHO MARGIN, SOUTH CORNWALL

5.1 Introduction

This chapter presents the results of the detailed structural mapping and observations, carried out on the critical coastal sections which straddle the northern margin of the Gramscatho Basin in south Cornwall (Fig. 5.1). A overview of the stratigraphy and sedimentology of the Devonian Meadfoot shales and Gramscatho sandstones, the only lithologies exposed in the study area, is presented (5.2), and followed by a review of the various igneous intrusives, both pre- and post-orogenic, which outcrop across this region (5.3). The small scale structures of the east coast, from Turbot Point [SX 2028, 0430] to Fowey [SX 2125, 0515] are detailed, and incorporated with the along strike structures recorded on the west coast, from Cligga Head [SW 1738, 0507] to Watergate Bay [SW 184, 065; 5.4 & 5.5]. Finally, the evidence for dextral transpression, and its regional implications, are discussed in 5.6.

Regional deformation chronology

Early structural analyses of S. Cornwall concentrated on relating the large-scale structures within this region to the emplacement of the Lizard ophiolite (e.g. Hendriks, 1959, 1971; Fig. 5.1). More recently, study of the variation in orientation, and the spatial distribution of minor structures led to the establishment of detailed local deformation chronologies (e.g. Smith, 1965; Dearman *et al.* 1980). A comprehensive account of the deformation chronology for S. Cornwall, which at that time was somewhat at variance with the established chronologies (e.g. Barnes *et al.* 1979), was offered by Rattey (1980a), who recorded five distinct deformation phases. This was subsequently revised (Rattey and Sanderson, 1984), whilst more recent revisions of the chronology, and the recognition of local variations, are given by Wilkinson (1990) and Shail (1992). The deformation chronology for S. Cornwall, summarised below, is based on Rattey (1980a,b), Rattey and Sanderson (1982, 1984) and Leveridge *et al.* (1990), with some minor modifications.

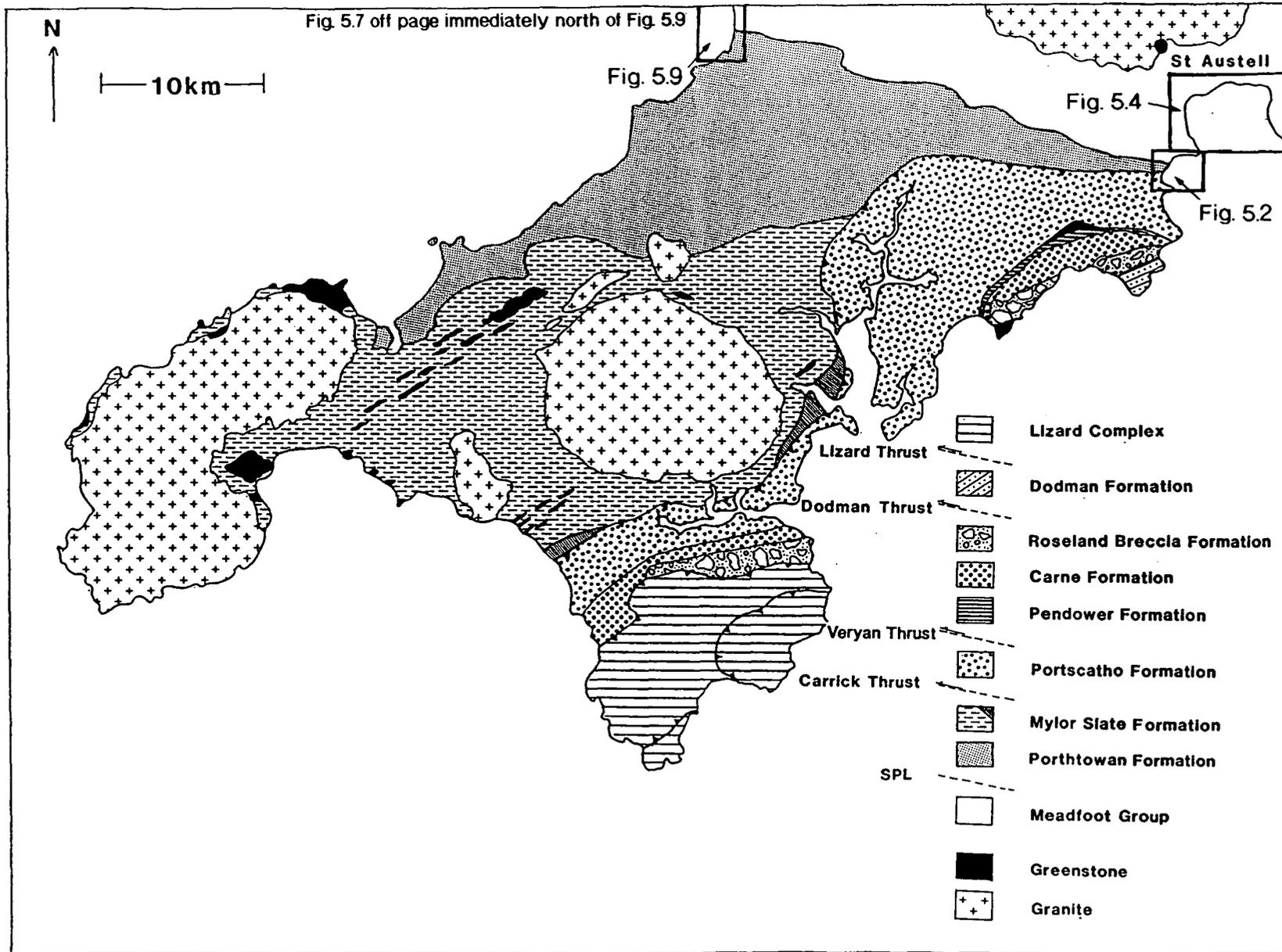


Figure 5.1 Simplified geological map of S. Cornwall, showing the distribution of the major tectonostratigraphic units and igneous intrusives. The east and west coast study areas are highlighted. After Shail (1992).

D1 Deformation: The first phase of deformation results in tight to isoclinal NNW-verging and facing F1 folds, which are sub-horizontal or plunge gently ENE/ WSW. The associated axial surfaces dip gently to the SSE. Fold styles are lithologically controlled, ranging from isoclines with sub-rounded hinges (e.g. in the Mylor Beds) to flattened chevron folds (e.g. in the Gramscatho Beds; Rattey, 1980b). An approximately axial planar regionally developed cleavage is associated with the folds and again there is a lithological control, such that a spaced cleavage is developed in most sandstones, detrital grains showing a flattening in the plane of the cleavage, whilst a slaty cleavage develops in the mudstones (Rattey & Sanderson, 1984). Variable local transection of the short overturned limbs of the folds is quite common, although transection angles are so small that the cleavage is considered to be axial planar (Rattey & Sanderson, 1984). Mineral lineations, where developed, plunge down the dip of the S1 cleavage, and are thought to lie parallel to the direction of thrust transport (NNW). However, along the northern margin of the Gramscatho Basin the bedding, and planar tectonic fabrics, steepen to sub-vertical, E-W striking, with the development of a marked sub-horizontal E-W stretching lineation (Holdsworth, 1989a).

Zones of oblique folding are observed to the north of the Lizard Complex where folds are aligned N-S (Dearman, 1969), and are considered to have developed during differential thrust sheet displacement associated with the emplacement of the ophiolite (Rattey & Sanderson, 1982). Wilkinson (1990) suggested that these folds formed by thrust propagation over lateral ramps. Leveridge *et al* (1990) interpreted the range of fold axial orientations, in the region north of the Lizard Complex, as sheath folds, which developed during emplacement of the Carrick Nappe, implying that D1 deformation was contemporaneous with large scale thrusting. Within the Tintagel High Strain Zone, small, gently plunging early folds with almost every axial trend have been recorded.

D2 Deformation: F1/F2 refolds are generally coaxial, except in areas of earlier oblique folding, where complex interference relationships are developed (e.g. Dearman *et al.* 1980). The second phase of deformation is superimposed on F1, and crenulates the S1 cleavage. It was first recognised between Hayle and Portreath by Smith (1965), and was subsequently described as a regional deformation (Sanderson, 1973; Rattey, 1980b). F2 folds are generally moderately to steeply inclined, possess ENE-WSW trending sub-horizontal axes and verge to the NNW. Axial surfaces dip steeply to gently SSE and are associated with an axial planar crenulation cleavage. Folds usually face upwards NNW except along the southern Cornish coast, where facing is downwards to the SSE on the short overturned limbs of earlier F1 folds (Rattey &

Sanderson, 1984). Orientation of D2 structures is locally variable, with a marked shallowing of the S2 dip towards the south (Rathey & Sanderson, 1984). F2 folds are often developed in the hangingwall to minor thrusts, which usually correspond to zones of high D2 strain, and bedding and S1 cleavage are locally transposed in such zones. Imbrication of the previously deformed sequence by D2 thrusts may have brought about much of the stratigraphic complexity on the west Meneage section, close to the Carrick Thrust (Wilkinson, 1990). Using the change in minor F2 vergence, from SSE to NNW, Rathey and Sanderson (1984) postulated the existence of the Godrevy Antiform, a large scale, upright, open F2 fold. However, Shail (1989), suggested that minor SSE verging folds in this region are F3, and are probably related to a D3 backthrusting event.

D3 Deformation: Rathey (1980b) described an intense, flat-lying 'S3' crenulation cleavage, which is axial planar to recumbent minor folds. These structures are thought to be confined to the roof zone and margins of the Cornubian granite batholith, leading Rathey (1980b) to suggest that they are related to granite emplacement. However, Leveridge *et al.* (1990) assigned a D5 age to these structures (see below) and ascribed minor SSE verging folds, with a well developed approximately axial planar north dipping cleavage, to F3. The intensity of D3 appears to diminish from the north to the south of S. Cornwall, and the geometry of the folds and associated faults led Leveridge *et al.* (1990) to suggest that the D3 event is related to backthrusting. As this is more commonly seen in the north of the Gramscatho region, it could be evidence for a structural buttress, possibly the bounding fault at the northern margin of the Gramscatho basin (Shail, 1992).

D4 Deformation: This has been recognised by Leveridge *et al.* (1990), who describe asymmetrical NW-verging folds. The associated S4 cleavage is generally axial planar to the folds and is either a crenulation cleavage or a fracture cleavage filled with quartz. Recent work (e.g. Shail, 1992) has failed to unequivocally show that this event is distinct from D2.

D5 Deformation: D5 is characterised by small-scale recumbent folds (F5), which refold all earlier structures, and are associated with an axial planar, intense, flat-lying crenulation cleavage (S5). F5 is generally only developed on the steep limbs of earlier structures. These structures appear to be localised around the Cornubian batholith, and they always verge away from the granites; there are also open domes or periclinal structures with axes parallel to the granite ridges (Rathey & Sanderson, 1984). Similar structures have been observed by Shail (1992), well away from the influence of the batholith and

contained within extensional fault zones which are sometimes bedding parallel. Vergence sense is variable and controlled by pre-existing bedding or cleavage orientations. Analogous structures are recorded by Holdsworth (1989b) in the Scottish Caledonides and ascribed to extensional collapse. A similar origin is favoured by Shail (1992), who considers these structures to have developed during late orogenic extension coeval with batholith intrusion.

5.2 Sedimentology and stratigraphy of the northern Gramscatho margin

5.2.1 Introduction

The study area in S. Cornwall straddles the northern margin of the Gramscatho Basin (Fig. 5.1). The only lithologies mapped north of this margin were monotonous Meadfoot shales, containing occasional sandstones and frequent intrusives. Although Dartmouth shales are exposed in the core of a large, open antiform in Watergate Bay [SW 1840, 0650] on the west coast (e.g. Reid & Scrivenor, 1906), this area was only briefly mapped and the lithologies not studied. Immediately to the south of the basin margin, are the deep water flysch sandstones and greywackes of the Gramscatho Group. Those exposed along the east coast, south of Pentewan [SX 2021, 0471], belong to the allochthonous Middle Devonian Portscatho Formation, whilst those exposed on the west coast, south of Carn Haut [SW 1761, 0560], belong to the parautochthonous Middle Devonian Porthtowan Formation (Holder & Leveridge, 1986b; see Fig. 2.5).

Brief lithological descriptions are given in the following sections (5.2.3, 5.2.4), although a detailed mapping of the lithostratigraphic and facies relationships was not carried out, as a comprehensive account of the provenance and facies relationships of the lithologies in this region has recently been completed by Shail (1992).

5.2.2 Previous work

The Gramscatho Basin is the most southerly of the Rhenohercynian basins exposed onshore in S. Britain, and has been the most extensively studied. A summary of the more pertinent advances in the understanding of the stratigraphic and tectonic evolution of this region has been given in section 2.2.2, whilst a recent and very comprehensive review of the literature concerning this basin is given by Shail (1992: Chap 4.2 for further detail). A general review of the literature on the Meadfoot and Dartmouth Groups is given in 2.2.2.

5.2.3 The Gramscatho Beds

i) The East Coast Exposures

The exact course of the Carrick Thrust across S. Cornwall is uncertain (see 5.4.3 below), and its position on the east coast has not been accurately described. Leveridge *et al.* (1984) originally placed it along the Gramscatho/Meadfoot boundary, implying that only the Portscatho Formation, the lowest formation within the Carrick Nappe, is exposed along the east coast. However, this model was subsequently revised (Leveridge & Holder, 1986b; Fig. 5.1), and a narrow zone of the parautochthonous Porthtowan Formation placed along the northern Gramscatho margin (the Treworgans Sandstone Member; see below). These Gramscatho lithologies were mapped north from Turbot Point [SX 2028, 0430] to Gamas Point [SX 2023, 0472], at the north end of Pentewan Beach, where the interbedded Meadfoot shales begin to dominate. These interbedded Gramscatho sandstones and shales strike ENE-WSW and dip steeply to the SSE.

Description

At Turbot Point, grey shales dominate and are interbanded with buff slates and occasional thick (<60-70cm) sandstone units (see Plate 5.6). The sandstones are pale greenish-grey and many exhibit a strong carious weathering. Moving northwards, these units remain subordinate, but vary in both thickness (down to 5-6cm) and frequency. The interbedded dominant shales become noticeably darker, such that at Portmellon [SX 2016, 0440] thin black shales and slates are common. Examples of soft sediment deformation are widespread, with 'flames' of black mudstone projecting into thick pale green sandstone layers. The sequence of black shales and interbedded thin sandstones continues northwards to Mevagissey Harbour [SX 2018, 0449], with limited graded bedding indicating a right way-up, southward younging sequence here. The marked increase in strain along this section (see 5.4.3) profoundly affects the lithologies, so that north of Mevagissey Harbour, highly deformed sequences of black shales with thin pale sandy lenticles, are all intensely quartz veined and tectonically disrupted. Northwards towards Polstreath [SX 2018, 0454], thick sandstones reappear, increasing from a few centimetres in thickness up to 1-2 metres. Rare grading again indicates a right way-up, southward younging sequence. Dark shales continue to dominate around Portgiskey [SX 2019, 0463], where they are interleaved with greenish-grey sandstones of variable thickness, and occasional thin green mudstones and coarse grit horizons up to 10 centimetres thick. Here, grading is variable, indicating both right way-up and inverted sequences, although a generally right way-up sequence is indicated by load structures and pebble horizons on the bases of some of the larger units [e.g. SX 2019, 0453]. Occasional thin, impersistent black chert

horizons occur along this section, and have been correlated with similar cherts at Perranporth (Reid & Scrivenor, 1906; Reid, 1907), although recent thrust tectonic reinterpretations (e.g. Leveridge *et al.* 1984) would render this correlation untenable.

A large exposure break occurs across Pentewan Beach, and on the north side the transition from Gramscatho sandstones into Meadfoot shales is thought to be preserved. The first outcrops immediately to the north of Pentewan harbour [SX 2021, 0471], are very thickly bedded, predominantly structureless, gritty sandstones, interlayered with thin horizons of greenish-grey banded silts and shales, and are assigned to the Treworgans Sandstone member of the Porthtowan Formation by Shail (1992). The coarse grit units range in thickness from 15-30 metres, dominating the cliff exposures. Fine conglomeratic bases and occasional discrete pebble horizons, allied to fining-up cycles, imply a right way-up sequence along this section. The basal conglomerates contain subrounded pebbles of quartz/feldspar, fragments of black shales and grains of a green/grey schistose rock. Large (metre scale) rip-up clasts, some of which are inverted (see Plate 5.11), attest to the effects of turbidity currents. Bedding, where obvious, occurs as vague plane parallel units. The sandstones exhibit a marked carious weathering and soft sediment 'flames' of black mudrocks are frequent (Plate 5.1). The transition from the coarse sandstone dominated Gramscatho Group to the pale mudstone dominated Meadfoot Group occurs in the cliffs at Gamas Point [SX 2023, 0472]. Here a WNW-ESE trending and NNE dipping (108/40NNE) fault coincides with the transitional boundary, but, contrary to the statement made by Holder & Leveridge (1986b), does not separate the lithologies. Plate 5.2 is an almost along-strike view of this fault and it can be seen that thickly bedded sandstones are overlain by quartz veined grey Meadfoot shales, and that their boundary is simply downfaulted to the north. Within the hangingwall, the Meadfoot shales clearly conformably overlie the thick, gritty sandstones, the transition occurring over ~100 metres.

Discussion

The Portscatho Formation is by far the thickest of the allochthonous Gramscatho units, comprising some 4.5 kilometres of alternating sandstone turbidites and dark grey slates, within a total stratigraphic thickness for the Gramscatho Group of 6.7 kilometres. The thickly bedded turbidite sandstones, displaying normal and multiple grading and parallel bedding, have been interpreted as Bouma sequences (Holder & Leveridge, 1986b). Sandstone petrographic analysis of the Porthscatho sandstones from this coastal section shows them to have a dissected continental magmatic arc provenance, with accumulation probably in a fore-arc basin (Floyd & Leveridge, 1987). Shail (1992) interpreted the Portscatho Formation as the result of sedimentation

in an entirely deep marine environment, with no evidence to support claims of widespread high energy wave action (*cf.* LeGall *et al.* 1985). Holder and Leveridge (1986b) suggested deposition in a deep water fan environment, with a progradation from outer- to mid-fan regimes, although Shail (1992) noted that the lack of palaeocurrent data makes the distinguishing of fan models difficult. The thick sandstones exposed north of Pentewan are the Treworgans Sandstone, the lowest member of the parautochthonous Porthtowan Formation (see 5.4.3), and may represent deposition in channels at the shelf margin (Shail, 1992).



Plate 5.1 Soft sediment deformation features in interbedded buff sandstones and black shales. Note the ripped up mudstone clasts and flame structures in the structureless sandstones. Gramscatho Group, north Pentewan [SX 2022, 0471].



Plate 5.2 Late brittle normal fault at Gamas Point [SX 2023, 0472]. This structure arbitrarily marks the boundary between the Meadfoot and Gramscatho Groups, but does not separate the lithologies, as thickly bedded Gramscatho sandstones and quartz veined grey Meadfoot shales occur in both the hanging- and footwalls. Viewed looking along strike (WNW).

ii) The West Coast Exposures

All of the Gramscatho Group in this region lies in the northern parautochthonous zone, beneath the Carrick Thrust, and is considered to belong to the Porthtowan Formation (Holder & Leveridge, 1986b). These lithologies were mapped along the western Cornish coast from their most northerly outcrop at Carn Haut [SW 1761, 0560], on Perran Beach, to Pen a Grader [SW 1733, 0528], south of the Cligga Head granite [SW 1737, 0537; Fig. 5.1, 5.9]. The effects of contact metamorphism, mineralisation and hydrothermal alteration are especially severe along the cliff sections south of Perran Bay (e.g. Hanover Cove, [SW 1737, 0532]), and this, allied to the inaccessibility of much of the exposure here, limited the study of these lithologies.

Description

The exposures at Carn Haut comprise pale green shales and green/grey silts interbedded with coarse, buff sandstone units, up to 1 metre thick. The sandstones thin southwards, remaining as thin (<10cm) impersistent horizons within the flaggy green shales. At the south end of Perran Bay, the shales are spotted with diffuse cordierite patches, which grow mimetically along bedding/cleavage intersections and mark the northern limit of the Cligga granite aureole [e.g. Cotty's Point; SW 1757, 0551]. South from here, the lithologies are highly altered and poorly preserved. They are dominated by pale yellow/green shales and silts, interbanded with thin dark pelites. At Pen a Grader, buff sandstones and blocky green shales dominate, the sandstones occasionally displaying grit bases and fining-upwards cycles. In the cliffs south of here are SE dipping thick sandstone units, which appear to conformably overlie dark shales and silts, although some of the thick 'sandstone' units are, in fact, granitic sills, fed by thin cross-cutting dykes.

Discussion

The sandstones noted at Carn Haut suggest, along with the green coloration of the interbedded shales, that these outcrops have greater affinities to the Gramscatho succession further south, rather than to the Meadfoot shales to the north, since sandstones of such exceptional thickness are never seen in the typically monotonous grey Meadfoot shales. The boundary between the base of the Gramscatho Group ('Perran Shales' of Reid & Scrivenor, 1906) and the overlying Meadfoot Group must lie beneath Penhale Sands to the north. Placing the boundary here follows the established convention of Reid and Scrivenor (1906), and refutes the recent alternative interpretation of Holder and Leveridge (1986b), who place the boundary further south, in the cliffs at Pen a Grader. In this locality, thick sandstones, interpreted by Holder and Leveridge (1986b) as the first exposures of the Gramscatho Group, overlie muddy

shales. This represents a relationship opposite to that seen at Gamas Point, on the east coast, and implies that the Gramscatho Group is younger than the Meadfoot Group.

Shail (1992) studied the Porthtowan Formation from Carn Haut to St. Ives Bay, and noted that, although it is largely mudstone dominated, sandstones become predominant in the uppermost parts of this 2500 metre thick, SE dipping and upwards younging succession. The 'Perran Shales', near the base of the sequence, probably represent mudstone dominated slope deposits (Shail, 1992), whilst the sandstones higher in the sequence probably represent channel sands within submarine fans (Holder & Leveridge, 1986b). Shail (1992) noted that the age of the transition from shelf to slope deposition is unclear, due to the lack of local palaeontological control, but by analogy with Liskeard it may be mid-Emsian (e.g. Burton & Tanner, 1986) or by analogy with S. Devon it could be Late Emsian (e.g. Evans, 1985).

5.2.4 The Meadfoot Beds

The Meadfoot Beds have been extensively studied around their type locality, and adjacent areas in S. Devon (e.g. Dineley, 1966; Richter, 1967; Seago & Chapman, 1988; Smith & Humphries, 1991), but little work has centred on the broad expanse of this lithology identified by the Geological Survey (Ussher *et al.* 1909) in S. Cornwall. Evans (1981) studied brachiopod assemblages from the Meadfoot shales around Fowey, whilst Burton and Tanner (1986) mapped isolated outcrops of the Staddon Grit Formation (Upper Meadfoot Group, see section 2.2.2), west of Liskeard. On the east coast, there has been no detailed palaeontological or stratigraphic study of the St. Austell Bay lithologies, apart from a brief study of key localities by Shail (1992).

i) The East Coast Exposures

The Meadfoot shales were mapped northwards from the conformable boundary with the Gramscatho Group at Gamas Point, across St. Austell Bay to Coombe Hawn [SX 2115, 0506], some 2 kilometres east of Gribbin Head [SX 2098, 0494; Fig. 5.1, 5.4]. The shales across the bay, which appear to be largely unaffected by thermal metamorphism associated with intrusion of the St. Austell granite, display variable strikes and dips. Along the west coast of St. Austell Bay, bedding dips gently to moderately to the north, and strikes E-W, whilst along the east coast, at Gribbin Head, it strikes ENE-WSW and dips moderately SSE.

Description

Northwards from Gamas Point, a monotonous grey mudstone dominated succession occurs, in which sandstones are reduced to thin (<0.8cm) discontinuous lenses. Sandstones are generally rare or absent in the succession as far north as Carlyon Bay [SX 2060, 0543]. Younging is difficult to ascertain, although asymmetric refraction of the primary cleavage suggests a right way-up, northwards younging sequence. Thin black interbedded mudstones are common and are often associated with dense aggregates of bright gold^{coloured} euhedral pyrites, e.g. at Gerrans Point, [SX 2040, 0487]. In the higher strain areas, the shales take on a black lustrous appearance and are intensely quartz veined, the veins often being sheared out to leave asymmetric quartz 'augen'. The significance and possible lateral continuity of these 'quartz-eyed slates' is discussed in 5.6.1. A narrow 'aureole' of hornfelsed slates surrounds the Black Head intrusive [SX 2040, 0480].

Thin grey interbedded bioclastic limestones are quite common within the mudstone dominated sequences north of Gerrans Point, and are characterised by marked carious weathering and disarticulated assemblages of brachiopods and sporadic solitary coral fauna. The corals in particular are markedly flattened and distorted within the main cleavage, although they rarely preserve a clear extension direction (see 5.4.4). A sequence of coarse-grained medium bedded sandstones, interbedded with pale grey/green shales, dominates the succession around Fishing Point [SX 2067, 0521; Plate 5.3]. Graded bedding and asymmetric cleavage refraction indicate a right way-up, northwards younging sequence in this region, although locally, younging evidence may be quite ambiguous. Medium bedded mudstone rich volcanoclastic debris flows, containing clasts of basalt up to 30 centimetres across, have been reported in this coastal section (Shail, 1992; Shail, *pers. comm.*), but were obscured by shingle at the time of mapping.

The Meadfoot beds exposed along the east side of St. Austell bay form a monotonous sequence of pale grey shales with thin black interbedded mudstones. A small ?faulted inlier of Dartmouth shales has been mapped in the cliffs at Polmear [SX 2090, 0535; Ussher *et al.* 1909] but this area is now extensively overgrown. At Gribben Head, gently folded thin pale sandstones are quite common, often displaying coarse gritty bases and fining upwards (see Plate 5.15).

A consistent feature throughout these shales is the presence of pale green/grey, very fine grained basic intrusives, very similar to those observed at Torcross, S. Devon (see Plate 5.12, and sections 3.4.1 and 5.3.2)



Plate 5.3 Rare coarse-grained, medium-bedded sandstones in the Meadfoot beds folded into reclined north verging but south facing primary folds. Fishing Point [SX 2067, 0521], viewed looking northwest.

Discussion

The Meadfoot shales mapped across St. Austell Bay are lithologically similar to those of the type area in S. Devon and are likely to therefore have been deposited in a mixed siliciclastic/carbonate shallow marine environment, characterised by periodic storm activity (Pound, 1983). The ubiquitous thin intrusives attest to contemporaneous igneous activity, and, together with the rare volcanoclastic debris flows, suggest that active rifting had begun (Durrance, 1985; Shail, 1992). The thick massive sandstones at Gamas Point (Treworgans Sandstone Member of the Porthtowan Formation) may represent shelf margin channel sands, suggesting that the shelf-slope transition approximately coincides with the Perranporth-Pentewan Line (Shail, 1992). The age of these shales is poorly constrained: Evans (1981) recorded late Seigenian brachiopod assemblages from the shales at Fowey, which are probably quite low in the succession, whilst Burton and Tanner (1986) suggested that the Staddon Grits west of Liskeard are mid-Emsian in age.

ii) The West Coast Exposures

The Meadfoot shales were mapped northwards along the west coast from Gravelhill Mine [SW 1764, 0575], at the north end of Perran Sands, to Towan Head [SW 1798, 0630], west of Newquay [SW 1820, 0620, Fig. 5.1, 5.7]. Only brief mapping was carried out from Towan Head to Stem Point [SW 1841, 0663], at the north end of Watergate Bay, in order to assess the regional structure. The bedding and primary cleavage dip moderately to steeply to the SSW across most of this section, but are subject to local variations.

Description

Much of the outcrop north of Perran Bay is highly deformed, with zones of medium to coarse-grained protophyllonites, and locally black phyllonites, interleaved with variably deformed Devonian metasediments. The assessment of younging is consequently difficult. The sequence as a whole is dominated by grey and grey/green mudstones with subordinate thin bedded quartzose sandstones.

The first outcrops of Meadfoot shales occur ~0.5 kilometres north of Carn Haut, in weathered rubbly exposures either side of a large felsite dyke which projects from the dune sands [SW 1762, 0566]. These are rather pale quartz rich pelites, which are clearly altered by the adjacent intrusive. At Gravelhill Mine [SW 1764, 0575], the Meadfoot Beds crop out as a sequence of curiously weathered and intensely quartz veined, stripey pelitic protophyllonites, the striping defined by mica-rich and mica-poor domains. Moving northwards to Penhale Point [SW 1757, 0593], there is a noticeable increase in strain (see section 5.5.4). Narrow zones of black phyllonite are interleaved with striped quartzose protophyllonites and highly sheared discontinuous intrusives, some of which can be mistaken for sandstones. The protophyllonites along here bear a strong resemblance to the greyschists in the Start Complex, the only obvious difference being the greater proportion of interbedded thin psammities, especially in the region around Ligger Point [SW 1756, 0581]. Younging is impossible to assess in most regions, even in the coarse psammities. Approaching Holywell Beach [SW 1765, 0595], the protophyllonites are reduced to 'rafts' in the dark phyllonite which dominates these cliffs, from Penhale Point [SW 1756, 0592] to north of Hoblyn's Cove [SW 1761, 0583]. Crossing Holywell Bay, the strain decreases rapidly, and pale green/grey weakly striped shales are dominant; there is no evidence of higher strain lithologies. From this point northwards to Crantock [SW 1780, 0610], a monotonous grey mudstone dominated sequence occurs, with interbedded thin buff sandstones and green/grey silt partings. Younging is generally difficult to assess along this section, and, whilst sporadic asymmetric refraction of the primary cleavage indicates a right-way-up, southwards younging sequence, this is far from unequivocal. However, the peninsula at Pentire Point is characterised by metre thick coarse sandstones, in which graded bedding and cleavage refraction clearly indicate a right way-up, southwards younging sequence. Grading in similar, curiously weathered medium grained sandstones along Fistral Bay [SW 1795, 0616] concur with this younging. There is abundant evidence of soft sediment deformation in the interbedded sandstones, black mudstones and thin intrusives in this region, especially at Porth Joke [SW 1770, 0606] and north Crantock [SW 1792, 0618].

The Meadfoot Beds at Towan Head are dominated by medium-grained green/brown sandstones, interlayered with fine grey and black mudstones. The

sandstones appear to be calcareous, although fossiliferous material is largely absent. Younging is uncertain, even in the coarse sandstone horizons, although sporadic, subtle grading does suggest a right way-up, southwards younging sequence.

Discussion

Despite the high degree of deformation displayed by the Meadfoot shales here, they are very similar to those already described from parts of S. Devon and Cornwall, and are thus similarly interpreted. However, their age has yet to be accurately palaeontologically constrained, and it is possible that the Meadfoot Beds are diachronous across this region (Shail, 1992). Recent palynomorph dating from the Dartmouth shales, exposed to the north of the study area across Watergate Bay (Davis, 1990), suggests a late Gedinnian-early Seigenian age, and thus the Meadfoot shales in the north of the study area must be at least mid-Seigenian in age.

5.2.5 Summary

The lithologies exposed along the northern margin of the Gramscatho basin display a transition from a monotonous grey/green mudstone dominated sequence in the north (Meadfoot Group), to a sandstone dominated shelf-slope sequence in the south (Gramscatho Group). The Meadfoot shales represent deposition in a mixed siliciclastic/carbonate shallow marine environment, and are probably Late Seigenian to mid/Late Emsian in age. Frequent thin basic intrusives are evidence of contemporaneous rift related volcanicity. The Gramscatho sandstones exposed along the east coast (Portscatho Formation) represent shelf margin channel sands transitional to a deep water marine fan environment, whilst those exposed on the west coast (Porthtowan Formation) represent slope deposits. The age of these basin margin lithologies is uncertain, although they may be mid- to Late Emsian.

5.3 Petrology of the igneous rocks

5.3.1 Introduction

The igneous rocks of S. Cornwall may be broadly divided into those emplaced prior to Variscan orogenesis, and those emplaced post-orogenically. The former group includes a range of volcanics, the Lizard ophiolite and its melange (see 2.2.2.). In the latter stages of the Variscan orogeny, the Cornubian granite batholith was emplaced. This is composed of multiple intrusions, in which highly fractionated acid members are subordinate to predominant two mica calc-alkaline granites. Shortly after granite

emplacement and regional uplift, late Carboniferous to early Permian post-orogenic volcanic episodes occurred, comprising both suprabatholithic acid volcanism, fed by late granite porphyry dykes (elvans), and mafic intrusives and extrusives, including various lamprophyres (Floyd *et al.* 1993b).

5.3.2 Pre-orogenic intrusives

All of the Devonian and Carboniferous volcanics in S. Cornwall appear to be submarine in origin, ranging from relatively shallow (reef/platform), through to deep water (basin and basin slope). Although dominated by basaltic material, early volcanics and intrusives represent a bimodal basic-acid suite of lavas, high level intrusives and volcanoclastics, often in close stratigraphic association. They have subsequently been extensively deformed, metamorphosed and altered, and rarely preserve their original mineralogy (Floyd *et al.* 1993b). No volcanics were noted in the short mapped sections of the Gramscatho sandstones, although such igneous bodies have been described in this formation further to the south (e.g. Floyd, 1982)

North of the Gramscatho margin, the Meadfoot shales are cut by frequent pale green/grey igneous bodies, which are very similar, both in field appearance and thin section, to the 'Torcross-type' intrusives noted in S. Devon (see 3.4.1). These fine grained intrusives form sheet like bodies that range in size from a few centimetres up to several metres thick. They are typically concordant with bedding (see Plate 5.12), and many of the larger examples display evidence of having been emplaced into soft, wet sediment, with pseudo-flame structures and fine interlamination. There is, however, no evidence for thorough mixing between the magma and sediment (e.g. peperite), and, whilst the adjacent sediments may be baked, there is no evidence for further alteration. Subaqueous pyroclastic flows have been described elsewhere in S. Cornwall (Floyd *et al.* 1993b), but could not be proven to occur in the regions studied. In places, (e.g. north Holywell), there is evidence of multiple intrusion of these sheets, with cross-cutting relationships and xenoliths of one dyke dispersed within an adjacent intrusive, both bodies preserving chilled margins. Later regional tectonism produces an intense foliation in the intrusives (see Plate 5.14), and the concomitant alteration has reduced many of these bodies to almost pure carbonate. The presence of xenoliths, pyrite fragments, feldspar phenocrysts and chlorite aggregates, which are flattened and elongated within the main cleavage, allows an accurate assessment of finite stretching directions (see 5.4, 5.5; see Plate 5.14).

The petrology of these examples is identical to that of the S. Devon bodies (see 3.4.1) and will not be repeated here. Geochemical analyses (XRF) were not carried out, as the highly altered condition of these units renders such techniques largely

invalid (see 3.4.2). Less altered basic intrusives across Cornwall have been analysed, and it can be shown that those north of the Start-Perranporth Line are generally alkali basalts, with low Zr/Nb ratios (Floyd *et al.* 1993b).

A second type of pre-orogenic intrusives, which are common across S. Cornwall, are the 'greenstones', a suite of basic to ultrabasic rocks. Three rather different greenstone bodies occur across the St. Austell Bay area (Fig. 5.1); the Black Head dolerite [SX 2035, 0480; Fig. 5.2], the 'Duporth Picrite' [SX 2036, 0512; Fig. 5.4], and a crudely foliated, almost schistose greenstone at Porthpean [SX 2032, 0502; Fig. 5.4]. None of these intrusives has been extensively studied, and each is only briefly described by Flett (in Ussher *et al.* 1909). Both the 'Duporth Picrite' and the outcrop at Porthpean strongly resemble the more massive and tuffaceous greenschists of the Start Complex, although the similarity in fresh hand specimen is less striking, the former being considerably coarser and less well foliated. Thin sections show that the picrite is now largely composed of serpentine with talc, chlorite, fine grained carbonate and occasional patches of white mica, although Flett (in Ussher *et al.* 1909) describes a second type of picrite containing unaltered augite. The Porthpean exposure is equally altered, the crude foliation being defined by aligned serpentine and talc, within a fine grained chloritic matrix. Occasional small subhedral albites impart a coarse grained appearance to this greenstone. The Black Head dolerite is included with the greenstones by Ussher *et al.* (1909), although it is a weakly foliated quartz diorite. Despite its fresh appearance in hand specimen, it is again much altered with large strained quartz phenocrysts set in a fine matrix of white mica and carbonate. Some of the quartz is fringed with fine grained polygonal aggregates of dynamically recrystallised grains. Chlorite is very common, frequently occurring with strain shadows around the obdurate quartz. Randomly orientated lath shaped opaques overprint the weak foliation.

5.3.3 Post-orogenic intrusives

For the purposes of S. Cornwall, this group includes the granites of the main Cornubian batholith, the associated late stage dykes, and the post-emplacment volcanics, of which only the lamprophyres occur in this area.

Cligga Head Granite and elvans

This is a small elliptical stock, exposed on the north Cornish coast between Perranporth and St. Agnes, and arises from a northerly projection of the Cornubian batholith (Fig. 5.9; Floyd *et al.* 1993b). The stock and adjacent metasediments are exposed in a large west facing cliff, where it can be seen that the northern contact is

intrusive whilst the southern contact appears to be faulted (Plate 5.4; Moore & Jackson, 1977). Cligga Head has been extensively studied, as it is a classic site for greisenisation, mineralisation and kaolinisation (see Floyd *et al.* 1993b for references). The aureole extends as far north as Perran Sands, as evidenced by cordierite spots of the 'Perran Shales' (see 5.5.4.), but granite emplacement has not significantly affected the structure of the adjacent metasediments (see 5.5.4). The petrography, structure and attendant mineralisation of this granite are fully described in Floyd *et al.* (1993b).



Plate 5.4 Series of steep faults coinciding with the southern contact of the Cligga Head granite [SW 1738, 0507], viewed looking east.

Associated with the granites are the 'elvans', thick granite porphyry dykes, which generally trend E-W, sub-parallel to regional jointing (Floyd *et al.* 1993b). One such example occurs in Perran Sands, although it is very poorly exposed and highly weathered (Fig. 5.9). On the east coast at Pentewan, a single NW-SE trending elvan occurs, which crops out on the beach at Polrudden [SX 2025, 0475; Fig. 5.2]. This granite porphyry, formerly quarried as 'Pentuan Stone' (Reid, 1907), also forms a 2-4 metre thick sill, cutting through the black shale cliffs north of the beach (Plate 5.5). In thin section it is composed of large phenocrysts of quartz, sporadic highly altered K-feldspars, and occasional white micas, set in a fine grained groundmass of similar composition. Small knots of chlorite probably represent altered biotites.



Plate 5.5 A 2-4 metre thick granite porphyry sill cutting through Meadfoot shales at Polrudden [SX 2025, 0475]. This 'Pentuan Stone' has been extensively mined further inland. Black shales here are highly deformed, the gently north dipping foliation intensely crenulated by north verging F2 folds.

Lamprophyres

During the late Carboniferous and early Permian, small volumes of volcanics were erupted, and include the lamprophyres, the Exeter Volcanic Series and rhyolites (Floyd *et al.* 1993b). Of these, only the lamprophyres occur within the study area, occurring at Holywell Bay and Towan Head. These lamprophyres are emplaced along the main foliation and are of the minette type, i.e. characterised by K-feldspar and biotite (Hawkes, 1981). An age of 291Ma (whole rock K-Ar) has been obtained by Hawkes (1981) for some of these intrusives, and this may be representative of the suite as a whole (Floyd *et al.* 1993b).

The lamprophyre at Holywell Bay [SW 1767, 0598; Fig. 5.7] varies from 2-3 metres in thickness, and forms a SE dipping sheet, sub-parallel to the main cleavage. It is a dark mottled grey/green colour, appears quite coarse grained and contains sporadic amygdales infilled with a white, weathered mineral (probably zeolite). In thin section, it comprises randomly orientated elongate laths of dark biotites, set in a dense biotite-K feldspar-chlorite-iron ore matrix. The larger iron ores produce diffuse brown haloes within the biotites. The presence of considerable chlorite attests to the highly altered state of this lamprophyre, a feature common to all such intrusives within this area. The Towan Head lamprophyres [SW 1800, 0628], some of which are up to 4 metres thick, display evidence of multiple intrusion, with several thin, foliation parallel, sheet like bodies chilled against one another and the adjacent shales. Narrow foliated 'rafts' of altered shale may be seen between individual sheets.

5.3.4 Summary

The igneous rocks exposed along the northern margin of the Gramscatho Basin range from pre- to post-orogenic. The pre-orogenic intrusives are submarine, predominantly basaltic and all highly altered, and may be broadly divided into the 'Torcross' types and the 'greenstones'. The 'Torcross' type intrusives are well foliated and now largely composed of carbonate, although they are important in that they frequently contain deformed markers. The 'greenstones' include a suite of basic to ultrabasic weakly foliated and coarse grained intrusives, which are now largely composed of serpentine, talc and chlorite.

The post-orogenic intrusives include the granites, their associated elvans, and the lamprophyre dykes. The single, much studied, granite stock at Cligga Head has not significantly affected the adjacent metasediments, its metamorphic aureole only just extending as far as Perran Sands. The thick elvans, rare along this coastal section, have altered granite compositions, and a consistent regional orientation, possibly reflecting basement fault control (see 5.6.1). The minette lamprophyres are clearly post-orogenic, as they are emplaced along the foliation and yield Westphalian K-Ar ages, although even these late stage intrusives are somewhat altered.

5.4 The structure of the eastern coastal sections

5.4.1 Introduction

The limits of the east coast study area are given in 5.2, and for the purposes of detailing structural characteristics, it is divided into the Mevagissey Bay and St. Austell Bay areas, the Black Head promontory forming a convenient division between the two regions (see Fig. 5.1). Access to the coastline is generally good, with excellent exposures both along and across strike.

Within the southern Mevagissey Bay area, the interbedded sandstones and shales display a transition moving northwards from primary (D1) dominated deformation to secondary (D2) dominated deformation, on approaching Mevagissey Harbour [SX 2017, 0448]. Northwards from here, the outcrops are characterised by narrow zones of high D2 strain which diminish in intensity on approaching Black Head [SX 2040, 0480]. The primary structure of this area, which is broadly representative of the regional pattern, is detailed first and then contrasted with the strong secondary overprint.

The northern St. Austell area is characterised by a monotonous sequence of low strain shales and slates, which are only locally affected by later D2 deformation, but which do display a significant late disruption due to extensional detachments and faults.

5.4.2 Previous work

There has been no detailed structural study within this area, since the publication of the survey memoirs (Reid, 1907; Ussher *et al.* 1909), although recently, attempts have been made to resolve the complex sequences within the Roseland area further to the south. Here, the supposed Ordovician "crush-breccias" (Reid, 1907) have been variously interpreted. Hendriks (1937) suggested that these breccias reflect regional thrusting, whilst Sadler (1973) considered them to be a tectonically imbricated condensed stratigraphic sequences. Barnes *et al.* (1979), however, reinterpreted the Roseland breccias as olistostromes, deposited ahead of, and deformed by, the overriding Lizard ophiolite. Barnes (1983) mapped the Roseland area, as far as Great Perhaver Beach [SX 2016, 0423], in considerable detail, and showed that the S. Cornish melanges preserve a distinct stratigraphy. Whilst no detailed structural work has been carried out within the Mevagissey/St. Austell area, recent tectonic evolutionary models have discussed the nature and possible significance of the northern margin of the Gramscatho Basin (the Perranporth-Pentewan line; Dearman, 1971), which is exposed north of Pentewan Beach. Leveridge *et al.* (1984) postulated the existence of the Carrick Thrust, and suggested that it enters Mevagissey Bay at Pentewan, and they further suggested that it can be traced offshore in seismic reflection profiles. Despite subsequent modification of this model by Holder and Leveridge (1986b), the exact course of the Carrick Thrust, and the nature of its outcrop on the east coast, is still uncertain (see 5.4.3). Most recently, it has been suggested that the Perranporth-Pentewan line marks a Devonian terrane boundary (the 'Start-Perranporth Line', Holdsworth, 1989b), reactivated as a transpression zone in the deforming cover sequence, or that it represents a series of faults controlling the development of the deep marine basin further to the south (e.g. Shail, 1992).

There has been no detailed structural analysis of the St. Austell Bay area, and the only published studies are regional tectonic/stratigraphic models (e.g. Hobson, 1976a). Hobson and Sanderson (1983) summarised facing directions across S.W. England and whilst they show a major facing reversal across a NW-SE trending fault through Gribben Head, they do not provide any further detail. Unfortunately this fault, and the Portnadler fault further to the east, precludes any along strike comparisons of the structure of St. Austell Bay, with that described by Tanner (1985) and Burton and Tanner (1986) from the Meadfoot shales south of Liskeard.

5.4.3 Mevagissey Bay

The low strain zone-Turbot Point to Mevagissey Harbour

The outcrop at Turbot Point is dominated by metre scale turbiditic sandstone units which are folded into open, upright primary folds (Plate 5.6). The primary cleavage (S1) fans around these structures, although it generally dips moderately to steeply to the south, striking ENE-WSW (Figs. 5.2, 5.3). Excellent grading in the sandstones indicates northwards and upwards facing in S1. Moving northwards, shales start to predominate, and as the primary cleavage becomes sub-parallel to bedding, younging reversals define a series of upright isoclinal primary folds. Just north of Turbot Point, the first evidence of a later cross cutting fabric appears as a sporadically developed, weak, north dipping S2 cleavage, locally crenulating S1 into a series of tight, centimetre-scale north verging folds. Across Colona Beach [SX 2026, 0432], bedding defines a series of open upright F1 folds, with cleavage refraction and graded bedding indicating a northwards and upwards facing sequence. S1-parallel quartz veins are gently crenulated by a steeply north dipping and ENE-WSW striking S2 cleavage, which is locally extensively quartz veined (Plate 5.7). Between here and Chapel Point [SX 2030, 0433], a series of north verging and facing primary folds are cut by a north dipping S2 cleavage of increasing intensity. Dextral shear bands are sporadically and weakly developed, whilst sinistral shears are largely absent. A fine lineation, observed on some S1 cleavage surfaces, appears to be a bedding/cleavage intersection, and whilst, there is a suggestion of grain flattening within S1 in the coarse graded sandstone bases, there is no visible evidence of grain elongation. Between Chapel Point and Rowards Quay [SX 2025, 0435], the S2 crenulation intensifies and locally tight, upright north verging F2 folds dominate the outcrop. Grading indicates north facing for these and the now isoclinal F1 folds. Immediately north of Rowards Quay, a sequence of NE-SW trending open to tight folds occurs, which are difficult to assign as F1 or F2 in the field (Plate 5.9). Thin sectioning of critical examples reveals a well developed early tectonic fabric (S1), of aligned phyllosilicates sub-parallel to bedding. This is gently crenulated by a later axial planar fabric (S2), with segregation of quartz and phyllosilicates into parallel domains. Between here and Portmellon [SX 2016, 0439], the strain diminishes and once again S1 becomes the dominant fabric. Folds of bedding vary from open upright structures through to intrafolial lenticles, consistently verging and facing north. Down-dip mineral stretching lineations, absent elsewhere in the section, become apparent locally [e.g. SX 2020, 0436], although there are rare examples of fine mineral lineations pitching gently E-W within the subvertical primary cleavage [e.g. SX 2018, 0438]. Shear bands are largely from these exposures, which are extensively faulted by NE-SW trending sub-vertical faults.

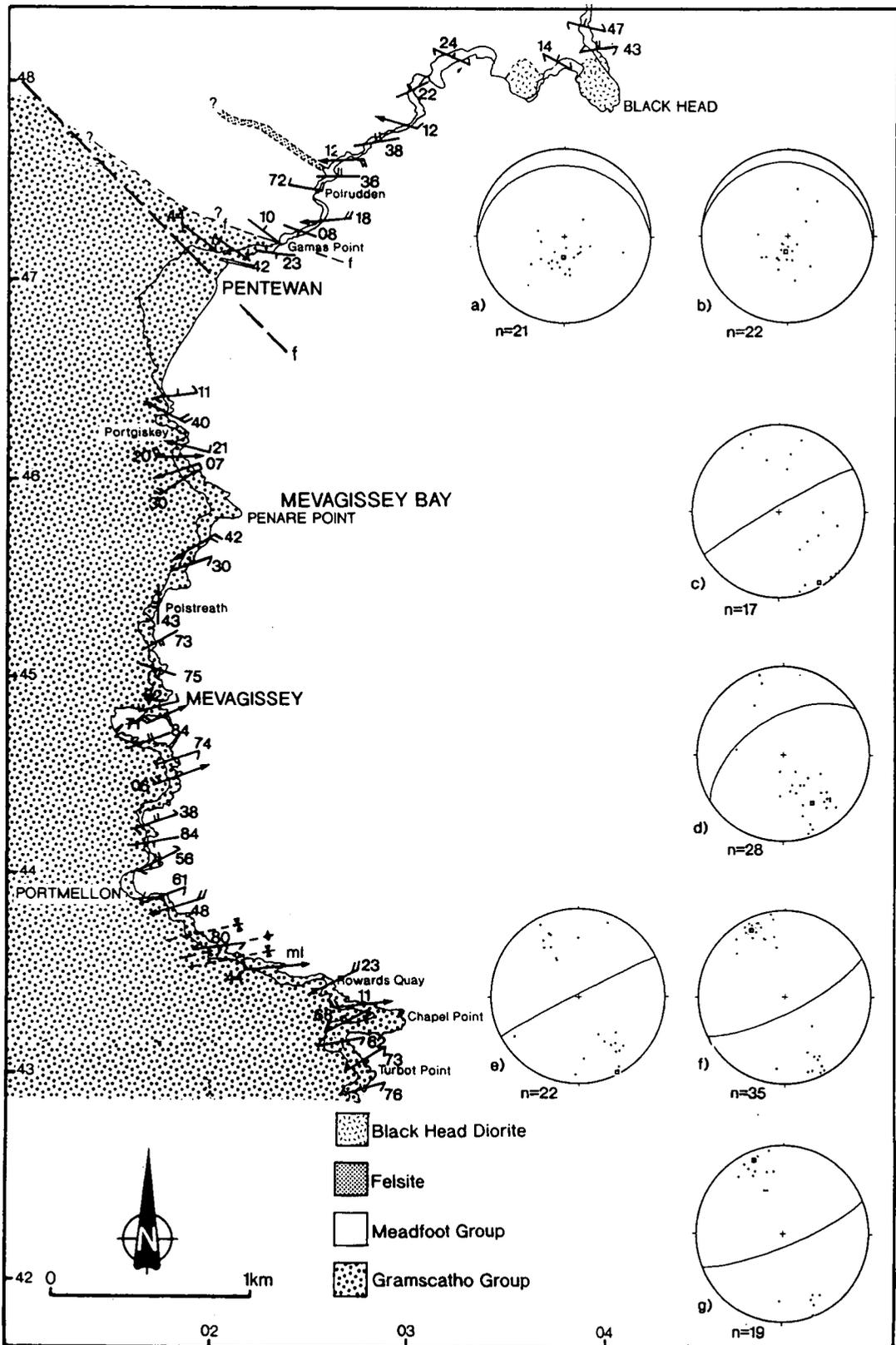


Figure 5.2 Detailed structure of the Mevagissey Bay area. Equal area stereograms of structural data (mean planes are given with their respective poles) from key areas are: a) bedding north of the Gramscatho/Meadfoot boundary; b) primary cleavage from the same area; c) foliation in the high strain area north of Mevagissey harbour; d) secondary cleavage (S2) from Turbot Point to Pentewan; e) late faults; f) primary cleavage and g) bedding from the same area. Refer to Fig. 4.3 for key to structural symbols.

At Portmellon Cove [SX 2016, 0439], bedding is reduced to intrafolial lenticles and the main cleavage (S1) is cut by an increasingly intense north dipping S2 fabric. Quartz veining is more apparent and the shales take on a 'shredded' schistose appearance, whilst the F2 folds become highly variable in orientation and style. At Polkirt beach [SX 2017, 0442], the shales are markedly disrupted by later shallow north dipping extensional faults. In the thicker psammites, the approximately bedding parallel S1 cleavage is crenulated by a strong north dipping S2 cleavage. Well preserved grading suggests a south facing sequence of primary folds, whilst bedding cleavage intersection lineations indicate that these F1 folds are gently ENE-WSW plunging whaleback structures. Tight, north verging F2 crenulations gradually start to dominate the exposures, until at Mevagissey Harbour [SX 2017, 0446] bedding is transposed into parallelism with S1, which is itself intensely crenulated by S2.



Plate 5.6 Medium- to thickly bedded coarse turbiditic sandstones folded into tight upright primary folds, which verge and face to the north. Gramscatho sandstones, Turbot Point [SX 2028, 0430], viewed looking west.

The high strain zone (D2)-Mevagissey Harbour to Pentewan Beach

The shales outcropping north of Mevagissey Harbour are highly deformed and display widespread evidence for pressure solution, as they are intensely quartz veined, and cut by fine anastomosing iron oxide coated seams (Plate 5.9). On Benny Island [SX 2018, 0449], the dominant F2 folds are highly variable in attitude and plunge. Although very steeply plunging in places, they do not appear to plunge through the vertical and there is no direct evidence for sheath folding in this high strain zone (see below). Also, vergence is difficult to assess and many of these folds give the impression of being the result of marked strain heterogeneity (e.g. see Plate 5.9; section 5.6.2).

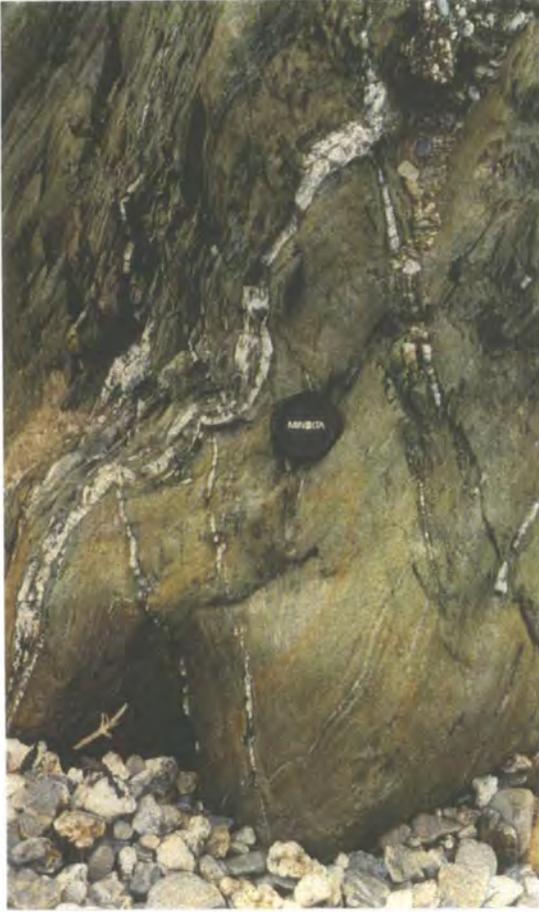


Plate 5.7 (left) S1 parallel quartz veins crenulated by steeply north dipping S2 parallel veins. Primary cleavage dips steeply to the south and primary folds verge and face to the north. F2 crenulations here are weak although intensify northwards. Gramscatho shales, Colona beach [SX 2026, 0432], viewed looking west.

Plate 5.8 (right) Sequence of tight upright folds at Rowards Quay [SX 2025, 0435] which appear to be primary. Thin sectioning reveals an early bedding parallel tectonic fabric, crenulated within fold hinges. Gramscatho sandstones, viewed looking southwest.



At Pentille [SX 2016, 0450], evidence for dextral shear is preserved, with numerous dextral shear bands associated with dextral kink bands and tension gash arrays. It is difficult to distinguish fabrics in these highly schistose pelites, although the occasional interbedded thick psammites, which appear to be less deformed, do allow an examination of the cleavage relationships. These suggest a series of tight north verging and facing primary (F1) folds, although the evidence for this is incomplete and sporadic. Large scale rotations of the fabric strike (e.g. the N-S strikes and steep westerly dips at Polstreath [SX 2016, 0454]) are common and, whilst oblique displacements on the many low angle extensional detachments in this area may be in part responsible, they cannot be the sole cause. These detachments also produce a local overprinting fabric, which can in places be difficult to distinguish from D2 structures (see 5.6.3). Approaching Penare Point [SX 2022, 0457], the strain diminishes rapidly and bedding again becomes apparent within the moderately north dipping main cleavage (S1). Approximately 400 metres of inaccessible coastline surrounds Penare Point, the next exposures occurring at Cockleridge beach [SX 2020, 0459], where a series of tight to isoclinal north verging and facing primary folds dominate the outcrop. S2 crenulations are only sporadically seen, but intensify locally and backsteepen the primary cleavage to dip steeply north. The effects of low angle, predominantly extensional, faults in both reorientating the fabric and causing local structural complexities is increasingly apparent along this section (see 5.6.3).

The exposures around Portgiskey beach [SX 2018, 0464] are characterised by metre scale sandstones with interbedded shales, deformed into a series of tight, gently northerly inclined and north verging folds (Plate 5.10), which are very similar to those noted at Rowards Quay. Again, representative examples were collected and examined in thin section. These proved the existence of an early bedding parallel fabric (S1), crenulated by a strong axial planar S2 cleavage. Primary vergence is difficult to assess, even in the thick sandstones, as bedding and S1 (seen as fine iron oxide coated pressure solution stripes) are sub-parallel. Grading is well developed, and the variability of facing across F2 folds may reflect younging reversals across earlier F1 isoclines. In general, the E-W trending gently whaleback F2 folds verge north and face south (Plate 5.10). Quartz veins are once again abundant in this high D2 strain zone. Pebbles in the graded sandstone bases are markedly flattened within the main cleavage, and some are elongated in a gently west plunging direction, e.g. north of Portgiskey [SX 2018, 0464].

Northwards from here to Pentewan Beach the strain diminishes, although ENE-WSW trending F2 folds are still the dominant structures at outcrop. Better exposed examples of F2 structures clearly refold a fine mineral stretching lineation, e.g. north of Portgiskey [SX 2018, 0464], although there does not appear to be a consistent refold

relationship. Shear bands appear in the rather weathered friable shales north of Sconhoe Rock [SX 2018, 0465], with dextral geometries dominant.



Plate 5.9 Intensely quartz veined black shales within the high strain zone at Mevagissey. Note the cross-cutting veins and disharmonic nature of folding. Gramscatho Group, Benny Island [SX 2018, 0449]



Plate 5.10 Small, tight F2 folds in thickly bedded Gramscatho sandstones at Portgiskey [SX 2018, 0464]. Excellent grading indicates that these structures face to the south. Fine mineral stretching lineations gently wrap clockwise around the fold cores.

There is no exposure across Pentewan Sands (~700m). A river valley runs NW inland from Pentewan, and has been the site of considerable tin working (R. Shail, *pers comm*). It is highly likely that this marks the site of a major NW-SE trending strike-slip fault, which enters Pentewan Sands just south of the harbour wall (see 5.6.1).

The first exposures north of Pentewan [SX 2022, 0472] are thick, metre scale sandstones, which dip gently to the WSW. These are highly disrupted by arrays of gently SW dipping extensional detachments, which bound several metre thick 'packages' of intensely quartz veined and shredded interbedded sandstones and shales. Tight centimetre scale folds in these rocks are clearly related to extensional displacements and earlier structure is largely overprinted. In the larger sandstone units, the problem is compounded by the effects of soft sediment deformation, where it appears that metre scale 'clasts' have been ripped up and inverted by turbidity currents, making the assessment of primary structural relationships difficult (Plate 5.11), evidence for inversion coming from flame structures, load casts, and downward grading. At Gamas Point, a moderately NE dipping normal fault conveniently offsets the concordant junction between the Gramscatho sandstones and the Meadfoot shales (see section 5.2.3; Plate 5.2). Northwards to Polrudden [SX 2025, 0475], the shales contain small folded sandy lenticles, defining tight flat lying and north verging F1 folds, occasionally cut by steeply north dipping S2 crenulations.



Plate 5.11 Large clast of sandstone within bedded grey shales, indicating the effects of turbidity currents. Gramscatho Group, north Pentewan [SX 2022, 0472], viewed looking west.

Across Polrudden beach, D2 structures are again dominant in the black, micaceous quartz riddled shales, where the gently north dipping foliation is intensely crenulated by tight north verging F2 folds. In these zones of high syn-D2 strain, the Meadfoot Beds are highly quartz veined black pelitic schists, the quartz veins often sheared out into

asymmetric quartz 'augen' with no consistent shear sense. Where the strain diminishes, shear bands appear and late extensional faulting is common, the displacements on these faults being evidenced by centimetre scale downslope verging folds and zones of detachment parallel extensional shear bands. The thick granite porphyry elvan at Polrudden, which is quarried inland to the NW, may mark the site of another NW-SE trending strike-slip fault, as there is evidence of syn-emplacement dextral shear (R. Shail, *pers comm*). Approaching Polgwyn Beach [SX 2030, 0480], the D2 structures die out, and large scale WNW-closing primary folds, locally cut by a weak S2 cleavage, dominate in the shales. The exposures around Drennick [SX 2035, 0479] and Black Head are dominated by the large basic intrusives, which are extensively faulted (see below).

Late extensional structures

A common feature across this section is the presence of frequent, metre to tens of metre scale, fabric parallel extensional detachments. These can produce overprinting fabrics, which may complicate local deformation chronologies (see 5.6.3). Across Mevagissey Bay, late faults consistently lie sub-parallel to the dominant foliation (compare Fig. 5.2e with 5.2f), suggesting that the dominant fabric is controlling the site of later extension. At Black Head, the large intrusives appear to be controlling the late extensional deformation, faults in this area bearing overprinted slickensides which allow a detailed kinematic history, involving episodes of compressional and extensional displacements, to be established (see 5.6.3).

Summary

The pattern across Mevagissey Bay is one of increasing strain from D1 dominated deformation in the south, to D2 dominated deformation north of Mevagissey Harbour. The structures south of Portmellon are characterised by open, north facing and verging primary folds, which are locally crenulated by a steep north dipping S2 cleavage of variable intensity. All fabrics strike ENE-WSW, as do later faults, which are predominantly low angle late extensional structures (Figs. 5.2, 5.3). Evidence for dextral shear, whilst not abundant, is apparent, and appears to increase northwards. Mineral stretching lineations are rarely observed and fluctuate from their regional down-dip orientation to an anomalous moderately E-W plunging position. The primary (S1) cleavage dips moderately to steeply to the south, and between Portmellon and Mevagissey Harbour there is the first evidence of this being backsteepened through the vertical to dip steeply north, this backsteepening being clearly associated with an intensification of the moderately north dipping S2 crenulation.

Between Mevagissey and Pentewan, is a broad zone of D2 dominated deformation, although local lower strain enclaves, possibly lithologically controlled, allow some assessment of the primary structure. The fabrics here have variable strikes and may in places trend N-S. The dominant F2 folds verge both north and south, reflecting variations across larger F2 folds. Their facing is similarly variable due to younging reversals across refolded F1 folds. From Pentewan to Black Head, the outcrops are dominated by quartz veined black schistose Meadfoot Beds, deformed into centimetre scale north and south verging F2 folds, which trend ESE-WNW. Extensional detachments are very common and locally disrupt the sequence so extensively that the measuring of structural orientations is effectively invalidated. At Black Head D1 deformation dominates and D2 is reduced to sporadic weak F2 crenulations.

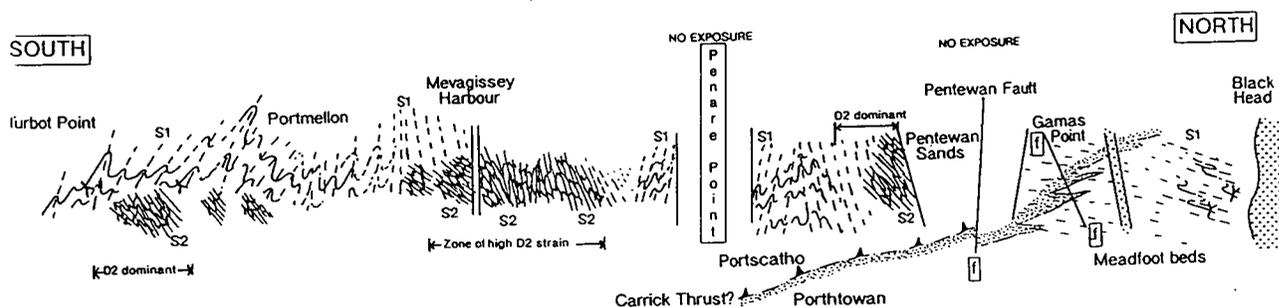


Figure 5.3 Schematic cross-section through Figure 5.2.

Discussion

The regional tectonic implications of the structural observations across Mevagissey Bay are discussed in 5.6, whilst a local interpretation of the structure is given here.

The junction between the Gramscatho sandstones and the Meadfoot shales has previously been interpreted as tectonic (Perranporth-Pentewan Line, Dearman, 1971; Carrick Thrust, Leveridge *et al.* 1984), and is of fundamental importance to tectonic evolutionary models of S. Cornwall. However, the junction, as exposed at Gamas Head, is gradational and whilst it may be offset by a normal fault, this structure does not separate the lithologies (see Fig. 5.3). There is no evidence for a major thrust, such as the Carrick Thrust, in this region, and it is possible that this thrust is faulted out against the proposed NW-SE trending dextral strike-slip fault, which runs through Pentewan beach (the 'Pentewan Fault'; see 5.6.1). The fault at Gamas Point parallels

the trend of the Meadfoot/Gramscatho boundary, as mapped inland (Reid, 1907), and could correspond to the surface representation of WNW-ESE trending faults at depth here. To the south, within the Carrick Nappe, structures trend ENE-WSW (Fig. 5.2) and it is thought that this reflects the influence of WNW-ESE trending extensional faults, which developed during Early Devonian extension across the Gramscatho Basin (e.g. Shail, 1992-Fig. 8.4.1; see 5.6.1). Local anomalies within the Carrick Nappe, e.g. the N-S strikes at Polstreath, may be due to lateral/oblique ramping or 'surge zones' during NNW transport.

The primary structures observed in the low strain regions are broadly similar to those seen in the Devonian shales immediately north of the Start Boundary Fault (SBF) in S. Devon, i.e. north verging but south facing primary folds, backsteepening of the primary fabric, switching of the mineral extension lineations and intense north dipping late crenulations. However, in S. Devon, this structural transition may be traced into, and appears to be controlled by, a steep E-W trending ductile fault zone, whilst in Mevagissey the structural transition may be traced into a broad D2 dominated zone, with the only obvious surface fault being the structure at Gamas Head, some way north of the highest D2 strain zone. Sheath folds, common along the SBF are absent from the Mevagissey exposures, although on Benny Island the F2 folds are very steeply plunging (see 5.6.2). It is quite probable that the zone of highest strain, seen straddling the SBF in S. Devon, has been faulted out at Mevagissey Bay (see 5.6.1).

5.4.4 St. Austell Bay

This section covers the exposures north of Black Head (Gerrans Point [SX 2040, 0487]) in the west, across Par Sands [SX 2083, 0530] in the north, to Gribbin Head in the east (Figs. 5.4, 5.5). The eastern side of Gribbin Head was also mapped in order to study the major facing reversal recorded in this area.

The sequence from Gerrans Point to Gwendra Point [SX 2034, 0494] is characterised by monotonous grey/green shales in which the primary cleavage dips moderately to gently to the NNE, and is only occasionally crenulated by a weak late cleavage. Typically these later fabrics are related to local extensional faults, which are weakly developed across this section. Frequent thin pale intrusives bear rusty pyrite aggregates which, although flattened within the primary cleavage, display no obvious stretch, e.g. north of Gerrans Point [SX 2039, 0489]. Deformed fossils, especially solitary corals, within calcareous shale horizons are similarly flattened within S1, but are unstretched, e.g. at Ropehaven [SX 2035, 0490]. Bedding is only ever seen as fine black mudstone lenticles, with the bedding/cleavage intersection plunging gently ENE. The vergence and facing of these primary folds is unknown.

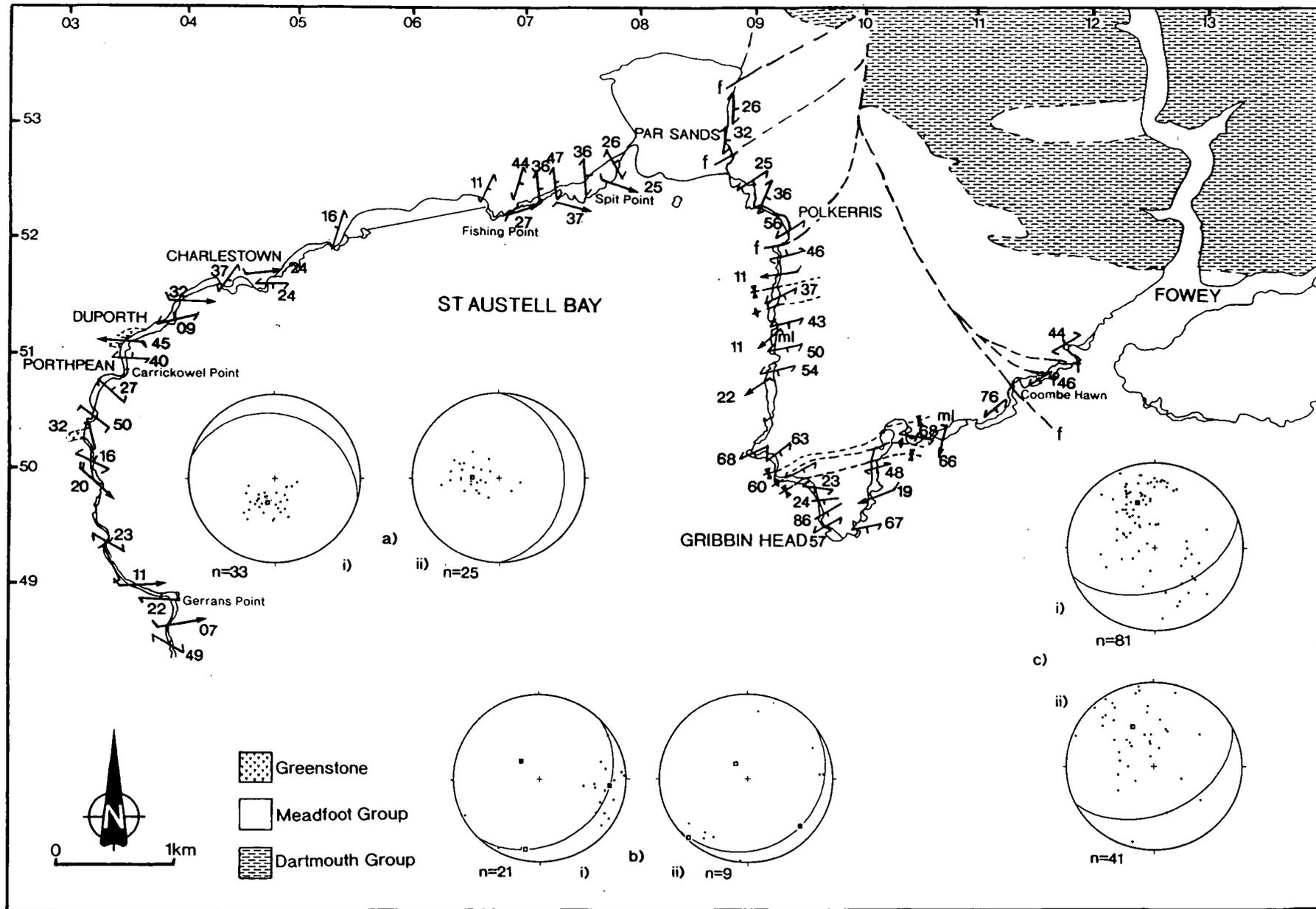


Figure 5.4 Detailed structure of the St. Austell Bay area. Equal area stereograms of structural data (mean planes are given with their respective poles) from key areas are: a) primary cleavage along the west coast (to Charlestown, i) and the north coast (to Spit Point, ii); b) bedding/cleavage intersections from the north coast (i) and the east coast (ii). Pole to great circle represents approximate axis of late fold (see text); and c) primary cleavage (i) and bedding (ii) from the Gribbin Head area. Refer to Fig. 4.3 for key to structural symbols.

A similarly monotonous sequence extends north to Charlestown [SX 2040, 0515], although bedding is occasionally preserved, revealing open north verging folds of uncertain facing direction. Locally, the fabric rotates to strike N-S, with bedding/cleavage intersections plunging to the NNE. Foliation parallel extensional detachments are increasingly common, especially at Polmear Island [SX 2038, 0513], and locally produce complex overprinting fabrics. Basic intrusives between Carrickowel Point [SX 2035, 0508] and Polmear Island contain flattened and elongate pale xenoliths and occasional weathered pyrite aggregates, which indicate variably plunging extension, from gently east to almost down-dip. North from Charlestown [SX 2039, 0515], the main cleavage is vary variable in orientation, due to rotation and refolding related to the large extensional detachments, which shred the sequence here. Rarely observed primary folds verge both south and north, e.g. at Appletree Beach [SX 2045, 0516], whilst facing is difficult to assess due to scant younging evidence.

Approaching the south end of Carlyon Bay [SX 2054, 0519], the fabric consistently strikes NNE-SSW and dips gently to the east. Much of the exposure here is highly altered and mineralised, the area having been extensively mined for silver and copper (Ussher *et al.* 1909). Much of the soft weathered outcrop across Carlyon Bay yields no useable structural data. A critical E-W section of highly anomalous structure extends from Fishing Point [SX 2067, 0522], at the east end of Carlyon Bay, to Polkerris harbour [SX 2093, 0521], east of Par Sands. The area was mapped in detail to assess its relationship to the more usual structure of this region. The transition from Charlestown to Carlyon Bay is poorly constrained due to coastal inaccessibility and general lack of exposure, and whilst the east coast transition from Polmear to Polkerris is well exposed, the outcrop is highly weathered and structure poorly preserved. From Fishing Point to Spit Point [SX 2077, 0525], the primary cleavage strikes N-S and dips moderately to the east, whilst bedding-cleavage intersections vary from plunging down the dip of the cleavage to the east, to plunging gently SSE. The primary vergence appears to be to the north, and this is largely confirmed by the thin basic intrusives, which are deformed into north verging and east plunging open folds (Plate 5.12). Facing is generally impossible to assess, although at Fishing Point, medium bedded sandstone units are folded into tight north verging primary folds, which, whilst offering ambiguous younging evidence, seem to suggest south facing (see Plate 5.3). Fabric parallel detachments are well developed, striking N-S and locally producing an intense crenulation of the primary cleavage (Plate 5.13). Evidence for D2 modification in this area is lacking, as is evidence for high strain, e.g. sheath folding, etc.



Plate 5.12 Thin, buff coloured pre-orogenic intrusives, now largely replaced by calcite, folded into north verging and east plunging folds. Meadfoot shales, Fishing Point [SX 2067, 0522].



Plate 5.13 Intense crenulation of the primary cleavage in interbedded sandstones and shales. Secondary crenulation is related to displacements on fabric parallel extensional detachments, which bound this 'package' of rock. Lower detachment is obscured by sand. Meadfoot Beds, Fishing Point [SX 2067, 0522], viewed looking west.

At Polmear, on the east side of Par Sands, a N-S striking primary cleavage dominates in the soft weathered shales. Little other structure may be observed and the fabric gradually rotates until at Polkerris it once again dips moderately to the north and NNW. Several major NNE-SSW trending faults run through here, especially at Little Hell [SX 2087, 0526] and south of Polkerris Harbour. Between Booley Bay [SX 2088, 0524] and Polkerris, an array of late extensional faults modify the primary cleavage orientation, and complicate the structural transition here.

Gribbin Head is characterised by open north verging primary folds which face south at Polkerris, although steepen rapidly to face north across the rest of the promontory. Basic intrusives, deformed into north verging primary folds (e.g. see Plate

5.12), bear xenoliths stretched in a down-dip direction (Plate 5.14), whilst occasional solitary corals within the adjacent calcareous shales also indicate down-dip extension, e.g. south of Polkerris [SX 2093, 0519]. Access at Gribbin Head is generally good and along its southern tip a sequence of primary north verging antiform/synform pairs may be mapped out and correlated along strike with some accuracy. A similar sequence is observed westwards from here, until at Combe Hawn [SX 2115, 0506] a major facing reversal occurs across a zone of NNW-SSE trending faults, the primary cleavage west of this dipping moderately to the SSE, whilst that to the east dips gently NNE. The fault zone is complex, and further disrupted by arrays of extensional collapse structures.



Plate 5.14 Cleavage surface (S1) within pre-orogenic intrusive, displaying down-dip stretched xenoliths and weathered pyrite fragments (now probably limonite). Meadfoot shales, Polridmouth [SX 2104, 0503].

Summary

The western coastline of St. Austell Bay is characterised by poorly developed open north verging primary folds, which appear to face south within the moderately north dipping primary cleavage, although younging evidence in these outcrops is generally ambiguous. Primary folds plunge gently ENE and the rarely observed mineral stretching lineations indicate mostly oblate strains with both down-dip and along strike extension. Late extensional faults increase in abundance and complexity northwards, locally causing intense disruption.

The northern edge of St. Austell Bay is characterised by an anomalous N-S striking and east dipping primary cleavage, with the bedding/cleavage intersection plunging down the dip towards the east and ENE. Primary folds verge north and the facing appears to be south. Local intense crenulations of the primary cleavage are clearly related to extensional fault movements. Eastwards from here the primary cleavage rotates to dip moderately northwards at Polkerris and then tips over to dip south at Gribbin Head. Primary folds at Gribbin Head verge and face north, and the well developed mineral stretching lineations consistently indicate down-dip extension. East of Gribbin Head the facing is seen to reverse across a NW-SE striking, strike-slip fault.

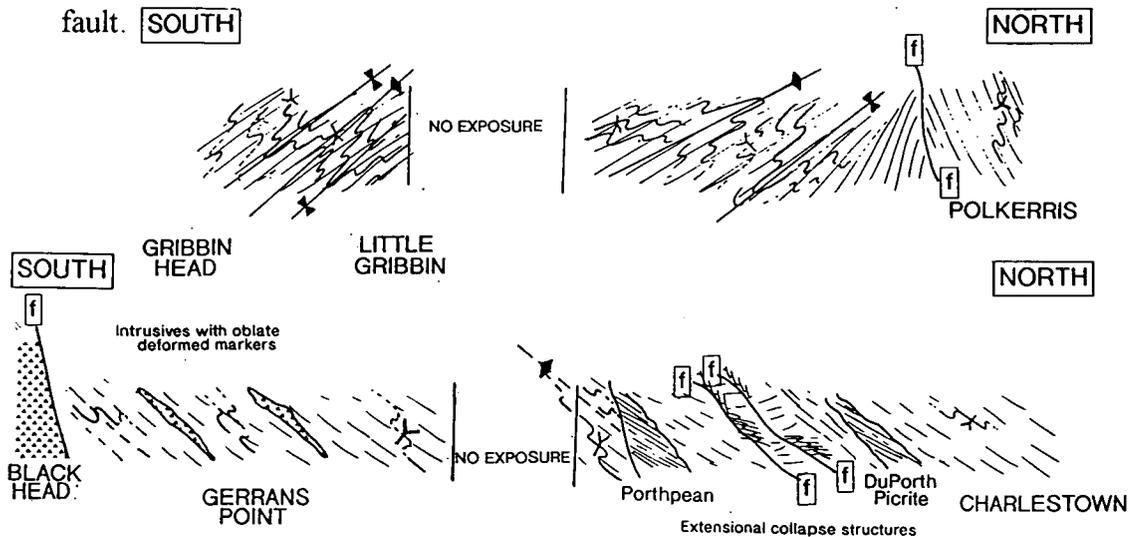


Figure 5.5 Schematic cross-section through Figure 5.4.

Discussion

Despite the apparently simple and monotonous structures observed across St. Austell Bay, there are a number of structural anomalies which need to be explained.

The first problem within this region concerns the gently north dipping S1 cleavage recorded across the western coast of St. Austell Bay. Vergence along here appears to be predominantly northward, and whilst facing evidence is rare, most examples indicate southwards facing. Evidence for large-scale secondary backfolding is lacking along this coast, and thus the anomalous structure cannot be simply accounted for by regional backsteepening of an originally northwards facing sequence (*cf.* the Mevagissey area). The northerly vergence within the north dipping and south facing primary cleavage is difficult to account for. Lane (1970) showed a zone of south facing and south verging primary folds running east of Gribbin Head, through Fowey, Lantic & Lantivet Bays. Reconnaissance mapping across Lantic and Lantivet confirmed that primary southwards facing and vergence do predominate. Mapping of St. Austell Bay has confirmed the presence of major NNW-SSE faults at Combe Hawn, across which the facing reverses, and NNE-SSW trending faults at Par Sands, including the

significant structures at Little Hell and Polkerris harbour, these probably being splays off the major N-S faults north of this region (Ussher *et al.* 1909). Lane (1970) accounts for the south facing east of Combe Hawn as the result of a simple 80° rotation of the original structure about a sub-horizontal ENE-WSW axis, the displacement flooring along a NNW transporting décollement. Evidence cited for this includes the change in fold vergence, the gradual backsteepening of the primary cleavage on approaching the south facing zone, and the reorientation of faults. Hobson (1976a), whilst agreeing with a limited amount of rotation, attributed the south facing at Lantic/Lantivet to a regionally consistent late open monoformal refold of the Dartmouth Antiform, the southern limb of this monoformal being sub-vertical or inclined to the north. He suggested that the moderate northerly inclination of the fabric at Lantic/Lantivet may indicate that the monoformal is tightest in this area. Although late modification of the primary structure is seen elsewhere in S.W. England (e.g. S. Devon), there is no direct evidence at Lantic/Lantivet or west St. Austell for this phenomenon, and the structural pattern at Lantic/Lantivet would necessitate the backsteepening of an inverted fold limb to give the south vergence. It is possible, and also highly speculative, that the structure observed across the west side of St. Austell Bay is on the inverted limb of a southwards transporting and verging primary overfold, with the Lantic and Lantivet structures representing the upper limb. The N-S faults either side of Gribbin Head effectively bound a regionally normal area of northwards verging and facing structures, which cannot be correlated westwards along strike. The northern and western limits of the inverted (St. Austell) limb are impossible to specify and again require the existence of a NW-SE trending strike-slip fault to the west of St. Austell Bay. Warr (1991) noted the importance of the pre-existing basin architecture, in controlling the deformation during basin inversion, and showed that broad zones of backthrusting and primary southwards facing could arise through fault controlled buttressing. Whilst this may account for the primary southwards facing along western St. Austell Bay, and the lack of secondary backsteepening structures, it cannot explain the primary northwards vergence. Also, along strike at Holywell Bay, the effects of buttressing diminish rapidly northwards from the northern Gramscatho margin (see 5.5.5), restricting the zone of southwards facing. Nonetheless, although it is difficult to account for the southwards transport invoked both in western St. Austell Bay, and east of Combe Hawn, the evidence clearly indicates that southerly transport must occur.

The second problem concerns the anomalous orientation of the primary cleavage at Fishing Point, which strikes ~N-S and dips to the east and ENE, almost orthogonal to the regional pattern, whilst bedding/cleavage intersection lineations

plunge in a regionally consistent ENE direction. This structure might be explained in a number of ways. It is possible that the cleavage measured at Fishing Point is a later fabric (S2) related to D2 modification, which would account for its orientation and the orthogonal relationship to the bedding/cleavage intersection. However, field evidence strongly suggests that the main cleavage here is primary, only locally crenulated by later fault related fabrics (e.g. see Plate 5.13). Alternatively, the structure at Fishing Point has simply been reclined by a later open NNW-SSE trending and north plunging synform. Evidence for this comes from the plunge of the F1 axes, which at Fishing Point plunge gently ESE, whilst at Polkerris they plunge to the SW and WSW. These axes all plot within a common great circle (Fig. 5.4b, i) & ii)), implying folding about an NNW-SSE trending axis. Whilst open domes and periclinal folds have been recorded parallel to granite ridges, possibly due to emplacement related extension (see 5.1, *D5 Deformation*), the influence of the St. Austell granite is not strong here, and evidence for such large scale open folds (e.g. S2 cleavages, etc.) is lacking both here and elsewhere within this region. Also, although the axial region of the proposed fold would be obscured within Par Sands, it is felt that such a large scale feature should produce a local deformation overprint. A third possibility is that the cleavage orientation reflects oblique/lateral ramping during primary transport, or even a 'surge zone' within the thrust sheet. In the latter case, the primary fold axes would be expected to rotate, and should show a consistent relationship to the primary cleavage either side of the 'surge zone' (Coward & Potts, 1983). Whilst such lateral transitions could not be tested, the E-W orientation of the primary folds at Fishing Point would tend to rule out this possibility. Lateral ramping does provide a reasonable explanation of the observed structure, and it is worth noting that several large N-S or NNW-SSE trending faults, sub-parallel to the primary cleavage here, have been mapped inland towards St. Austell (Ussher *et al.* 1909). These may be influenced by local N-S trending basement structures, which have also controlled the lateral ramping. Thus the available evidence indicates that the anomalous N-S striking cleavage in this region may be attributed either to basement controlled lateral ramping or possibly extension related late open refolding.

5.5 The structure of the western coastal sections

5.5.1 Introduction

The west coast of S. Cornwall was mapped on scales of 1:2500 and 1:5000, from Pen a Grader, south of Cligga Head, to Towan Head, west of Newquay, whilst brief reconnaissance mapping was carried out from Newquay to Stem Point, at the northern limb of the Watergate Bay antiform (Fig. 5.1).

Access to the coast north of Perran Bay is excellent and the exposure is almost 100%. South of Perran Bay, the precipitous cliffs from Droskyn Point to Pen a Grader greatly hinder access, although the steps adjacent to old mine workings at Cligga Head do allow access to the foreshore at Hanover Cove. Whilst the sand dunes along Perran Bay obscure approximately 1 kilometre of critical exposure, this is the only significant gap within an almost complete north-south coastal traverse.

The study area is divided into a northern zone, Holywell (north) to Towan Head, characterised by primary (D1) dominated deformation, and a southern zone, Holywell (south) to Perranporth/Cligga, characterised by D2 dominated deformation, and whilst the transition between the two areas is gradational, an arbitrary division is placed through Holywell Bay (Fig. 5.1).

5.5.2 Previous work

The early work on the Perranporth/Newquay area is summarised in Reid and Scrivenor (1906) and is largely concerned with establishing a formal stratigraphy, with few structural observations. MacAlister (in Reid & Scrivenor, 1906) detailed the major mine workings in the area, and noted that the extensively studied 'Perran Iron Lode' is "...characterised by large amounts of brecciated material..." (p100), suggesting that this structure has been the site of significant fault displacements.

More recently, Ripley (1965) recognised several 'fold phases', and described two cross-cutting sets of F1 folds, the first set plunging 10-30° NE-SW or ESE-WNW, whilst the second set plunge N-S. He also noted a progressive change in the inclination of the primary cleavage, from sub-horizontal in the north, to ~45° south of Holywell Bay. He noted that associated with this steepening is a tightening of folds moving southwards, such that at Holywell Bay the average interlimb angle is 15°. Following this, Henley (1970, 1973) and Sanderson (1971), produced conflicting models for the large scale structure of this area. Both observed tight, gently southerly inclined and north facing primary folds, cut by a moderate north dipping S2 fabric. They disagreed over primary vergence, Sanderson (1971) suggesting that it is southwards, and thus implying that the structure here lies on an inverted fold limb, whilst Henley (1973) argued that southwards vergence is restricted to a 100 metre

zone at Cotty's Point, an inverted minor fold limb on a major north verging structure. Both also noted a zone of steep fabrics straddling the Meadfoot/Gramscatho boundary. Henley (1973) also noted that within a 1-2 kilometre wide zone, both F1 and F2 folds tighten and cleavages approach the vertical. Sanderson (1971) suggested that within the steep zone, F2 folds verge north, whilst on the shallower limbs the F2 vergence is south.

In order to account for the steep zone, Sanderson (1971) invoked a model involving a large, monoformal south facing F2 re-fold of originally flat lying inverted primary folds, the limbs of which were probably separated by an F1 slide, which formed a distinct tectonic junction between the Meadfoot and Gramscatho beds (Fig. 5.6). Henley (1973) questioned various aspects of this model, including the style of the F2 monoform, the use of ambiguous younging from coastal sections, the extent of the inverted sequence and the need to place a slide at the Meadfoot/Gramscatho boundary. He suggested, based on better younging evidence from further inland, that the sequence is the right way up, and that the steep zone is due to post- F2 movement of the Perran Iron Lode. Both authors also recognised sub-horizontal F3 folds, possibly related to granite emplacement.

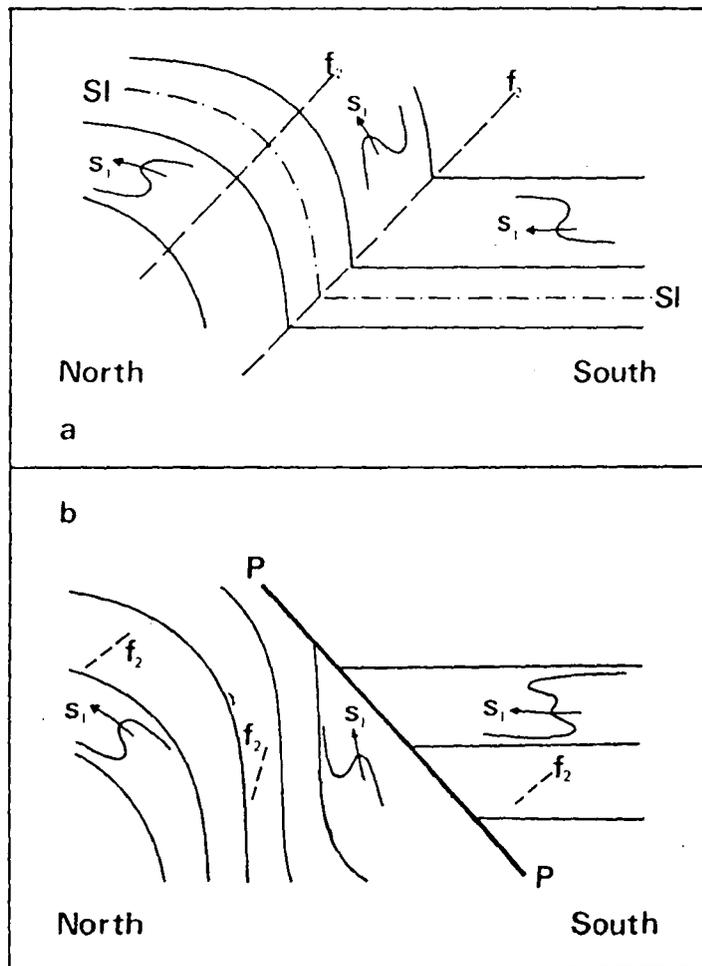


Figure 5.6 Two alternative interpretations of the structure of the Newquay/Perranporth area: a) shows a steep zone resulting from F2 monoformal refolding, SI-SI is the proposed slide separating the inverted and right-way-up limbs (after Sanderson, 1971). b) shows the possible effects of N-S compression about the Perran Iron Lode (P-P), which separates the steep zone in the north, caused by post -F2 compression, from the relatively undeformed rocks to the south (after Henley, 1973).

Hobson (1976a), in detailing the structure of the Dartmouth Antiform, regarded the Perranporth-Pentewan Line as tectonically significant, and possibly only one of several, as yet unrecognised, E-W trending faults controlling deformation in S.W. England. Primmer (1983, 1985) studied the evidence of low grade regional metamorphism across the Perranporth-Pentewan Line, and concluded that two different metamorphic environments exist, and that the more southerly area is in many ways comparable to the Start Complex (see 5.6.1). This emphasised the link between the Perranporth-Pentewan Line and the Start boundary, and tended to refute any link between the Start Complex and the Lizard/Dodman areas.

Leveridge *et al.* (1984), proposed a tectonic history for S. Cornwall, and agreed with Henley (1973) in disputing the existence of an F1 slide marking the western outcrop of the Perranporth-Pentewan Line. They also noted that the main structural feature of the Perranporth area is a steep zone of secondary transposed cleavage, and suggested that this cannot be traced east to Mevagissey Bay. Holdsworth (1989a) cited evidence for dextral shear, lineation switching, sheath folding and backsteepening of primary fabrics in the Perranporth area, and suggested that this indicates dextral transpression due to oblique shortening across the Perranporth-Pentewan Line (his 'Start-Perranporth Line'). Most recently, Shail (1992) offered a possible tectonic evolutionary history for the Gramscatho Basin, based on a study of provenance and sedimentology. He suggested that, although its northern margin trends ESE-WNW, indicating that the main basin bounding faults are also of this orientation, the main structures probably controlling basin evolution and inversion are NE-SW trending extensional faults. Evidence for the orientation of the basement architecture comes from the trends of igneous bodies and mineral lodes (See 5.6.1).

5.5.3 The low strain (D1-dominated) zone-Towan Head to Holywell Bay

Across Towan Head [SW 1800, 0630], the coarse sandy Meadfoot shales are strongly foliated, with the primary cleavage here dipping moderately to the south and SSE (Fig. 5.7, 5.8). Bedding is reduced to thin pale discontinuous horizons or intrafolial lenticles, and defines a series of tight, north verging primary folds of uncertain facing. Almost orthogonal to these structures, a set of N-S striking, and west overturning, centimetre scale folds of the main cleavage are developed. These are very variable in style although predominantly appear to be brittle, with open asymmetric angular hinge regions and markedly fractured axial surfaces. In places, these brittle folds pass laterally and vertically into zones of intense conjugate kink folds, whilst in others they pass into open sub-cylindrical folds with no axial planar fabric or obvious vergence.

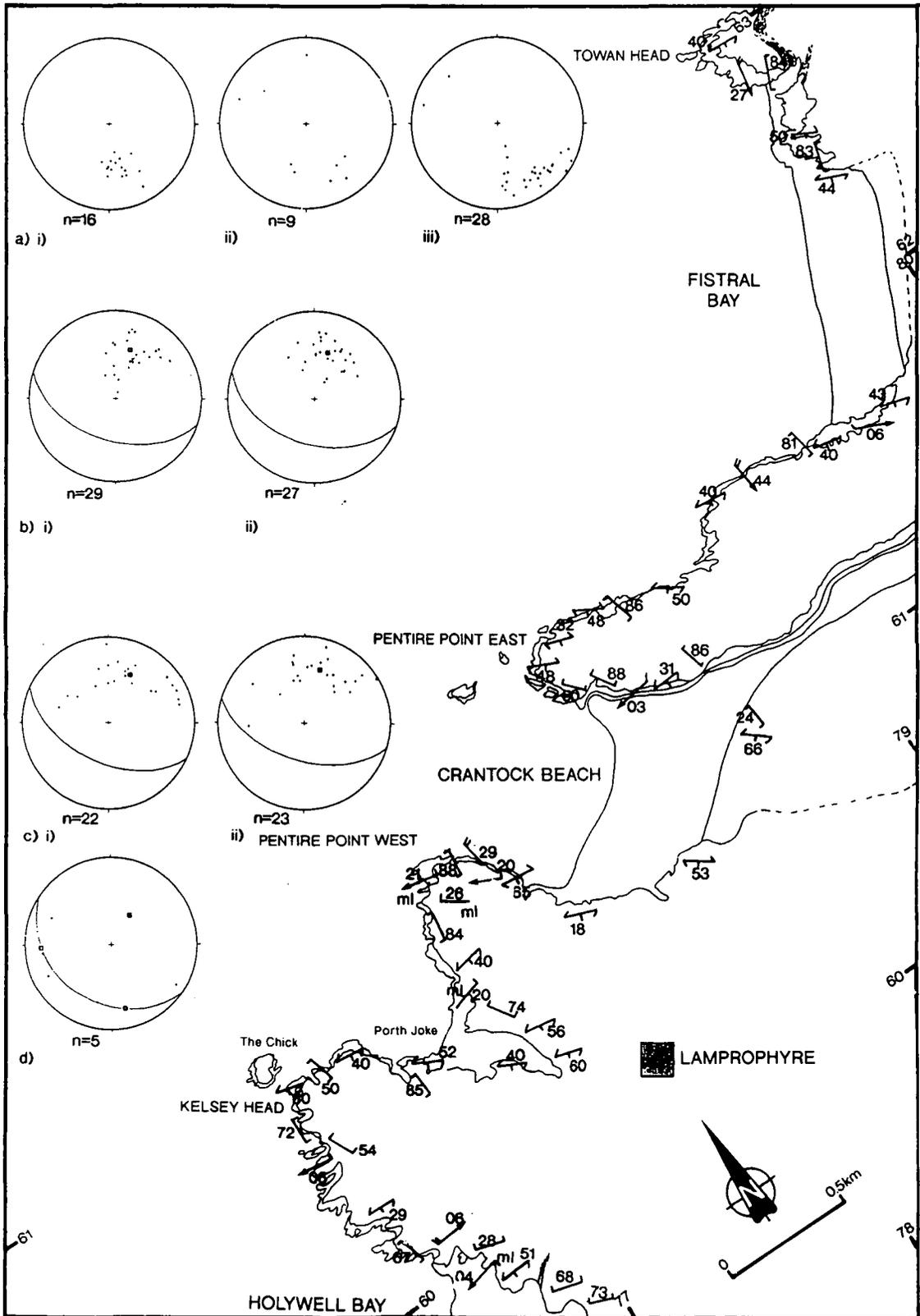


Figure 5.7 Detailed structure of the Holywell/Fistral Bay area. Equal area stereograms of structural data (mean planes are given with their respective poles) from key areas are: a) post main cleavage age fold axes. Note the rotation and shallowing from i) Towan Head (46/168), through ii) Crantock (28/155) to iii) Porth Joke (21/143); b) primary cleavage at i) Crantock and ii) Porth Joke; c) late faults at i) Crantock and ii) Porth Joke; d) mineral stretching lineations at Porth Joke. Refer to Fig. 4.3 for key to structural symbols and Fig. 5.2 for key to lithologies.

A very similar structural pattern is seen moving along strike, from south of Fistril Beach [SW1795, 0616] to Pentire Point East [SW 1780, 0615], except that in the latter region, evidence for dextral shear becomes very apparent. The relationship between the shear bands, extensional detachments and the late folds is difficult to resolve with certainty. The shear bands preserve dextral oblique slickenfibres, as do the larger extensional detachments, which are associated with centimetre scale downslope verging ductile folds. The N-S brittle folds are not reorientated by the shears here, nor do they refold the ductile folds, and it would appear that the structures are synchronous. Also, the orientation of the N-S brittle folds is very variable, this variation appearing to be largely primary. Moving westwards to Pentire Point East, the evidence for oblique extension increases and there are several examples of downslope facing centimetre scale sheath folds, indicating that a high degree of shear strain has localised along these detachments. Exposure is insufficient to test whether these structures display hinge line variations.

The western tip of Pentire Point East, is a large open north facing synclinal fold of coarse sandstones and interbedded shales. The northern limb of this primary structure is dissected by extensional faults, whilst just to the south of here a series of north verging and facing primary folds is similarly disrupted. Here are some of the best examples of foliation parallel extensional 'packages' within the entire field area and are discussed in 5.6.3.

The south face of Pentire Point is characterised by anastomosing arrays of NW-SE striking dextral shear bands, with excellent slickenfibres and asymmetric folds indicating oblique dextral extensional movements (Fig. 5.7; see 5.6.2). Again, the relationship between these shears and the post main cleavage age folds is difficult to assess, and whilst there are demonstrable examples of brittle N-S folds being rotated adjacent to shear bands, a consistent deformation chronology for these post-main cleavage age folds could not be established. Crossing Crantock beach [SW 1780, 0610], occasional poorly exposed outcrops in the sand dunes reveal tight north verging primary folds of unknown facing, cut by a steep axial planar fabric, e.g. beneath Rushy Green [SW 1785, 0610]. Sporadic dextral shear bands, often quartz infilled, cross this outcrop.

Moving along strike to Pentire Point West [SW 1773, 0614], a variably south dipping primary cleavage is axial planar to isoclinal north verging folds. Occasional thin sandstones (<10 cm) give ambiguous younging, although the facing in general appears to be to the north. Again, dextral shear bands, with excellent displacement fibres, are common, these striking NW-SE and dipping moderately to the SW (e.g. see Fig. 5.7; 5.6.2). Evidence for late extension is common, and it appears that the

extensional displacements along here have localised along the many thin intrusives, indicating that these may have formed zones of weakness.

From Pentire Point West to Porth Joke [SW 1770, 0607], a sequence of very uniform and monotonous slates occurs, in which occasional thin sandstones are deformed into tight north verging and facing primary folds. Deformed markers within pale, bedding-parallel intrusives indicate gently WSW plunging extension directions. Once again, N-S to NW-SE trending late folds are common, and occur on all scales from centimetre up to tens of metre scale. These vary from brittle angular kink folds, often in conjugate sets, through to open sub-cylindrical folds, which, whilst having no axial planar fabric, do display markedly fractured axial surfaces. Some of these folds may be the product of displacements on the larger extensional detachments, although this cannot account for the majority of these anomalously orientated structures. At Porth Joke, the shallow detachments are cut by later sub-vertical normal faults. A whole range of orientations of later folds, from N-S to ESE-WNW, occurs, and it appears that there is no correlation between fold style and orientation, although the majority of the N-S orientated structures do appear to be more brittle in appearance. Also, there is no correlation between fold style and development of axial planar fabrics, some of the sub-cylindrical structures being intensely shattered by axial planar fractures. These fractures can in places completely dissect the outcrop, effectively precluding measurement of structural data.

Approaching Kelsey Head [SW 1765, 0609], later folds completely dominate the outcrop, and primary folds are reduced to north verging intrafolial lenticles. The across strike section from Kelsey Head to north Holywell Bay [SW 1768, 0597] is very monotonous, with broad expanses of variably south dipping slaty cleavage cut by thick, pre-orogenic basic intrusives. Isoclinal primary folds consistently verge north and plunge gently ESE [e.g. SW 1764, 0604]. The steep cliffs along here are cut by extensive arrays of large, gently south dipping detachments, which appear to have oblique dextral displacements, as evidenced by offsets of pre-orogenic intrusives and occasional thin sandstones [e.g. SW 1766, 0600]. Frequent NW-SE trending and SW dipping dextral shear bands also display oblique slickenfibres, and it thus appears that a complete spectrum of synchronous structures occurs, from centimetre scale dextral oblique shears up to tens of metre scale fabric parallel detachments with dextral oblique displacements. The orientation of the late folds here is highly variable, some of this being a consequence of subsequent rotation by shears, although the majority of it appears to be a primary feature. Clear examples of these N-S folds being rotated by dextral oblique shears occurs south of Kelsey Head [e.g. SW 1764, 0608], whilst at Holywell Bay the shears and folds appear to be more or less coeval.

Across the north edge of Holywell Bay, thin graded sandstone units are folded into open north verging and facing primary folds, which are crenulated by the NW-SE trending brittle folds [e.g SW 1767, 0598]. These post main cleavage age folds have a well developed axial planar fabric, which here is no longer cut by brittle fractures, but appears to be a more ductile crenulation. Moving southwards, this crenulation fabric, which has rotated to dip variably to the north and NNE, intensifies, although isoclinal north verging F1 folds are still preserved [e.g. SW 1767, 0597]. Excellent younging in sandy horizons indicates primary north facing. Frequent pale green intrusives, some of which display evidence for multiple intrusion, bear deformed markers elongate in a ESE-WNW direction. Dextral shear bands here are reduced to millimetre to centimetre spaced surfaces, with limited offsets. Similarly, large extensional detachments are rare, although small (10's cm) ductile shears are still seen.

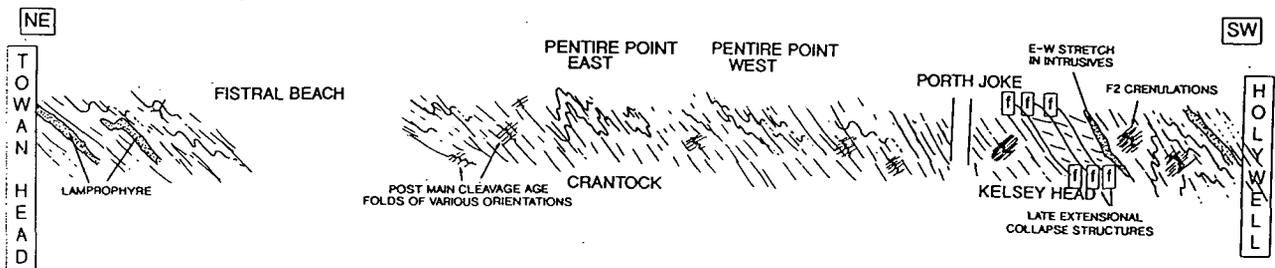


Figure 5.8 Schematic cross section through Figure 5.7

Summary

In the north of the area, the primary cleavage dips moderately to the SSE and tight primary folds verge north (Fig. 5.8). South plunging post main cleavage age folds predominantly overturn to the west, although these are locally highly variable in orientation, style and attitude. Moving southwards, dextral shear increases dramatically and is most intense in the cliff exposures north and south of Crantock beach. Shear bands and the larger scale extensional detachments all display oblique displacement fibres and have variable relationships to the post main cleavage age folds, although may post date folding. Approximate E-W extension is indicated in the primary cleavage by deformed markers in frequent pale basic intrusives, many of which act as zones of weakness during later extension. Grading in occasional interbedded thin sandstones gives a consistent northwards facing. Approaching Holywell Bay, primary folds tighten and are locally reduced to intrafolial isoclines. The primary cleavage here is intensely crenulated by a moderate NE dipping later fabric.

5.5.4 The high strain (D2-dominated) zone-Holywell Bay to Perran Bay

The first exposures south of Holywell beach [SW 1764, 0582] are characterised by high strain, with bedding generally transposed into parallelism with the primary cleavage, and folds reduced to rare attenuated intrafolial isoclinal folds. The main fabric dips steeply to the south, this backsteepening probably related to the intensifying, north dipping S2 fabric. The primary vergence appears to be to the north (Fig. 5.9). It is impossible to assess facing in these high strain outcrops. These grey/green striped shales are crenulated by a later fabric of variable orientation and intensity, which is very similar to that noted to the north of the bay. The nature of the main fabric here, and its relationship to the secondary crenulations, is discussed below.

Moving southwest towards Penhale Point [SW 1756, 0592], the secondary crenulation intensifies and locally starts to dominate the outcrop. This crenulation is variable in orientation (N-S to ESE-WNW) and attitude (sub-recumbent to sub-horizontal), and has a generally brittle appearance, resembling intense kink banding (Plate 5.15). Locally, these late crenulations show a marked variability of plunge [e.g. 1763, 0592], steepening up to the sub-vertical, although their hinge lines cannot be demonstrated to pass through the vertical. The vergence of these folds does not appear to follow any systematic pattern, and the refolding of the early mineral lineation by these structures is similarly variable. Also, zones of steeply plunging folds are all demonstrably bounded by shears, and it seems that these structures probably do not represent sheath folds, but have simply been reorientated into their steep positions. As the late strain increases, the north verging primary isoclinal folds are gradually lost, and the primary fabric appears to be completely transposed. The vergence of the secondary crenulations, although locally variable, especially where the late fabric is sub-horizontal, appears to be predominantly northwards. Dextral shear bands are reduced to very fine, millimetre spaced discrete surfaces with small brittle offsets. There is no evidence for larger scale dextral shears or late extensional structures.

Just to the east of Penhale Point, is a steep north dipping late fault, in the hangingwall to which the main fabric steepens to dip steeply north (Plate 5.16). In the footwall, there is a broad zone (>60m wide) of ductile fault rock, which appears to have originally been quartz veined Devonian shale, and is now reduced to a coarse phyllonite. Asymmetric quartz augen within this high strain zone consistently indicate dextral shear (Plate 5.17). Occasional 'rafts' of pale grey shale and thin buff layers,

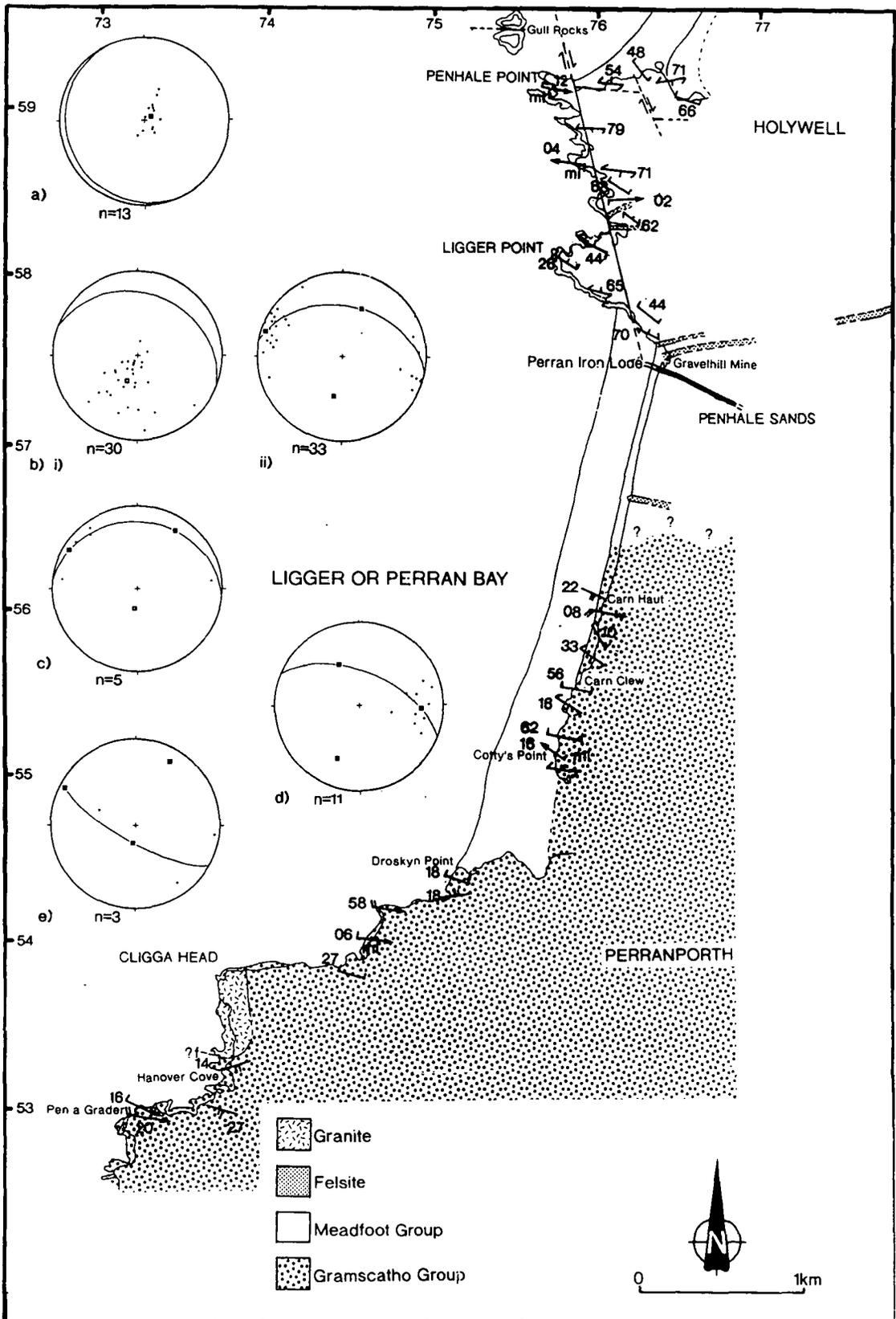


Figure 5.9 Detailed structure of the Holywell/Perran Bay area. Equal area stereograms of structural data (mean planes are given with their respective poles) from key areas are: **a)** foliation from the high strain area along the north of Perran Bay; **b)** secondary fold axial planes (i) and axes (ii) from Penhale to Cotty's Point; **c)** mineral stretching lineations from the same area; **d)** secondary folds from Droskyn Point to Pen a Grader; **e)** mineral stretching lineations from the same area. Refer to Fig. 4.3 for key to structural symbols.

presumably originally intrusive, are similarly dextrally sheared. The zone as a whole is intensely crenulated by the secondary fabric (see Plate 5.17), indicating that the dextral shear zone originated at an early stage in the deformation history (syn-D1?). Whilst the northern margin to this shear zone is a brittle fault, the southern margin is gradational, as similar 10's metre wide zones of dextrally sheared phyllonites are interleaved with lower strain striped pelites for some distance to the south. A major NNW-SSE trending fault, extensively mineralised and marked by a linear series of mines across Penhale (Fig. 5.9), prevents further access along the coast. This fault appears to dextrally offset the major E-W shear zone, and as it cannot be seen along strike to the east, it is likely that a further NW-SE trending fault displaces it dextrally south of Holywell Bay (Fig. 5.9). It is possible that the obvious E-W gully which bisects Gull Island [SW 1755, 0595], is a westwards continuation of this large shear zone.

The cliff top exposures at Penhale Point are characterised by highly deformed green/grey pelites, interleaved with narrow zones of black sheared protomylonite. The pelitic fabric, which in places becomes quite phyllonitic, dips steeply north and is intensely crenulated by the NW-SE to NNW-SSE striking secondary fabric. These secondary folds plunge variably to the ENE-WSW, and locally appear to refold a fine lineation of uncertain origin [e.g. SW 1759, 0588]. Stretched xenoliths in frequent highly sheared pale intrusives indicate ~E-W extension [e.g. SW 1758, 0587]. Moving southwards from Penhale Point, the secondary folds give a consistent NE vergence, and again display marked plunge variations. The presence of occasional refolded primary isoclinal suggests that the strain is decreasing. Accompanying this is a reduction in the number of phyllonite zones, which die out altogether on approaching Hoblyn's Cove [SW 1761, 0583]. Here, thick intrusives, bearing WSW plunging stretched pyrite fragments, are folded into metre scale open north verging primary folds. The secondary crenulations here verge north and clearly refold a lineation, which in some of the coarser units appears to be a mineral stretching lineation, plunging gently ESE-WNW around the E-W plunging secondary folds.

From Ligger Point [SW 1756, 0581] to Gravelhill Mine [SW 1764, 0576], a quite monotonous pelitic sequence occurs in which the steeply north dipping main fabric is crenulated by north verging angular secondary folds that appear to have two differently orientated axial planar fabrics, although this is very difficult to prove. At Gravelhill Mine, north verging intrafolial isoclinal are more apparent, and are refolded by somewhat more ductile looking secondary crenulations. Here, the Perran Iron Lode is exposed in a series of NNW-SSE trending caves, which run inland towards a disused quarry and mine openings in Penhale Sands. Although this undoubtedly represents a major fault, the orientation and kinematics of this structure are impossible to assess.



Plate 5.15 (left) Intense late crenulation of the transposed fabric in Meadfoot shales on the south side of Holywell Bay [SW 1764, 0592].

Plate 5.16 (right) Steep north dipping normal fault on south side of Holywell Bay [SW 1760, 0591]. Note that fabric in hangingwall dips steeply north. In the footwall is the dextrally sheared phyllonite (see Plate 5.17).

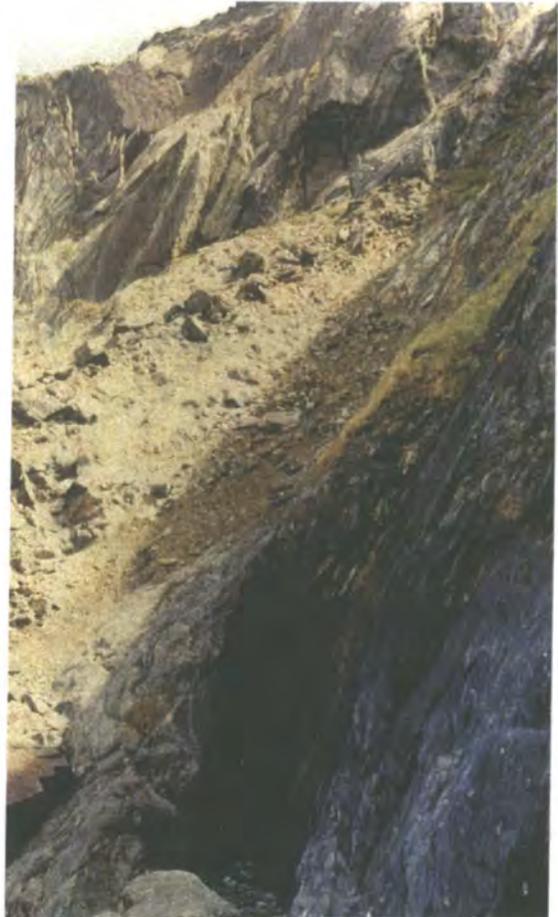




Plate 5.17 Intensely sheared black shales in which quartz veins are sheared into dextrally asymmetric 'augen'. Note the thin buff horizons, which are presumably sheared intrusives, and the late (F3) crenulation of the mylonitic fabric. Protomylonite, Holywell Beach [SW 1760, 0591].

There is no exposure for approximately 800 metres south of Gravelhill Mine, until a single large E-W trending elvan outcrops, cutting through poorly exposed quartz rich pelites. No structure could be recorded in these rubbly outcrops. although the main fabric here dips gently to the SW. Some 600 metres of no exposure occurs from here to Carn Haut [SW 1761, 0566], where foliated and quartz veined metre thick buff psammities and interbedded pale grey pelites of the Gramscatho Group are exposed. Here, the main fabric dips gently to the SSW, and is again crenulated by a NE dipping later cleavage (Plate 5.18), which, on the basis of orientation and style, appears to be the same as that to the north of Perran Sands. These secondary folds verge to the SW, although locally overturn to verge NE in the inverted limbs of metre to 10's metre scale folds. Approximately 100 metres south of Carn Clew [SW 1759, 0555], north closing primary isoclines are seen for the first time in this sequence (Plate 5.19). Rare bedding/cleavage relationships indicate that these gently ESE-WNW plunging folds verge north, but it is impossible to assess the facing in these coastal outcrops. At Cotty's Point [SW 1757, 0551], the strain diminishes and large north verging early folds become more apparent. The primary cleavage here dips very gently to the SSW, although large scale open secondary folding has locally modified this to dip W and NW. The secondary folds at Cotty's Point display sharp antiformal hinges and open synformal hinges (e.g. Sanderson, 1971), although this is only seen at a few localities and is not thought to be representative of the larger scale structure (Plate 5.20). South of Carn Clew, the slates lie within the aureole of the Cligga Head granite, and are variably cordierite spotted, these spots being elongate in a NW-SE direction [e.g. SW 1757, 0552], although this may reflect mimetic growth along intersection fabrics

related to the late crenulation. Spots recording a genuine extension indicate a gently plunging ESE-WNW stretch. Just south of Cotty's Point, is the first appearance of a later, gently south dipping fabric. This crenulates all earlier structures, is thus at least D3 in age, and has been attributed to later backthrusting (Leveridge *et al.* 1990). The evidence at Cotty's Point suggests that extensional displacements along original thrusts are responsible for this late crenulation, e.g. these late crenulations are only seen in the hangingwalls to extensionally reactivated secondary thrusts. There is no evidence for the large scale fabric parallel detachments seen further to the north.



Plate 5.18 Late crenulations of the primary fabric in Perran Bay, which are similar to those seen in Holywell Bay (e.g. see Plate 5.15). Gramscatho shales, Carn Haut [SW 1761, 0566], viewed looking east.



Plate 5.19 North closing primary isoclines refolded by moderately north dipping and south verging F2 folds. Gramscatho shales, Carn Clew [SW 1759, 0555], viewed looking east.



Plate 5.20 Asymmetric F2 folds, with rounded antiformal hinges and sharp synformal hinges. These verge south and locally re-fold north verging and facing primary folds. Gramscatho shales, Cotty's Point [SW 1757, 0551].

Crossing the south end of Perran beach, the quality of outcrop deteriorates rapidly, with the exposures around Chapel Rocks [SW 1755, 0546] and Droskyn Point [SW 1752, 0544] being highly altered buff coloured psammites. Occasional tight secondary folds here dip steeply NE and verge S, whilst the ubiquitous cordierite spots record a gently plunging E-W extension.

The cliff top exposures south of Droskyn Point are generally very poor and are characterised by tight south verging folds of the main cleavage, which dip moderately to steeply to the NE. Limited shoreline access, e.g. below Perran St. George Mine [SW 1746, 0538], shows the main cleavage to be intensely crenulated by a steeply NE dipping fabric, which would appear to be the same as that seen along Perran Bay, and is probably not related to nearby granite emplacement. Stretched cordierite spots again lie parallel to conspicuous fabric intersections. The shales exposed within Hanover Cove [SW 1737, 0532], are too highly altered to yield any useable data, although a crude fabric here dips gently to the S and SSE. In the foreshore beneath Wheal Prudence mine [SW 1732, 0529], centimetre scale tight to isoclinal folds of the main cleavage plunge gently to the E and ESE and verge predominantly southwards, although the strain does appear to be very heterogeneous here. Their axial planar cleavage dips gently to the N and NNE, and clearly crenulates at least one earlier fabric. Occasional examples of folds with somewhat curvilinear hinge lines would appear to be due to displacements on fabric parallel oblique extensional detachments, and it is possible that these folds are late stage, related to granite induced extension.

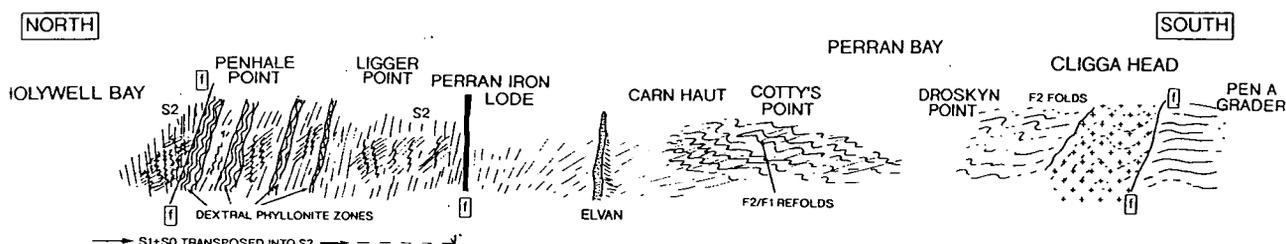


Figure 5.10 Schematic cross-section through Figure 5.9

Summary

South of Holywell Bay, the strain is considerably higher than to the north, with bedding transposed into parallelism with the upright south dipping main cleavage. This cleavage is crenulated by a N and NE dipping secondary fabric, which intensifies southwards to dominate the outcrops (Fig. 5.10). Dextral shear bands are discrete millimetre spaced surfaces with limited offsets, and there is no evidence of large scale extensional structures. Narrow zones of variably plunging secondary folds are not thought to represent sheath folding, and are simply a consequence of intense local shearing (see 5.6.2). A broad zone (~800m) of intensely sheared and quartz veined shales extends for some distance south of Holywell Bay, and appears to represent a long lived E-W dextral shear zone in this region.

Approaching Perran Bay, the strain diminishes, and north verging intrafolial isoclines, similar to those seen north of Holywell Bay, are apparent. The secondary, crenulation rotates to dip variably NNE. After approximately 1.5 kilometres of no exposure, the main fabric at Carn Haut dips gently south, and is crenulated by a NE dipping secondary fabric. At Cotty's Point primary folds verge, and possible face, north, whilst secondary folds verge south. Here, the secondary crenulation steepens to sub-vertical, and cordierite spots indicate ESE-WNW extension. A weak local D3 overprint, possibly related to late extension, is occasionally seen.

The exposures south of Perran Bay are poor, although they preserve a similar structural pattern to that seen further north, and it is thought that emplacement of the Cligga granite has not produced a significant structural overprint, although the effects of hydrothermal alteration are widespread.

5.5.5 Discussion

The coastal exposures of this area display a transition from low strain deformation in the north, to a broad zone of high strain straddling Holywell and Perran Bays. The structure is locally complex, and presents a number of problems, the most significant being the correlation of the post main cleavage age folds. At Towan Head, these structures appear to be brittle, essentially kink type folds, plunging moderately to the SSE. They gradually rotate to plunge gently to the SE along north Holywell Bay, and are locally highly variable in style and orientation. Whilst true N-S kink bands do occur in the north of the area, and were attributed to D4 deformation by Sanderson (1973), these are rare and clearly post-date the late brittle folds. It appears that all of the post main cleavage age folds are of the same generation, and, as they clearly deform the primary cleavage and are locally rotated by ductile shears, they are at least D2 in age. South of Holywell Bay, similar late folds are seen to intensify southwards and crenulate the high strain main fabric. In places, this high strain fabric appears to be simply bedding transposed into the primary cleavage, with tight intrafolial isoclinal folds apparently indicating primary north vergence. However, approaching Penhale it appears that the primary fabric itself has been transposed (e.g. Plate 5.15), and that the main fabric is at least D2 in age, although thin sectioning of critical samples proved to be largely inconclusive. Thus, it appears that there is a very rapid transition from D1 dominated deformation north of Holywell Bay, to D2 dominated deformation in the south. Accompanying this, a marked backsteepening of the main cleavage and a rotation of the late crenulation to a more ESE-WNW orientation occurs. Whilst the variability in orientation and attitude of the late crenulations may be due to the influence of pre-existing anisotropy (e.g. Treagus & Treagus, 1981) and is thus primary, it does also seem that these folds are rotated into parallelism with the intense main fabric in zones of high strain such as Penhale, or south of Porth Joke beach.

At Perran Bay, there is a gradual fall in the strain and a shallowing of the main cleavage moving southwards. The crenulating fabric is locally D2, as it clearly refolds primary isoclinal folds, and thus the high strain zone is restricted to the exposures from Penhale to Gravelhill mine. Unfortunately, the transition, both northwards and southwards, from high D2 strain to D1 strain, is obscured by beach, and it is quite likely that the high strain zone is fault or shear bound.

The marked localisation of high strain and the concomitant structural changes in this area have been variously interpreted (see 5.5.2). Amongst these, Henley (1973) indicated the significance of basement faults by suggesting that reverse movements along the Perran Iron Lode produced the local structural patterns. It is quite possible that the Perran Iron Lode is just one of a series of major ESE-WNW trending basement faults in this area, which acted as buttresses during Variscan shortening to

induce locally anomalous structural patterns in the cover sequence. These anomalies include: a backsteepening of the primary cleavage (from gently south dipping at Towan Head, to steeply north dipping south of Holywell Bay), a tightening of primary folds (from open folds at Towan Head, to intrafolial isoclinal folds at Holywell), a change in the vergence of the secondary folds (N and NE verging at Towan Head, to S and SSE verging in Perran Bay), and the imposition of a strong ESE-WNW structural 'grain' in this area. Accompanying these shortening effects, dextral simple shear appears to have partitioned into linear E-W zones. These are narrow and focused within the high strain zone (e.g. the dextrally sheared phyllonites south of Holywell) and broad and diffuse in the north (e.g. the pervasive dextral shear bands across Pentire Point East). The effects of partitioning into zones of pure and simple shear may be seen in the rotation of structural elements along north Crantock, where the effects of dextral shear increase dramatically. Here, the primary fabric and late crenulations have rotated $\sim 8^\circ$ clockwise, whilst approaching Penhale, the structures rotate apparently anticlockwise into parallelism with the zones of pure shear. This partitioning is locally complicated by pervasive shear bands and considerable evidence for later oblique extensional displacements. Thus, the structure along the northern Gramscatho margin is anomalous with respect to the rest of S. Cornwall, and is characterised by steep structural dips, anomalous lineation orientations and considerable evidence for dextral E-W shear.

5.6 Discussion and evidence for dextral transpression

5.6.1 The northern margin of the Gramscatho Basin

Structure

The northern margin of the Gramscatho Basin, as mapped on sheets 346, 347 and 353/354 (Reid & Scrivenor, 1906, Reid, 1907; Ussher *et al* 1909), forms a gently folded contact running ESE-WNW. On the west coast, it runs conspicuously sub-parallel to the Perran Iron Lode and several thick, laterally extensive felsite dykes (see Fig. 5.9). Towards the east coast, it again runs sub-parallel to large felsite dykes and several mineral lodes, until at Sticker [SX 1980, 0502], it appears to turn SE, entering Mevagissey Bay at Gamas Point (see Fig. 5.2). Evidence cited in 5.4.3, strongly suggests that a major NW-SE trending dextral strike slip fault runs through Pentewan beach. If this proposed fault (the 'Pentewan Fault') is extended northwestwards inland to Sticker, it can be seen that several large felsite dykes terminate abruptly against it. Also, further inland several NW-SE trending mineral lodes lie along the approximate position of the proposed fault, which may then be traced into an abrupt 2 kilometre long offset in the margin of the St. Austell granite. To the east, the St. Austell granite

appears to be dextrally offset along several sub-parallel structures. This suggests that the Pentewan Fault is a major dextral feature, with a significant offset (>2km?).

Mapping of the Mevagissey and St. Austell Bay area has shown that the intense high strain zone noted at Holywell Bay is not seen along this coastal section. Whilst the shales north of Polrudden are quite highly strained and display sheared quartz veins, they nowhere resemble the dextrally sheared phyllonite belt seen on the west coast. However, during the construction of the Sticker bypass, ductile phyllonites with abundant dextral shear criteria were noted in the fresh roadcuts, these approximately coinciding with the Meadfoot/Gramscatho boundary (R. Shail, *pers comm*). Thus it appears that the zone of ductile dextral shear may be traced eastwards as far as Sticker, before being faulted out into Mevagissey Bay along the Pentewan Fault.

On the west coast, a second set of mineral lodes, trending from NE-SW to ENE-WSW, occur within the Gramscatho Basin, some of which apparently cross-cut the ESE-WNW northern margin. Shail (1992) suggests that these are evidence for early Devonian extensional faults, which controlled the basin evolution, and that they may have subsequently influenced the shortening cover sequence. The interaction between these extensional faults and the basin bounding structures would produce complex deformation in the cover sequence. Shail (1992) suggests that within the Gramscatho Basin these ENE-WSW faults would be almost orthogonal to the NNW directed Variscan shortening, and would thus induce backsteepening in the cover sequence. At Perranporth, it appears the ESE-WNW basin bounding faults (e.g. the Perran Iron Lode), which are oblique to the NNW directed Variscan shortening, are dominant, as shown by the strong evidence for dextral shear and the ESE-WNW orientation of the structural grain. Significant structural changes occur across or near to the Perran Iron Lode, indicating that this may be the most important of the basin bounding faults. Thus, at Gravelhill Mine, the high strain fabric dips steeply to the north, whilst at Carn Haut the considerably lower strain fabric dips gently southwards. Also, at Gravelhill Mine, secondary crenulations verge variably north and south, whilst at Carn Haut they consistently verge southwards, suggesting backfolding.

In Mevagissey Bay, it is probable that the ENE-WSW extensional faults are the influential structures, as there is an ENE-WSW grain to the markedly backsteepened structures and only limited evidence of dextral shear. This is compatible with the basin bounding fault here being originally further to the north (e.g. at Sticker) and now faulted out into Mevagissey Bay. The sequence north of Polrudden, which thus lies somewhere to the north of the high strain zone, has a marked ESE-WNW structural orientation, and secondary crenulations (locally D2) are sub-vertical or dip steeply SSW.

Holdsworth (1989a) suggested that the high angle obliquity of the Variscan shortening against the ESE-WNW trending faults, which presumably form the northern margin of the Gramscatho Basin, would induce transpression in the deforming cover sequence, and termed this zone of anomalous structures the Start-Perranporth Line (see 5.6.1 and 5.6.2). The SPL appears to be a broad zone of partitioned strain (Start-Perranporth Zone-Shail, 1992), comprising a central zone of high secondary strain (the dextral phyllonites), flanked by broad external zones of partitioned compression and strike-slip deformation. On the west coast, these structural transitions may be mapped out, although are broken by expanses of obscured critical exposure, whilst on the east coast, most of the central zone of high secondary strain is faulted out to leave the transitional external zones exposed north and south of Pentewan Sands.

Metamorphism and Microstructure

Primmer (1983, 1985) carried out a detailed study of the change in metamorphism across the Perranporth-Pentewan Line, and was able to show, using illite crystallinity and b_0 determinations, that two distinct metamorphic environments exist, with epizone metamorphism in the north and upper anchizone metamorphism in the south. He suggested that the later readjustments of the b_0 parameters in the north were related to a regional D2 backfold in this region, which runs E-W from Cornwall to S. Devon. He also used this to further emphasise the link between the Perranporth-Pentewan Line and the Start boundary, where a marked change in metamorphic grade has long been known (e.g. Tilley, 1923; Marshall, 1965; Robinson, 1981).

Warr *et al.* (1991) used illite crystallinity data to divide S.W. England into six zones, and were able to show that epizone metamorphism, broadly equivalent to greenschist facies, was restricted to the subsurface limits of the S.W. England batholith. This, they suggested, reflected updoming of high grade regionally metamorphosed rocks, rather than contact metamorphic effects. They showed that the metamorphism in N. Cornwall is diastathermal (extension related), due to sedimentary burial and high heat flow, whilst to the south it is related to tectonic burial during northwards directed thrust thickening. Metamorphic grade inversions within the Gramscatho Basin are broadly compatible with a thrust burial cause for metamorphism (Warr *et al.* 1991).

Dating of this metamorphism also serves to show a link between the Start Complex and S. Cornwall. The whole-rock K-Ar ages of Dodson and Rex (1971), now somewhat modified (e.g. Warr *et al.* 1991), indicate a zone of anomalously young ages broadly coincident with the northern Gramscatho margin and the Start boundary. This has been attributed to persistent ductile backfolding (e.g. Coward & McClay, 1983)

although is more probably the result of continued dextrally transpressive reactivation along the SPZ (e.g. Holdsworth, 1989a)

The metamorphism across the SPZ has been comprehensively discussed by Shail (1992), who noted that the recognition of diagnostic mineral assemblages in low grade siliciclastic sedimentary rocks is complex. Barnes and Andrews (1981) have also provided a detailed petrographic study of basic lithologies in the Lizard and Roseland/Gorran areas, and describe mineral assemblages indicative of pumpellyite-actinolite regional grade metamorphism. A study of the deformation microstructures was carried out across the SPZ and representative examples given below. Much of this study centred on thin sandstone horizons, as shale and slates generally yield little deformation microstructural evidence (e.g. see 4.4.5).

North of the high D2 strain zone, the fine sandstones are crudely foliated, this foliation defined by aligned white micas and sub-parallel anastomosing solution seams, occasionally coated in opaque oxides. The quartz grains are all flattened, display a marked undulose extinction, and the larger grains are frequently sub-grained, suggesting that significant amounts of recovery have occurred. Also, sub-grain margins are often fringed with dynamically recrystallised new grains, indicating that higher temperature sub-grain rotation is an important deformation mechanism. There does not appear to be any evidence for significant amounts of grain boundary migration. Evidence for fluid assisted DMT is abundant, with zones of large quartz veins abruptly truncated against seams of polygonal recrystallised quartz aggregates, and zones of fine grained intergrown quartz and white mica. The truncation surfaces, which would appear to represent sites of considerable dissolution, are often coated in opaque oxides, whilst evidence of precipitation of material is given by widespread short stubby pressure shadows of quartz and chlorite, fringing partially altered twinned plagioclases. Further evidence of DMT is seen in some of the finer shales from south of the high strain zone. In these, quartz and white micas are segregated in discrete sub-parallel domains, imparting a strong foliation. Within the quartz domains, there is little evidence of higher temperature deformation, the flattened quartz grains displaying a mild undulose extinction and limited recovery.

Similar shales recovered from within the high strain zone bear evidence of transposed early foliations and late crenulation overprints. The main foliation is strongly developed and has almost completely transposed the previous fabric, with DMT processes producing a marked segregation into quartz and mica domains. Within the quartz domains, there is little evidence of higher temperature deformation, although some of the larger grains do appear to have undergone sub-grain rotation. In the coarser sandstones, however, there is abundant evidence of recovery, and higher temperature dynamic recrystallisation, principally through sub-grain rotation. A

number of samples, principally those recovered from the shear zones, do attest to higher temperature (mid to upper greenschist) dynamic recrystallisation as an important mechanism, whilst the suggestion of higher temperature grain boundary migration may reflect the very high strains localised within the partitioned transpressive zone. In thin section, the dextral phyllonite comprises lensoid aggregates of highly strained and flattened quartz grains, wrapped by intergrown white micas and chlorite (Plate 5.21). The larger relic quartz grains bear considerable evidence of recovery, with dense polygonal aggregates of dynamically recrystallised new grains. Evidence of sub-grain rotation is abundant, and there would also appear to be limited evidence of grain boundary migration. The phyllonite is cut by crudely aligned solution sinuous seams, which define a weak fabric, and attest to the importance of DMT. Thin, discontinuous quartz and/or chlorite veins are evidence of precipitation, as are the chlorite pressure shadows fringing fractured and displaced relic quartz grains. Thus, it would appear that even in the high strain zone some of the lithologies have deformed principally by lower temperature processes, especially fluid assisted DMT, and that the high strain appearance of many of these is simply due to superimposed foliations and marked mineralogical segregation.

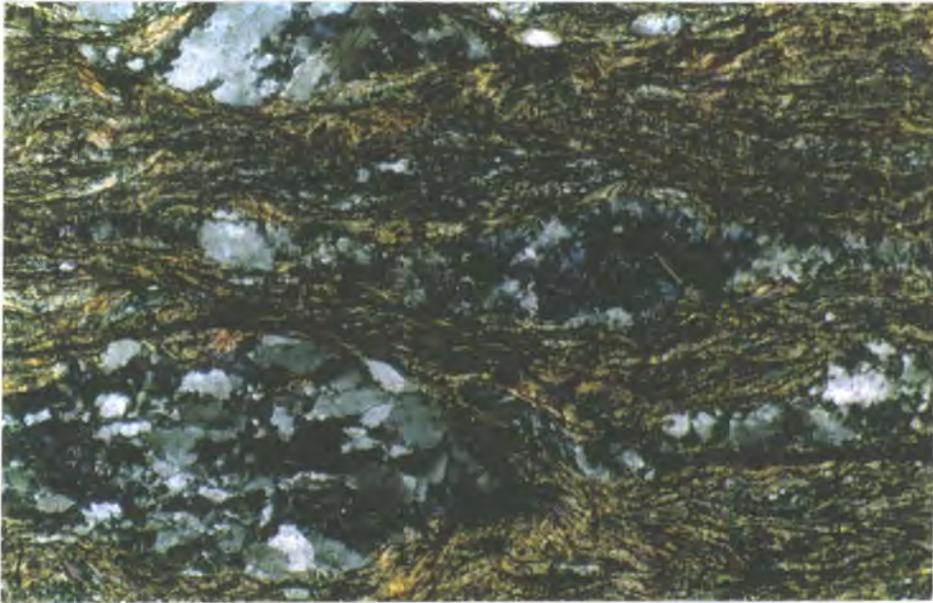


Plate 5.21 Thin section of dextrally sheared phyllonite from the high strain zone on the west coast at Holywell Bay. Relic quartz grains are highly strained and now extensively sub-grained. Dynamic recrystallisation is principally by sub-grain rotation, e.g. large clast at bottom left, whilst limited evidence for grain boundary migration, e.g. clast at top left, attests to higher temperature deformation. Phyllonite, Holywell Bay [SW 1759, 0592]. Under X4 magnification, XP. FoV= 6x3.7mm.

5.6.2 Evidence for Transpression

i) Lineation Switching

A switch in the direction of maximum extension, from regional down-dip to along strike, was recorded in S. Devon (see 4.6.1), and was attributed to transpressive

strains. A similar switch in the extension direction is seen on both the east and west coasts of S. Cornwall. On the east coast, flattened, but unstretched grits, within sandstone units at Chapel Point, indicate that no single extension direction predominates within the XY plane (i.e. $K=0$, oblate strain), although down-dip extension directions have been recorded from further south in the Gorran region. Moving north towards the high strain zone at Mevagissey Harbour, both down-dip and along strike lineations are recorded, again suggesting that these outcrops lie within the 'switching zone', where no single partitioned strain predominates. In the high strain rocks north of Mevagissey Harbour, mineral lineations were difficult to observe, although again deformed coarse grits within sandstone bases indicate approximate E-W extension. The zone of most intense E-W shear appears to have been faulted out, and the transition zone of oblate strain is seen again north of Gerrans Point, where markers within thin intrusives are flattened in S1, but unstretched. Moving northwards, these intrusives bear variably plunging stretched markers, ranging from sub-horizontal (E-W) to sub-vertical, although across the north of St. Austell Bay and into Gribbin Head, all the stretched pyrite and xenoliths clearly indicate down-dip extension (e.g. see Plate 5.14).

On the west coast, the switch is less clear, due to the nature of the outcrop (e.g. highly altered by granite emplacement south of Perranporth), the development of later apparently elongate markers (cordierite spots growing mimetically along fabric intersections at Perran Bay), and the presence of well developed crenulation lineations of various generations. However, approximate along strike lineations were recorded from Crantock southwards to the high strain zone, usually within pre-orogenic intrusives. Evidence from Pentire Point West indicates that the intrusives are zones of weakness, with late extension focusing into these. These later displacements are not the cause of the lineations orientations, as displacement related fibres here are markedly oblique to the recorded extension directions. The fact that these intrusives are zones of weakness (e.g. some of the intrusives in S. Devon are reduced to calc mylonite by foliation parallel shear), suggests that they will be a focus for strain, and would therefore be expected to record the switch from shortening to strike-slip dominated deformation (as seen along western St. Austell Bay and north of Challaborough, S. Devon).

The possible mechanisms for orogen parallel extension are discussed in 4.6.1, and, once again, it appears that the evidence from S. Cornwall is consistent with a transpressive switch in the maximum finite extension, due to the partitioning of transpressive strain components.

ii) Refolded lineations

In S. Devon, it was noted that the measurement of linear data from slates/shales is liable to suffer from a strong azimuth bias (see 4.6.1), due to the two dimensional nature of the outcrop. This problem is also encountered in S. Cornwall, although within the thicker bedded Gramscatho lithologies and the high strain 'schists', three dimensional exposure did allow an examination of the lineation/fold relationships.

On the east coast, very little evidence of lineation obliquity was seen. At Rowards Quay, the mineral lineation, often overprinted by fabric intersections, appears to lie sub-parallel to the metre scale F2 folds, whilst just to the north, mineral stretching lineations on S2 cleavage surfaces lie sub-parallel to the S1/S2 intersection. At Polkerris, however, there is evidence from many of the well exposed F2 folds for a small degree of clockwise lineation obliquity, this rarely exceeding 10° (e.g. Plate 5.10). This may suggest a clockwise rotation of the primary lineation, due to partitioned dextral shear, prior to refolding by F2. The occasional F2 folds recorded across St. Austell Bay, did not appear to bear any refolded lineations.

On the west coast, the lineations are much more variable in orientation, and no consistent relationship could be established. This is, in part, due to the extreme variability of the secondary folds/crenulations in the northern part of the area, and, as yet, no satisfactory mechanism can be offered to account for the initiation of secondary folds in these anomalous orientations. However, between Penhale Point and Gravelhill Mine, where secondary crenulations of the transposed fabric are dominant, there is evidence of a consistent clockwise lineation relationship, although overprinting by the ubiquitous crenulation lineation is extensive. Although the obliquity is small, indicating limited dextral shear, there is considerable evidence from this area of intense long lived dextral shear.

The transpressive model of Sanderson and Marchini (1984), shows that the maximum incremental extension direction switches within the XY-plane (slaty cleavage), and thus the sub-horizontal lineations in S. Cornwall will lie within the main cleavage. As the orientation of this cleavage appears to be controlled by the orientation of underlying local basement faults, the azimuth of the gently plunging lineations in S. Cornwall will be locally variable, e.g. ENE-WSW plunges in Mevagissey Bay, ESE-WNW plunges in Perran Bay.

Thus, in summary, the relationship between late folds and the primary lineation across the SPZ is variable, and only in a limited number of localities may a consistent clockwise relationship be seen. This reflects the variability of late fold orientations, the ubiquitous overprinting crenulation lineation, a probable initial curvature of the lineation (e.g. see 4.6.2), and the effects of measuring linear data in two dimensional

surfaces. However, evidence for early dextral shear is abundant, and the probability of primary structural elements being rotated clockwise prior to refolding is not in doubt.

iii) Sheath folds

Sheath folds are common along the high strain zone straddling the Start Boundary Fault in S. Devon (4.6.4), and it is reasonable to expect similar structures to occur within the high strain zone of the SPZ in S. Cornwall. Evidence for markedly curvilinear fold hinges is scant, however, and no true 'eyed' sheath folds were observed.

On the east coast, a zone of tight, steeply plunging folds and folded quartz veins, was seen on Benny Island, at Mevagissey Harbour. Here, the fold vergence patterns are not consistent with hinge line curvilinearity, and no examples of folds plunging through the vertical could be located. However, these folds plot consistently within a sub-vertical ENE-WSW trending plane, which suggests that there is a control on their orientation (Fig. 5.11). Similar zones of steeply plunging folds are also seen south of Holywell Bay (see 5.5.4), and again here vergence patterns, refolded lineation relationships and three dimensional examination of the outcrops suggests that they are not a consequence of marked hinge line curvilinearity. One obvious controlling factor is the proximity of large, quartz veined shears, cutting through the outcrops and quite probably locally reorientating originally gently plunging structures. The rotational models of Fossen and Tikoff (1993; see Fig. 1.10) show that, where a high degree of pure shear shortening is superimposed on strike-slip simple shear, gently plunging fold axes rotate towards a vertical plunge, rather than rotating into parallelism with the shear zone. It appears that the structures observed at Holywell Bay are consistent with this model (see Fig. 5.12), and thus these locally reorientated folds may be suggestive of transpressive strains. The structures on Benny Island are more problematic, possibly reflecting intense pure shear superimposed on simple shear.

The lack of obvious sheath folds on the east coast may simply be due to the high strain zone being faulted out. On the west coast, however, the high strain zone is well exposed and, given the evidence for intense ~E-W dextral shear, it is reasonable to expect to see sheath folding. It is probable, by analogy with the SBF, that the folds are localised along the obvious surface expression of the controlling basement faults, which here appears to be the Perran Iron Lode. The southern margin of this fault is obscured by ~1.5 kilometres of beach, whilst to the north, the dextral shear has partitioned into very narrow, and very high strain, phyllonite zones, where it is unlikely that folds would be preserved.

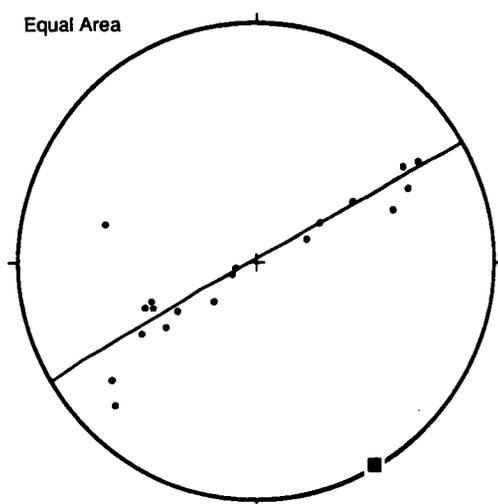
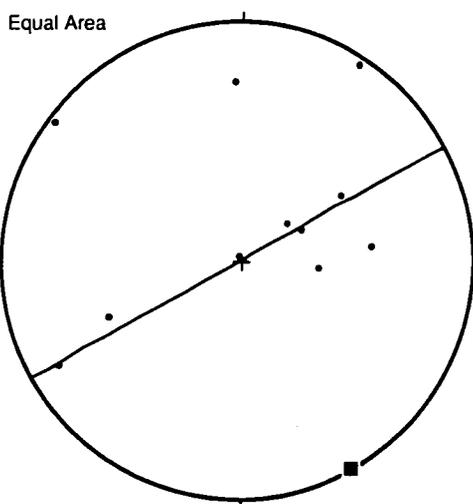


Figure 5.11 (left) Equal area stereograms of steeply plunging folds, which all lie within a sub-vertical plane (240/89N), in the high strain zone at Mevagissey Bay, Benny Island [SX 2018, 0449].

Figure 5.12 (right) Equal area stereograms of steeply plunging folds from the high strain zone at Holywell Bay. Folds all fall within a sub-vertical zone (241/89N) and appear to have been rotated from the horizontal into the steeper orientations (see also Fig. 1.10; Fossen & Tickoff, 1993).



iv) Shear bands

As noted in 5.4, shear bands are poorly developed within the Mevagissey and St. Austell Bay areas, although there is an obvious increase in their frequency on approaching the high strain zone at Mevagissey Harbour. Dextral shear bands here strike, on average, $110/290^\circ$ and predominantly dip steeply to the NNE (Fig. 5.13a). Sinistral shear bands are rare, isolated examples, although generally striking NNE-SSW, and dipping steeply both east and west. Virtually none of the east coast shears bore displacement fibres, and those which were measured indicate only a slight dip-slip component.

On the west coast, shear bands are very well developed and in certain areas, e.g. Crantock, they completely shred the outcrop. Here, dextral shears, which outnumber sinistral by almost 5:1, strike, on average, $130/310^\circ$, and dip steeply SW (Fig. 5.13b). Sinistral shear bands strike, on average, $027/207^\circ$, and predominantly dip moderately to steeply to the ESE (Fig. 5.13c). The obtuse bisectrix, as calculated by stereographic projection and which represents the direction of maximum shortening (Variscan transport), trends 347° . This is very similar to the orientation obtained for S. Devon, so it would appear that there is no significant change in the direction of

regional transport. The smaller angle subtended by the shears in S. Cornwall, compared to their S. Devon equivalents, indicates that rotation (both active and passive) may be less significant. The long lived nature of the dextral shears, overlapping with late extensional collapse, does suggest that active rotation should be important, although the histograms of shear distributions from key localities are symmetrical. The lithological control over shear rotation, observed in the Start Complex, does not apply in S. Cornwall.

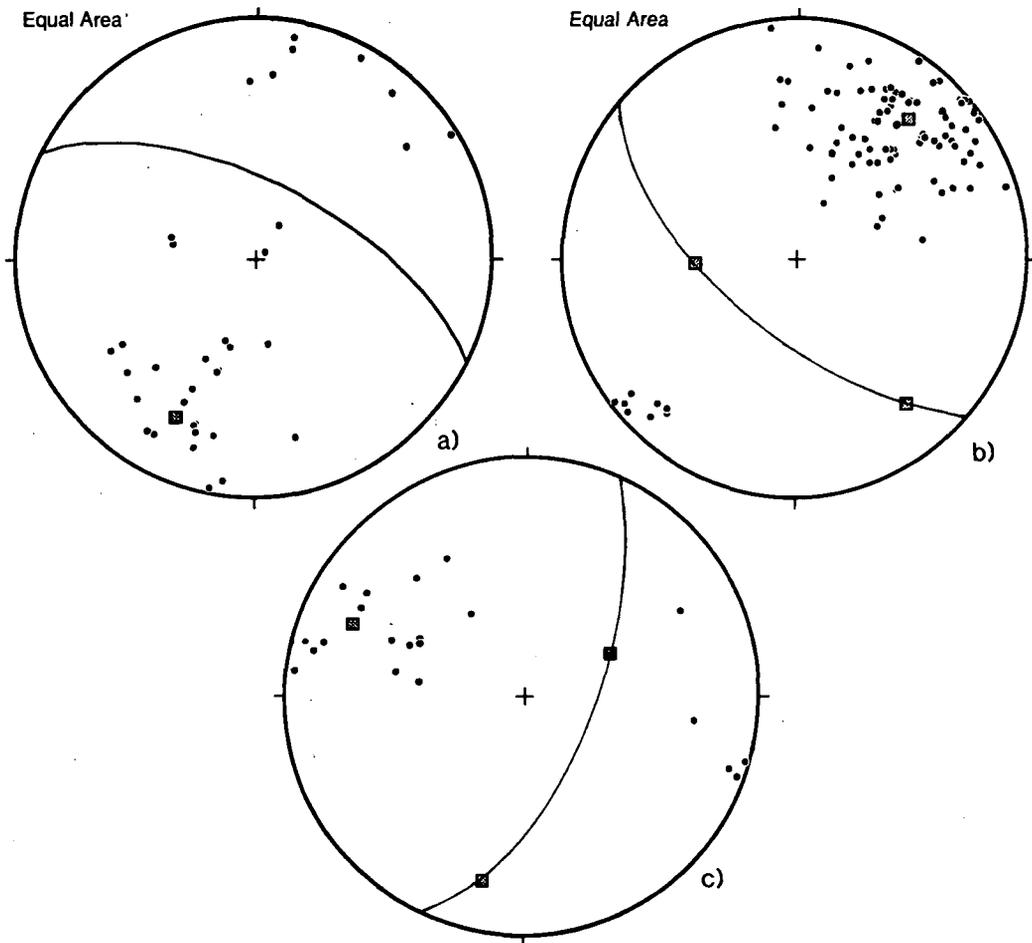


Figure 5.13 Equal area stereograms of shear bands within the Start-Perranporth Zone: a) dextral shear bands from Mevagissey and St. Austell Bays, mean shear plane 297/64N; b) dextral shear bands from the west coast, mean shear plane 130/63SW; c) sinistral shear bands from the west coast, mean shear plane is 027/69SE.

It appears that the dip and azimuth of the dextral shear bands is largely controlled by the orientation of the main cleavage, in that the east coast shears all dip NNE, as does the dominant fabric across much of Mevagissey and St. Austell Bays, whilst on the west coast, the shears dip SW, sub-parallel to the main fabric (see Figs. 5.2 & 5.7).

The shears bands exposed on the west coast display excellent displacement fibres, which indicate dextral oblique offsets. Similar dextral oblique offsets are seen in the larger fabric parallel extensional detachments (see 5.6.3 below), indicating that

dextral shear is long lived, and that extensional movements initiated whilst dextral shear was still going on. Thus, the extension is not significantly post-orogenic. Although detailed mapping was not carried out across Watergate Bay, reconnaissance mapping here showed that the extensional detachments here have almost pure dip-slip fibres, and that tracing these southwards, the degree of dextral displacement increases, ultimately dominating at and south of Crantock beach (Fig. 5.14). The rotation of displacement fibres clearly demonstrates the localised nature of the dextral shear, and again may be used to constrain the zone of dextral shear. The presence of these fibres, on both the dextral and sinistral shears, allowed their geometrical relationships to be tested by stereographic projection (Fig. 5.15). The mean shear planes, with their respective fibre pitches, subtend an angle of 88° , the bisectrix of which trends 347° (see above). The intersection between the mean planes is the XZ plane, within which the displacement fibres should lie, assuming the structures are related. It can be seen on Figure 5.15 that the mean recorded dextral shear band fibre plunges $30/295$, whilst the intersection between this surface and the XZ plane plunges $24/298$, strongly suggesting that the structures are related. Very few displacement fibres were recorded on sinistral shears, although the data was largely consistent with the above conclusion. This confirms the impression from S. Devon and S. Cornwall that these shears are coeval and related to the same bulk strain.

Evidence from S. Devon (e.g. refolded lineations), indicates that dextral shear becomes important immediately post primary folding, and the evidence from the west Cornish coast appears to confirm this, e.g. post main cleavage age folds crenulating the high strain dextral shear zones.

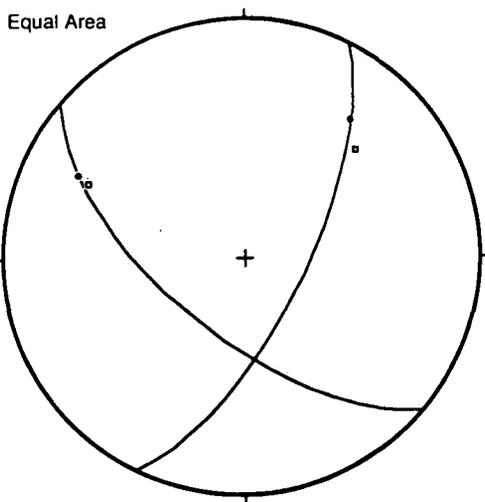


Figure 5.15 Summary equal area stereogram of the shears and displacement fibres from the west Cornish coast. The mean recorded displacement fibres (open squares) approximately coincide with the expected position (dots), as calculated from planar intersections (see text for details).

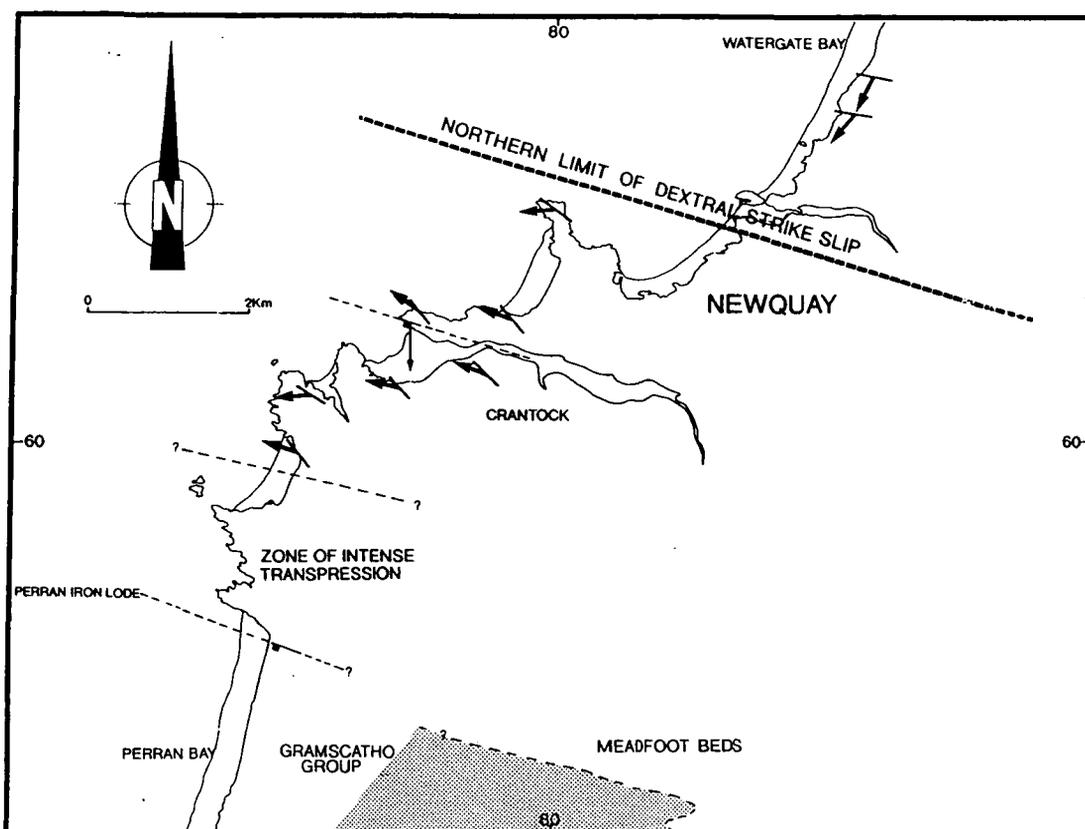


Figure 5.14 Map of the Newquay/Perran Bay area. Slickenfibres along brittle/ductile shear band C-surfaces indicate a change in the displacement from dip-slip in the north, to dextral oblique slip in the south, on approaching the zone of most intense dextral transpression.

5.6.3 Post-orogenic extensional collapse

The evidence for post-orogenic extensional collapse is becoming increasingly recognised within S. Cornwall (e.g. Holdsworth *et al.* 1992), and, as it is the subject of detailed ongoing research (A. Styles, Durham; A. Alexander, C.S.M), it will only be briefly discussed here.

All the coastal sections mapped across the SPZ in Cornwall bear evidence of late extension, this often occurring along foliation parallel detachments, varying from centimetre spaced up to several metre spaced. Kinematic indicators within these, such as downslope verging ductile folds, downslope closing sheath folds, imbricate systems of shear bands (all presumably related to second-order compressional and extensional tip strains during displacement propagation) and displacement fibres, suggest down-dip extensional displacements. The ductile folds associated with these structures may locally confuse deformation chronologies, especially where the discrete bounding surfaces are not readily apparent (e.g. see Plate 5.13). Also, rotations associated with

extensional displacements along these surfaces can reorientate the pre-existing structure, and in places effectively preclude meaningful measurement of structure (e.g. Carrickowel Point & Landrion Point).

Along the east coast, the extension appears to be mainly dip-slip (*cf.* the west coast), and its timing is accurately constrained by reactivated compressional structures. For example, across Polgwyn beach, southwest of Black Head, large D2 thrusts with quartz slickensides and displacement fibres have displaced pre-orogenic intrusives, which now have a net normal offset. This is confirmed by overprinted fibres, which indicate D2 thrusting towards 350° , followed by considerable extension towards 264° . Further east at Black Head, the extensional detachments display variable offsets according to orientation. Here, the faults vary from NNE-SSW trending, with variable degrees of sinistral oblique fibres, through to ENE-WSW trending, with pure dip-slip fibres (Fig. 5.16). This gradual change in the extensional displacement with orientation suggests that NNE-SSW directed sinistral shear, presumably related to the overall dextral shear regime, is important during extension. On the west coast, long-lived dextral shear is clearly involved in later extension (e.g. Fig. 5.14), and thus it appears that these strike-slip movements 'straddle' the transition from compression to extension. Indeed, evidence suggests that extension has begun quite early during the deformation, e.g. syn- to post- local D2, and thus these structures are not strictly post-orogenic.

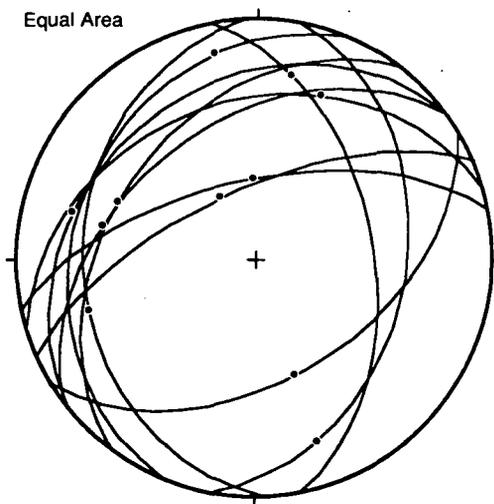


Figure 5.16 Equal area stereogram of late extensional faults at Black Head. Displacements vary from sinistral strike-slip (NNE-SSW trending), through to dip-slip (ENE-WSW trending).

It appears that extensional detachments are highly influenced by pre-existing anisotropy, and thus the dominant cleavage controls their siting along the east coast, whilst it appears that the pervasive dextral shear bands influence the siting of extension along the west coast. In places, the low angle detachments are cut by steeper, demonstrably normal faults, which appear to be related to later stages of extension.

Preliminary investigations (e.g. Holdsworth *et al.* 1992) suggest that early crustal collapse may be linked to magma ascent, with extension possibly creating the space for large scale emplacement of late-post tectonic granitoid magma. This is a difficult concept to test in S. Cornwall, where the individual granite plutons have created locally complex structural overprints. Certainly, evidence from the western Cornish coast does indicate that extension is important quite early on in the deformation history, and could quite probably facilitate granitoid emplacement.

The increased recognition of significant extension in S. Cornwall necessitates a review of traditionally accepted deformation chronologies, in which all structures are directly attributable to crustal thickening processes. Also, the detailed mapping of major zones of extension may lead to a greater understanding of the distribution of metamorphic zones across S.W. England.

CHAPTER 6

SUMMARY AND CONCLUSIONS: THE EVOLUTION OF THE START-PERRANPORTH ZONE

6.1 Introduction

The structural pattern of S.W. England, while superficially resembling a simple fold and thrust belt, has many local complexities, thought to reflect the influence of the pre-existing basin architecture. The region is characterised by NNW verging primary folds, whose axial planes dip gently to the SSE. A mineral stretching lineation lies down dip of the primary cleavage, approximately parallel to Variscan transport. Within this overall pattern there are a number of approximately E-W trending linear zones of complexity, including the Tintagel High Strain Zone (Sanderson & Dearman, 1973), the Padstow Facing Confrontation (Andrews *et al.* 1988) and the Start-Perranporth Line (Holdsworth, 1989a; Start-Perranporth Zone, Shail, 1992)). All of these structures are thought to reflect the influence of the pre-existing basin architecture on the deforming basinal sequences. They attest to the importance of major E-W trending faults which control the siting, evolution and inversion of the S.W. England basins. Of these zones, the SPZ is the most laterally extensive and probably the most significant. As defined by Holdsworth (1989a), it straddles the SBF in S. Devon and continues westwards along strike to approximately coincide with the northern margin of the Gramscatho Basin in S. Cornwall. These two areas have been examined in great detail, and will be discussed individually below, prior to a regional synthesis.

6.2 The Start Complex

The SPZ in S. Devon coincides with the Start Boundary Fault, which separates the Start Complex from the Lower Devonian shallow marine mudrocks to the north. The Start Complex comprises an interlayered sequence of highly altered green and grey schists, which are petrologically very uniform and which are all highly deformed, primary fabrics having been overprinted by extensive later recrystallisation and tectonism.

The age of the Start Complex is uncertain, although the strong geochemical similarities between the Start greenschists and the Lizard dykes (*c.* 375Ma; Davis, 1984), which both have a strongly depleted N-type MORB signature, suggests that they are of similar age and derivation (Floyd *et al.* 1993a). It appears, based on their thickness, lateral persistence and uniform mineralogical composition that the greenschists were originally either basaltic volcanoclastics or lava flows, although their finely laminated appearance is strongly suggestive of the former. These metavolcanics may represent incipient ocean crust flooring a small basin (the 'Start Basin'), in which a thick pelitic sequence accumulated (the greyschists). Geochemically, the greyschists are broadly similar to other Rhenohercynian basinal sequences (e.g. Gramscatho, Giessen) in terms of tectonic environment, further emphasising the link between the Start Complex and S. Cornwall.

The Start Basin was subsequently inverted and deformed during Variscan orogenesis. Whilst the Start schists and the Devonian sequence to the north share a common deformation history, the metamorphic segregation fabric in the Start Complex predates the first phase of folds common to both, and may reflect sub-oceanic ridge related deformation, similar to that observed in the Lizard (e.g. Vearnecombe, 1979).

Some controversy surrounds the large scale structure of the Start Complex and its deformation history relative to the rest of South Hams. Early workers (e.g. Marshall, 1962) recognised two main fold phases, which were directly correlated with those of the Devonian sequence (e.g. Hobson, 1977), and which are cut by later recumbent folds and kink bands, apparently of regional significance. The overall structure was interpreted as a large F2 antiform plunging and closing westwards through Bolt Tail. However, this study has confirmed the ideas of Seago and Chapman (1988), and proven the existence of a unique early deformation fabric in the Start schists, which has no direct equivalent in the Devonian. The later recumbent folds are shown to be a consequence of local strain heterogeneity. These are only seen in complexly deformed greyschists sequences, cannot be proven to refold earlier structures, and may not be reasonably correlated with the recumbent folds of S. Cornwall (*cf.* Hobson, 1977). The overall structure is reinterpreted as an F3 fold, closing westwards through Bolberry Down.

The small scale structures of the Start Complex, e.g. mineral lineations, minor fold axes, pervasive brittle/ductile shear bands, sheath folds, etc. are all consistent with dextral simple shear being superimposed on the shortening during deformation, a pattern also seen in the Devonian shales to the north. This superimposition produces a focussed E-W trending zone of dextral transpression between 5-8 kilometres wide. Further evidence, e.g. lineations switching from regional down-dip to locally along-

strike, confirm that partitioned dextral transpression is the predominant deformation mechanism in this region, and that it appears to localise along the Start Boundary Fault. This structure is manifest as a series of steep north dipping normal faults, intimately associated with large volumes of highly altered and sheared 'Torcross' type intrusive. Fault rocks indicate a long movement history, with mylonitisation followed by several brecciation phases.

Adjacent to the SBF, regional fabrics, which normally dip S/SSE, are steepened to the vertical, main phase folds tighten and the along strike shear strain intensifies, all suggesting that the SBF acted as a buttress to the deforming cover sequence, and that it is only the surface representation of a much larger and significant basement fault (e.g. Holdsworth, 1989a). Gravity data (e.g. Bott & Scott, 1964) is consistent with a series of steep faults in this area, and further serve to disprove the theory that the SBF is a backsteepened ductile thrust (e.g. Coward & McClay, 1983). The regional implications of this structure are discussed in 6.4.

6.3 The northern Gramscatho margin

The northern margin of the Gramscatho Basin is characterised by a sedimentary transition from coarse, turbiditic Gramscatho sandstones, to the finely interbedded green and grey Meadfoot shales and mudstones. On the west coast, this transition is largely obscured by Perran Sands, whilst on the east coast at Gamas Head a gradual transition may be seen, where Meadfoot shales are interbedded with and overly the coarse Treworgans sandstones of the parautochthonous Porthtowan Formation. These coarse grained marginal facies, of probable Late Emsian age, have been interpreted as channel sands, transitional to a deep water marine fan environment (e.g. Shail, 1992). The stratigraphic relationships at Gamas Head suggest that the Meadfoot shales overly, and are thus apparently younger than, the Gramscatho sandstones. However, this is probably a consequence of an interdigitated margin between the palaeontologically dated older Meadfoot Group and the younger Gramscatho Group.

Recent thin-skinned fold and thrust models for S. Cornwall (e.g. Leveridge *et al.* 1984; Holder & Leveridge, 1986b; Leveridge *et al.* 1990) have invoked the presence of a major thrust, the Carrick Thrust, to account for the stratigraphic relationships observed along the east coast. The exact course of this structure is not specified, and it does not obviously emerge at Pentewan. The change in sedimentology and structure across Pentewan beach is, however, suggestive of a fault here, and it is highly likely that the Carrick Thrust is faulted out along the proposed Pentewan fault.

Deformation on both coasts is characterised by a backsteepening of the primary fabrics on approaching the Gramscatho margin, accompanied by a tightening of primary folds, the intensification of later crenulations, a switch in the finite extension direction, a change in vergence of secondary folds from north verging in the north to south verging in the south, and an increase in the evidence for ductile dextral shear. Whilst some of small scale structural relationships observed in the Start Complex (e.g. refold relationships of folds, clockwise obliquity of mineral lineations, etc.) are rarely seen, or are difficult to prove in S. Cornwall, the evidence is still consistent with partitioned dextral transpression being the dominant deformation mechanism. However, in S. Cornwall the structure is more cryptic and the deformation, appears to have partitioned into discrete zones (e.g. the dextral phyllonite zone at Holywell Bay). Also, the obvious surface faults seen in S. Devon (e.g. the SBF) do not occur across S. Cornwall, although many of the mineral lodes probably reflect significant basement fractures (e.g. Perran Iron Lode). This, allied to the extensive breaks in critical exposure (e.g. Perran Sands), and the effects of later NW-SE dextral strike-slip faults (e.g. Pentewan Fault) means that the very high strain features characteristic of the immediate SBZ in S. Devon, are not so well displayed in S. Cornwall. Again it appears that the local structural complexity along the Gramscatho margin reflects the influence of basement faults of various orientations (see 6.4).

6.4 The evolution of the Start-Perranporth Zone

Evidence presented above suggests that the Start schists and the Gramscatho Group are along strike equivalents, and represent thick basinal sequences deposited south of a major basin controlling fault of regional significance. Whilst some workers (e.g. Franke *et al.* 1989) suggest that these units were deposited in a single Late Silurian to Early Devonian 'Cornwall Basin', it is more likely that a sequence of smaller basins existed. The suggestion that NW-SE trending dextral strike slip faults acted as oblique linking fault segments (e.g. Holdsworth, 1989a) may not be the case as these faults are usually considered to be later features, although they may possibly reflect reactivated earlier structures. The exact position and areal extent of the basins is still uncertain. The limited offshore extent of the Start type schists (e.g. Fig. 3.6) does suggest that the Start Basin was relatively small, a fact borne out by the limited thickness of incipient ocean floor in this area (the greenschists) relative to that in S. Cornwall (the Lizard). Evidence from S. Cornwall suggests that the Gramscatho Basin was similarly restricted in size, e.g. the isotopic data from the Lizard (Davis, 1984) suggests restricted development of oceanic crust, whilst the provenance and

sedimentology of the Gramscatho Group also indicates a basin of limited size (Shail, 1992).

The development of these basins is also uncertain, with three larger scale models having been offered to account for the evolution of S. Cornwall. These are: a) back-arc rifting related to northerly directed subduction (e.g. Leeder, 1982), b) an intracratonic strike-slip system in which transtensional basins developed (e.g. Barnes & Andrews, 1986; Holdsworth, 1989a), and c) an ocean basin with southerly directed subduction beneath an active arc (e.g. Holder & Leveridge, 1986b). Shail (1992) rejected the latter fore-arc model as the arc material is probably pre-Devonian in age. He also noted that it is difficult to test whether the Gramscatho Basin (and the Start Basin) is the product of transtensional opening, since the model of Barnes and Andrews (1986) is largely dependant on the orientation of the Lizard dykes, and the observed orientations of basic volcanics in the Mylor Slate Formation indicate that dextral transtension would be unlikely during that later basin history. It is unlikely that unequivocal evidence for the original basin configuration will emerge from the Start Complex, and the evidence from S. Cornwall is inconclusive. However, the presence of a single steep basement fault, which clearly forms the northern margin to the Start and Gramscatho Basins, and which has exerted a profound influence during inversion, does tend to favour the intracratonic strike-slip model of Barnes and Andrews (1986), rather than orthogonal extensional models. Whilst the shear sense of this system of transtensional basins remains uncertain, the geometry of the inversion is more readily understood.

Holdsworth (1989a) invoked a model of oblique fault buttressing (E-W) during NNW directed Variscan shortening to account for the anomalous structural features along the SPZ. He envisaged a narrow central zone of D2 dominated deformation, characterised by intense along strike shear and the local development of sheath folds, flanked by external zones of lower strain transitional to the regional structural pattern. This study has shown that whilst this model is largely applicable to the structures straddling the SBF, it does not appear to apply so well in S. Cornwall. On the east and west Cornish coasts the switch in the finite extension directions is less well constrained, sheath folds are largely lacking, and areas of critical exposure are either obscured or faulted out. The dextral offset of the SPZ across S. Cornwall (see Holdsworth, 1989a-Fig. 1) reflects dextral strike-slip displacements along the Pentewan fault, and related sub-parallel structures. The course of the SPZ across S. Cornwall is well constrained from Pentewan, through Sticker to the west Cornish coast (see 5.6.1).

The suggestion that the major deformation controlling fault represents a Devonian terrane boundary (e.g. Holdsworth, 1989a) is to some extent incorrect, as there is clear sedimentary continuity across it during the Devonian (Shail, 1992), although it is quite probably pre-Devonian in age. The presence of this structure, and the dextral transpression it induces in the cover sequence, provides the best explanation for the observed small scale structures and transitions through S. Devon and S. Cornwall, e.g. switching the maximum finite extension directions, changes in secondary fold vergence, backsteepening of primary fabrics, tightening of primary folds, etc. It also best accounts for the remarkable along strike consistency of this anomalous zone.

REFERENCES

- AHRENDT, H., CLAUER, N., HUNZIKER, J.C. & WEBER, K. 1983. Migration of folding and metamorphism in the Rheinische Schiefergebirge deduced from K-Ar and Rb-Sr age determinations. *In*: MARTIN, H. & EDER, F.W. (eds) *Intracontinental Fold Belts*. Springer Verlag, 323-338
- ALLEN, J.R.L. 1979. Old Red Sandstone in external basins, with particular reference to Southern Britain. *In*: HOUSE, M.R., SCRUTTON, C.T. & BASSETT, M.G. (eds) *The Devonian System*. Special Paper in Palaeontology, **23**, 65-80.
- ANDERTON, R., BRIDGES, P.H., LEEDER, M.R. & SELWOOD, B.W. 1979. *A dynamic stratigraphy of the British Isles*. George Allen & Unwin, London.
- ANDREWS, J. R., BARKER, A. J. & PAMPLIN, C. F. 1988. A reappraisal of the facing confrontation in north Cornwall: fold- or thrust- dominated tectonics. *Jl. geol. Soc. Lond.*, **145**, 777-788.
- AYDIN, A. & NUR, A. 1985. The types and roles of step-overs in strike-slip tectonics. *In*: BIDDLE, K.T. & CHRISTIE-BLICK, N. (eds) *Strike-slip deformation, basin formation and sedimentation*. Society of Economic Palaeontologists and Mineralogists Special Publication, **37**, 35-43.
- BADHAM, J.P.N. 1982. Strike-slip orogens-an explanation for the Hercynides. *Jl. geol. Soc. Lond.*, **139**, 493-504.
- BADHAM, J.P.N. & HALLS, C. 1975. Microplate tectonics, oblique collisions and the evolution of orogenic systems. *Geology*, **3**, 373-376.
- BARNES, R.P. 1983. The stratigraphy of a sedimentary melange and associated deposits in south Cornwall, England. *Proc. Geol. Ass.* **94**, 219-229
- BARNES, R.P. 1984. Possible Lizard-derived material in the underlying Meneage Formation. *Jl. geol. Soc. Lond.*, **141**, 79-85.
- BARNES, R.P., ANDREWS, J.R. & BADHAM, J.P.N. 1979. Preliminary investigations of south Cornish mélanges. *Proc. Ussher Soc.* **4**, 262-268.
- BARNES, R.P. & ANDREWS, J.R. 1981. Pumpellyite-actinolite regional grade metamorphism in south Cornwall. *Proc. Ussher Soc.* **5**, 139-146
- BARNES, R.P. & ANDREWS, J.R. 1984. Hot or cold emplacement of the Lizard Complex? *Jl. geol. Soc. Lond.*, **141**, 37-40.
- BARNES, R.P. & ANDREWS, J.R. 1986. Upper Palaeozoic ophiolite generation and obduction in south Cornwall. *Jl. geol. Soc. Lond.*, **143**, 197-124.
- BEESE, AP. 1982. The argillite facies of the Middle Devonian succession in north Cornwall. *Proc. Ussher Soc.* **5**, 321-323.
- BEESE, A.P. 1984. Stratigraphy of the Upper Devonian argillite succession in north Cornwall. *Geol. Mag.* **121**, 61-69.
- BELL, A.M. 1981. Vergence: an evaluation. *J. Struct. Geol.*, **3**, 197-202.
- BERTHÉ, D., CHOUKROUNE, P. & JEGOUZO, P. 1979. Orthogneiss, mylonite and non-coaxial deformation of granite: the example of the South Armorican Shear Zone. *J. Struct. Geol.*, **1**, 31-42.
- BIRPS & ECORS 1986. Deep seismic reflection profiling between England, France and Ireland. *Jl. geol. Soc. Lond.*, **143**, 45-52.
- BLUCK, B.J., HAUGHTON, P.D.W., HOUSE, M.R., SELWOOD, E.B. & TUNBRIDGE, I.P. 1988. Devonian of England, Wales and Scotland. *In*: McMILLAN, N.J., EMBRAY, A.F. & GLASS, D.J. (eds) *Devonian of the World*. Memoir, Canadian Society of Petroleum Geologists, **14**, 305-324.

- BONNEY, T.G. 1884. On the Geology of the South Devon Coast from Torcross to Hope Cove. *Jl. geol. Soc. Lond.*, **40**, 1-26.
- BORRADAILE, G.J. 1978. Transected folds: a study illustrated with examples from Canada and Scotland. *Bull. geol. Soc. Am.*, **89**, 481-493.
- BOTT, M.H.P., DAY, A.A. & MASSON-SMITH, D. 1958. The geological interpretation of gravity and magnetic surveys in Devon and Cornwall. *Phil. Trans. R. Soc. Lond.*, **251A**, 161-191.
- BOTT, M.H.P. & SCOTT, P. 1964. Recent geophysical studies in south-west England. In: HOSKING, K.F.G. & SHRIMPTON, G.J. (eds). *Present views on some Aspects of the Geology of Cornwall*. Royal Geological Society of Cornwall, 25-44.
- BRETZ, J.H. 1940. Solution cavities in the Joliet limestone of northeastern Illinois. *J. Geol.*, **50**, 675-711.
- BROOKS, M., DOODY, J.J. & AL-RAWI, F.R.J. 1984. Major crustal reflectors beneath SW England. *Jl. geol. Soc. Lond.*, **141**, 97-103.
- BROOKS, M. & LE GALL, B. 1992. Discussion on the crustal evolutionary model for the Variscides of Ireland and Wales from SWAT seismic data. *Jl. geol. Soc. Lond.*, **149**, 681-682.
- BURG, J.P., BALE, P., BRUN, J.P. & GIRARDEAU, J. 1987. Stretching lineation and transport direction in the Ibero-Armorican arc during the siluro-devonian collision. *Geodynamica Acta*, **1**, 71-87.
- BURTON, C.G. & TANNER, P.W.G. 1986. The stratigraphy and structure of the Devonian rocks around Liskeard, east Cornwall, with regional implications. *Jl. geol. Soc. Lond.*, **143**, 95-105.
- BUTLER, R.W.H. 1989. The influence of pre-existing basin structure on thrust system evolution in the Western Alps. In: COOPER, M.A. & WILLIAMS, G.D. (eds) *Inversion Tectonics*. Geological Society, London, Special Publication, **44**, 105-122.
- CARRERAS, J., ESTRADA, A. & WHITE, S. 1977. The effect of folding on the *c*-axis fabrics of quartz mylonite. *Tectonophysics*, **39**, 3-24.
- CHADWICK, R.A. 1993. Aspects of basin inversion in southern Britain. *Jl. geol. Soc. Lond.*, **150**, 311-322.
- CHADWICK, R.A., KENOLTY, N. & WHITTAKER, A. 1983. Crustal structure beneath southern England from deep seismic reflection profiles. *Jl. geol. Soc. Lond.*, **140**, 893-911.
- CHAPMAN, T.J. 1989. The Permian to Cretaceous structural evolution of the Western Approaches Basin (Melville sub-basin), UK. In: COOPER, M.A. & WILLIAMS, G.D. (eds) *Inversion Tectonics*. Geological Society, London, Special Publication, **44**, 177-200.
- CHAPMAN, T.J., FRY, R.L. & HEAVEY, P.T. 1984. A structural cross-section through SW Devon. *Jl. geol. Soc. Lond.*, **143**, 113-118.
- CHEADLE, M.J., CZUCHRA, B.L., BYRNE, T., ANDO, C.J., OLIVER, J.E., BROWN, L.D., KAUFMAN, S., MALIN, P.E. & PHINNEY, R.A. 1986. The deep crustal structure of the Mojave desert, California, from COCORP seismic reflection data. *Tectonics*, **5**, 293-320.
- COBBOLD, P.R. & QUINQUIS, H. 1980. Development of sheath folds in shear regimes. *J. Struct. Geol.*, **2**, 119-126.
- COPE, J.C.W. 1987. Pre-Devonian geology of south-west England. *Proc. Ussher Soc.* **7**, 468-473.

- COWARD, M.P. & McCLAY, K.R. 1983. Thrust tectonics in S.Devon. *Jl. geol. Soc. Lond.*, **140**, 215-228.
- COWARD, M.P. & POTTS, G.J. 1983. Complex strain patterns developed at the frontal and lateral tips to shear zones and thrust zones. *J. Struct. Geol.*, **5**, 383-399.
- COWARD, M.P. & SMALLWOOD, S. 1984. An interpretation of the Variscan tectonics of SW Britain. In: HUTTON, D.H.W. & SANDERSON, D.J. (eds) *Variscan tectonics of the North Atlantic Region*. Geological Society, London, Special Publication, **14**, 89-102.
- CROWELL, J.C. 1982. The Violin Breccia, Ridge Basin, southern California. In: CROWELL, J.C. & LINK, M.H. (eds) *Geological history of Ridge Basin, southern California*. Society of Economic Palaeontologists and Mineralogists, Pacific Section, Guidebook, 89-97.
- CURTIS, M.L. 1993. The structural and kinematic evolution of an upper crustal transpression zone: The Lusitanian Basin, Portugal. *Unpub. Ph.D thesis, University of Durham*.
- DARBYSHIRE, D.P.F. & SHEPHERD, T.J. 1985. Chronology of granite Magmatism and associated mineralisation, SW England. *Jl. geol. Soc. Lond.*, **142**, 1159-1197.
- DAVIS, G.R. 1984. Isotopic evolution of the Lizard Complex. *Jl. geol. Soc. Lond.*, **141**, 3-14.
- DAVIS, P.G. 1990. A late Gedinnian-early Siegenian palynomorph assemblage from the Dartmouth Beds of north Cornwall. *Proc. Ussher Soc.*, **7**, 307.
- DAY, G.A. & EDWARDS, J.W.F. 1983. Variscan thrusting in the basement of the English Channel and SW Approaches. *Proc. Ussher Soc.*, **5**, 432-436.
- DE LA BECHE, H.T. 1839. Report on the geology of Cornwall, Devon and west Somerset. *Mem. Geol. Surv., G.B.*
- DEARMAN, W.R. 1963. Wrench faulting in Cornwall and south Devon. *Proc. Geol. Ass.*, **74**, 265-287.
- DEARMAN, W.R. 1966. Superposed structures in the Trebursye Beds of Launceston, Cornwall. *Proc. Ussher Soc.*, **1**, 241-245.
- DEARMAN, W.R. 1969. Tergiversate folds from South-west England. *Proc. Ussher Soc.*, **2**, 112-115.
- DEARMAN, W.R. 1971. A general view of the structure of Cornubia. *Proc. Ussher Soc.*, **2**, 220-236.
- DEARMAN, W.R. & FRESNEY, E.C. 1966. Repeated folding at Boscastle, North Cornwall, England. *Proc. Geol. Ass.*, **77**, 199-215.
- DEARMAN, W.R., FRESNEY, E.C., SELWOOD, E.B., SIMPSON, S., STONE, M. & TAYLOR, R.T. 1971. Symposium on the structure of SW England. *Proc. Ussher Soc.*, **2**, 220-263.
- DEARMAN, W.R., LEVERIDGE, B.E., RATTEY, R.P. & SANDERSON, D.J. 1980. Superposed folding at Rosemullion Head, South Cornwall. *Proc. Ussher Soc.*, **5**, 33-38.
- DINELEY, D.L. 1966. The Dartmouth Beds of Bigbury Bay, south Devon. *Jl. geol. Soc. Lond.*, **122**, 187-219.
- DINELEY, D.L. 1986. Cornubian quarter century: advances in the geology of south-west England, 1960-1985. *Proc. Ussher Soc.*, **6**, 275-290.
- DINELEY, D.L. 1992. Devonian. In: DUFF, P.McL.D. & SMITH, A.J. (eds) *Geology of England and Wales*. Geological Society, London, Special Publication, **51**, 179-205.

- DODSON, M.H. & REX, D.C. 1971 Potassium-argon ages of slates and phyllites from southwest England. *Jl. geol. Soc. Lond.*, **126**, (for 1970) 465-499.
- DONATO, J.A. 1988. Possible Variscan thrusting beneath the Somerton Anticline, Somerset. *Jl. geol. Soc. Lond.*, **145**, 431-438.
- DOODY, J.J. & BROOKS, M. 1986. Seismic refraction investigations of the structural setting of the Lizard and Start complexes, SW England. *Jl. geol. Soc. Lond.*, **143**, 135-140.
- DUBEY, A.K. & COBBOLD, P.R. 1977. Noncylindrical flexural slip folds in nature and experiment. *Tectonophysics*, **38**, 223-239.
- DURNEY, D.W. & RAMSAY, J.G. 1973. Incremental strains measured by syntectonic crystal growths. In: DeJONG, K.A. & SCHOLTEN, R. (eds) *Gravity and Tectonics*. John Wiley & Sons, New York, 67-96.
- DURRANCE, E.M. 1985. Lower Devonian acid igneous rocks in south Devon: Implications for Variscan plate tectonics. *Proc. Ussher Soc.*, **6**, 205-210
- EDMONDS, E.A. 1974. Classification of the Carboniferous rocks of south-west England. *Report, Institute Geological Sciences*, 74/13.
- EDMONDS, E.A., MCKEOWN, M.C. & WILLIAMS, M. 1975. *British Regional Geology: South-West England* (4th Edition). London, HMSO for British Geological Survey.
- EDWARDS, J.W.F., DAY, G.A. & LEVERIDGE, B.E. 1989. Thrusts under Mount's Bay and Plymouth Bay. *Proc. Ussher Soc.*, **7**, 131-135.
- ELLIOTT, T. 1976. Upper Carboniferous sedimentary cycles produced by river dominated elongate deltas. *Jl. geol. Soc. Lond.*, **132**, 199-208.
- ELLIS, M. & WATKINSON, A.J. 1987. Orogen-parallel extension and oblique tectonics—the relation between stretching lineations and relative plate motions. *Geology*, **15** (11), 1022-1026.
- ENGEL, W. & FRANKE, W. 1983. Flysch sedimentation; its relations to tectonism in the European Varicides. In: MARTIN, H. & EDER, F.W. (eds) *Intracontinental Fold Belts*. Springer Verlag, 191-184.
- EVANS, C.D.R. 1990. *United Kingdom offshore regional report: the geology of the western English Channel and its western approaches*. British Geological Survey, HMSO, London
- EVANS, K.M. 1981. A marine fauna from the Dartmouth Beds (Lower Devonian) of Cornwall. *Geol. Mag.*, **118**, 519-523.
- EVANS, K.M. 1985. The brachiopod fauna of the Meadfoot Group (Lower Devonian) of the Torbay area, south Devon. *J. Geol.*, **2**, 81-90.
- EXLEY, C.S. 1966. The granitic rocks of Haig Fras. *Nature*, **210**, 365-367
- FIELITZ, W. 1992. Variscan transpressive inversion in the northwestern central Rhenohercynian belt of western Germany. *J. Struct. Geol.*, **14**, 547-565.
- FLETT, J.S. 1946. Geology of the Lizard and Meneage (sheet 359). *Mem. Geol. Surv., G.B. (2nd edn)*.
- FLOYD, P.A. 1982. Chemical variations of Hercynian basalts relative to plate tectonics. *Jl. geol. Soc. Lond.*, **139**, 505-520.
- FLOYD, P.A. 1984. Geochemical characteristics and comparison of the basic rocks of the Lizard Complex and the basaltic lavas within the Hercynian troughs of SW England. *Jl. Geol. Soc. London*, **141**, 61-70.
- FLOYD, P.A. & LEVERIDGE, B.E. 1987. Tectonic environment of the Devonian Gramscatho basin: framework mode and geochemical evidence from turbiditic sandstones. *Jl. geol. Soc. Lond.*, **144**, 531-542.

- FLOYD, P.A., SHAIL, R.A., LEVERIDGE, B.E. & FRANKE, W. 1991. Geochemistry and provenance of Rhenohercynian synorogenic sandstones: implications for tectonic environment discrimination. In: MORTON, A.C., TODD, S.P. & HAUGHTON, P.D.W. (eds) *Developments in Sedimentary Provenance Studies*. Geological Society, London, Special Publication, **57**, 173-188.
- FLOYD, P.A., HOLDSWORTH, R.E. & STEELE, S.A. 1993a. Geochemistry of the Start greenschists: Rhenohercynian MORB? *Geol. Mag.*, **130**, 345-352.
- FLOYD, P.A., EXLEY, C.S. & STYLES, M.T. 1993b. *Igneous rocks of South-West England*. Chapman & Hall, London.
- FOSSON, H. & TIKOFF, B. 1993. The deformation matrix for simultaneous simple shearing, pure shearing and volume change, and its application to transpression-transension tectonics. *J. Struct. Geol.* **15**, 413-422.
- FRANKE, W. & ENGEL, W. 1982. Variscan sedimentary basins on the continent and relations with south-west England. *Proc. Ussher Soc.*, **5**, 259-269.
- FRANKE, W., FLOYD, P.A., HOLDER, M.T. & LEVERIDGE, B.E. 1989. The Rhenohercynian Ocean revisited. *Terra (abs)*, **1**, 365.
- FRESHNEY, E.C. 1965. Low-angle faulting in the Boscastle area. *Proc. Ussher Soc.*, **1**, 195-180.
- FRESHNEY, E.C. & TAYLOR, R.T. 1980. The variscides of south-west Britain. In: OWEN, T.R. (ed) *United Kingdom: Introduction to general Geology and Guide to excursions 002, 055, 093, and 151*. International Geological Congress, Paris 1980. Institute of Geological Sciences, London.
- FRITZ, H. & NEUBAUER, F. 1993. Kinematics of crustal stacking and dispersion in the southeastern Bohemian Massif. *Geol. Rund.*, **82** (3), 556-565.
- GAUSS, G.A. & HOUSE, D.M. 1972. The Devonian successions in the Padstow area, North Cornwall. *Jl. geol. Soc. Lond.*, **128**, 151-192.
- GAYER, R. & JONES, J. 1989. The Variscan foreland in South Wales. *Proc. Ussher Soc.*, **7**, 197-199.
- GOODE, A.J.J. & MERRIMAN, R.J. 1987. Evidence of crystalline basement west of the Land's End Granite, Cornwall. *Proc. Geol. Ass.*, **98**, 39-43.
- GOODE, A.J.J., MERRIMAN, R.J. & DARBYSHIRE, D.P.F. 1987. Evidence of crystalline basement east of the Lands End Granite, Cornwall. *Proc. Geol. Ass.*, **98**, 272.
- GREEN, D.H. 1964. The petrogenesis of the high temperature peridotite intrusion in the Lizard area, Cornwall. *J. Pet.*, **5**, 134-188.
- HANCOCK, P.L. (ed) 1983. *The Variscan Fold Belt in the British Isles*. Hilger, Bristol.
- HANCOCK, P.L., DUNNE, W.M. & TRINGHAM, M.E. 1983. Variscan deformation in Southwest Wales. In: HANCOCK, P.L. (ed) *The Variscan Fold belt in the British Isles*, Hilger, Bristol, 47-73.
- HARLAND, W.B. 1971. Tectonic transpression in Caledonian Spitsbergen. *Geol. Mag.* **108**, 27-42.
- HARRIS, A.L., BRADBURY, H.J. & MCGONIGAL, M.H. 1976. The evolution and transport of the Tay Nappe. *Scott. J. Geol.*, **12**, (2), 103-113.
- HARTLEY, A.J. 1993. Silesian sedimentation in South-West Britain: Sedimentary Responses to the Developing Variscan orogeny. In: GAYER, R.A., GREILING, R.O. & VOGEL, A.K. (eds) *Rhenohercynian and Subvariscan Fold Belts*. Viewig Publishing, Wiesbaden, 159-196.
- HARTLEY, A.J. & WARR, L.N. 1990. Upper Carboniferous foreland basin evolution in SW England. *Proc. Ussher Soc.*, **7**, 212-216.

- HARWOOD, G.M. 1976. The Staddon Grits- or Meadfoot Beds? *Proc. Ussher Soc.*, **3**, 333-338.
- HAWKES, J.R. 1981. A tectonic "watershed" of fundamental consequence in the post-Westphalian evolution of Cornubia. *Proc. Ussher Soc.*, **5**, 128-131.
- HAYWARD, A.B. & GRAHAM, R.H. 1989. Some geometrical characteristics of inversion. In: COOPER, M.A. & WILLIAMS, G.D. (eds) *Inversion Tectonics*. Geological Society, London, Special Publications, **44**, 17-39.
- HEMPTON, M.R. 1983. Evolution of thought concerning sedimentation in pull-apart basins. In: BOARDMAN, S.J. (ed) *Revolution in Earth Sciences*. Kendal/Hunt, 167-180.
- HEMPTON, M.R. & DUNNE, L.A. 1984. Sedimentation of pull-apart basins: active examples in Eastern Turkey. *J. Geol.*, **92**, 513-530.
- HENDRIKS, E.M.L. 1937. Rock succession and structure in south Cornwall: a revision. With notes on the Central European facies and Variscan folding there present. *Jl. geol. Soc. Lond.*, **93**, 322-367.
- HENDRIKS, E.M.L. 1939. The Start-Dodman-Lizard Boundary Zone in relation to the Alpine structure of Cornwall. *Geol. Mag.*, **76**, 385-402.
- HENDRIKS, E.M.L. 1951. Geological succession and structure in western south Devonshire. *Trans. R. Geol. Soc. Cornwall*, **18**, 255-295.
- HENDRIKS, E.M.L. 1959. A summary of the present views on the structure of Devon and Cornwall. *Geol. Mag.*, **96**, 253-257.
- HENDRIKS, E.M.L. 1971. Facies variations in relation to tectonic evolution in Cornwall. *Trans. R. Geol. Soc. Cornwall*, **20**, 114-151.
- HENLEY, S. 1970. The geology and geochemistry of an area around Perranporth, Cornwall. *Unpub. Ph.D thesis, University of Nottingham*.
- HENLEY, S. 1973. The structure of the Perranporth area, Cornwall. *Proc. Ussher Soc.*, **2**, 521-524.
- HENLEY, S. 1974. Geochemistry of the Devonian sediments around the Perranporth area, Cornwall. *Proc. Ussher Soc.*, **3**, 128-136.
- HIGGS, R. 1984. Possible wave-influenced sedimentary structures in the Bude Formation (Lower Westphalian, south-west England), and their environmental implications. *Proc. Ussher Soc.*, **6**, 88-94.
- HIGGS, R. 1991. The Bude Formation (Lower Westphalian), SW England: siliciclastic shelf sedimentation in a large equatorial lake. *Sedimentology*, **38**, 445-469.
- HILL, J.B. & MacALISTER, D.A. 1906. The geology of Falmouth and Truro and the mining district of Camborne and Redruth. *Mem. Geol. Surv., G.B.*
- HILLIS, R.R. & CHAPMAN, T.J. 1992. Variscan structure and its influence on post-Carboniferous basin development, Western Approaches Basin, SW UK Continental Shelf. *Jl. geol. Soc. Lond.*, **149**, 413-417.
- HOBBS, B.E., MEANS, W.D. & WILLIAMS, P.F. 1976. *An Outline of Structural Geology*. Wiley International.
- HOBSON, D.M. 1976a. The structure of the Dartmouth antiform. *Proc. Ussher Soc.*, **3**, 320-332
- HOBSON, D.M. 1976b. A structural section between Plymouth and Bolt Tail, south Devon. *Proc. Geol. Ass.*, **87**, 27-43.
- HOBSON, D.M. 1977. Polyphase folds from the Start complex. *Proc. Ussher Soc.*, **4**, 102-110.

- HOBSON, D.M. & SANDERSON, D.J. 1975. Major early folds at the southern margin of the Culm synclinorium. *Jl. geol. Soc. Lond.*, **131**, 337-352.
- HOBSON, D.M. & SANDERSON, D.J. 1983. Variscan deformation in Southwest England. In: HANCOCK, P.L. (ed) *The Variscan Fold belt in the British Isles*, Hilger, Bristol, 108-129.
- HOLDER, M.T. & LEVERIDGE, B.E. 1986a. Correlation of the Rhenohercynian Variscides. *Jl. geol. Soc. Lond.*, **143**, 141-147.
- HOLDER, M.T. & LEVERIDGE, B.E. 1986b. A model for the tectonic evolution of south Cornwall. *Jl. geol. Soc. Lond.*, **143**, 125-134.
- HOLDER, M.T. & LEVERIDGE, B.E. 1987. Discussion on the correlation of the Rhenohercynian Variscides. *Jl. geol. Soc. Lond.*, **144**, 353-354.
- HOLDSWORTH, R.E. 1988. The stereographic analysis of facing. *J. Struct. Geol.* **10**, 219-223.
- HOLDSWORTH, R.E. 1989a. The Start-Perranporth line: a Devonian terrane boundary in the Variscan orogen of SW England? *Jl. geol. Soc. Lond.*, **146**, 419-421.
- HOLDSWORTH, R.E. 1989b. Late brittle deformation in a Caledonian ductile thrust wedge: new evidence for gravitational collapse in the Moine Thrust sheet, Sutherland, Scotland. *Tectonophysics*, **190**, 19-28.
- HOLDSWORTH, R.E. & STRACHAN, R. 1991. Interlinked system of ductile strike-slip and thrusting formed by Caledonian sinistral transpression in northeastern Greenland. *Geology*, **19**, 510-513.
- HOLDSWORTH, R.E., BUTLER, C.A., SHAIL, R.S. & STEELE, S.A. 1992. Late stage gravity collapse structures in orogenic belts: Field examples from the British Isles. *23rd Ann. Meeting, T.S.G. (abs)*.
- HOLL, H.B. 1868. On the Older Rocks of S.Devon and Cornwall. *Jl. geol. Soc. Lond.*, **24**, 438-406.
- HOUSE, M.R., RICHARDSON, J.B., CHALONER, W.G., ALLEN, J.R.L. HOLLAND, C.H. & WESTOLL, T.S. (eds) 1977. *A correlation of the Devonian rocks of the British Isles*. Geological Society, London, Special Publication, 7.
- HUNT, A.R. 1892. On certain affinities between the Devonian rocks of South Devon and the metamorphic schists. *Geol. Mag.*, **9**, 241-247.
- HUTTON, D.H.W. 1987. Strike-slip terranes and a model for the evolution of the British and Irish Caledonides. *Geol. Mag.*, **124**, 405-425.
- HUTTON, D.H.W. & SANDERSON, D.J. (eds) 1984. *Variscan Tectonics of the North Atlantic Region*. Geological Society, London, Special Publication, 14.
- ISSAC, K.P., TURNER, P.J. & STEWART, I.J. 1982. The evolution of the Hercynides of central South west England. *Jl. geol. Soc. Lond.*, **139**, 523-531.
- JOHNSON, T.E. 1991. Nomenclature and geometric classification of cleavage-transected folds. *J. Struct. Geol.*, **13**, 261-274.
- JONES, R.H. 1991. A seismic reflection survey as part of the geophysical investigation of the Carnmellis Granite. *Proc. Ussher Soc.*, **7**, 418-420.
- KELLING, G. 1988. Silesian sedimentation and tectonics in the South Wales Basin: a brief review. In: BESLEY, B.M. & KELLING, G. (eds) *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe*. Blackie, Glasgow, 38-42.
- KENOLTY, N., CHADWICK, R.A., BLUNDELL, D.J. & BACON, B. 1981. Deep seismic reflection survey across the Variscan Front of southern England. *Nature*, **293**, 451-452.

- KERR, A.C. 1993 The geochemistry and petrogenesis of the Mull and Morvern Tertiary lava succession, Argyll, Scotland. *Unpub. Ph.D thesis, University of Durham.*
- KERR, D.R. & KIDWALL, S.M. 1991. Late Cenozoic sedimentation and tectonics, Western Salton Trough, California. In: WALAWENDER, M.J. & HANAN, B.B. (eds) *Geological excursions in southern California and Mexico*. Guidebook, Annual Meeting, Geological Society of America, San Diego, California.
- KIRBY, G.A. 1979. The Lizard Complex as an ophiolite. *Nature*, **282**, 58-61.
- KOSSMAT, F. 1927. Gliederung des varistischen Gebirgsbaues. *Abhandlungen des sächsischen geologischen Landesamtes*, **1**, 39p.
- LANE, A.N. 1970. Possible Tertiary deformation of Armorican structures in south east Cornwall. *Proc. Ussher. Soc.*, **2**, 197-204
- LANG, W.H. 1929. On fossil wood (*Dadoxylon hendriksi*, n. sp) and other plant remains from the clay slates of S. Cornwall. *Ann. Botany*, **43**, 663-683.
- LEEDER, M.R. 1976. Sedimentary facies and the origin of basin subsidence along the northern margins of the supposed Hercynian ocean. *Tectonophysics*, **36**, 167-199.
- LEEDER, M.R. 1982. Upper Palaeozoic basins of the British Isles; Caledonide inheritance versus Hercynian plate margin processes. *Jl. geol. Soc. Lond.*, **139**, 479-491.
- LEEDER, M.R. 1988. Recent developments in Carboniferous geology: a critical review with implications for the British Isles and NE Europe. *Proc. Geol. Ass.*, **99**, 73-100.
- LEEDER, M.R. 1992. Dinantian. In: DUFF, P.McL.D. & SMITH, A.J. (eds) *Geology of England and Wales*. Geological Society, London, Special Publication, **51**, 237-238.
- LE GALL, B. 1990. Evidence of an imbricate crustal thrust belt in the southern British Varicides: contributions of South-western Approaches traverse (SWAT) deep seismic reflection profiling recorded through the English Channel and the Celtic Sea. *Tectonics*, **9**, 283-302.
- LE GALL, B. 1991. Crustal evolutionary model for the Varicides of Ireland and Wales from SWAT seismic data. *Jl. geol. Soc. Lond.*, **148**, 759-774.
- LE GALL, B. 1992. The deep structure of the Ardennes Variscan thrust belt from structural and ECORS seismic data. *J. Struct. Geol.*, **14**, 531-546.
- LE GALL, B., LE HERISSE, A. & DEUNFF, J. 1985. New palynological data from the Gramscatho Group of the Lizard Front (Cornwall): palaeogeographical and geodynamical implications. *Proc. Geol. Ass.*, **96**, 237-253.
- LEVERIDGE, B.E. 1974. The tectonics of the Roseland coastal section, south Cornwall. *Unpub. Ph.D thesis, University of Newcastle.*
- LEVERIDGE, B.E., HOLDER, M.T. & DAY, G.A. 1984 Thrust nappe tectonics in the Devonian of South Cornwall and the western English Channel. In: HUTTON, D.H.W. & SANDERSON, D.J. (eds) *Variscan tectonics of the North Atlantic Region*. Geological Society, London, Special Publication, **14**, 148-165.
- LEVERIDGE, B.E. & HOLDER, M.T. 1985. Olistostromic breccias at the Mylor/Gramscatho boundary, south Cornwall. *Proc. Ussher. Soc.*, **6**, 147-154.
- LEVERIDGE, B.E., HOLDER, M.T. & GOODE, A.J.J. 1990. Geology of the country around Falmouth. *Mem. Geol. Surv., G.B.*
- LISLE, R.J. 1985. The facing of faults. *Geol. Mag.*, **122**, 249-251.
- LISTER, G.S & SNOKE, A.W. 1984. S-C Mylonites. *J. Struct. Geol.*, **6**, 617-638.
- MARSHALL, B. 1962. The small structures of Start Point, south Devon. *Proc. Ussher. Soc.*, **1**, 19-21.
- MARSHALL, B. 1965. The Start Boundary Problem. *Unpub. Ph.D Thesis, University of Bristol.*

- MARSHALL, B. & FALLA, J. 1966. The mineralisation at Hallsands, South Devon. *Proc. Ussher Soc.*, **1**, 9-11.
- MANN, P., HEMPTON, M.R., BRADLEY, D.C. & BURKE, K. 1983. Development of pull-apart basins. *J. Geol.*, **91**, 529-554.
- MATTHEWS, S.C. 1977. The Variscan Foldbelt in southwest England. *N. Jb. Geol. Paläont. Abh.*, **154**, 94-127.
- MATTHEWS, S.C. 1984. Northern margins of the Varicides in the North Atlantic region: comment on the tectonic context of the problem. In: HUTTON, D.H.W. & SANDERSON, D.J. (eds) *Variscan Tectonics of the North Atlantic Region*. Geological Society, London. Special Publication, **14**, 71-85.
- McCROSS, A.M. 1986. Simple constructions for deformation in transpression/ transtension zones. *J. Struct. Geol.*, **8**, 715-718.
- MEANS, W.D. 1987. A newly recognised type of slickenside striation. *J. Struct. Geol.*, **9**, 585-590.
- MEISSNER, R., BARTELTSEN, H. & MURAWSKI, H. 1981. Thin-skinned tectonics in the Rhenish Massif, Germany. *Nature*, **290**, 339-401.
- MELVIN, J. 1986. Upper Carboniferous fine grained turbiditic sandstones from southwest England: a model for growth in an ancient, delta fed subsea fan. *J. Sed. Pet.*, **56**, 19-34.
- MÉNARD, G. & MOLNAR, P. 1988. Collapse of the Hercynian Tibetan Plateau into a late European Basin and range province. *Nature*, **334**, 235-237.
- MILLER, J.A. & GREEN, D.H. 1961. Age determinations of rocks in the Lizard (Cornwall) area. *Nature*, **192**, 1175-1176.
- MOODY, J.D. & HILL, M.J. 1956. Wrench fault tectonics. *Bull. geol. Soc. Am.*, **67**, 1207-1246.
- MOORE, J. McM. & JACKSON, N. 1977. Structure and mineralisation in the Cligga granite stock, Cornwall. *Jl. geol. Soc. Lond.*, **133**, 467-480.
- NAYLOR, M.A., MANDL, G. & SIJPESTEIJN, C.H.K. 1986. Fault geometries in basement-induced wrench faulting under different initial stress states. *J. Struct. Geol.*, **8**, 737-752.
- NICHOLSON, R. 1991. Vein morphology, host rock deformation and the origin of the fabrics of en echelon mineral veins. *J. Struct. Geol.*, **13**, 635-641.
- NICHOLSON, R. & POLLARD, D.D. 1985. Dilation and linkage of echelon cracks. *J. Struct. Geol.*, **7**, 583-590.
- NORRIS, R.J., KOONS, R.O. & COOPER, A.F. 1990. The obliquely-convergent plate boundary in the South Island of New Zealand: implications for ancient collision zones. *J. Struct. Geol.*, **8**, 737-752.
- PAMPLIN, C.F. 1988. A re-examination of the Tintagel High Strain Zone and the Padstow Facing Confrontation, north Cornwall. *Unpub. Ph.D thesis, University of Southampton*.
- PAMPLIN, C.F. 1990. A model for the tectono-thermal evolution of north Cornwall. *Proc. Ussher Soc.*, **7**, 206-211.
- PAMPLIN, C.F. & ANDREWS, J.R. 1988. Timing and sense of shear in the Padstow Confrontation Zone, north Cornwall. *Proc. Ussher Soc.*, **7**, 73-76.
- PASSCHIER, C.W. & SIMPSON, C. 1986. Porphyroclast systems as kinematic indicators. *J. Struct. Geol.*, **8**, 831-843.
- PETIT, J.P. 1987. Criteria for the sense of movement on fault surfaces in brittle rocks. *J. Struct. Geol.*, **9**, 597-608.

- PHILLIPS, F.C. 1964. Metamorphic rocks of the sea floor between Start Point and Dodman Point, S.W. England. *J. Marine Biol. Ass., U.K.*, **44**, 655-663.
- PLATT, J.P. & VISSERS, R.L.M. 1980. Extensional structures in anisotropic rocks. *J. Struct. Geol.*, **2**, 397-410.
- POUND, C.J. 1983. The sedimentology of the Lower-Middle Devonian Staddon Grits and Jenyncliff Slates on the east side of Plymouth Sound, Devon. *Proc. Ussher Soc.*, **5**, 465-472.
- POWELL, C.M. 1989. Structural controls on Palaeozoic basin evolution and inversion in southwest Wales. *Jl. geol. Soc. Lond.*, **146**, 439-446.
- POWELL, C.McA. 1974. Timing of slaty cleavage during folding of Precambrian rocks, northwest Tasmania. *Bull. geol. Soc. Am.*, **85**, 1043-1060.
- PRATT, W.T & FITCHES, W.R. 1993. Transected folds from the western part of the Bala Lineament, Wales. *J. Struct. Geol.*, **15**, 55-68.
- PRIMMER, T.J. 1983. Low-grade regional metamorphism across the Perranporth-Pentewan Line, Cornwall. *Proc. Ussher Soc.*, **5**, 421-427.
- PRIMMER, T.J. 1985. A transition from diagenesis to greenschists facies within a major Variscan fold/thrust complex in south-west England. *Min. Mag.*, **49**, 365-374.
- RAISIN, C.A. 1887. Notes on the Metamorphic Rocks of South Devon. *Jl. geol. Soc. Lond.*, **43**, 715-733.
- RAMSAY, D.M. & STURT, B.A. 1963. A Study of Fold Styles, Their Associations and Symmetry Relationships, from Sørøy, North Norway. *Norsk. Geol. Tidsskr.*, **43**, 411-430.
- RAMSAY, J.G. 1962a. Interference patterns produced by the superposition of folds of similar type. *J. Geol.*, **70**, 466-481.
- RAMSAY, J.G. 1962b. The Geometry of Conjugate Fold Systems. *Geol. Mag.*, **99**, 516-526.
- RAMSAY, J.G. 1967. *The Folding and Fracturing of Rocks*. McGraw-Hill. New York.
- RAMSAY, J.G. 1980. Shear zone geometry: a review. *J. Struct. Geol.*, **2**, 83-89.
- RAMSAY, J.G. & GRAHAM, R.H. 1970. Strain variations in shear belts. *Can. J. Earth Sci.*, **7**, 786-813.
- RAMSAY, J.G. & HUBER, M.I. 1983. *The Techniques of Modern Structural Geology. Volume 1, Strain Analysis*. Academic Press, London.
- RAMSAY, J.G. & HUBER, M.I. 1987. *The Techniques of Modern Structural Geology, Volume 2: Folds and Fractures*. Academic Press, London.
- RAST, N. 1983. Variscan Orogeny. In: HANCOCK, P.L. (ed) *The Variscan Foldbelt in the British Isles*. Hilger, Bristol, 1-19.
- RATTEY, R.P. 1980a. Structural studies in southwest Cornwall. *Unpub. Ph.D Thesis, Queen's University, Belfast*.
- RATTEY, R.P. 1980b. Deformation in south-west Cornwall. *Proc. Ussher Soc.*, **5**, 39-43.
- RATTEY, R.P. & SANDERSON, D.J. 1982. Patterns of folding within nappes and thrust sheets: examples from the Variscan of southwest England. *Tectonophysics*, **88**, 247-267.
- RATTEY, R.P. & SANDERSON, D.J. 1984. The structure of SW Cornwall and its bearing on the emplacement of the Lizard Complex. *Jl. geol. Soc. Lond.*, **141**, 87-95.
- READING, H.G. 1975. Strike-slip fault system; an ancient example from the Cantabrians. *Extracts, IX Internat. Congr. Sediment.*, 287-291.
- REID, C. 1907. The geology of the country around Mevagissey. *Mem. Geol. Surv., G.B.*

- REID, C. & SCRIVENOR, J.B. 1906. The geology of the country near Newquay. *Mem. Geol. Surv., G.B.*
- RICHTER, D. 1967. Sedimentology and facies of the Meadfoot Beds (Lower Devonian) in South-east Devon (England). *Geol. Rund.*, **56**, 543-561.
- RIPLEY, M.J. 1965. Structural studies between Holywell Bay and Dinas Head, North Cornwall. *Proc. Ussher Soc.*, **1**, 194-195.
- ROBERTS, J.L. & SANDERSON, D.J. 1971. Polyphase development of slaty cleavage and the confrontation of facing directions in the Devonian rocks of north Cornwall. *Nature: Physical Science*, **230**, 87-89.
- ROBERTS, J.L. 1974. The structure of the Dalradian rocks in the SW Highlands of Scotland. *J. geol. Soc. Lond.*, **130**, 93-124.
- ROBINSON, D. 1981. Metamorphic rocks of an intermediate facies series juxtaposed at the Start boundary, southwest England. *Geol. Mag.*, **118**, 297-301.
- SADLER, P.M. 1974. An appraisal of the 'Lizard-Dodman-Start Thrust' concept. *Proc. Ussher Soc.*, **3**, 71-81.
- SANDERS, L.D. 1955. Structural observations on south-east Lizard. *Geol. Mag.* **92**, 231-240.
- SANDERSON, D.J. 1971. Superposed folding at the northern margin of the Gramscatho and Mylor Beds, Perranporth, Cornwall. *Proc. Ussher Soc.*, **2**, 266-269.
- SANDERSON, D.J. 1973. Correlation of fold phases in SW England. *Proc. Ussher Soc.*, **2**, 525-528.
- SANDERSON, D.J. 1979. The transition from upright to recumbent folding in the Variscan fold belt of southwest England: a model based on the kinematics of simple shear. *J. Struct. Geol.*, **1**, 191-180.
- SANDERSON, D.J. 1984. Structural variations across the northern margin of the Varicides in NW Europe. In: HUTTON, D.H.W. & SANDERSON, D.J. (eds) *Variscan Tectonics of the North Atlantic Region*. Geological Society, London, Special Publications, **14**, 148-165.
- SANDERSON, D.J. & DEARMAN, W.R. 1973. Structural zones in Variscan fold belt in SW England, their location and development. *Jl. geol. Soc. Lond.*, **129**, 527-536.
- SANDERSON, D.J. & MARCHINI, W.R.D. 1984. Transpression. *J. Struct. Geol.*, **6**, 449-458.
- SANDERSON, D.J., ROBERTS, S., MCGOWAN, J.A. & GUMIEL, P. 1991. Hercynian transpressional tectonics at the southern margin of the Central Iberian Zone, west Spain. *Jl. geol. Soc. Lond.*, **148**, 493-498.
- SCOTESE, C., BAMBACH, R.K., BARTON, C., VAN DER VOO, R. & ZEIGLER, P.A. 1979. Palaeozoic base maps. *J. Geol.*, **87**, 217-278.
- SCRUTTON, C.T. 1977. Facies variations in the Devonian limestones of eastern South Devon. *Geol. Mag.*, **114**, 165-193.
- SEAGO, R.D. 1991. Tectonics and sedimentation in the Devonian and Carboniferous rocks of S.W. Devon, England. *Unpub. Ph.D thesis, Polytechnic South-West*.
- SEAGO, R.D. & CHAPMAN, T.J. 1988. The confrontation of structural styles and the evolution of a foreland basin in central SW England. *Jl. geol. Soc. Lond.*, **145**, 789-800.
- SEDGWICK, A. & MURCHISON, R.I. 1839. Classification of the older stratified rocks of Devonshire and Cornwall. *Phil. Mag.* **4**, 241-260.
- SELWOOD, E.B. 1990. A review of basin development in central south-west England. *Proc. Ussher Soc.*, **7**, 199-205.

- SELWOOD, E.B. & DURRANCE, E.M. 1982. The Devonian rocks. *In*: DURRANCE, E.M. & LAMING, D.J.C. (eds) *The geology of Devon*, University of Exeter.
- SELWOOD, E.B., STEWART, I.J. & THOMAS, J.M. 1985. Upper Palaeozoic sediments and the structure in north Cornwall -a reinterpretation. *Proc. Geol. Ass.*, **129**, 129-141.
- SELWOOD, E.B. & THOMAS, J.M. 1986a. Upper Palaeozoic successions and nappe structures in north Cornwall. *Jl. geol. Soc. Lond.*, **143**, 75-82.
- SELWOOD, E.B. & THOMAS, J.M. 1986b. Variscan facies and structure in central SW England. *Jl. geol. Soc. Lond.*, **143**, 199-207.
- SHACKLETON, R.M. 1957. Downward-facing structures of the Highland Border. *Jl. geol. Soc. Lond.*, **113**, 363-392.
- SHACKLETON, R.M. 1984. Thin-skinned tectonics, basement control and the Variscan Front. *In*: HUTTON, D.H.W. & SANDERSON, D.J. (eds) *Variscan Tectonics of the North Atlantic Region*. Geological Society, London, Special Publication, **14**, 148-165.
- SHACKLETON, R.M., RIES, A.C. & COWARD, M.P. 1982. An interpretation of the Variscan structures in SW England. *Jl. geol. Soc. Lond.*, **139**, 533-541.
- SHACKLETON, R.M. & RIES, A.C. 1984. The relation between regionally consistent stretching lineations and plate motions. *J. Struct. Geol.*, **6**, 111-117.
- SHAIL, R.K. 1989. Gramscatho-Mylor facies relationships; Hayle, south Cornwall. *Proc. Ussher Soc.*, **7**, 125-130.
- SHAIL, R.K. 1992. The provenance of the Gramsatho Group, S. Cornwall *Unpub. Ph.D. thesis, University of Keele*.
- SIMPSON, C. & SCHMIDT, S. 1983. An evaluation of criteria to deduce the sense of movement in sheared rocks. *Bull. Geol. Soc. Am.*, **94**, 1281-1288.
- SKJERNAA, L. 1989. Tubular folds and sheath folds: definitions and conceptual models for their development, with examples from the Grapesvare area, northern Sweden. *J. Struct. Geol.*, **6**, 689-703.
- SMITH, M.A.P. 1965. Repeated folding between Hayle and Portreath. *Proc. Ussher Soc.*, **1**, 170-171.
- SMITH, O. 1991. Geological notes and details for 1:10,000 sheets SX07, SW, NW, NE (part), SW 97 NE, NW (part), SW 98 SE, SW (part), SX 08 SE (part), SX 07 SE (part). (St. Minver, St. Kew, St. Tudy areas, North Cornwall District. *Open File Report, BGS*).
- SMITH, S.A. & HUMPHREYS, B. 1991. Sedimentology and depositional setting of the Dartmouth Group, Bigbury Bay, south Devon. *Jl. geol. Soc. Lond.*, **148**, 235-244.
- SMYTHE, D.K. 1984. Discussion on thrust tectonics of S. Devon. *Jl. geol. Soc. Lond.*, **141**, 593-594.
- SOMERVAIL, A.R. 1887. On metamorphic and associated rocks of the extreme south of Devonshire. *Trans. Devon. Assoc.*, **19**, 419-437.
- SOPER, N.J. 1986. Geometry of transecting, anastomosing solution cleavage in transpression zones. *J. Struct. Geol.*, **8**, 937-940.
- SOPER, N.J. & HUTTON, D.H.W. 1984. Late Caledonian sinistral displacements in Britain: implications for a three-plate collision model. *Tectonics*, **3**, 781-794.
- SOPER, N.J., WEBB, B.C. & WOODCOCK, N.H. 1987. Late Caledonian (Acadian) transpression in north-east England: timing, geometry and geotectonic significance. *Proc. Yorks. Geol. Soc.*, **46**, 195-192.

- SOPER N.J. & WOODCOCK, N.H. 1990. Silurian collision and sediment dispersal patterns in southern Britain. *Geol. Mag.* **127** (6), 527-542.
- SOPER, N.J., STRACHAN, R.A., HOLDSWORTH, R.E., GAYER, R.A. & GREILING, R.O. 1992. Sinistral transpression and the Silurian closure of Iapetus. *Jl. geol. Soc. Lond.*, **149**, 871-880.
- STILLE, H. 1924. *Grundfragen der Vergleichenden Tectonik*. Gebrüder Borntraeger, Berlin.
- STRACHAN, R.A., TREOLAR, P.J., BROWN, M. & D'LEMONS, R.S. 1989. Cadomian terrane tectonics and magmatism in the Armorican massif. *Jl. geol. Soc. Lond.*, **146**, 422-426.
- STRACHAN, R.A., HOLDSWORTH, R.E., FRIDERICHSEN, J.D. & JEPSEN, H.F. 1992. Regional Caledonian structure within an oblique convergence zone, Dronning Louise Land, NE Greenland. *Jl. geol. Soc. Lond.*, **149**, 359-371.
- STYLES, M.T. & KIRBY, G.A. 1980. New investigations of the Lizard Complex, Cornwall, England and a discussion of an ophiolite model. In: PANAYIOTOU, A. (ed) *Proceedings International Ophiolite Symposium*. Ministry of Agriculture and Natural Resources Geological Survey Department, Cyprus, 519-526.
- STYLES, M.T. & RUNDLE, C.C. 1984. The Rb-Sr isochron age of the Kennack Gneiss and its bearing on the age of the Lizard Complex, Cornwall. *Jl. geol. Soc. Lond.*, **141**, 15-19.
- SUN, S.S. & McDONOUGH, W.F. 1989. Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and process. In: SAUNDERS, A.D. & NORRY, M.J. (eds) *Magmatism in the Ocean Basins*. Geological Society, London, Special Publication, **42**, 313-345.
- SYLVESTER, A.G. 1991. Geological structure, transpression and neotectonics of the San Andreas fault in the Salton Trough, Part 1: Palm tree structure in the central Mecca Hills. In: WALAWENDER, M.J. & HANAN, B.B. (eds) *Geological excursions in southern California and Mexico*. Guidebook, Annual Meeting, Geological Society of America, San Diego, California.
- TANNER, P.W.G. 1985. Structural history of the Devonian rocks south of Liskeard, East Cornwall. *Proc. Ussher Soc.*, **6**, 155-164.
- TAYLOR, R.T. & WILSON, A.C. 1975. Notes on some igneous rocks of west Cornwall. *Proc. Ussher Soc.*, **3**, 255-262.
- TAYLOR, S.R. & McLENNAN, S.M. 1985. *The continental crust; its composition and evolution*. Blackwell Scientific Publications, Oxford.
- TILLEY, C.E. 1923. The petrology of the metamorphosed rocks of the Start area (South Devon). *Jl. geol. Soc. Lond.*, **79**, 172-204.
- TOBISCH, O.T. & PATERSON, S.R. 1988. Analysis and interpretation of composite foliations in areas of progressive deformation. *J. Struct. Geol.*, **10**, 745-754.
- TOMBS, J.M.C. 1977. A study of the space form of the Cornubian granite batholith and its application to detailed gravity surveys in Cornwall. *Min. Reconm. Report, IGS*, **11**.
- THOMAS, J.M. 1988. Basin history of the Culm Trough of Southwest England. In: BESLY, B. & KELLING, G. (eds) *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe*. Blackie, Glasgow, 24-37.
- TREAGUS, S.H. 1983. A theory of finite strain variation through contrasting layers, and its bearing on cleavage refraction. *J. Struct. Geol.*, **5**, 351-368.
- TREAGUS, S.H. & TREAGUS, J.E. 1981. Folds and the strain ellipsoid: a general model. *J. Struct. Geol.*, **3**, 1-17.

- TREAGUS, S.H. & TREAGUS, J.E. 1992. Transected folds and transpression: how are they associated? *J. Struct. Geol.*, **14**, 361-367.
- TUCKER, M.E. 1969. Crinoidal turbidites from the Devonian of Cornwall and their palaeogeographic significance. *Sedimentology*, **13**, 281-290.
- TUNBRIDGE, I.P. 1980. The Yes Tor Member of the Hangman Sandstone Group (North Devon). *Proc Ussher Soc.*, **5**, 7-12.
- TUNBRIDGE, I.P. 1986. Mid-Devonian tectonics and sedimentation in the Bristol Channel Area. *Jl. geol. Soc. Lond.*, **143**, 107-115.
- TURNER, P.J. 1986. Stratigraphical and structural variations in central SW England: a critical appraisal. *Proc. Geol. Ass.*, **97**, 331-345.
- TURNER, R.E., TAYLOR, R.T., GOODE, A.J.J. & OWENS, B. 1979. Palynological evidence for the age of the Mylor Slate, Mount Wellington, Cornwall. *Proc Ussher Soc.*, **4**, 274-283.
- TWISS, R.J. & MOORES, E.M. 1992. *Structural Geology*. W.H. Freeman & Co., New York
- UNDERHILL, J.R. & WOODCOCK, N.H. 1987. Faulting mechanisms in high porosity sandstones; New Red Sandstone, Arran, Scotland. In: JONES, M.E. & PRESTON, R.F. (eds) *Deformation of Sediments and Sedimentary Rocks*. Geological Society, London, Special Publication, **29**, 91-105.
- USSHER, W.A.E. 1892. The British Culm Measures. *Proc. Somerset Archaeol. Nat. Hist. Soc.*, **38**, 111-219.
- USSHER, W.A.E. 1904. Geology of the country around Kingsbridge and Salcombe. *Mem. Geol. Surv., G.B.*
- USSHER, W.A.E., BARROW, G. & MacALISTER, A.R.S.M. 1909. The geology of the country around Bodmin and St. Austell. *Mem. Geol. Surv., G.B.*
- VAN STRAATEN, P. & TUCKER, M.E. 1972. The Upper Devonian Saltern Cove is an intraformational slump. *Palaeon.* **15**, 430-438.
- VEARNECOMBE, J.R. 1979. The Lizard ophiolite and two phases of suboceanic deformation. In: PANAYIOTOU, A. (ed) *Proceedings of the International Ophiolite Symposium*. Ministry of Agriculture and Natural Resources Geological Survey Department, Cyprus, 527-537.
- WARR, L.N. 1991. Basin inversion controls in the Variscan of SW England. *Unpub. Ph.D thesis, University of Exeter*.
- WARR, L.N. 1993. Basin inversion and foreland basin development in the Rhenohercynian Zone of South-West England. In: GAYER, R.A., GREILING, R.O. & VOGEL, A.K. (eds) *Rhenohercynian and Subvariscan Fold Belts*. Vieweg Publishing, Wiesbaden, 197-224.
- WARR, L.N., PRIMMER, T.J. & ROBINSON, D. 1991. Very low-grade metamorphism in southwest England: a diastathermal and thrust-related origin. *J. Met. Geol.*, **9**, 751-764.
- WEBER, K. & BEHR, M.J. 1983. Geodynamic Interpretation of the mid-European Varicides. In: MARTIN, H. & EDER, F.W. (eds) *Intercontinental Fold Belts*. Springer Verlag, 427-469.
- WHALLEY, J.S. & LLOYD, G.E. 1986. Tectonics of the Bude Formation, north Cornwall-the recognition of northerly directed décollement. *Jl. geol. Soc. Lond.*, **143**, 83-88.
- WILCOX, R.E., HARDING, T.P. & SEELY, D.R. 1973. Basic Wrench Tectonics. *Bull. Am. Ass. Pet. Geol.*, **57**, 74-96.

- WILKINSON, J.J. 1990. The origin and evolution of Hercynian crustal fluids, south Cornwall, England. *Unpub. Ph.D thesis, University of Southampton.*
- WILKINSON, J.J. & KNIGHT, R.R.W. 1989. Palynological evidence from the Porthleven area, south Cornwall: implications for Devonian stratigraphy and Hercynian structural evolution. *Jl. geol. Soc. Lond.*, **146**, 739-742.
- WILLIAMS, G.D. & CHAPMAN, T.J. 1986. The Bristol-Mendip Foreland Thrust Belt. *Jl. geol. Soc. Lond.*, **143**, 63-73.
- WILLIAMS, G.D., POWELL, C.M. & COOPER, M.A. 1989. Geometries and kinematics of inversion tectonics. In: COOPER, M.A. & WILLIAMS, G.D. (eds) *Inversion Tectonics*. Geological Society, London, Special Publication, **44**, 3-15.
- WILSON, G. 1951. The tectonics of the Tintagel area, north Cornwall. *Jl. geol. Soc. Lond.*, **106**, 393-432.
- WILSON, G. 1953. Mullion and Rodding Structures in the Moine Series of Scotland. *Proc. Geol. Ass.*, **64**, 118-151.
- WILSON, A.C. & TAYLOR, R.T. 1976. Stratigraphy and sedimentation in West Cornwall. *Trans. Royal Geol. Soc., Cornwall*, **20**, 246-259.
- WOOD, D.A. 1980. The application of a Th-Hf-Ta diagram to problems of tectonomagmatic classification and to establishing the nature of crustal contamination of basaltic lavas of the British Tertiary Volcanic Province. *Earth and Planetary Sci. Let.*, **50**, 11-30.
- WOODCOCK, N.H., AWAN, M.A., JOHNSON, T.E., MACKIE, A.H. & SMITH, R.D.A. 1988. Acadian tectonics in Wales during Avalonia/Laurentia convergence. *Tectonics*, **7**, 483-495.
- ZEIGLER, P.A. 1982. *Geological Atlas of Central and Western Europe*. Elsevier.
- ZIEGLER, P.A. 1987. Evolution of the Western Approaches Trough. *Tectonophysics*, **137**, 341-346.
- ZIEGLER, P.A. 1990. *Geological Atlas of Central and Western Europe (2nd Edition)*. Drukkerij Verweij B.V., Mijdrecht, Holland, 239p.

APPENDIX 1

ANALYTICAL GEOCHEMISTRY

A1.1 Sample preparation

Crushing

All of the Start Complex metasediments (greyschist) samples were prepared and analysed at Durham University. Initially, weathered surfaces were removed from the samples with the aid of a rock splitter and a clipper saw. After washing to remove any surface contamination, a substantial proportion of the sample was coarsely crushed (~1cm diameter), using a *Fritsch Pulverisette* jaw crusher. At this stage, the crushed rock was inspected and any visible altered or weathered material removed, before final crushing to fine material (<0.5cm diameter). The plates of the jaw crusher were cleaned between preparation of each sample with a wire brush and absolute alcohol. Approximately 150g of each sample was then reduced to a fine powder, using an agate ball mill.

Fusion disc preparation

4-5 g of rock powder was accurately weighed out, and heated in porcelain crucibles at 900°C for 2 hours. The ignited powders were then allowed to cool in a desiccator, and then reweighed to determine the loss on ignition (LOI). Approximately 0.45g of ignited rock powder was accurately weighed out, along with exactly five times the amount of dried lithium metaborate and lithium tetraborate flux (Johnson Matthey 'Spectroflux 100B'). After thorough mixing, this 1:5 rock:flux mixture was placed in a platinum crucible (with lid) in a furnace at 1050°C for 15-20 minutes. The molten mixture was then removed from the furnace, swirled briefly in the platinum crucible to ensure thorough mixing, and then poured into graphite moulds on a hot plate set at 250°C. A stainless steel plunger was then used to form the disc within the mould. Following slow cooling for 20 minutes, the disc was labelled, bagged and stored in a desiccator. During this latter procedure, great care was taken to ensure that the analytical surface of the disc did not come into contact with skin, as this would be liable to cause considerable contamination with Na₂O, P₂O₅ and K₂O (Kerr, *pers comm*).

Pressed powder pellet preparation

Approximately 8g of rock powder was added with ~10 drops of PVA 'Mowiol' organic binding solution, in a small beaker and mixed thoroughly. The mixture was then subjected to 10 tonnes of pressure for half a minute, within a stainless steel mould. The pellets were dried in an oven at 100°C overnight, labelled and bagged.

A1.2 XRF analysis

This analysis was carried out on a Philips PW 1400 spectrometer, fitted with a PW 1500/10 automatic sample changer. A rhodium tube was used to provide the X-ray source (accelerating potential=80kV, 35mA, or 55kV, 50mA when analysing REE's). Philips X40 software has been used to process the raw count rates of the peaks and backgrounds.

The machine was calibrated by analysing the elemental count rates on various international standards of known composition. For each element being analysed, the X40 program (after correction for interelemental effects and possible line overlaps) generates a plot of count rates vs. known concentration of the standards, and calculates a best fit straight line through the data. The operator can delete those standards which plot far from the regression line, in order to optimise the calibration. In order to ensure precision, virtually all the samples were analysed twice, each time on a different calibration, and the results of the two runs have been averaged.

Geochemical data for the Start Complex greyschists (SPS) and greenschists (SPV), are given in the following pages.

Appendix I

	sps2	sps3	sps4	sps5	sps7	sps8	sps9	sps10	sps11	sps12	sps13	sps14
SiO2	69.78	91.68	91.98	71.93	89.41	67	60.71	62.01	62.04	70.54	65.19	68
TiO2	0.77	0.79	0.7	0.72	0.24	1.08	1.18	1.34	1.19	0.82	0.81	0.81
Al2O3	16.17	4.94	4.93	14.89	4.9	18.99	16.92	16.56	16.03	17.07	15.67	17.63
Fe2O3	5.46	1.06	1.14	5.94	2.53	7.62	8.84	8.6	8.38	6.62	5.93	6.82
MnO	0.03	0.01	0	0.07	0.09	0.07	0.13	0.13	0.11	0.07	0.07	0.06
MgO	2.22	0	0.04	0.6	0.39	0.8	4.28	3.88	4.14	1.47	1.18	1.45
CaO	0.23	0.21	0.07	0.23	0.09	0.08	1.35	0.84	2.19	0.19	0.17	0.2
Na2O	1.88	0	0.02	1.98	0.01	0	3.88	4.32	4.72	0.7	0.5	0.69
K2O	2.7	0.37	0.56	2.52	1.3	3.36	1.53	1.59	0.45	3.12	3.01	3.04
P2O5	0.17	0.17	0.05	0.21	0.08	0.14	0.19	0.21	0.18	0.15	0.13	0.16
LOI	4.4	1.53	6.5	4.15	1.85	6.25	3.7	3.3	3.12	3.8	3.11	3.35

Ba	601.11	84.93	127.8	393.41	223.47	578.97	383.45	388.56	93.41	568.3	466.55	578.64
Ce	48.05	76.1	59.3	52.21	28.44	68.56	35.89	41.19	36.77	50.41	33.01	49.91
Cr	127.6	36.3	34.8	104.8	48.6	153.8	149.3	152.7	149	103	76.4	105.7
Ga	18.4	5.8	9	21	9.5	22.1	18.5	19.2	18.6	19.1	19.3	19.4
La	23.94	34.21	27.29	30.04	13.28	45.66	17.18	24.69	23.99	28.29	20.29	28.41
Nb	11.9	17.1	15.8	13	5.5	16.1	11.1	12.2	12.1	15	12	15
Nd	21.16	31.7	22.13	19.92	10.56	37.85	16.78	24.38	20.56	22.87	14.14	22.12
Ni	38.5	12.5	7	45.5	44.5	36.6	57.6	44.3	40.2	33.9	31.7	30
Rb	103.4	15.3	23.6	90.5	51.8	121.7	43.9	41.6	11.9	136.8	105.5	130.5
Sr	84.6	92.3	59	97.9	52.7	18.4	40.6	45.1	161.1	92.3	58.2	119.2
V	146.4	47.8	46.5	202.7	120.6	247.8	244.5	241.3	225.9	124.8	97.4	132.1
Y	21.8	35	27.1	23.6	14.6	34.2	30.8	37.1	42.7	29.7	21.8	26.1
Zr	172.4	828.9	617.5	157	49	230.2	205.3	229.5	204.3	274.2	171.7	185.9
Sc	12.5	4.1	2.4	12.4	2.2	18.4	21.9	20.5	15.2	10.6	11.6	16

	sps15	sps17	sps18	sps19	sps20	sps21	sps22	sps23	sps24	sps25	sps26	sps27
SiO2	76.42	69.49	60.1	46.34	56.2	59.02	52.53	67.38	68.08	86.75	68.96	78.98
TiO2	0.61	0.86	0.99	1.51	1.24	1.15	1.27	0.71	0.82	0.34	1.34	0.49
Al2O3	14.09	17.07	17.86	33.06	24.94	22.6	27.6	15.36	17.84	7.3	16.56	9.9
Fe2O3	5.6	6.16	9.27	10.04	8.4	7.56	8.93	7.88	6.82	1	8.6	3.98
MnO	0.05	0.05	0.18	0.09	0.1	0.06	0.07	0.38	0.06	0.01	0.13	0.07
MgO	1.22	1.43	4.21	2.41	1.91	2.33	2.17	3.33	1.54	0.48	3.88	1.18
CaO	0.19	0.18	0.27	0.2	0.25	0.21	0.24	0.18	0.22	0.19	0.84	0.16
Na2O	0.82	0.2	1.53	1.89	1.44	1.45	1.27	0.06	2.14	0.04	4.32	0.1
K2O	2.65	3.56	3.49	6.02	4.76	4.84	4.87	3.69	3.12	1.99	1.59	2.61
P2O5	0.13	0.12	0.18	0.12	0.16	0.15	0.15	0.09	0.17	0.11	0.21	0.11
LOI	3.9	3.1	4.3	5.3	4.34	4.1	4.8	3.3	4.11	2.75	2.2	4.17

Ba	483.37	668.3	599.2	936	831.15	877.44	879.26	332.03	647.3	306.1	282.65	351.63
Ce	55.12	53.46	46.29	77.74	70.06	61.34	80.39	46.37	73.45	46.97	53.88	36.27
Cr	77.9	103.9	231.7	198	151.5	167.2	164.5	130.7	210.3	75.5	132	96.3
Ga	16.9	20.6	22.7	34	28.7	26.2	30.3	19.9	25.3	12.7	16.6	13.6
La	30.13	34.41	31.3	44.57	45.34	38.29	61.88	28.06	62.77	23.31	35.54	16.9
Nb	12	15.7	11.5	24.7	20.4	18	21.6	14.4	21.9	8.4	10.4	9.6
Nd	27.48	28.81	24.7	36.17	36.55	34.26	46.83	19.45	47.55	19.29	22.47	12.53
Ni	28.5	31.6	48.3	31	26.5	30	25.9	117.9	88	9.3	73.9	28.8
Rb	108.6	156.4	126	242.6	200.6	193.1	212.3	130	239.6	81.4	80.1	104.2
Sr	63.3	72.7	25.4	230.7	137.1	126.7	189.2	14.6	54.2	22.2	18.3	24.2
V	110.9	137.3	236.9	210.5	173.6	182.9	185.7	153.6	229.4	233.3	160.4	201.8
Y	21.3	28.6	32.9	38.5	38.7	40	32.8	20.8	44.6	13.8	20.4	19.3
Zr	143.5	213.2	154.7	275.8	338.9	293.4	276.7	139.1	238.4	69.9	106.2	92.2
Sc	12	14.1	25.4	23	19.8	21.1	16.2	10.4	18.8	7.8	8.4	10.6

Appendix 1

	sps28	sps30	sps31	sps32	sps33	sps34	sps35	sps36	sps37	sps38	sps39	sps40
SiO2	57.84	65.86	80.75	78.58	58.98	56.65	78.17	60.55	48.25	49.11	62.89	75.56
TiO2	0.82	0.66	0.39	0.66	1.07	1.06	0.43	1.04	1.32	1.53	0.9	0.64
Al2O3	16.61	18.33	10	11.55	23.58	23.28	12.4	22.79	31.04	29.77	18.34	13.86
Fe2O3	6.79	6.48	4.08	4.74	7.84	7.59	4.14	8.38	8.94	9.34	6.76	4.46
MnO	0.05	0.07	0.15	0.05	0.07	0.07	0.04	0.04	0.05	0.06	0.06	0.04
MgO	1.4	1.63	0.75	1.1	1.89	1.86	0.98	1.76	2.21	2.21	1.61	1.04
CaO	0.18	0.24	0.8	0.2	0.23	0.25	0.22	0.07	0.09	0.07	0.23	0.15
Na2O	0.77	0.76	0.03	0.19	1.02	1.53	0.22	0.76	1.47	0.97	0.7	0.29
K2O	3.23	3.54	2.05	2.01	4.43	4.12	2.35	3.9	5.65	6.05	3.86	2.88
P2O5	0.17	0.15	0.11	0.12	0.16	0.15	0.14	0.08	0.09	0.09	0.16	0.12
LOI	3.65	5.04	2.77	3	2	3.28	3.74	5.64	2.88	3.1	4.14	2.58
Ba	644.38	665.61	484.19	484.73	762.46	724.18	480	738.92	783.83	1026.63	938.3	516.47
Ce	59.02	42.42	44.83	55.79	68.28	72.81	45.87	70.95	78.93	86.27	91.68	54.35
Cr	115	107.1	61	72.9	141.2	143.5	64	143.6	178.2	178.8	124.1	81.4
Ga	23.2	22.2	12.8	15.1	26.7	27.5	15.7	23.6	30.9	30.7	21.8	17.5
La	35.05	25.99	22.74	30.48	42.7	46.13	19.26	40.16	39.58	59.46	63.29	26.17
Nb	16.6	14.2	6.7	10.6	21.1	18	7.3	17.4	28.6	28.6	14.1	14.4
Nd	25.59	21.82	17.74	23.04	36.82	39.59	15.92	27.43	32.99	49.32	48.88	20.81
Ni	19.8	25.3	19.9	20	27.6	32.2	22.5	35.9	32.9	34.2	24.8	23.9
Rb	145	151	91.2	91.1	185.3	173.1	98.8	158.6	226.2	250.6	165.9	120.3
Sr	125.8	112	53.4	52	138.6	171.9	60.8	130.7	221.2	154.7	94	53.9
V	146.9	145	100.9	96.4	174.1	176.2	105.8	176.1	233.9	220.1	165.2	122.5
Y	28.1	21.5	12.6	21	30.4	28.8	16.9	31.8	39.4	46.4	23.2	19.6
Zr	194	99.9	115.3	275.3	193.3	179.5	81.7	165.8	192.6	300.1	184	185.6
Sc	18.6	13.8	11.6	7.2	18.2	17.4	11.8	17.7	20.2	22.8	19.2	10

Appendix 1

	spv1	spv2	spv3	spv4	spv5	spv6	spv7	spv8	spv9	spv10	spv11	spv12
SiO2	48.8	46.89	49.09	47.65	47.81	47.62	47.24	48.63	46.43	52.59	51.24	46.68
TiO2	1.31	1.27	1.24	1.25	1.33	1.32	1.19	1.39	1.4	1.35	1.37	1.58
Al2O3	14.15	15.34	15.04	15.7	15.74	14.99	16	14.35	15.7	15.76	16.4	15.69
Fe2O3	10.93	11.87	11.39	11.76	11.2	11.52	10.56	11.65	12.2	9.86	10.56	12.7
MnO	0.16	0.17	0.17	0.14	0.16	0.17	17	0.16	0.17	0.16	0.19	0.15
MgO	8.04	9.08	9.17	8.32	7.87	8.52	7.73	7.95	5.15	3.14	3.54	7.61
CaO	11.91	9.4	6.75	7.35	10.79	10.8	13.2	10.23	14.55	13.35	11.88	9.76
Na2O	1.89	2.15	3.66	2.3	2.15	1.86	1.65	2.83	1.92	1.73	2.74	2.24
K2O	na	0.38	0.13	0.72	na	na	na	0.12	0.11	0.2	0.15	0.27
P2O5	0.11	na	na	0.11	0.1	0.11	na	0.1	0.16	0.14	0.13	0.1
LOI	2.38	3.19	3.18	3.87	2.8	2.98	2.43	1.89	1.55	1.5	1.43	2.96
FE*MG	1.22	1.18	1.12	1.27	1.28	1.22	1.23	1.32	2.13	2.83	2.68	1.5
Ba	2	20	14	22	5	2	5	15	2	15	4	2
Ce	4	6	4	5	2	6	5	3	9	4	10	10
Cr	291	391	322	318	326	328	333	350	233	316	312	394
Cu	83	56	64	62	84	82	86	74	76	31	68	58
Ga	9	17	14	17	20	13	12	12	19	21	20	19
La	1	1	1	1	1	1	1	1	2	1	2	1
Nb	5	4	4	4	5	5	5	5	7	4	4	4
Nd	16	10	11	12	12	10	11	15	13	13	18	15
Ni	80	123	87	97	80	94	102	86	76	100	101	96
Pb	15	18	12	11	9	11	12	7	13	13	13	13
Rb	6	10	7	28	6	5	3	8	3	8	6	8
Sr	149	109	54	102	134	118	159	119	256	311	226	144
V	274	274	278	284	279	281	255	305	360	214	233	299
Y	37	31	27	33	34	36	30	35	52	33	34	34
Zn	120	116	262	108	88	91	79	86	84	70	71	109
Zr	89	78	73	73	81	82	69	85	93	107	112	98
Cs	0.46		0.53		0.12			0.41		0.31		
Hf	2.3		2		2.3			2.2		2.6		
Sc	41.6		39		42.6			43.8		38.7		
Ta	0.39		0.14		0.226			0.2		0.41		
Th	na		na		na			na		na		
U	na		0.18		na			na		0.16		
La	2.39		1.53		2.24			2.61		2.87		
Ce	9.33		7.32		7.75			7.96		10.1		
Pr	1.42		1.22		1.33			1.45		1.79		
Nd	9.89		7.89		8.2			8.99		11.09		
Sm	3.07		2.44		2.82			3.01		3.31		
Eu	1.14		0.93		1.07			1.11		1.23		
Gd	4.73		3.79		4.35			4.58		4.7		
Dy	5.47		4.4		5.05			5.21		5.07		
Ho	1.25		0.99		1.14			1.19		1.12		
Er	3.38		2.49		3.43			3.49		3.18		
Yb	3.43		2.65		3.18			3.24		2.9		
Lu	0.53		0.42		0.5			0.51		0.45		

Appendix I

	spv14	spv15	spv16	spv17	spv18	spv23	spv24	spv26	spv27	spv28	spv37	spv38
SiO2	48.41	47.07	47.03	47.94	46.45	49.77	47.82	47.88	48.19	42.15	47.68	48.67
TiO2	1.52	1.36	1.34	1.41	1.65	1.14	1.22	1.58	1.36	1.71	1.42	1.45
Al2O3	14.18	15.31	16.43	16.91	14.5	18.8	17.12	16.44	16.47	17.55	16.63	15.3
Fe2O3	12.11	11.92	11.31	10.26	13.2	8.79	9.79	11.18	10.75	12.16	10.26	10.37
MnO	0.18	0.17	0.16	0.14	0.14	0.12	0.15	0.21	0.14	0.15	0.14	0.18
MgO	8.61	6.85	8.33	7.8	9.35	6.86	7.95	8.13	8.13	11.84	7.58	7.96
CaO	8.51	12.25	7.92	8.87	7.95	5.79	9.16	6.37	6.79	5.58	8.92	10.29
Na2O	3.17	2.05	2.76	3.1	2.39	4.93	3.22	3.37	3.2	2.69	2.44	2.77
K2O	0.32	na	0.59	0.44	0.25	0.25	0.34	0.7	0.94	0.25	1.14	0.14
P2O5	0.13	0.11	0.11	0.13	0.15	na	0.11	0.17	0.12	0.14	0.24	0.13
LOI	2.72	2.47	3.57	3.22	3.96	3.37	3.05	3.7	3.61	5.3	3.3	2.67
FE*MG	1.27	1.57	1.22	1.18	1.27	1.15	1.11	1.25	1.2	0.92	1.22	1.17
Ba	2	5	24	6	9	16	7	28	24	2	16	25
Ce	4	5	5	9	7	9	8	13	5	10	10	8
Cr	200	286	454	305	246	248	301	333	383	522	387	392
Cu	26	45	184	71	67	33	57	149	111	145	76	73
Ga	19	19	13	19	15	13	14	13	16	21	21	16
La	2	1	1	2	2	1	1	2	7	1	2	1
Nb	4	5	4	4	5	3	3	5	4	4	6	5
Nd	11	4	15	16	16	8	15	16	20	16	11	16
Ni	60	82	161	78	80	139	82	126	134	195	109	72
Pb	26	13	13	12	12	12	10	12	13	10	24	15
Rb	17	5	15	21	10	8	9	28	25	8	17	7
Sr	98	131	147	206	115	155	222	146	163	158	204	251
V	383	312	262	238	333	163	194	287	267	260	273	242
Y	41	38	25	28	38	21	31	41	30	52	35	30
Zn	710	98	235	98	130	72	74	149	962	349	94	74
Zr	93	85	87	107	111	82	95	118	95	116	113	101
Cs	53.6	0.63	8.1			0.64			30.7		3.4	
Hf	2.5	2.3	2.4			2			2.4		2.7	
Sc	44.3	43.8	38.7			29.5			40.1		39.9	
Ta	0.16	0.129	0.12			0.14			0.118		0.38	
Th	na	na	na			0.109			0.105		0.37	
U	0.39	0.12	na			0.11			0.17		0.22	
La	2.46	2.33	2.88			2.75			5.26		5.78	
Ce	9.09	8.57	7.93			8.8			10.8		14.88	
Pr	1.62	1.41	1.69			1.4			2.33		2.42	
Nd	10.79	9.29	9.59			8.59			12.08		12.09	
Sm	3.51	3.12	3.17			2.44			3.94		3.52	
Eu	1.3	1.17	1.19			0.96			1.47		1.27	
Gd	5.31	4.82	4.56			3.5			5.61		4.73	
Dy	6.18	5.61	5.11			3.62			6.2		4.96	
Ho	1.4	1.27	1.14			0.81			1.4		1.11	
Er	3.9	3.66	3.52			2.21			4.15		3.35	
Yb	3.76	3.46	3.11			2.02			3.62		2.89	
Lu	0.58	0.54	0.48			0.32			0.58		0.45	

Appendix I

	spv39	spv40	spv41	spv42	spv43	spv44	spv47	spv48	spv49	spv50	spv51	spv52
SiO2	46.6	45.51	47.86	48.87	49.05	48.05	47.84	47.62	48.77	47.86	44.2	45.67
TiO2	1.82	1.7	1.28	1.22	1.24	1.32	1.31	1.53	1.39	1.25	0.98	1.12
Al2O3	18.69	17.44	15.66	14.45	14.66	14.98	19.46	15.19	14.42	15.19	14.77	16.19
Fe2O3	11.59	13.83	11.3	11.4	11.04	11.6	12.12	12.71	11.69	12.13	9.06	11.53
MnO	0.12	0.16	0.16	0.14	0.17	0.17	0.15	0.18	0.17	0.15	0.12	0.13
MgO	6.05	7.14	7.45	8.08	7.5	8.69	4.62	6.91	8.38	8.09	6.86	7.15
CaO	7.17	6.95	11.17	9.68	11.77	8.72	5.67	10.39	9.16	10.45	13.12	10.44
Na2O	2.57	2.23	1.93	2.68	2.25	2.51	3.09	2.05	2.49	1.07	2.38	2.62
K2O	1	1.17	na	0.65	na	0.63	2.24	0.22	0.27	0.22	0.7	0.7
P2O5	0.23	0.17	0.1	0.11	0.11	na	0.12	0.13	0.15	0.1	na	na
LOI	3.94	3.96	2.79	2.57	2.36	2.92	3.49	2.83	2.77	3.48	7.41	3.9
FE\MG	1.72	1.74	1.37	1.27	1.32	1.2	2.36	1.66	1.26	1.35	1.19	1.45
Ba	29	83	2	13	2	14	99	5	5	2	39	2
Ce	15	9	6	5	5	6	8	9	7	2	4	9
Cr	239	341	377	362	350	358	493	340	363	306	270	450
Cu	68	83	69	55	71	57	26	40	40	76	68	119
Ga	15	19	21	12	12	17	14	18	16	17	13	17
La	2	5	1	1	1	1	1	1	1	1	1	1
Nb	10	7	4	5	3	5	5	6	4	5	3	5
Nd	20	18	12	12	9	12	10	18	11	6	11	5
Ni	120	99	80	88	88	90	96	87	84	92	52	67
Pb	10	16	12	8	13	15	10	12	13	13	14	15
Rb	26	41	5	15	3	26	38	8	12	10	26	21
Sr	392	155	125	106	135	121	149	180	112	151	160	230
V	190	318	275	302	270	263	218	354	338	316	242	272
Y	28	39	33	32	33	34	29	38	33	34	25	27
Zn	111	122	85	95	97	130	98	110	106	117	109	112
Zr	154	116	81	74	78	78	78	95	80	80	63	70
Cs		27.7			0.12			1.41		6.2		7.2
Hf		3.1			2.1			2.4		2.1		1.9
Sc		47.9			45.7			48.7		43		44.6
Ta		0.28			0.18			0.18		0.11		0.1
Th		0.36			0.128			na		na		na
U		0.31			0.27			0.15		0.13		0.17
La		5.69			2.56			2.81		2.38		2.42
Ce		13.55			7.77			8.59		7.13		8
Pr		2.3			1.41			1.64		1.36		1.34
Nd		11.77			8.69			9.59		8.09		8.49
Sm		3.67			2.92			3.4		2.7		2.77
Eu		1.34			1.09			1.24		1		1.12
Gd		5.19			4.5			4.98		4.18		4.06
Dy		5.85			5.16			5.75		4.9		4.65
Ho		1.31			1.17			1.3		1.14		1.03
Er		4.04			3.52			4.07		3.42		3.12
Yb		3.53			3.22			3.63		3.17		2.78
Lu		0.55			0.51			0.56		0.51		0

APPENDIX 2

SIMPLE CONSTRUCTIONS IN TRANSPRESSION ZONES

A simple geometric construction for determining the orientations of the principal axes of the infinitesimal strain ellipsoid in tectonic zones, where the boundaries are obliquely convergent or divergent, was given by McCoss (1986). The construction, which assumes volume conservation, may be produced in two different ways.

Method A: determination of the orientation of principal axes

If the direction of the zone boundary displacement (\underline{S}) is known, then the orientation of the principal axes of the infinitesimal sectional strain ellipse may be determined according to the following construction (Fig. A2.1):

- i) Draw the zone boundaries (or boundary) in such an orientation that \underline{S} , the displacement vector points up the page.
- ii) Draw a circle centred on the zone boundary. The radius is arbitrary, although half the normal separation places the constructed principal axes in the centre of the zone.
- iii) Construct a line normal to the zone boundary from the centre of the circle.
- iv) Identify the two points at the top and base of the circle which lie on a vertical diameter, i.e. parallel to \underline{S} .
- v) Project lines from the top and base of the circle through the intersection of the circle and the zone normal. The line from the top of the circle parallels the maximum principal axis of the infinitesimal sectional strain ellipse, and that from the base parallels the minimum axis.

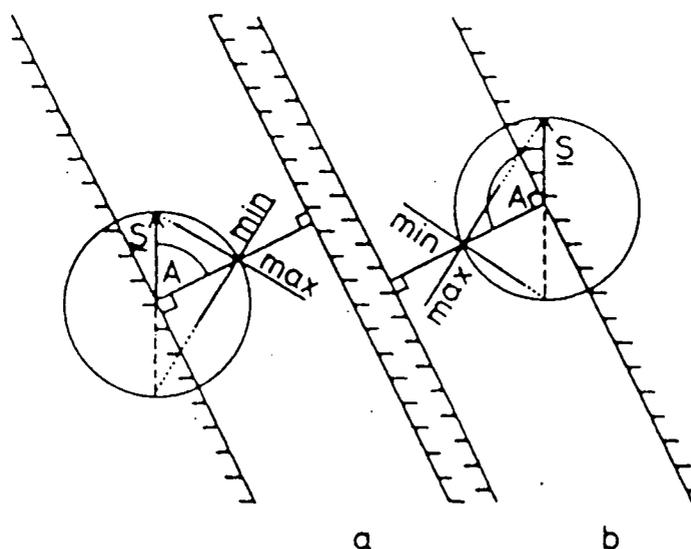


Figure A2.1 A geometrical construction for a) transpression, and b) transtension, which facilitates the identification of the maximum and minimum principal axes of the infinitesimal strain ellipse (after McCoss, 1986).

Method B: determination of the orientation of the zone boundary displacement vectors

If the orientation of the maximum and/or minimum principal axes of the infinitesimal sectional strain ellipse can be inferred from first incremental structures, e.g. tension gashes, stylolites, late folds and faults, etc. then the orientation of the zone displacement vectors may be determined as follows (Fig. A2.2).

- i) Draw a circle, of arbitrary radius, centred in the zone boundary.
- ii) Project a line normal to the zone from the centre of the circle.
- iii) Draw lines parallel to the maximum and minimum principal axes of the infinitesimal sectional strain ellipse, so that their intersection coincides with that of the circle and zone normal.
- iv) Project the maximum and minimum axes so that they intersect the circle, and draw a line between these intersections, through the circles centre. This line parallels the displacement vector. The vector sense is from the intersection with the minimum axis, towards that with the maximum axis.

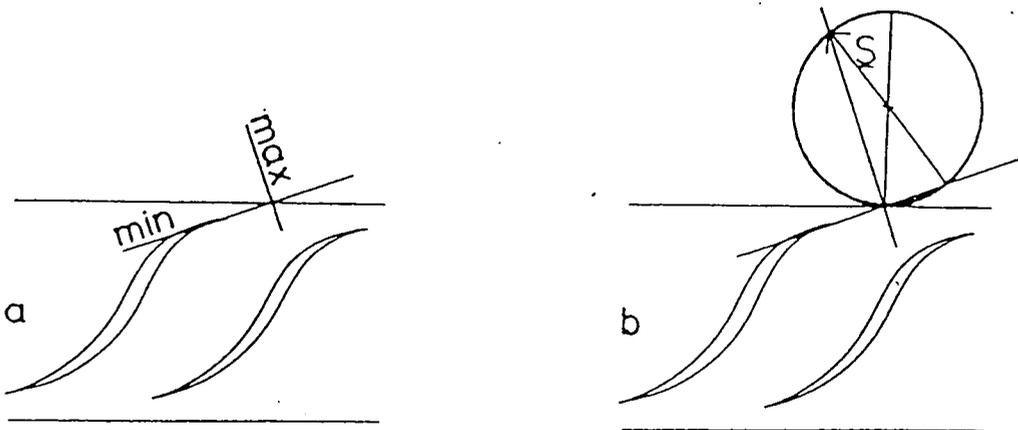


Figure A2.2 The use of sigmoidal tension gashes to infer the minimum and maximum principal axes of the infinitesimal strain ellipse (a) and the resultant constructed zone displacement vector (b). After McCoss (1986).

The relative magnitudes of the principal axes of the infinitesimal strain ellipsoid may be determined using the geometrical construction given in A2.1. This is done by measuring the angle, (A) between \underline{S} and the zone normal. The infinitesimal relative magnitudes and ellipsoid shapes are a function of A , and are given in Table 1. McCoss (1986) recognised eight precise values of A , which bound distinctly different regimes, using this to tighten the definition of vague terms such as compression, wrench and extension (see 1.2.2, *i*) *constant incremental strain* for further detail).

Table 1. The relative magnitudes, implied tectonic regimes, and ellipsoid shapes for given values of A , as $\underline{S} \rightarrow 0$, in a vertical tectonic zone. After McCoss (1986)

A	Relative Magnitudes	Tectonic regimes	Ellipsoid shape
Acute	see below	Transpression	Oblate
Obtuse	see below	Transtension	Prolate
0	$\lambda_v > \lambda_1 = 1 > \lambda_2$	Plane compression	Planar
$0 < A < 70.5^\circ$	$\lambda_v > \lambda_1 > 1 > \lambda_2$	General compression	Oblate
70.5°	$\lambda_v = \lambda_1 > 1 > \lambda_2$	ASTP	AS Oblate
$70.5^\circ < A < 90^\circ$	$\lambda_1 > \lambda_v > 1 > \lambda_2$	General wrench (c-field)	Oblate
90°	$\lambda_1 > \lambda_v = 1 > \lambda_2$	Plane wrench	Planar
$90^\circ < A < 109.5^\circ$	$\lambda_1 > 1 > \lambda_v > \lambda_2$	General wrench (e-field)	Prolate
109.5°	$\lambda_1 > 1 > \lambda_v = \lambda_2$	ASTT	AS Prolate
$109.5^\circ < A < 180^\circ$	$\lambda_1 > 1 > \lambda_2 > \lambda_v$	General extension	Prolate
180°	$\lambda_1 > \lambda_2 = 1 > \lambda_v$	Plane extension	Planar



BRITISH GEOLOGICAL SURVEY
1:10,000 Series
SHEET SX 73 NE

HEIGHTS IN METRES

ORDNANCE SURVEY

SHEET SA 73 NE
& part of 73 SE

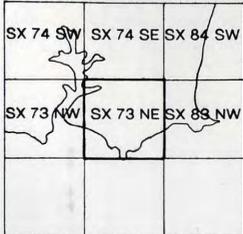
Based on large scale surveys between 1952-65, revised 1994
Contour interval 5m
© 1991

INDEX AND EXPLANATION

- DRIFT
- ALLUVIUM
 - HEAD
- LANDSLIP
-
- SHEET COMPLEX
- METROSANT
 - MICA SCHIST

- INCLINED FAULT, DIP IN DEGREES
CROSSMARKS DENOTE PHASE (e.g. D₁)
- INCLINED CLEAVAGE, DIP IN DEGREES
CROSSMARKS DENOTE PHASE (e.g. D₂)
- INCLINED AXIAL PLANE OF MINOR FOLD, DIP IN DEGREES
CROSSMARKS DENOTE PHASE (e.g. D₃)
- PLUNGING AXIS OF MINOR FOLD, PLUNGE IN DEGREES
CROSSMARKS DENOTE PHASE (e.g. D₃)
- PLUNGING MINERAL LINEATION, PLUNGE IN DEGREES
CROSSMARKS DENOTE PHASE (e.g. D₃)
- PLUNGING LINEATION, PLUNGE IN DEGREES
CROSSMARKS DENOTE PHASE (e.g. D₃)
- DIRECTION OF VERGENCE OF FOLD
- GEOLOGICAL BOUNDARY, DRIFT
- GEOLOGICAL BOUNDARY, SOLID
- FAULT, CROSSMARK ON DOWNTHROW SIDE
ARROW IN METRES, DIP IN DEGREES
- SENSE OF RELATIVE MOVEMENT
- THRUST FAULT, BENDS TO HANGINGWALL
- AXIAL TRACE OF MAJOR ANTIFORM
- AXIAL TRACE OF MAJOR SYNFORM
- AXIAL TRACE OF MINOR ANTIFORM
- AXIAL TRACE OF MINOR SYNFORM

DIAGRAM SHOWING ADJOINING 1:10,000 NATIONAL GRID SHEETS



BRITISH GEOLOGICAL SURVEY

1:10,000 Series

SHEET SX 63 NE

HEIGHTS IN METRES

ORDNANCE SURVEY

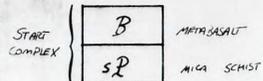
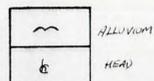


SHEET SX 63 NE

Based on large scale surveys in 1850, revised 1860-88
Contour interval 2m
1880 SOUTH HAMPS DISTRICT

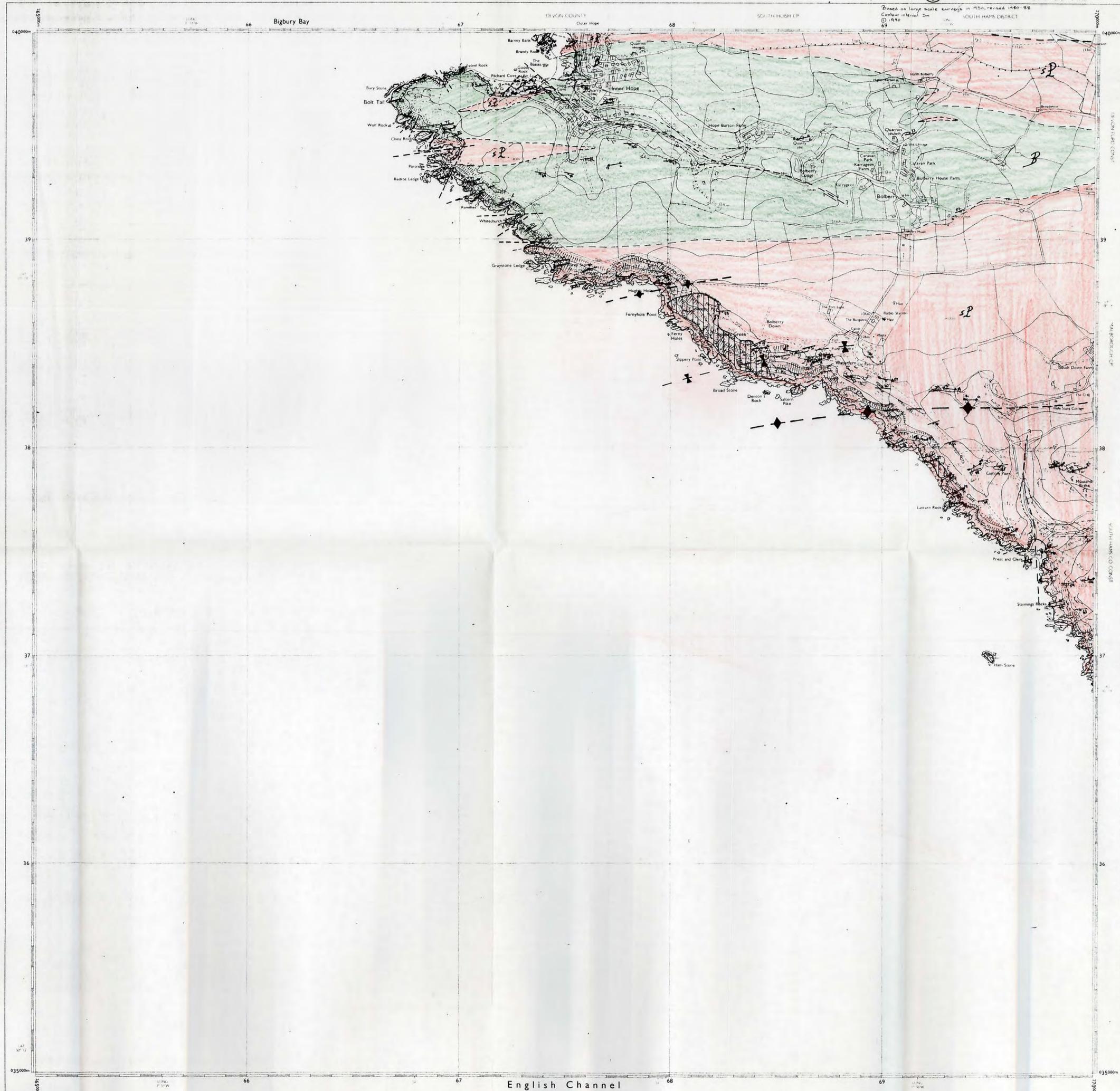
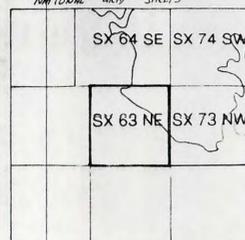
INDEX AND EXPLANATION

DRIFT



- INCLINED FOLIATION, DIP IN DEGREES
- INCLINED CLEAVAGE DIP IN DEGREES, CROSS MARKS DENOTE PHASE (C₁, D₁)
- INCLINED AXIAL PLANE OF MAJOR FOLD, DIP IN DEGREES
- PLUNGING AXIS OF MINOR FOLD, PLUNGE IN DEGREES, CROSS MARKS DENOTE PHASE (C₁, D₁)
- PLUNGING AXIAL LINEATION PLUNGE IN DEGREES, CROSS MARKS DENOTE PHASE (C₁, D₁)
- PLUNGING LINEATION PLUNGE IN DEGREES, CROSS MARKS DENOTE PHASE (C₁, D₁ ROLLING)
- DIRECTION OF VERGENCE OF FOLD
- GEOLOGICAL BOUNDARY, DRIFT
- GEOLOGICAL BOUNDARY, SOLID
- FAULT, CROSS MARK ON DOWNTHROW SIDE, THROW IN METRES, DIP IN DEGREES
- SENSE OF RELATIVE MOVEMENT
- THRUST FAULT, BARBS TO HANGING WALL
- AXIAL TRACE OF MAJOR UNIFORM
- AXIAL TRACE OF MAJOR SHARP

DIAGRAM SHOWING ADJOINING 1:10,000 NATIONAL GRID SHEETS



English Channel

SX 63 NE

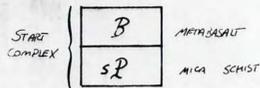
BRITISH GEOLOGICAL SURVEY

1:10,000 Series

SHEET SX 63 NE

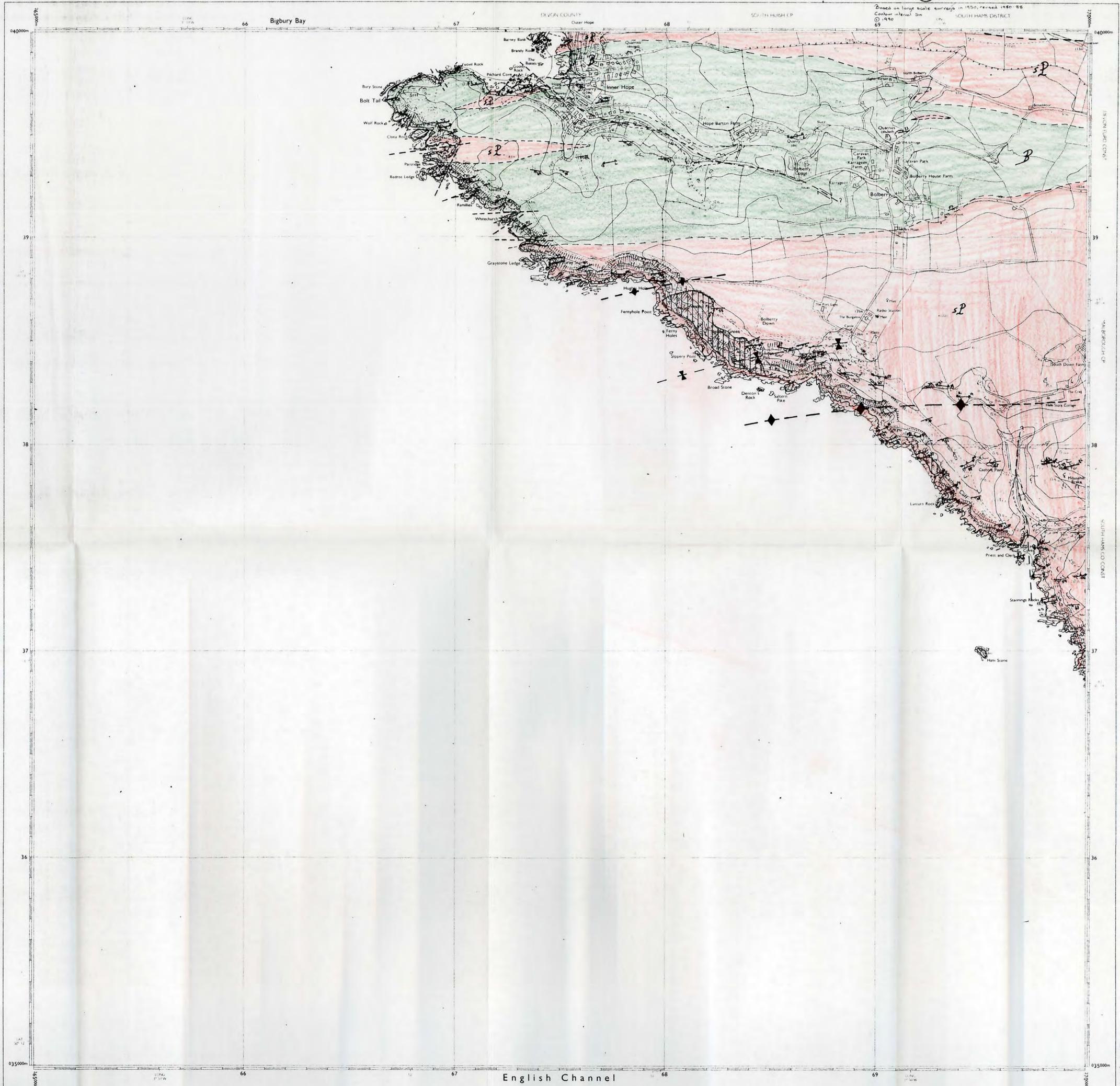
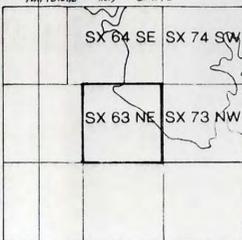
INDEX AND EXPLANATION

DRIFT



- INCLINED PLANE, DIP IN DEGREES
- INCLINED PLANE, DIP IN DEGREES, CROSS-MARKS DENOTE PHASE (e.g. D₂)
- INCLINED AXIAL PLANE OF MAJOR FOLD, DIP IN DEGREES
- PLUNGING AXIS OF MAJOR FOLD, PLUNGE IN DEGREES, CROSS-MARKS DENOTE PHASE (e.g. D₂)
- PLUNGING MINERAL LINEATION PLUNGE IN DEGREES, CROSS-MARKS DENOTE PHASE (e.g. D₂)
- PLUNGING LINEATION PLUNGE IN DEGREES, CROSS-MARKS DENOTE PHASE (e.g. D₂)
- DIRECTION OF VERGENCE OF FOLD
- GEOLOGICAL BOUNDARY, DRAFT
- GEOLOGICAL BOUNDARY, SOLID
- FAULT, CROSS-MARK ON DOWNTHROW SIDE, THROW IN METRES, DIP IN DEGREES
- SENSE OF RELATIVE MOVEMENT
- THRUST FAULT, BARBS TO HANGINGWALL
- AXIAL TRACE OF MAJOR ANTIFORM
- AXIAL TRACE OF MAJOR SYNFORM

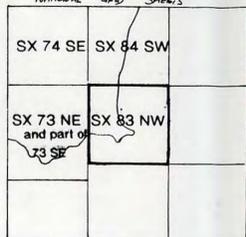
DIAGRAM SHOWING ADJOINING 1:10,000 NATIONAL GRID SHEETS



INDEX AND EXPLANATION

- DEIFT**
- ALLUVIUM
 - ROAD
 - LANDSLIP
- START COMPLEX**
- METABASALT
 - MICA SCHIST
- INCLINED FOLIATION, DIP IN DEGREES**
- INCLINED CLEAVAGE, DIP IN DEGREES, CROSSMARKS DENOTE PHASE (e.g. D₂)
 - INCLINED AXIAL PLANE OF MINOR FOLD, DIP IN DEGREES
 - PLUNGING AXIS OF MINOR FOLD, PLUNGE IN DEGREES, CROSSMARKS DENOTE PHASE (e.g. D₂)
 - PLUNGING MINERAL LINEATION, PLUNGE IN DEGREES, CROSSMARKS DENOTE PHASE (e.g. D₂)
 - PLUNGING LINEATION, PLUNGE IN DEGREES, CROSSMARKS DENOTE PHASE (e.g. D₂ RODDING)
- A** DIRECTION OF VERGENCE OF FOLD
- — —** GEOLOGICAL BOUNDARY, DEIFT
- — —** GEOLOGICAL BOUNDARY, SOLID
- F** FAULT, CROSSMARKS ON DOWNTHROW SIDE, ARROW IN METRES, DIP IN DEGREES
- SENSE OF RELATIVE MOVEMENT
- T** THRUST FAULT, BARBS TO HANGWALL
- AXIAL TRACE OF MAJOR ANTIFORM
- AXIAL TRACE OF MAJOR SYNFORM
- AXIAL TRACE OF MINOR ANTIFORM
- AXIAL TRACE OF MINOR SYNFORM

DIAGRAM SHOWING ASSIGNING 1:10,000 NATIONAL GRID SHEETS



BRITISH GEOLOGICAL SURVEY

1:10,000 Series

SHEET SX 64 SE

HEIGHTS IN METRES

ORDNANCE SURVEY



SHEET SX 64 SE & part of 64 SW

INDEX AND EXPLANATION

DRIFT



ALLUVIUM

SAND

SHEET COMPLEX



MENDIPS

M19 EAST

Inclined foundation, dip in degrees

Inclined axial plane of minor fold dip in degrees, crossmarks denote phase (g, d)

Plunging axis of minor fold plunge in degrees, crossmarks denote phase (g, d)

Direction of widening of fold

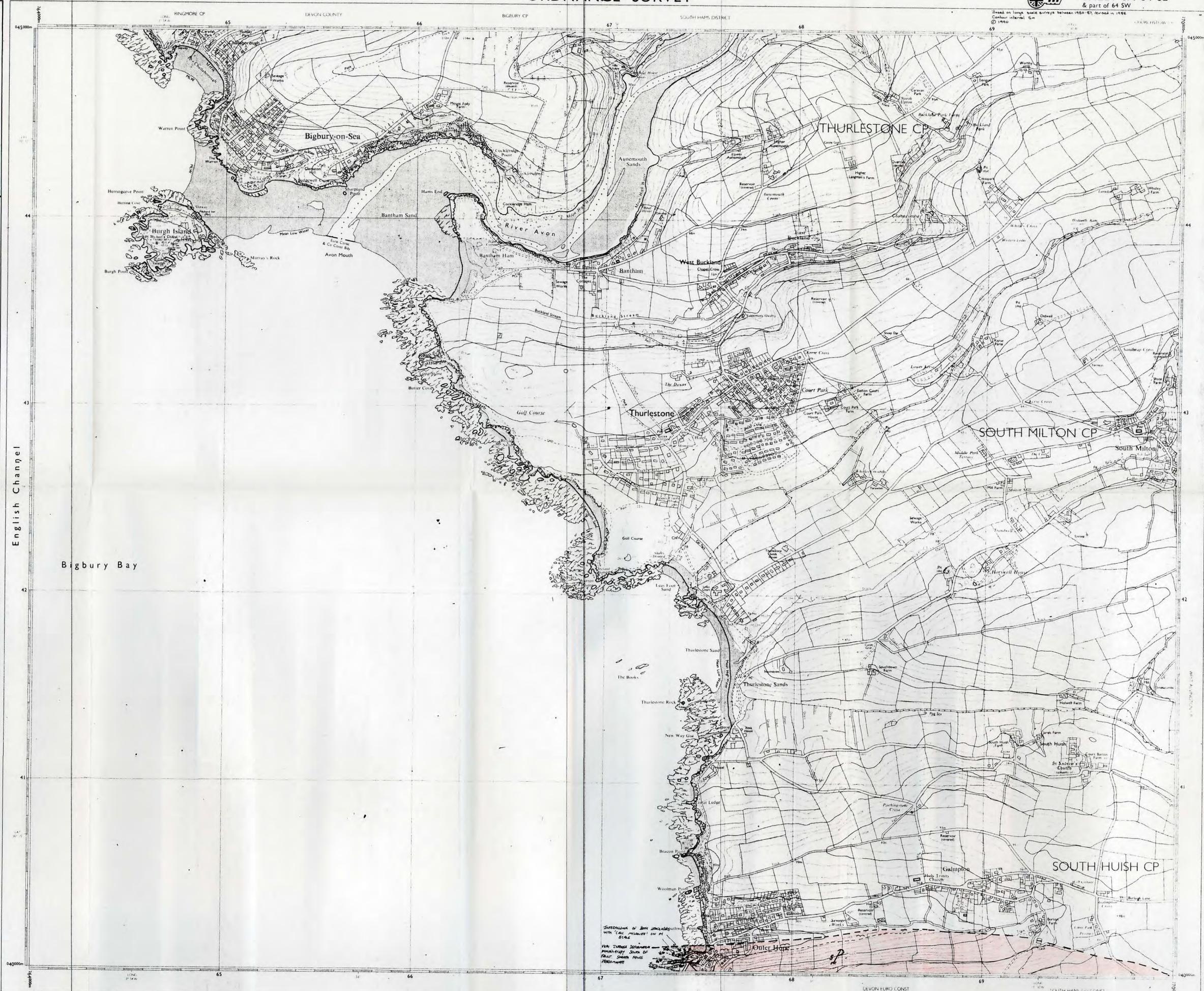
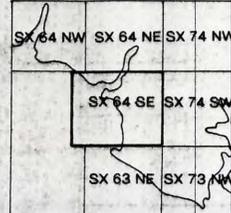
Geological boundary, drift

Geological boundary, solid

Fault, crossmark on downthrow side, arrow in metres, dip in degrees

Sense of relative movement

Diagram showing automatic 1:10,000 National Grid sheets



INTERPOLATED 20m contours (contour interval 10m) with 10m contours on 1:10,000 scale. MAIN contour interval 10m. MINOR contour interval 5m. FAULT shown with crossmark on downthrow side. ARROW in metres, DIP in degrees. SENSE OF RELATIVE MOVEMENT.

SX 64 SE & part of 64 SW