Geol. Mag. **152** (2), 2015, pp. 242–268. © Cambridge University Press 2014
doi:10.1017/S0016756814000260

Structure of the Dingle Peninsula, SW Ireland: evidence for the nature and timing of Caledonian, Acadian and Variscan tectonics

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(*Received* 27 February 2014; *accepted* 13 May 2014; *first published online* 14 July 2014)

Abstract – The Palaeozoic rocks of the Dingle Peninsula provide a record of the evolution of the Caledonides, Acadides and Variscides. The succession ranges from Early Ordovician deep-water sediments, through Silurian shallow marine to non-marine sediments and volcanic rocks to an Old Red Sandstone (ORS) succession topped by Carboniferous marine shales. Comparison of structural styles in the unconformity-bounded groups, together with a detailed analysis of fault zones, allows the tectonic history to be deduced. The older rocks record Caledonian processes on the margin of Avalonia during Early Ordovician time and convergence then soft collision with Laurentia during Silurian time. The Dingle Basin was developed during the late Silurian – Early Devonian transtension in the Iapetus suture zone and was inverted in the latest Emsian Acadian orogenic episode. Post-Dingle Group ORS groups in the north of the peninsula were deposited in a syn-rift footwall block to the main Munster Basin. The Acadian transpressional and Munster Basin extensional structures were reactivated or overprinted in the Variscan deformation such that Acadian folds are transected by Variscan cleavage in both plan and vertical views. After Iapetus closure, changes in the tectonic regime are believed to be a result of adjustments in the geometry of subduction of the Rheic Ocean.

Keywords: transpression, Devonian, suture, mylonites, cleavage, reactivation.

1. Introduction

The Dingle Peninsula lies within the Iapetus suture zone of the Caledonides, marking the transition of Caledonian terranes associated with Laurentia and those of Avalonian affinity (Todd, Murphy & Kennan, 1991; Figs 1, 2). Ordovician and Silurian rocks provide a record of the closure of Iapetus in the Caledonian cycle. A late Silurian – early Carboniferous Old Red Sandstone succession brackets the Acadian orogenic episode. Moreover, the peninsula lies just north of the Variscan Front, the supposed northern boundary of more intense deformation associated with the closure of the Rheic Ocean and collision of Gondwana with Laurentia (Fig. 1). The Palaeozoic succession, which ranges in age from Tremadoc (Todd *et al.* 2000) to Mississippian (Pracht, 1996), therefore contains stratigraphic and structural evidence of progressive continental assembly of Pangaea during the Caledonian, Acadian and Variscan tectonic cycles.

Much has been published on the sedimentary record of the peninsula (e.g. Todd, Boyd & Dodd, 1988; Todd, Williams & Hancock, 1988; Todd, 1989, 2000; Boyd & Sloan, 2000; Richmond & Williams, 2000). Less has been published about the structure of the area, although S. P. Todd (unpub. thesis, 1989) and S. D. King (unpub. thesis, 1989) presented structural studies within unpublished PhD dissertations. Meere & Mulchrone (2006) provided strain data from the Old Red Sandstone (ORS) lithologies of the peninsula to conclude that the main penetrative cleavage and intense strain was of Acadian origin, and was developed in a *dextral* transpressive regime. This was in contrast to Todd (1989) who presented structural data from an Acadian fault zone (Minard Head Fault Zone) in the southeast of the peninsula to add to sedimentary observations supporting *sinistral* transpressive tectonics. This paper presents observations, data and interpretations of the structure of the entire Palaeozoic succession. By distinguishing structure styles in the major tectonostratigraphic units which are bounded by unconformities, it is possible to delineate the nature and timing of the tectonic evolution of the area. Particular attention is paid to fault zones, some of which include cataclasites and mylonites. This regional picture in turn provides evidence of the sequence of plate/terrane tectonic evolution of this part of Pangaea.

A review of the stratigraphy of the peninsula is presented first. This is followed by an outline of the main phases of deformation, starting with the youngest Variscan and then addressing older phases that affect the older stratigraphic units. The paper describes the tectonic events in reverse order to illuminate the overprinting effects of younger phases on existing structures. Further detail on the structure of fault zones is then presented. Finally, these observations are combined into a tectonostratigraphic history for the area which, in conclusion, is placed in a regional context.

2. Stratigraphy

The oldest proven rocks on the peninsula are the Tremadoc to Arenig shales, sandstones and mélange of the Annascaul Formation (Fig. 3; Todd *et al.* 2000).

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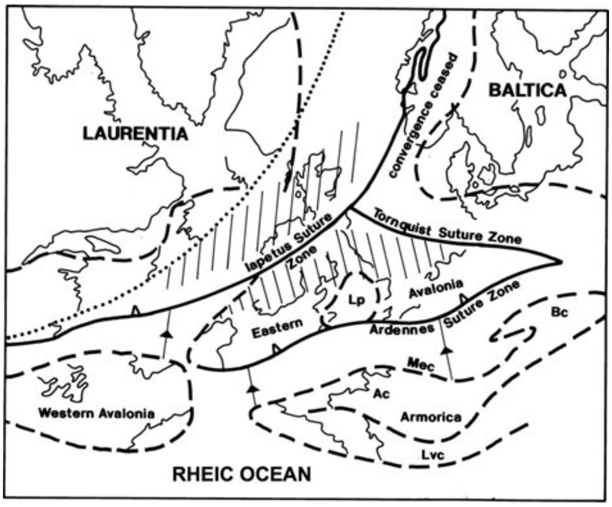

Figure 1. Sketch of the palaeogeography of the British and Irish Caledonides during Early Devonian time (from Soper & Hutton, 1984). Lp – London platform; Ac – Amorican craton; Bc – Bohemian craton; Lvc – Ligurian–Vosgian cordillera.

These Ordovician deep-water sediments have been correlated with the Ribband Group of SE Ireland and, as such, are assigned a Leinster terrane or Avalonian affinity (Todd *et al.* 2000). The Annascaul Formation is confined to two inliers in the east of the peninsula, the most extensive being the Annascaul inlier (Fig. 4). The Annascaul Formation is overlain unconformably by the mudstones, tuffs and limestones of the Wenlock Ballynane Formation (Dunquin Group, Fig. 3; Parkin, 1976; Todd *et al.* 2000). The Ballynane Formation is overlain, apparently conformably, by the mudstones/siltstones of the Caherconree Formation and the sandstones/mudstones of the Derrymore Glen Formation (Dunquin Group, Fig. 3). Both these formations are proven to be Ludlow in age through graptolite and shelly faunas (Parkin, 1976).

This Silurian succession in the eastern Bull's Head, Annascaul and Derrymore Glen inliers is quite different in facies from the main Llandovery–Ludlow succession in the Dunquin inlier, some 20–40 km to the west (Fig. 4; Holland, 1969; Sloan & Bennett, 1990; Sloan & Williams, 1991; Boyd & Sloan, 2000; Higgs & Williams, 2011). The oldest rocks in the west are the slates of the Coosglass Formation, now proven to be of Llandovery – early Wenlock age through palynomorphs (Higgs & Williams, 2011). These rocks are separated by a fault from the rest of the Wenlock–Ludlow Dunquin Group, which comprises non-marine to shallow-marine mudstones, siltstones, sandstones, various volcaniclastic deposits and lavas (Sloan & Bennett, 1990; Sloan & Williams, 1991; Boyd & Sloan, 2000).

The Dingle Group, a succession of mudstones, sandstones and conglomerates, was deposited in a non-marine basin (Todd, Boyd & Dodd, 1988). A conformable transition from the uppermost shallow-marine sediments of the Dunquin Group is inferred in two localities. On the mainland, a partially exposed sequence from the upper Ludlovian Croaghmarhin Formation (Dunquin Group) to the Bull's Head Formation (Dingle Group) is interpreted to be the result of the progradation of a wave-dominated barrier shoreline (Boyd & Sloan, 2000). A better-exposed section crops out on Inishabro, one of the Blasket Islands off the western tip of the Dingle Peninsula. This section also appears to be transitional from the Silurian Inishnabro Formation to the Bull's Head Formation (Todd, 1991). In other places, the contact between the Dunquin and Dingle groups is clearly faulted or unconformable (Todd, Williams & Hancock, 1988). The variation between unconformable and apparently conformable relationships is interpreted to be the consequence of contrasting subsidence styles

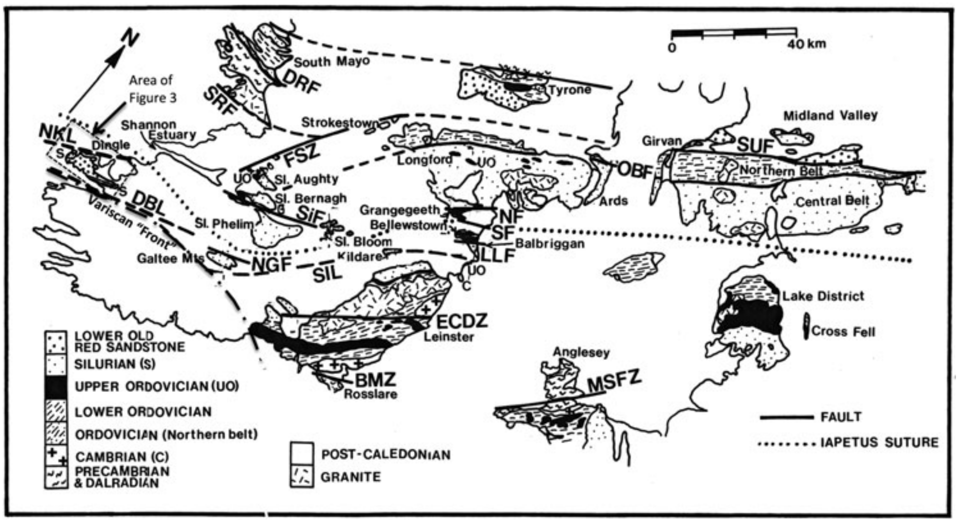

Figure 2. Geological map of the Iapetus suture zone of Ireland and Britain (from Todd, Murphy & Kennan, 1991). BMZ – Ballycogly Mylonite Zone; DBL – Dingle Bay Lineament; DRF – Doon Rock Fault; ECDZ – East Carlow Deformation Zone; FSZ – Fergus Shear Zone; LLF – Lowther Lodge Fault; MSFZ – Menai Straits Fault Zone; NF – Navan Fault; NGF – North Galtees Fault; NKL – North Kerry Lineament; OBF – Orlock Bridge Fault; SF – Slane Fault; SIF – Silvermines Fault; SIL – South Ireland Lineament; SRF – Skird Rocks Fault; SUF – Southern Uplands Fault.

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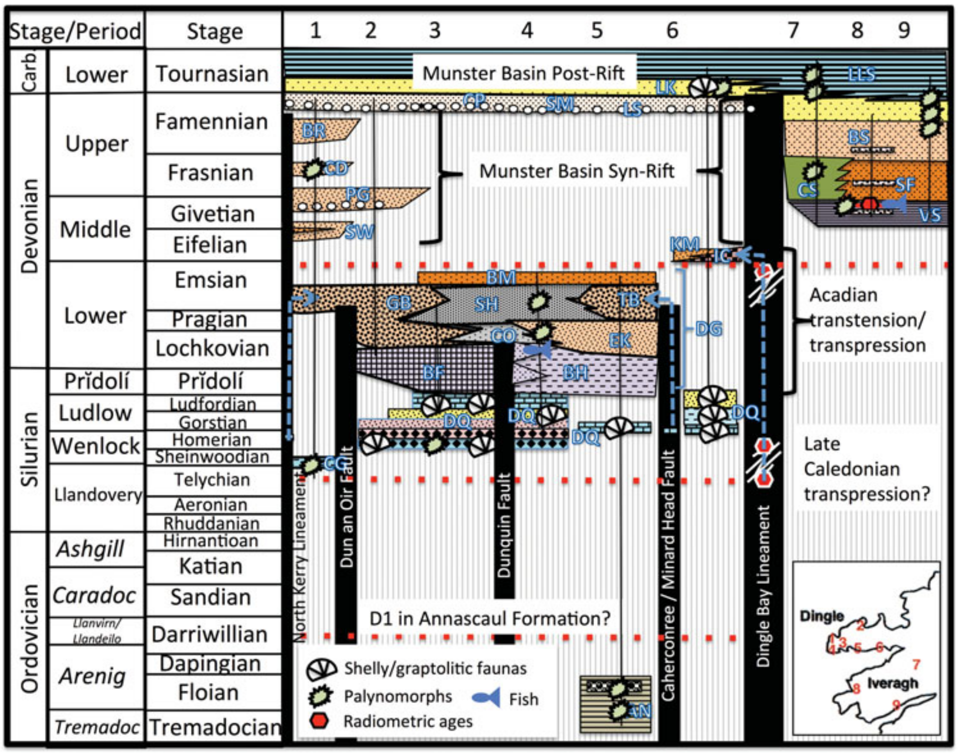

Figure 3. Chronostratigraphy of the Palaeozoic rocks of the Dingle Peninsula (updated from Todd 2000, fig. 3). Compiled from various sources by area. 1. Todd, Williams & Hancock (1988), Pracht (1996), Richmond & Williams (2000) and Higgs & Williams (2011); 2. Todd, Boyd & Dodd (1988), Todd, Williams & Hancock (1988) and Richmond & Williams (2000); 3. Holland (1988), Todd, Boyd & Dodd (1988) and Todd, Williams & Hancock (1988); 4. Holland (1987, 1988), Todd, Boyd & Dodd (1988), Todd, Williams & Hancock (1988), Todd (1989, 1991) and Higgs (1999); 5. Parkin (1976), Todd, Boyd & Dodd (1988) and Todd *et al.* (2000); 6. Horne (1974), Parkin (1976), Higgs & Russell (1981) and Pracht (1996); 7. Higgs & Russell (1981), Pracht (1996) and Williams *et al.* (1997, 2000); 8. Russell (1978), Higgs & Russell (1981) and Williams *et al.* (1997, 2000); 9. Higgs & Russell (1981) and Williams *et al.* (2000). AN – Annascaul Formation; BF – Ballyferriter Formation; BH – Bull's Head Formation; BM – Ballymore Formation; BR – Ballyroe Group; BS – Ballinskelligs Sandstone Formation; CD – Carrigduff Group; CG – Coosglass Formation; CO – Coumeenoole Formation; CP – Cappagh Sandstone Formation; CS – Chloritic Sandstone Formation; DQ – Dunquin Group; EK – Eask Formation; GB – Glashabeg Conglomerate Formation; IC – Inch Conglomerate Formation; KM – Kilmurry Sandstone Formation; LK – Lack Formation; LLS – Lower Limestone Shales; LS – Lough Slat Conglomerate Formation; PG – Pointagare Group; SF – St Finan's Sandstone Formation; SH – Slea Head Formation; SM – Slieve Mish Group; SW – Smerwick Group; TB – Trabeg Conglomerate Formation; VS – Valentia Slate Formation. The time scale is from Gradstein, Ogg & Schmitz (2012). The Welsh-derived stage names commonly used in the UK for the Ordovician are provided as well as the international standard.

between the basin depocentre, intra-basin highs and the margins of the basin (Fig. 3; Todd, Boyd & Dodd, 1988; Boyd & Sloan, 2000).

The age of the Dingle Group is constrained by three fossil finds. K. Higgs and colleagues (pers. comm. 2014) have discovered fish fragments and spores from the Trabane Member in the upper part of the Bull's Head Formation. These fossils not only support a lacustrine sand flat margin origin for these deposits, but indicate a middle–late Lochkovian age for the upper part of the Bull's Head Formation (lower Dingle Group; Fig. 3). A palynological assemblage obtained from grey siltstones in the alluvial apron deposits of the Eask Formation is ascribed an early late Pragian age (Higgs, 1999; Fig. 3). A further assemblage from a medial to upper position in the Dingle Group, within the axial fluvial deposits of the Slea Head Formation which is the lateral equivalent of the alluvial fan/apron facies of the Trabeg and Glashabeg formations (Todd, Boyd & Dodd, 1988; Todd, 1989, 2000), has been interpreted to be early–?middle Emsian in age (Higgs, 1999; Fig. 3). Biostratigraphic evidence therefore shows that most of the Dingle Group is Early Devonian in age. However, the age of the base of the Dingle Group still remains uncertain as no biostratigraphical data have yet been found in the basal Heterolithic Member of the Bull's Head Formation. The Silurian–Devonian boundary most likely occurs in the condensed lacustrine sequence in the lower part of the Bull's Head Formation (Higgs, 1999, p. 195).

The north limb of the Feohanagh Anticline, along the north coast of the peninsula, contains a distinct succession of unconformity-bounded ORS groups (Figs 3, 4). The Dingle Group is represented only by a variable thickness of mudstones and conglomerates of the Glashabeg Formation (Todd, Williams & Hancock, 1988). The Smerwick Group overlies the Dingle Group, either in faulted or unconformable contact (Todd, Williams & Hancock, 1988). Richmond & Williams (2000) proposed that the Smerwick Group is of Lower Devonian age like the Dingle Group, and was translated into place as an allochthonous terrane. In the

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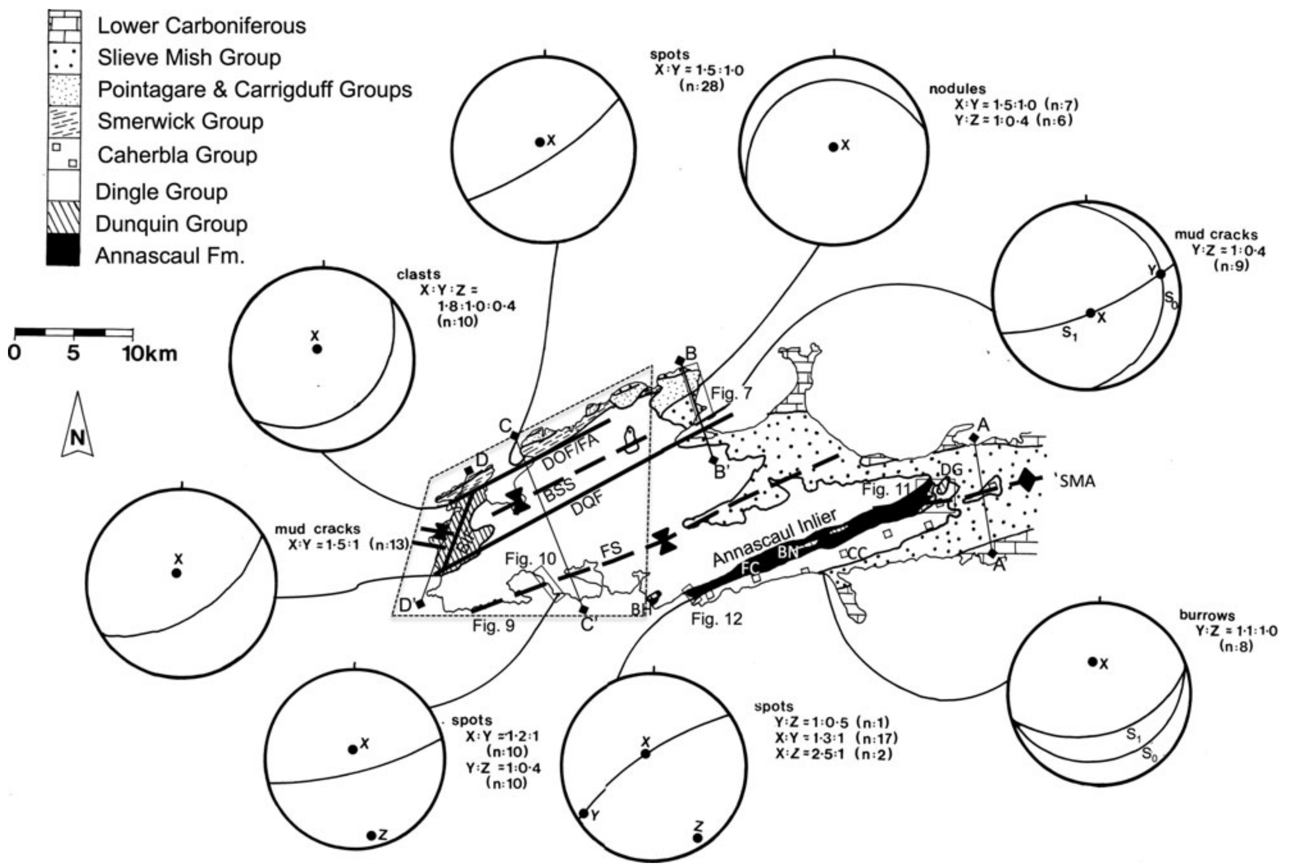

Figure 4. Simplified geological map of the Dingle Peninsula showing the main structural elements and the locations of the maps and cross-sections shown in later figures. The lower-hemisphere equal-area projections provide orientations and dimensions of strain data collected during field mapping of the peninsula. BH – Bull's Head inlier; BIF – Ballyickeen Fault; BN – Ballynane; BSS – Ballysitteragh Syncline; CC – Caheracruttia; DG – Derrymore Glen inlier; DOF/FA – Dún an Oír Fault/Feohanagh Anticline; DQF – Dunquin Fault; FC – Farranacarriga; FS – Fahan Syncline; SMA – Slieve Mish Anticline.

absence of definitive age data and the compelling evidence of a major terrane-bounding fault, the simpler interpretation is that the Smerwick Group is younger than the Dingle Group with the boundary as an angular unconformity that has been faulted in some places (Todd, Williams & Hancock, 1988). Four successive unconformity-bounded groups follow the Smerwick Group (Todd, Boyd & Dodd, 1988; Richmond & Williams, 2000). Three of these are confined to the northern Dingle Peninsula; the youngest, the Slieve Mish Group, oversteps the Dingle Group and older rocks. The Carrigduff Group has yielded spores in a solitary and rather sparse assemblage evaluated by K. T. Higgs as early–middle Frasnian in age (L. K. Richmond, unpub. thesis, 1998; Richmond & Williams, 2000). The Slieve Mish Group, cropping out across the Dingle Peninsula, is the youngest ORS group and passes conformably into marine sandstones and mudrocks with Carboniferous fauna and flora (Todd, Boyd & Dodd, 1988; Higgs, Clayton & Keegan, 1988).

On the south side of the Dingle Peninsula the ORS Caherbla Group contains breccio-conglomerates of the Inch Conglomerate Formation and sandstones of the Kilmurry Sandstone Formation. This group is not directly dated, but is generally believed to be of Middle Devonian age (Fig. 3; Todd, Boyd & Dodd, 1988; Williams *et al.* 1997; Todd, 2000; Morrissey *et al.* 2011).

3. Variscan deformation

The expression and amount of end-Carboniferous Variscan strain in the Dingle Peninsula is best gauged by structural assessment of the Slieve Mish Group and other upper Palaeozoic formations which clearly postdate Caledonian and Acadian events. Major reverse faulting on the offshore Dingle Bay and North Kerry lineaments is inferred from structural geometries and stratigraphic separations recorded in onshore outcrop in the Dingle and Iveragh peninsulas (Price & Todd, 1988; Todd, 1989). In the Dingle Peninsula, shortening has produced open to closed, upright folds in the Slieve Mish Group and overlying Carboniferous shales and limestones (Shackleton, 1940; Capewell, 1951, 1965). These folds have an east–west trend in the east of the peninsula but take on a more NE–SW orientation in the west and have wavelengths of up to several kilometres (Figs 4, 5). They developed by buckling, facilitated by interlayer flexural slip as indicated by bed-parallel quartz slickenfibres with lineations oriented normal to fold axes. Rudimentary line length measurement of unbalanced cross-sections indicates that the buckling achieved shortening of about 14 % (Fig. 5).

Throughout the Irish Variscides there is a single cleavage that is typically axial planar to Variscan folds. In the quartz-rich sandstones and conglomerates of the Slieve Mish Group in the Dingle Peninsula, the cleavage is formed as smooth, disjunctive, variably spaced (a few centimetres) and anastomosing pressure solution seams (Fig. 6). In mudrocks, the cleavage is more tightly spaced (less than 1 cm). The cleavage forms strongly convergent fans about anticlines in the Slieve Mish Group; these fans are symmetrical about the axial plane of the folds (Fig. 7) and are therefore axial planar, *sensu lato*, with no transection in plan view. Cleavage fans like this are believed to have formed early in the deformation, prior to buckling (e.g. Cooper *et al.* 1984).

Comparative strain data, collected on a spot basis during field mapping in this research, are presented in Figure 4. A roughly NNW-oriented Z direction with the XY plane vertical is usually indicated by markers such as reduction spots, diagenetic nodules, mudcracks and burrows. A small number of measurements in mudrocks of the Slieve Mish Group suggest that ductile deformation associated with cleavage development has accomplished a strain ratio in the yz plane of the strain ellipsoid, R_s yz, of *c.* 2.5 in the NNW direction (Fig. 4). This appears to have been accompanied by some vertical stretching so R_s xz ~ 3.75. Bulk strain of Slieve Mish Group sandstones appears to have been more heterogeneous with some shortening achieved by the largely unquantifiable effects of volume loss during pressure solution (but see Fig. 6c for an unusual case where volume loss can be observed). An R_f/Ø analysis (a graphical approach for estimating finite strain of deformed elliptical objects) of clasts in sandstones and conglomerates in the Slieve Mish Group at Bull's Head gave R_s yz values in the range of 1.1–1.3 (Meere & Mulchrone, 2006). This approach tends to provide a minimum estimate of bulk strain, as the process is not homogenous as theoretically assumed (Meere & Mulchrone, 2006). *Taenidium barretti* burrow tops exposed on the gently SE-dipping Caherbla Group sandstones in the SE of the peninsula have an average long:short axis ratio of 1:0.7 and are always parallel to structural strike, discordant with palaeocurrent directions. They are therefore considered to have been originally subcircular and deformed by tectonic strain (Morrissey *et al.* 2011). A minimum (assuming X is vertical) R_s yz strain for the burrows in these post-Acadian rocks is *c.* 1.42.

There therefore appears to be a dichotomy between bulk strain between mudrocks and sandstones/conglomerates in the Slieve Mish Group (and other irrefutable post-Acadian groups) which appears to be best explained by assuming that the R_s yz strain in the mudrocks of *c.* 2.5 is homogeneous and is matched in the sandstones/conglomerates through a heterogeneous combination of strain (*c.* 1.1–1.4) and volume loss on cleavage folia (Fig. 6c).

Bulk strain (including a cleavage) and layer buckling therefore account for the Variscan deformation of the Slieve Mish Group and other post-Acadian ORS groups. The former apparently accounts for at least twice as much total strain as the latter, which is similar to findings in the main Variscan Belt to the south (Cooper *et al.* 1984, 1986; Ford, 1987). One implication is that considerable Variscan bulk shortening (probably more than 40 %) must have overprinted Acadian and older structures in the older strata. This includes the Variscan cleavage, present in even the most

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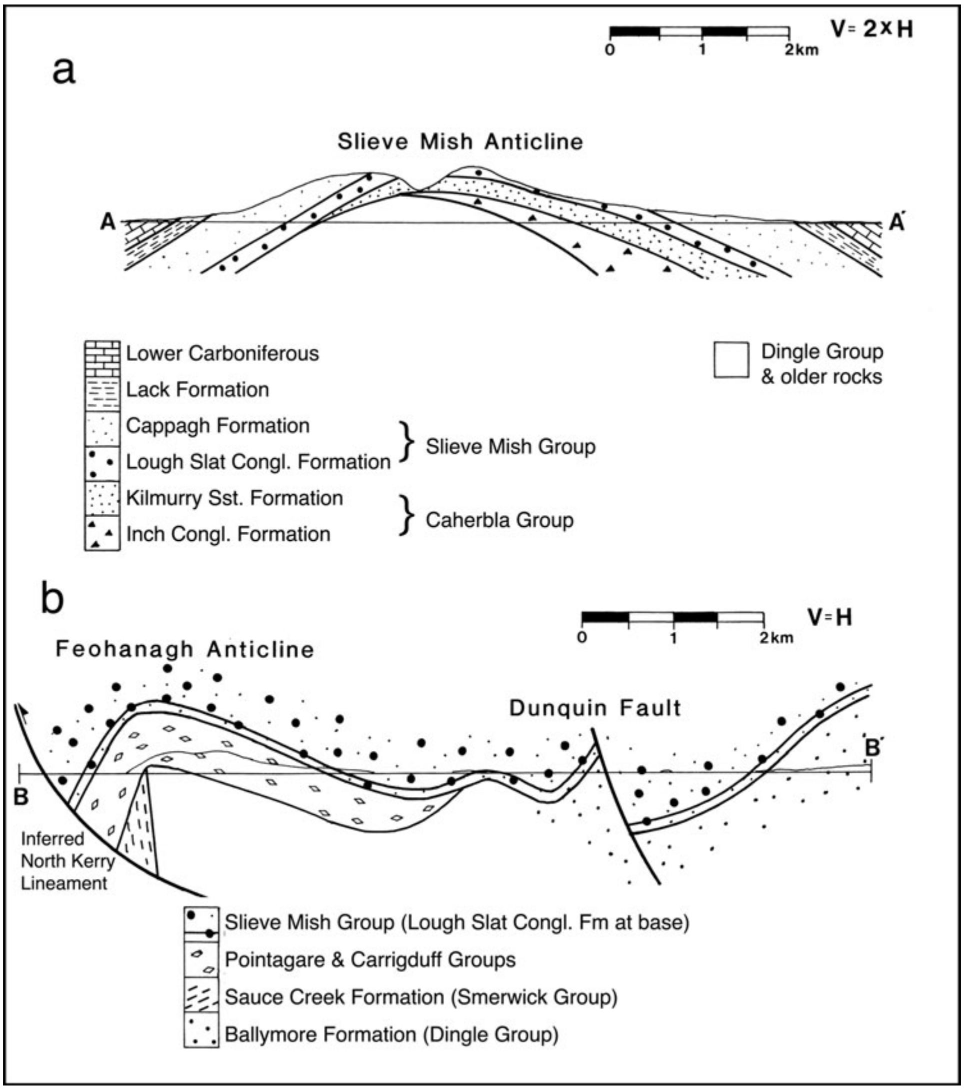

Figure 5. Structural cross-section through the Variscan structures of the Slieve Mish Group. (a) Slieve Mish Anticline (after Capewell 1951). (b) Structural cross-section through the area of Brandon Bay. For locations of the cross-sections see Figure 4.

competent, quartz-rich lithologies in the Slieve Mish Group, passing downwards into older pre-Late Devonian rocks (Fig. 6d, e).

4. Late Devonian extension

The effects of Late Devonian extension of the Dingle area are manifest in faults and fault-block rotations that occurred prior to deposition of the latest Devonian Slieve Mish Group (Horne, 1977; Todd, Boyd & Dodd, 1988; Todd, Williams & Hancock, 1988; Todd, 1989). For example, several ENE-trending normal faults cut the Caherbla Group, but are overstepped by the Slieve Mish Group. A series of such faults form the northern boundary of the Caherbla Group outcrop and the southern boundary of the Annascaul inlier (Fig. 4). For example, the Caherconree Fault displaces the Caherbla Group by several tens of metres but fails to affect the Slieve Mish Group (Horne, 1977). The Minard Head Fault Zone, at the opposite SE end of the outcrop strip, is a Caledonian/Acadian fault zone that was reactivated in a normal down-to-the-south motion (Todd, 1989; see Section 8.a). In the northern footwall of the fault, the Caherbla Group was removed by erosion prior to overstep by the Slieve Mish Group. In the hanging wall, the Caherbla Group was tilted northwards before unconformable deposition of the Slieve Mish Group; it now dips less steeply southwards than the Slieve Mish Group on the southern limb of the Slieve Mish anticline. The product of erosion of the Caherbla and older groups of the Dingle Peninsula may have been deposited as three distinct conglomerate units observed in the otherwise finer-grained Late Devonian alluvium of NW Iveragh (Capewell, 1975; Todd, Boyd & Dodd, 1988; Williams *et al.* 1989, 1997, 2000; Williams, 2000).

Similar extensional faulting and fault-block rotations (cf. Jackson & McKenzie, 1983; Barr, 1987) are believed to have affected the geometry of the post-Dingle Group unconformity-bounded ORS groups in the Feohanagh Anticline (Figs 3, 4, 5; Todd, Boyd & Dodd, 1988; Todd, Williams & Hancock, 1988, Todd, 1989). The Smerwick Group crops out only on the northern limb of the anticline and is inverted, dipping steeply towards the SE (Fig. 8). Successive groups lie

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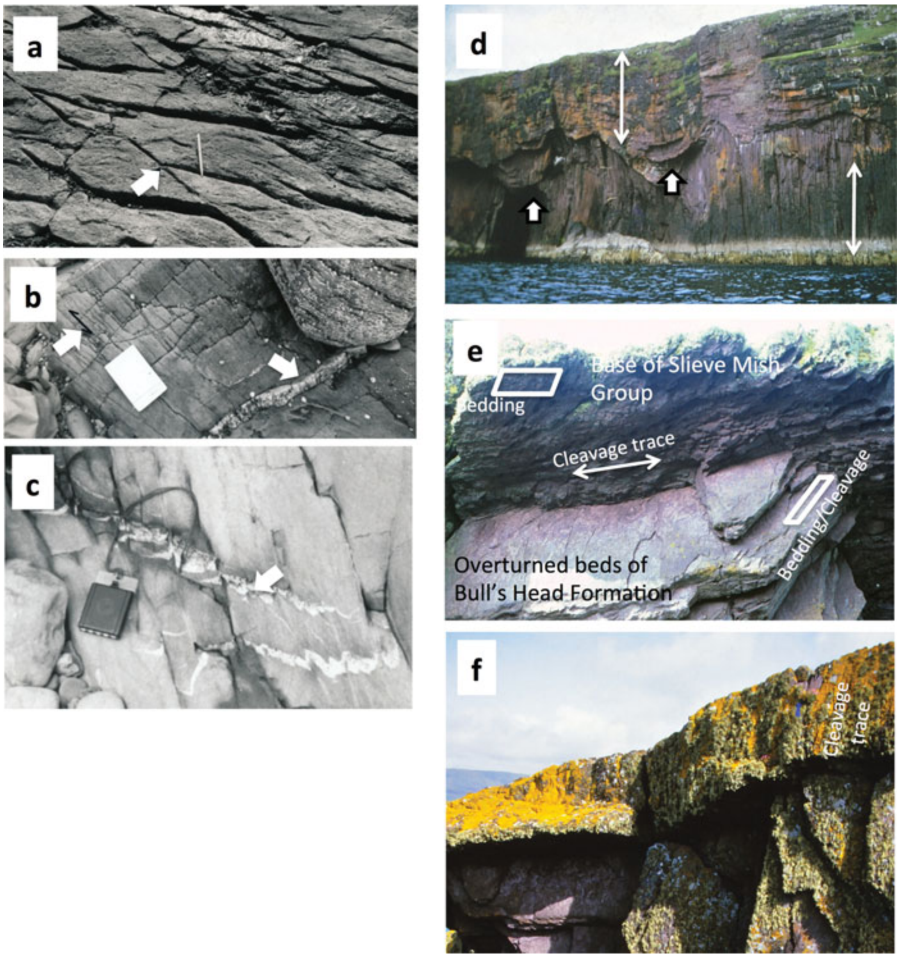

Figure 6. (Colour online) Photographs of the Variscan pressure solution cleavage in sandstones of the Slieve Mish and Dingle groups. (a) Curvi-planar anastomosing cleavage on a sandstone bedding plane (e.g. arrows) at Pointnarianna, west of Bull's Head (Fig. 4; V 491 979). The pen for scale is 14 cm long. (b) Spaced planar cleavage (marked by arrow in left of picture) cut by joints and quartz veins (marked by arrow in right) in a sandstone bed at Caher Point, Brandon Bay (notebook for scale is 20 cm long). (c) Spaced planar cleavage cutting quartz veins in a sandstone bed of the Cappagh Formation at Caher Point, Brandon Bay (closed compass clinometer is 10 cm long). Note the ptygmatic folding and offsets of the quartz vein by the cleavage folia, evidence of substantial volume loss. (d) Cliff face at Pointnarianna showing classic unconformity between gently dipping Slieve Mish Group and underlying steeply dipping and overturned Bull's Head Formation (Dingle Group). The cliff is *c.* 40 m high. Note the channels in the unconformity picked out by the block arrows and the cleavage in the sandstones of the Slieve Mish Group and bedding/cleavage fabric in the Bull's Head Formation picked out by double-headed line arrows. (e) Unconformable contact between base of Slieve Mish Group and Bull's Head Formation at Pointnarianna, viewing underside of bedding plane of sandstone bed at base of Slieve Mish Group. The field of view is *c.* 5 m wide. Note the trace of closely spaced pressure solution cleavage on the base of the Slieve Mish Group, parallel to the intersection of the combined cleavage/bedding fabric in the Bull's Head Formation. (f) Cross-sectional view of same basal unit (*c.* 1.8 m thick) of Slieve Mish Group at Pointnarianna as in (e), with cleavage trace on vertical joint/erosion surfaces.

unconformably on the previous groups, dipping north at progressively lower angles until the Slieve Mish Group oversteps all and extends across the peninsula. Todd, Williams & Hancock (1988) interpreted these geometries as the result of the Feohanagh anticline being developed as a Late Devonian rollover anticline into a NE–SW-trending extensional fault off the current north coast (the North Kerry Lineament; Todd, 1989). This rollover or tilted fault-block lay in the regional footwall of the Munster Basin separated by the Dingle Bay Lineament, across which there was more than 5 km displacement during Late Devonian time (Price & Todd, 1988; Todd, 1989). Successive pulses of extension are believed to have sequentially caused rotation of the pre-existing strata in the Feohanagh Anticline, uplift and erosion of the Dingle regional footwall to the Munster Basin and then subsidence and accommodation of the next group. Note that in Figure 3 the three conglomerate units in Late Devonian NW Iveragh, thought to be products of recycling of the Dingle footwall during uplift, are tentatively time-correlated with the rotational unconformities on the northwest limb of the Feohanagh Anticline (Fig. 8a).

5. Acadian deformation of the Dingle Group

The angular unconformity between the Dingle and Slieve Mish groups, particularly at the classic locality of Pointnarianna (V 491 979; Irish National Grid references are provided for key localities not specifically

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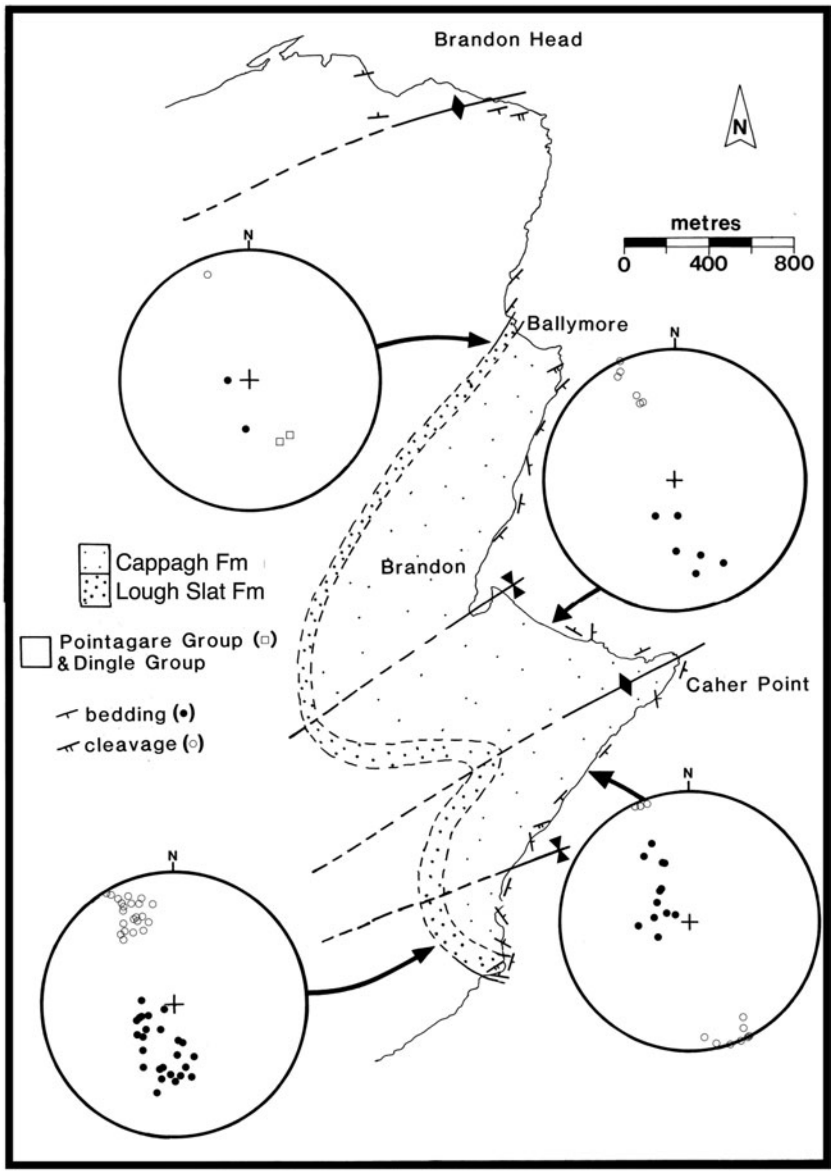

Figure 7. Geological map of the Brandon Bay area to illustrate the Variscan structure of the area. Lower-hemisphere equal-area projections show the variation of bedding and cleavage across macroscopic folds with cleavage forming fans around axial planes, maintaining a high angle to bedding irrespective of position in the fold. For location see Figure 4.

located in figures) west of Bull's Head, has often been cited as the unconformity marking a late Caledonian deformation event associated with the Iapetus closure; this is now interpreted as a distinct event, the Acadian orogeny, that occurred in the Early Devonian some time after the closure of Iapetus (Fig. 6d, e; Shackleton, 1940; Soper, Webb & Woodcock, 1987; Todd, 1989, 2000). At this location the gently dipping conglomerates and sandstones of the Slieve Mish Group lie unconformably on the Bull's Head Formation (lowermost Dingle Group), inverted and dipping steeply NNW on the southern limb of the Fahan Syncline (Figs 6, 8, 9). The Fahan Syncline, like other macroscopic folds in the Dingle Group, is much tighter than any folds in the Slieve Mish Group (Fig. 8a; compare with Fig. 5). The single cleavage in the Dingle Group is steeper than the axial plane of the Fahan syncline and other folds, and often transects the fold in plan view in an anticlockwise direction (Fig. 10) in contrast to the symmetrical cleavage in folds in the Slieve Mish Group described in Section 3 (see also Figs 5, 7). This observation was first used by Shackleton (1940) to argue that the folds in the Dingle Group, including the Fahan syncline, were initiated in the 'Caledonian orogeny' (now distinguished as Acadian) and overprinted by Variscan compression. This interpretation is upheld here (but see Meere & Mulchrone, 2006).

The macroscopic folds in the Dingle Group trend ENE–WSW and have axial planes that dip SSE. The folds verge NNW and short limbs are overturned in that direction (Figs 8a & 9). From south to north across the peninsula, there is a decrease in fold wavelength and amplitude (Fig. 8a). The folds are cut by a number of strike faults, both normal and reverse. Folding and faulting together account for *c.* 40 % shortening in the Dingle Group, estimated by rudimentary

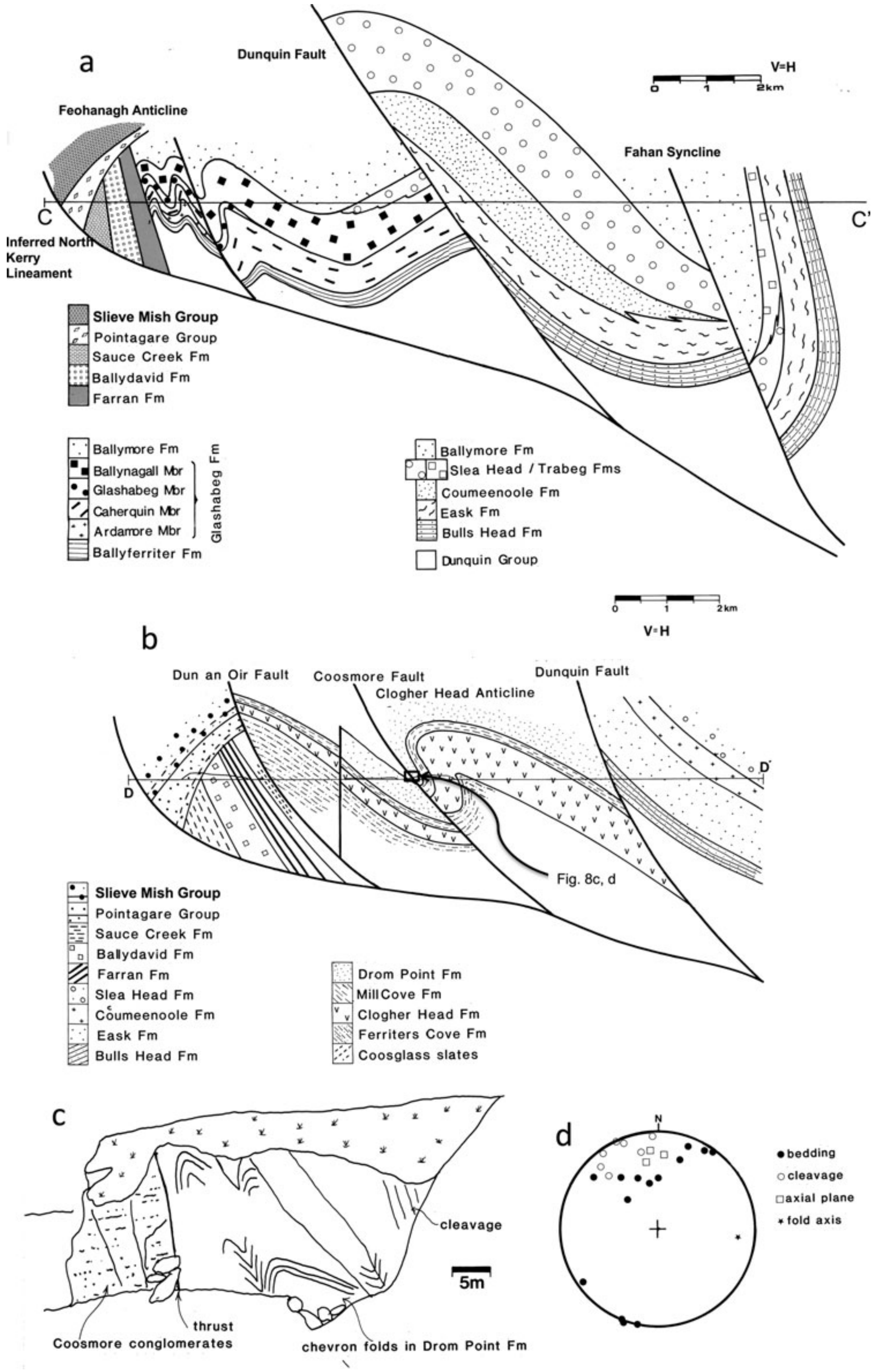

Figure 8. Acadian/Variscan structural style of the Dingle and Dunquin groups. For locations of the cross-sections see Figure 4. (a) Structural cross-section through the Acadian/Variscan structure of the Dingle Group. (b) Structural cross-section through the west coast of the Dingle Peninsula, illustrating the dominantly Acadian structure of the Dunquin Group. (c) Field sketch of the view (looking east) of the east cliff of Coosmore that exposes a cross-section through the faulted core of the Ballyferriter Syncline including chevron folds in the Drom Point Formation (hanging wall) and regularly dipping conglomerates of Glashabeg Formation (footwall). The cliff is *c.* 30 m high. (d) Lower-hemisphere equal-area stereographic projection depicting structural data for the folds in the Drom Point Formation. Note that the single Variscan cleavage strongly transects the axial planes of these Acadian folds.

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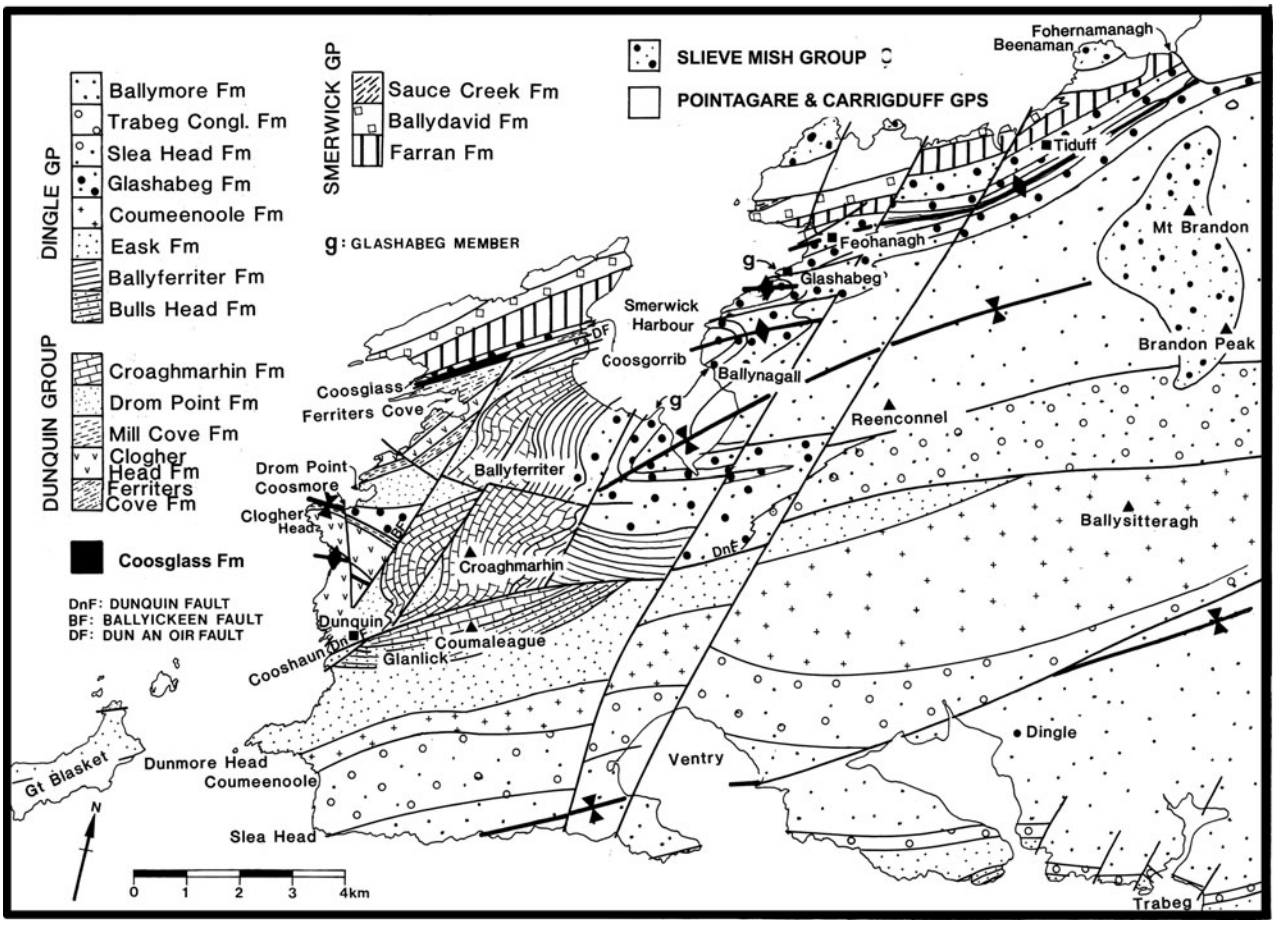

Figure 9. Geological map of the western Dingle Peninsula, illustrating the macroscopic Acadian and Variscan folds and faults.

line length measurement of unbalanced cross-sections.

The Dingle Group is cut by a single, spaced pressure solution cleavage, which is closely spaced to slaty in lithic arenites and mudrocks. Considerable flattening accompanied the development of the cleavage (Fig. 4; Meere & Mulchrone, 2006). Strain observations in mudrocks such as reduction spots suggest an R_s yz strain of 2.5–3.5 in a NNW–SSE direction (Fig. 4; Meere & Mulchrone, 2006). Similarly to the Slieve Mish mudrocks, there was vertical stretching with an R_s xz strain of *c.* 3.8 (Meere & Mulchrone, 2006). Strain in sandstones and conglomerates is more heterogeneous, involving a cleavage partitioned to the matrix with volume loss difficult to quantify and vertical stretching evidenced by mica beards on sandstone clasts and ladder veins (Meere & Mulchrone, 2006, fig. 4). The comparison of strain above and below the Acadian unconformity is fraught with uncertainties, including: debate on the level of the Acadian unconformity in the north of the peninsula (see above); paucity of mudrocks in the irrefutable post-Acadian ORS in comparison with mudrocks in the Dingle Group; contrasting sandstone/conglomerate lithologies ranging from lithic wacke sandstones and petromict conglomerates (with labile clasts) in the Dingle Group to quartz arenite sandstones and oligomictic conglomerates (with indurate clasts) in the Slieve Mish Group; and the effect of different bedding orientations and structural locations. A compilation of the available quantitative data indicates that the Dingle Group was considerably shortened by folding and faulting during the Acadian event and these folds were probably tightened further by Variscan bulk shortening (Table 1). Bulk strain appears to be mostly Variscan in origin with, leaving aside lithological differences and volume loss estimates, R_s yz strain differences between pre-Acadian and post-Acadian strata ranging from 12 % in mudrocks to 25 % in sandstones/conglomerates. The Variscan cleavage is therefore axial planar (in fans) to folds in the Slieve Mish Group, forming early in the Variscan deformation prior to buckling, but is non-axial planar and superimposed on Dingle Group folds to transect axial planes in both plan and cross-section.

6. Acadian deformation of the Dunquin Group

There is also evidence that some Acadian structures began to develop through deposition of the Dingle Group. For example, in the northwest of the peninsula at coastal exposures in Coosmore, Coosgorrib and Coosglass, the Dunquin and Dingle Group succession is attenuated compared to thicker sequences to the south (Fig. 9; Todd, Williams & Hancock, 1988). The thinned Dingle Group sequence is composed of conglomerates of the Glashabeg Formation and

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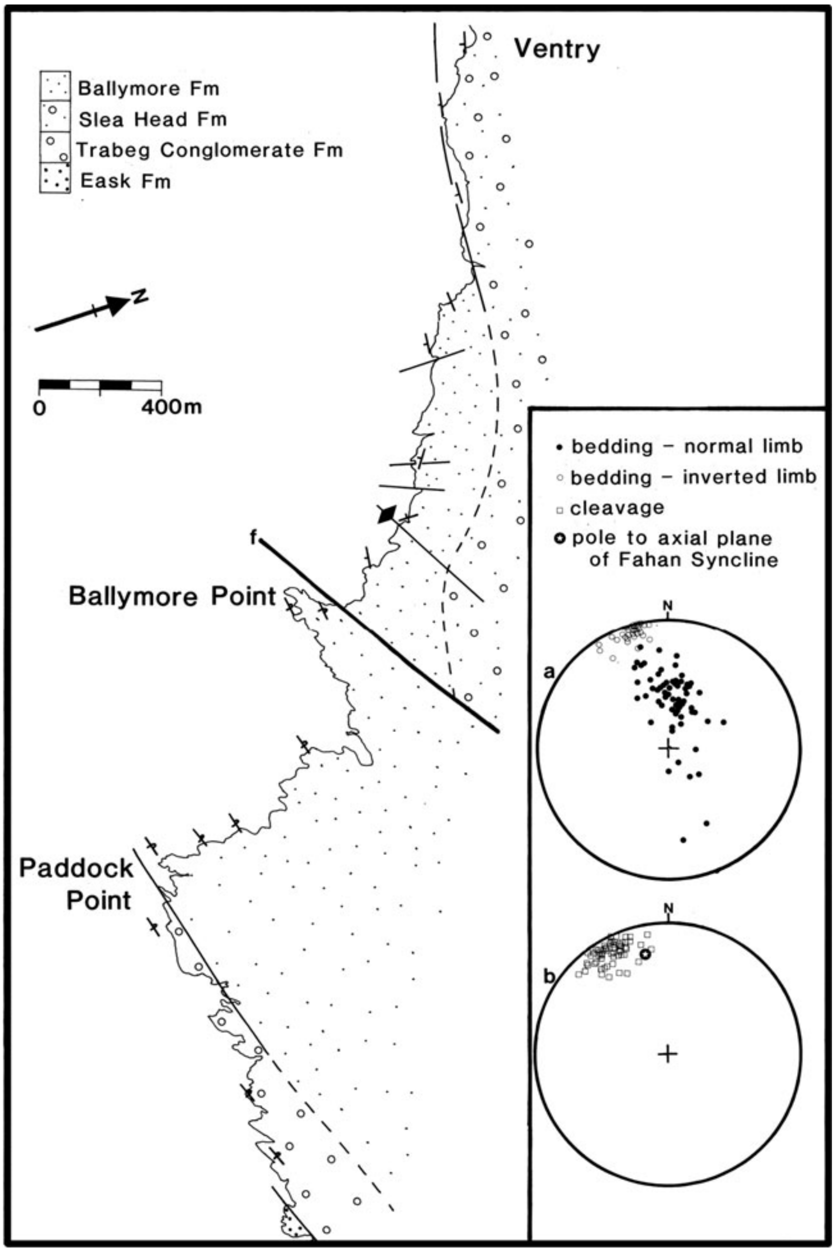

Figure 10. Geological map of the Ballymore area to illustrate the Acadian/Variscan style of the faulted core of the Fahan Syncline. Lower-hemisphere equal-area stereographic projections depict (a) poles to bedding and cleavage and (b) the pole to the axial plane of the Fahan Syncline. Note how the Variscan cleavage has a steeper dip and strikes anticlockwise of the Acadian-initiated Fahan Syncline, and is much more consistent in orientation compared to its development in purely Variscan folds (Fig. 7). For location see Figure 4.

unconformably oversteps onto older Dunquin Group strata in a north and west direction (Fig. 9; Todd, Williams & Hancock, 1988). This same area is bounded to the east by the largely inferred NNE-trending Ballyicken Fault and to the south by the Dunquin Fault (Figs 4, 9; Todd, Williams & Hancock, 1988). The macroscopic folds of this domain have an orientation and morphology that contrasts with the folds in the Dingle Group to the east (Figs 4, 8b, 9). The folds here have a WNW orientation, are tighter than folds in the Dingle Group and are markedly discordant to the single ubiquitous cleavage (Fig. 8b–d). From the angular unconformity and the anomalous folds in the Dunquin Group, it is inferred that this area was being deformed and uplifted in early Dingle Group times; the ancestral Ballyickeen and Dunquin Faults form the structural boundaries, dividing this area from the main depocentres of the Dingle Basin to the south and east where the much thicker Dingle Group follows conformably from the Dunquin Group. These basin margin effects are interpreted to have caused the coarser influxes of sand and paraconglomerate into the lacustrine Bull's Head Formation and recycling of Dunquin Group lithotypes into the Dingle Group (Todd, Boyd & Dodd, 1988; Boyd & Sloan, 2000; Todd, 2000). The Dingle Group conglomerate formations contain sedimentary evidence of differential subsidence, episodic fluvial plain tilting and the strike-lengthening of a major fan lithosome by

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Table 1. Summary of strain data illustrating shortening of the Dingle Group (pre-Acadian) and Caherbla, Pointagare and Slieve Mish groups (post-Acadian) in a NNW–SSE direction, approximately parallel to the *YZ* plane.

Phase	Acadian and Variscan	Variscan only
Group	Dingle Group	Slieve Mish, Pointagare and Caherbla groups
Folding and faulting	*c.* 40 %; Fig. 8	*c.* 14 %; Fig. 5
Bulk strain in mudrocks (reduction spots, mud cracks and nodule)	R_s *yz* strain *c.* 2.5, Fig. 4; R_s *yz* strain c. 2.5–3.1 (Meere & Mulchrone, 2006)	R_s *yz* strain *c.* 2.5, Fig. 4
Bulk strain in sandstones (burrow tops)		R_s *yz* strain *c.* 1.1, Fig. 4 R_s *yz* strain *c.* 1.4 in Caherbla Group (Morrissey *et al.* 2011)
Clasts in sandstones	R_s *xz* strain 1.3–1.5 in lithic wackes (Meere & Mulchrone, 2006), so R_s *yz* strain estimated to be *c.* 1.2	R_s *xz* strain 1.1–1.2 in quartz arenites (Meere & Mulchrone, 2006), so R_s *yz* strain estimated to be *c.* 1.1
Clasts in conglomerates	R_s *xz* strain 1.5–1.9 in petromict conglomerates (Meere & Mulchrone, 2006)	R_s *xz* strain *c.* 1.25 in oligomict conglomerates (Meere & Mulchrone, 2006)
Volume loss in sandstones	Cleavage folia partitioned to matrix; vertical stretching indicated by ladder veins (Meere & Mulchrone, 2006)	Pressure solution of cleavage folia (Fig. 6)
Summary	40 % shortening through folding/faulting and R_s *yz* strain of *c.* 2.8 accomplished homogenously in mudrocks and heterogeneously in sandstones/conglomerates by strain (*c.* 1.5) and volume loss	14 % shortening through folding/faulting and R_s *yz* strain of *c.* 2.5 accomplished homogenously in mudrocks and heterogeneously in sandstones/conglomerates by strain (*c.* 1.2) and volume loss

syn-sedimentary strike-slip displacement of the feeder system (Todd, Boyd & Dodd, 1988; Todd, 1989; Todd & Went, 1991).

A picture of ongoing late Silurian – early Devonian deformation, subsidence and finally inversion of the Dingle Basin to produce the Acadian structure emerges from the combination of stratigraphic and structural observations. The contemporaneous development of compressional structures adjacent to areas of rapid subsidence suggests a transtensional/transpressive regime (Todd, Boyd & Dodd, 1988; Todd, 1989, 2000; Boyd & Sloan, 2000).

7. Caledonian/Acadian deformation of the Annascaul Formation

The early Ordovician Annascaul Formation provides evidence of Caledonian deformation that is pre-Wenlock in age, as well as bearing the overprints of Acadian deformation, Devonian extension and Variscan compression. There are at least three cleavages observed together in the Annascaul Formation, and the presence of a fourth is inferred from collation of structural styles and local overprinting relationships (Figs 11, 12).

Annascaul Formation mudrocks contain a bedding-parallel slaty cleavage labelled S* (after the definition of Powell & Rickard, 1985), deformed by at least two, possibly three, later cleavages. S* is developed by the preferred dimensional orientation (PDO; Elliott & Williams, 1988) of illite and chlorite grains. The fabric post-dates, but mimetically overgrows, a depositional fabric of illite and chlorite grains. The fabric is very typically parallel to bedding, occasionally divergent (<15°). S* augens competent clasts in the mélange units, but passes through shale clasts. Early slump folds (Parkin, 1976) are also cut by S*. S* is interpreted to be the product of burial metamorphism and is also observed in the corollary parts of the Ordovician Ribband Group in east and southeast Ireland (Shannon, 1979; Murphy, 1987).

F_1 folds are distinguished in Minard Bay, the best exposure of the Annascaul Formation structure, as tight to isoclinal Class 1C folds that have upright axial planes and axes that plunge SW or NE (Fig. 12). One macroscopic F_1 fold is mapped in Foilaniska where an upright synclinal core is exposed (Fig. 12). South of the fold hinge, later superimposed F_2 folds are downward facing. The F_1 fold is overturned to the NW and has a wavelength greater than 1.6 km. S* is crenulated in the hinges of F_1 folds where it is cut by S_1, which is axial planar to F_1 and forms a disjunctive cleavage with some crenulation of S*. On the limbs of F_1 folds a very strong phyllitic, papery fabric parallels bedding with S_1 parallel to and enhancing S*, forming disjunctive pressure solution folia.

Annascaul D_2 structures are much more widespread and easy to identify than D_1. F_2 folds are steeply inclined to upright, steeply to moderately plunging and tight to isoclinal. In pelites, the folds have a Class 1C profile; isolated sandstone units form Class 1B folds with rounded hinges; packages of interbedded greywacke and shale form chevron folds. These folds are parasitic to larger macroscopic F_2 folds and have wavelengths of 1–6 m. The macroscopic antiforms have wavelengths of 20–300 m (Fig. 12).

An S_2 cleavage, approximately axial planar to F_2 folds, is generally restricted to (or at least discernible in) the hinge zones of F_2 folds or the intensely deformed D_2 shear zones described in more detail in the following section. S_2 tends to be parallel to the axial surfaces of F_2 folds, but slightly transects the fold hinge in an anticlockwise sense in some places.

In the Kilteenban–Caherconree area (Fig. 11), S* and S_2 are generally not parallel and S* has an unusual orientation dipping NE. F_2 fold axes plunge east or west

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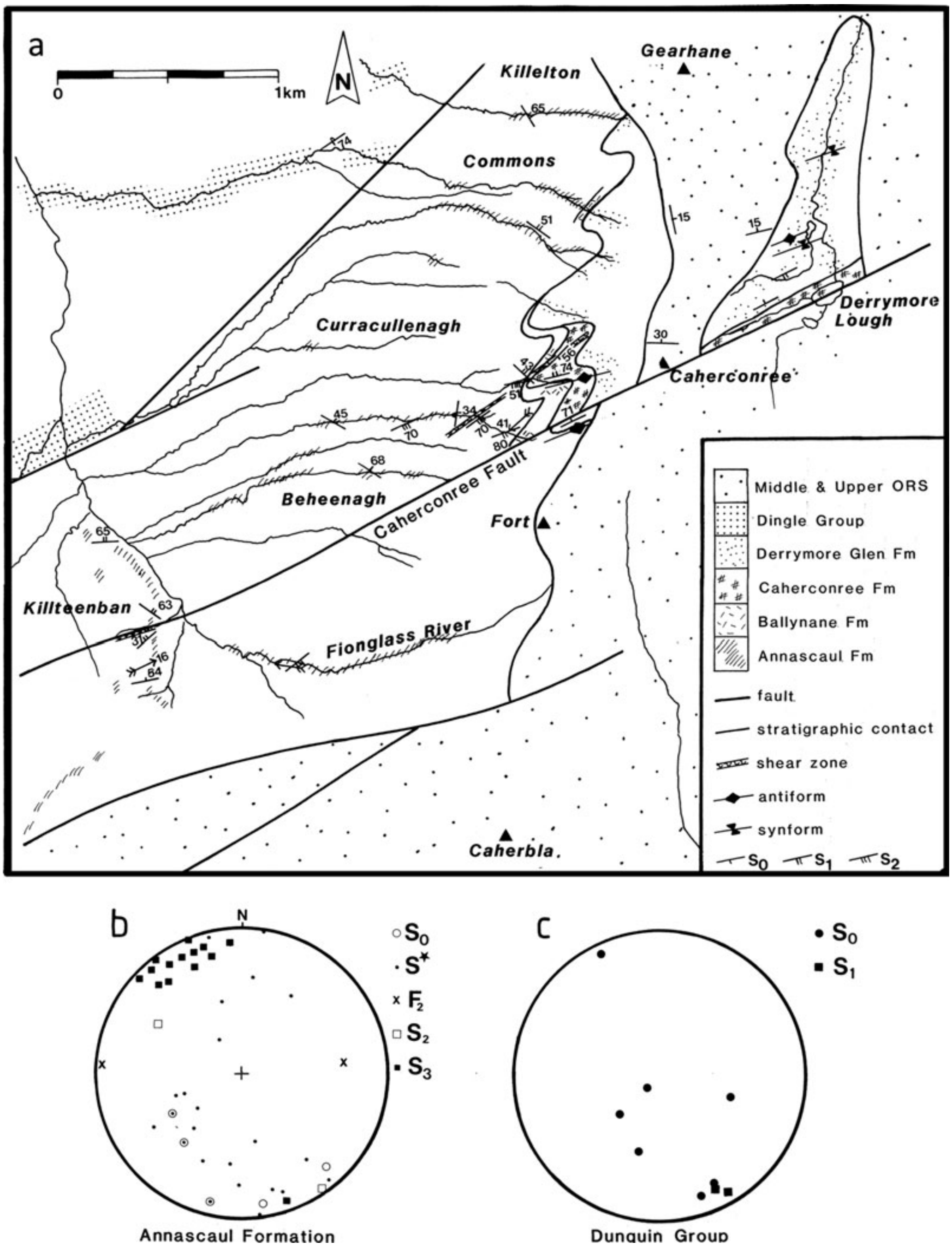

Figure 11. (a) Geological map of the Caherconree area. Lower-hemisphere equal-area projections depicting structural attributes of the (b) Annascaul Formation and (c) Dunquin Group.

rather than northeast or southwest as in Minard Bay. S_2 also has an atypical morphology and orientation. In some places S_2 appears as a conjugate crenulation cleavage and is overprinted by a second crenulation cleavage, S_3.

8. Structure of fault zones

The structure of faults provides enhanced evidence of the nature and timing of deformation. The effects of Acadian faulting are most prominent in the southeast of the Dingle Peninsula and die out to the northwest.

8.a. Minard Head Fault Zone

One of the best-exposed fault zones in the area occurs on the northern side of Minard Head as an outcrop strip *c.* 40 m wide (Fig. 12). This fault zone was referred to as the Caherconree Fault Zone by Todd (1989), following Horne (1970, 1974, 1977). Subsequent mapping has shown the Caherconree Fault at Caherconree is however a separate structure to that at Minard Head; the latter was therefore renamed the Minard Head Fault Zone (MHFZ). Rocks of the Annascaul and Bull's Head Formations are both deformed in the zone. To the southeast, on the southern side of Minard Head, the Inch Conglomerate Formation (Caherbla Group) is largely

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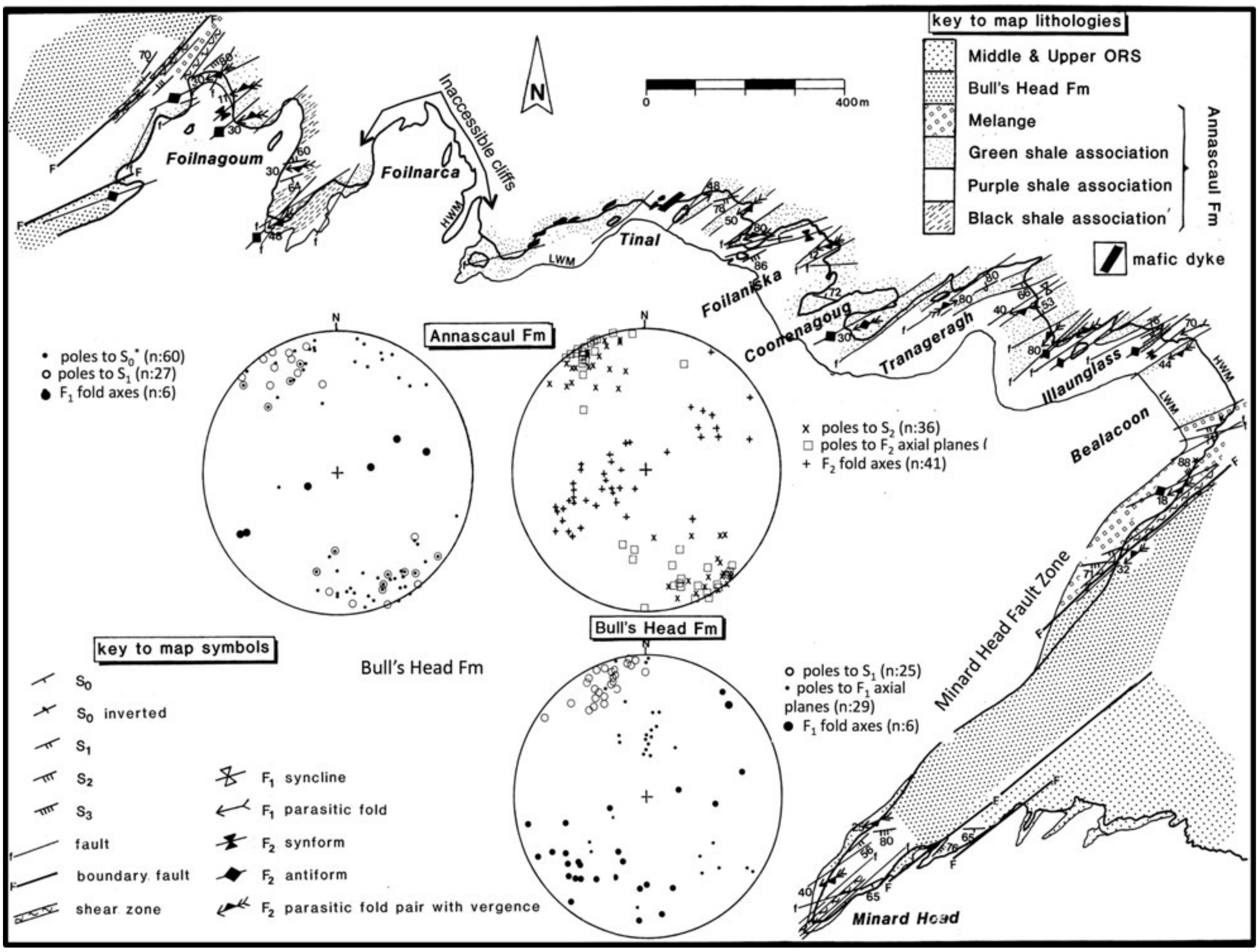

Figure 12. Geological map of Minard Bay, with insets of lower-hemisphere equal-area projections depicting orientations of the structures in the Annascaul Formation and Bull's Head Formation. Middle and Upper ORS refers to the Caherbla and Slieve Mish groups, respectively.

unaffected by the fault zone except for some simple normal faulting (Fig. 12).

Most discrete faults and fault-rock fabrics within the zone dip steeply SE. In the footwall, the Annascaul Formation is a mélange of black shale matrix and clasts of greywacke, shale, vein quartz, dolomite and brecciated lava.

Three zones with irregular and gradational boundaries can be recognized in the deformed footwall rocks of the Annascaul Formation (Fig. 13). In Zone 1, the main fault fabric is an intensification of the S_2 cleavage in the Annascaul Formation outside the fault zone. Within the mélange, S_2 crenulates S* which is present in both the matrix and clasts, particularly shale clasts; more competent blocks such as greywacke clasts are augened by S*. In some places, blocks containing S* are cut or augened by S_2, indicating some cataclasis and formation of clasts post-S* but pre-S_2.

The rocks are cut by numerous discrete faults, some of which are mineralized with quartz and sulphide minerals. In plan most steep faults trend anticlockwise of the S_2 cleavage, which is deflected into the faults; some faults lie clockwise of the cleavage (Fig. 13). The dominance of the anticlockwise relationship between cleavage and shear planes indicates a predominant component of sinistral shear in the zone. Although the majority of the individual strike-slip faults in the MHFZ are sinistral, some are dextral. Sinistral faults are aligned mainly about 060°, whereas dextral faults are aligned about 080° (Fig. 13; Todd, 1989). Similar preferred orientations are observed elsewhere in the Annascaul inlier, outside the MHFZ (Fig. 14; Todd, 1989).

Other mesoscopic indicators of shear sense in Zone 1 include porphyroclasts that range in diameter from a few millimetres to 10 m, but are generally less than 50 mm long. These clasts, and the clasts of sedimentary origin, are highly flattened parallel to subparallel to the S_2 fault fabric (Fig. 13). The porphyroclasts commonly have long axes that are oriented at a slight, oblique angle to the S_2 fault fabric (Fig. 13). Some of the augened clasts possess asymmetric pressure-shadow tails which in plan are mostly clockwise of the fabric, indicating back-rotation of the clast in sinistral transpressive shear (e.g. Lister & Williams, 1979; Lister & Snoke, 1984). In cross-sections perpendicular to the fault fabric, similar asymmetric tails indicate down-to-the-north shear. Clasts in the mélange are flattened and in agreement with strain indicators outside the fault zone, indicating an approximately NE–SW-oriented *Z* direction with *Y* horizontal and *X* vertical. From these indicators it is

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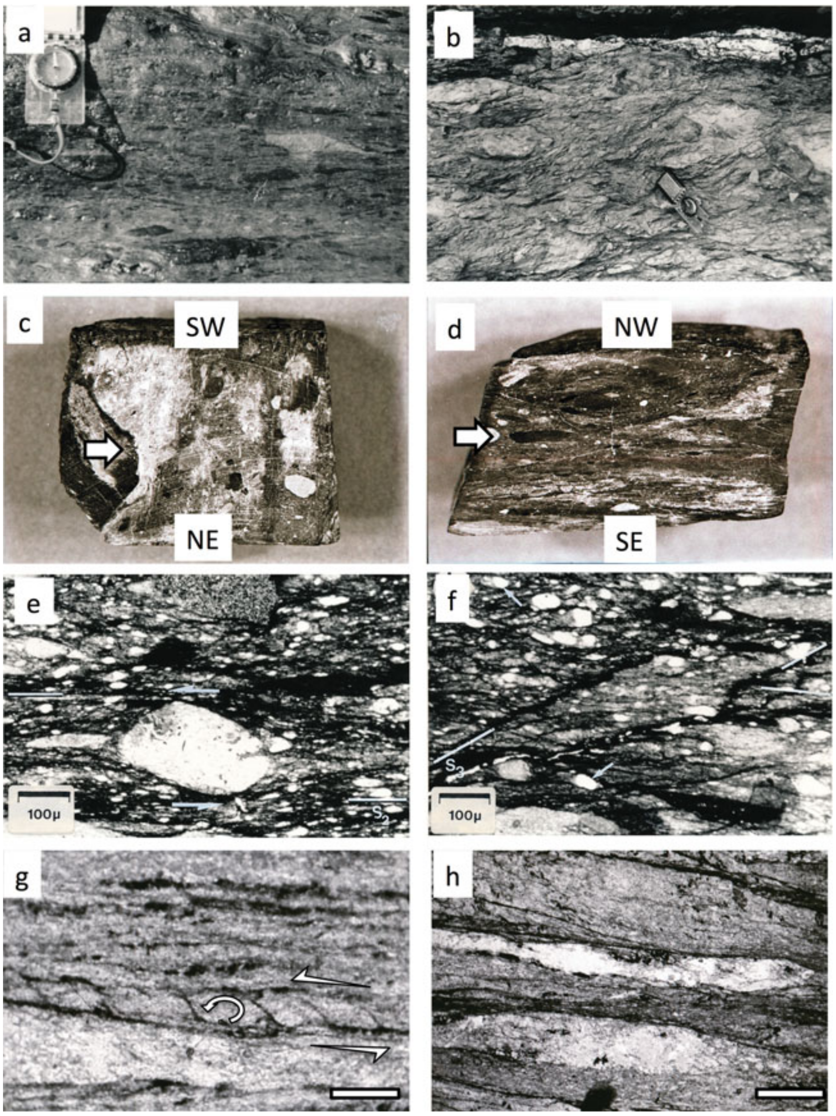

Figure 13. (Colour online) Fault rocks in the Minard Head Fault Zone and at Farranacarriga. (a) MHFZ Zone 1 rocks: plan view of mélange/cataclasite in Annascaul Formation with penetrative S_2 cleavage (the compass clinometer is 10 cm long). (b) MHFZ Zone 3 rocks: plan view of foliated cataclasite in Annascaul Formation protolith adjacent to the principal fault plane marked by quartz-fibre sheets at the top of the photograph (compass-clinometer is 18 cm long). (c) Slabbed, oriented specimen of Zone 1 fault rocks of the MHFZ cut parallel to the main fault fabric. Note the equidimensional shapes of the clasts. (d) Same sample as (c), but cut perpendicular to fault fabric. (e) Photomicrograph of the Zone 1 rocks in the MHFZ. The section is cut in the horizontal plane, perpendicular to the main fabric (S_2) picked out by haematite-rich cleavage seams. A large porphyroclast in the centre of the field of view displays asymmetric pressure-shadow tails of comminuted grains, the asymmetry indicative of sinistral shear as marked by the half arrows. (f) MHFZ Zone 1 cataclasite foliated by S_2 cut by a later S_3 cleavage. Both (e) and (f) photographs taken in plane-polarized light. (g) Plan view photomicrograph of the Farranacarriga Fault rocks developed in the Annascaul Formation showing boudins if phyllonite within the principal fault fabric are bounded by small shear planes. The anticlockwise rotation of the boudins indicates sinistral shear. (h) Plan view photomicrograph of the Farranacarriga Fault rocks developed in the Annascaul Formation showing a quartz ribbon mylonite in which the light areas are quartz with a little feldspar and the dark areas are graphitic phyllonite. In both (g) and (h), the photographs were taken in plane-polarized light and the scale bar is 0.1 mm long.

concluded that shortening across the zone was more important than simple shear (Sanderson & Marchini, 1984).

Field observations of Zone 1 fault rocks are supported by microscopic analysis. S_2 forms a pervasive crenulation cleavage that in Zone 1 achieves transposition of S^*. Some of the phyllitic clasts still contain S^* which has been typically rotated clockwise of the S_2 foliation, again indicative of sinistral shear (Fig. 13e, f). Other cataclasized arenite clasts are rounded to sub-angular and often contain tails of comminuted material; some of the tails are obliquely oriented to the fault fabric. Thin veins (*c.* 100 μm thick) of recrystallized mosaic quartz occur in some quartz and arenite clasts. These veins, plus the undulose extinction of quartz with the clasts and comminuted grains (cf. Watts & Williams, 1979; Anderson & Oliver, 1986) indicate limited quartz deformation followed by grain recovery by recrystallization (Sibson, 1977; Wise *et al.* 1984). Nevertheless, the dominance of brittle

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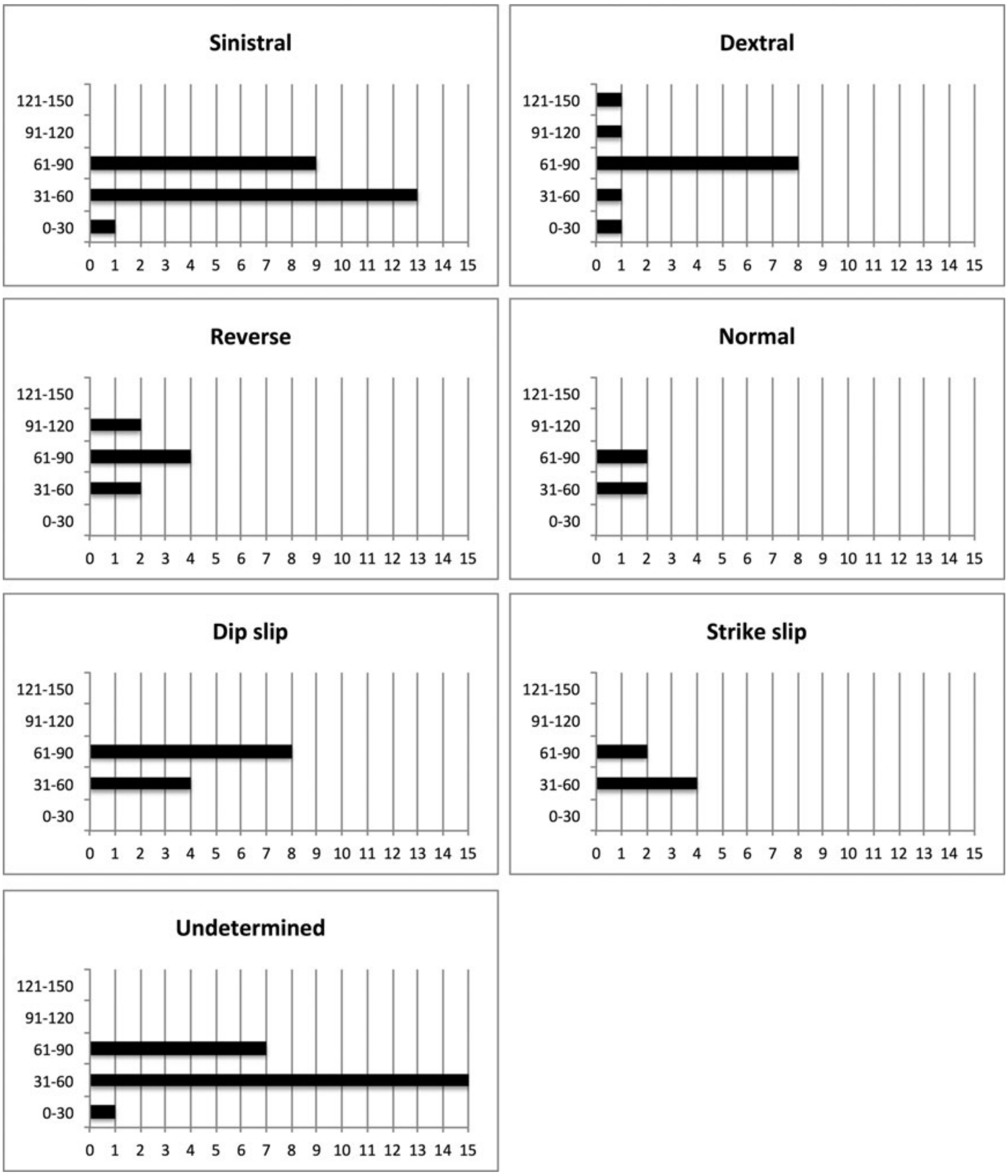

Figure 14. Histogram of the frequency of occurrence of mesoscopic fault strikes in the Annascaul inlier. Vertical axis in each histogram is the strike angle in degrees.

cataclastic processes over plastic deformation suggests that these rocks should be termed foliated cataclasites (Chester, Friedman & Logan, 1985).

In many places in Zone 1 of the MHFZ S_2 is cut by later cleavages. A gently dipping crenulation cleavage is apparently of minor significance. Another spaced pressure solution cleavage occurs more widely, and is correlated on the basis of geometry and form with S_3 observed in the Annascaul Formation outside the fault zone and with the main cleavage observed in the Bull's Head Formation in the hanging wall of the MHFZ (Figs 12, 13f).

Zone 2 fault rocks form elongate lenses a few metres wide. Although they are not necessarily more strained than Zone 1, Zone 2 fault rocks are distinctive in being finer grained and possessing abundant quartz veins. They are possibly the product of deformation of fine-grained black shales without abundant clasts. The quartz veins display a variety of geometries. Simple lenticular extension veins lie at a variety of angles to the S_2 cleavage. The veins may be oriented clockwise, suggesting dextral shear, or anticlockwise to the fabric, suggesting sinistral with the latter predominating. Some quartz veins are folded into tight to isoclinal folds with variable vergence. Others are pulled apart or boudined to form isolated vein quartz clasts. These different types of quartz vein may be the products of the same progressive simple transpressive shear.

Zone 3 fault rocks of the MHFZ are also derived from the Annascaul Formation and occupy a band 3–4 m

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wide adjacent to the principal fault plane (Fig. 13b). The yellow phyllitic rocks are intensely foliated and contain numerous shear planes. The foliation is domainal and anastomosing. Some resistant competent blocks of sandstones and conglomerate persist, but generally the grain size is small.

The principal fault plane of the MHFZ, separating Zone 3 rocks developed from the Annascaul Formation from Zone 4 rocks developed from the Bull's Head Formation (Dingle Group), is marked by a composite series of quartz-fibre sheets (Fig. 13b). These sheets cut all the previously described fault fabrics on the Annascaul Formation. The sheets generally dip SE and bear lineations indicative of oblique slip. The risers of the accretionary steps on the lineated quartz slickenfibre surfaces face downwards on the footwall indicating normal, down-to-the-south displacement combined with a subordinate component of dextral strike-slip. The quartz-fibre sheets are laterally extensive and largely undeformed, although in some places are cut and displaced by steep reverse faults that dip more steeply SE, are buckled into gentle inclined folds or are boudined.

Two zones may be defined in the Bull's Head Formation deformed within the MHFZ. Zone 4 is *c.* 1–2 m wide and lies parallel to the principal fault plane. It is characterized by an intensification of the brecciation, shearing and cleavage that occurs less commonly in the less-deformed rocks of Zone 5 to the east. Thinly bedded sandstones of the Bull's Head Formation, their usual purple hue altered to a bright green colour, have been sheared and boudined. The product is a cataclasite containing phacoids of resistant sandstone in a finely brecciated, foliated matrix.

A few metres southeast of the principal fault plane, cataclasis is reduced and Zone 5 is characterized by thinly bedded purple sandstones and siltstones of the Bull's Head Formation deformed by mesoscopic folds and faults (Fig. 12). Only limited cataclasis has occurred on some of the fault planes. The folds have either tight chevron-shaped profiles with angular hinges, or more open Class 1B profiles with rounded hinges. The folds have wavelengths of *c.* 5–15 m. They are variably oriented; tighter folds tend to be more steeply plunging and moderately to steeply inclined while open folds are generally more upright. The variability of the folds is taken as evidence of two phases of folding (Fig. 12; Todd, 1989).

A ubiquitous cleavage cross-cuts most of the inclined tight folds and is axial planar only to the open upright folds (Fig. 12) and is interpreted to be Variscan in age (see further discussion in Section 9.a). In one place on the north side of the Minard Head sandstones of the Bull's Head Formation contain two cleavages: (1) a closely spaced cleavage trending NE–SW and dipping moderately SE; and (2) a spaced cleavage cuts the NE–SW cleavage, trends ENE and dips steeply SSE. This is one of the few places where the Dingle Group contains more than one cleavage.

Numerous faults cut the folds. Some of them are accompanied by brecciation, but most are sharp, unmineralized planar and curvi-planar shear planes. A silt matrix supports angular arenite clasts up to 5 cm long in the breccia bands. A cleavage is commonly developed in the matrix of the breccias and, in some places, this fabric is deflected asymptotically into the walls of the fault-breccia band. In several places, particularly in the region of the southwest promontory of Minard Head, faults bound packages of little-deformed sandstone beds. Bedding within these packages or duplexes (Woodcock & Fischer, 1986) dips steeply and is often deflected asymptotically from the centre of the package towards the bounding fault (see fig. 10 of Todd, 1989). The faults trend approximately NE–SW and sinistral offset is most common where offset is demonstrable or where kinematic indicators such as deflected cleavage or bedding are present (Fig. 13). The sense of shear on the faults is concomitant with shearing of the short limbs of the folds, and the shearing and folding appears to have been contemporaneous. In some places fold hinges are completely cut out; abrupt changes in younging of the beds therefore occurs across the faults that bound these duplexes.

The east–west trend of the axial planes of the earlier folds in the Bull's Head Formation represents north–south shortening which is aligned anticlockwise of the NE-trending faults. This arrangement would be expected from the development of the folds and faults contemporaneously in a regime of sinistral transpression (Sanderson & Marchini, 1984; Todd, 1989).

8.b. Foilnagoum Fault Zone

This fault zone forms the northern boundary of the southwest end of the Annascaul inlier (Fig. 12). It is exposed in Foilnagoum in Minard Bay and inland in Glanminard (V 538 994). Faults and fault fabrics in the zone strike NE–SW and generally dip SE. In Foilnagoum, D_2 structures are prevalent in the Annascaul Formation and both mesoscopic and macroscopic fold closures are present. Further to the NW, in the cliffs above the cove where exposures are limited to the sides of a steep winding path, the rocks are intensely sheared. Sandstone beds have been disrupted into rhombs and lenses that float in a foliated argillite matrix. Widespread quartz and sulphide mineralization is present. Brecciation and cataclasis occurred prior to the superposition of S_2, which crenulates the locally preserved S^* cleavage.

The contact with the Bull's Head Formation is not exposed but sandstones of the younger formation close to the contact are discoloured to pale green. In places the sandstones are brecciated. Subordinate siltstones contain one discernible cleavage.

Inland from Foilnagoum, the contact between the Bull's Head and Annascaul formations trends across the stream valley of Glanminard. Here the contact itself is not exposed, but rocks of both the northern wall (Bull's Head Formation) and the southern wall (Annascaul Formation) are exposed close to the inferred trace

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of the contact. The sandstones of the Bull's Head Formation are brecciated and sheared so that arenite clasts up to a few centimetres in diameter are enveloped in a foliated matrix. In plan view, a well-developed foliation contains quartz and arenite porphyroclasts with asymmetric tails of comminuted grains indicative of sinistral shear. There is a second, spaced pressure solution cleavage which trends parallel to the early slaty fabric. This is considered similar to the area of the MHFZ, where the Bull's Head Formation contains two cleavages.

8.c. Coosatorrig

In the wave-cut platform of Coosatorrig within the Bull's Head inlier (Fig. 4; V 493 979), green and purple shales and thin greywacke sandstones of the Annascaul Formation are cut by a vertical NE-striking fault zone where D_2 structures include the progressive folding, flattening and shearing of more competent horizons such as silicified siltstones and quartz wacke. Arenites often form lenses or rhombs that lie within a phyllitic matrix in trains or isolation parallel or oblique to the main S_2 cleavage fabric. These boudins possess pressure shadows of comminuted grains that form asymmetric tails which trend clockwise to S_2. Vertical faults cutting the arenite bands have sinistral offsets, and trend anticlockwise of the cleavage. These features are consistent with sinistral transpression.

8.d. Farranacarriga

There are several exposures of Annascaul Formation and Dunquin groups in the Annascaul inlier that are affected by shear zones thought to generally be present throughout the inlier, as well as affecting the northwest and southeast boundaries described in Sections 8.a and 8.b. For example, on a road cut by the main Dingle to Annascaul road in the townland of Farranacarriga (Fig. 4; Q 578 020), there is an exposure of fault rocks formed by dark grey slates. The slates contain a very strong, steep, penetrative fabric that in places is banded and mylonitic. The fabric strikes about 080–100°. In addition to millimetre-thick banding, pinched quartz stringers up to 10 cm long lie parallel to the fabric. Larger quartz segregations, which form rhombs or lenses, are enveloped in the foliated phyllitic matrix. Some of these rhomboidal segregations possess tails that are oblique to the fabric back rotation in sinistral shear; others have tails indicative of dextral shear.

In thin-section the quartz stringers prove to be bands of mosaic quartz with sutured grain boundaries and undulose extinction (Fig. 13g, h). Groups of finer-grained recrystallized quartz grains occur in patches on these quartz ribbons. The quartz ribbons are enclosed in a finer-grained phyllonitic matrix of aligned illitic and graphitic micas. The strong mylonitic foliation and the evidence of crystal-plastic deformation indicate that these rocks should be termed quartz-ribbon mylonites.

A number of types of microscopic shear-sense indicators occur in these mylonites, including C-S fabrics, en échelon shear bands and asymmetric pressure-shadow tails on porphyroclasts. All the kinematic indicators show that strike-slip was pervasive, but both sinistral and dextral indicators are present. Fabric relationships indicate that dextral shear post-dated sinistral and, in this outcrop, dextral fabrics are more abundant than sinistral fabrics.

8.e. Ballynane

The southern boundary of the outlier of Ballynane Formation is faulted against the Annascaul Formation at Ballynane (Fig. 4; Q 624 026). The fault rocks, which are foliated cataclasites, occupy a zone up to 8 m wide. In the Annascaul Formation south of the fault zone, the green slates are affected by microfolds that deform S*. A crenulation cleavage, S_2, is axial planar to these folds. S_2 becomes more pronounced as the NE-aligned fault zone is approached. The cataclasites contains clasts of shale, siltstone and sandstone. The strong penetrative fabric, which is an intensification of S_2 in the Annascaul Formation, is composed of cleavage folia enhanced by opaque minerals. Porphyroclasts within the cataclasite often have pressure-shadow tails of comminuted grains that are asymmetric and indicate a sinistral shear sense. In vertical cross-section, the long axes of the porphyroclasts plunge north at an acute angle and anticlockwise to the fault-fabric. This relationship indicates down-to-the-north displacement in the zone. In plan view, the deflection of S_2 into closely spaced shear planes indicates sinistral shear within the zone, so in combination sinistral transpression is manifest.

The Ballynane Formation siltstones are deformed within the fault zone, although less intensely so than those of the Annascaul Formation. Only one cleavage is discernible in the Ballynane Formation, although it grades from being a closely spaced penetrative slaty fabric close to the fault zone to a spaced, planar pressure solution cleavage further to the north away from the fault zone.

8.f. Caheracruttia

A very small inlier of the Annascaul Formation crops out within the Inch Conglomerate Formation of the Caherbla Group and is exposed in the stream that flows through the townland of Caheracruttia (Fig. 4; Q 726 067). The inlier forms a strike-parallel strip enveloped by breccio-conglomerates of the Inch Conglomerate Formation. The Annascaul Formation is here represented by green, glossy phyllonites with a well-developed penetrative slaty cleavage. Locally, patches of cataclasites with clasts of phyllite are developed. There are abundant isolated quartz pods interspersed in a matrix of comminuted clasts of phyllite (Fig. 15a). In plan view the long axes of quartz clasts trend at an acute

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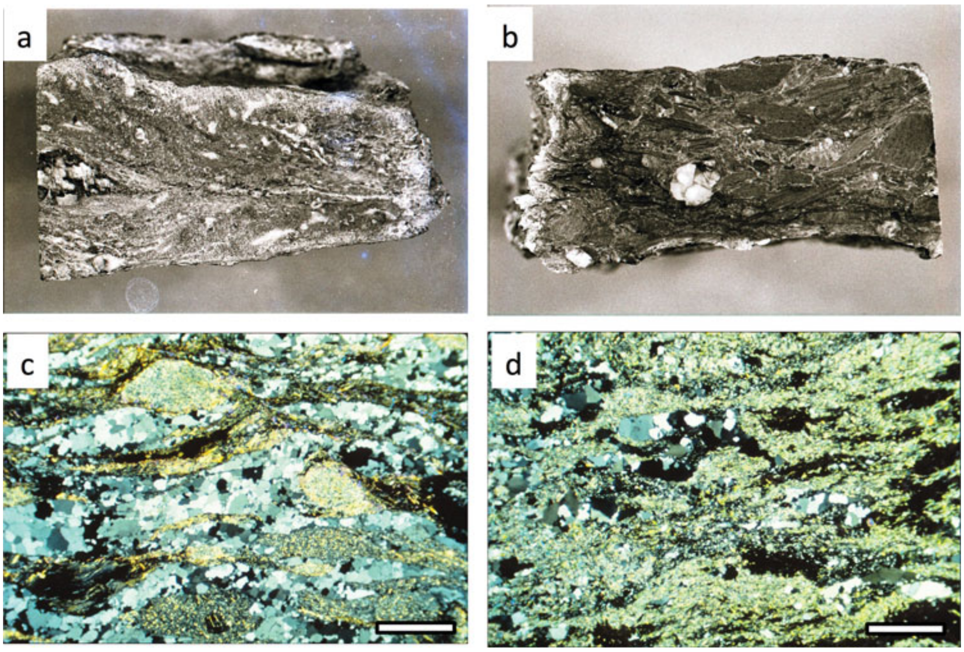

Figure 15. Fault rocks associated with the Inch Conglomerate Formation. (a) Slabbed specimen of cataclasite developed in the Annascaul Formation exposed in the stream bed at Caheracruttia. (b) Slabbed specimen of foliated breccio-conglomerate of the Inch Conglomerate Formation (lithosome 2 of Todd 2000, fig. 15) that directly overlies the cataclasites illustrated in (a). Photomicrographs, taken in cross-polarized light, of two clasts of mylonite from lithosome 1 (Todd, 2000) showing the spectrum of mylonitization of schist/gneiss protoliths including (c) quartz ribbon mylonite and (d) ultramylonite. The slabs in (a, b) are *c.* 15 cm long and the scale bars in (c, d) are 0.1 mm long.

angle, anticlockwise of the cleavage. The cleavage is cut and deflected by shear planes that also indicate sinistral shear. In vertical cross-section the shear fabrics indicate down-to-the-north displacement. In common with the vast majority of the fault rocks of the southeast Dingle Peninsula, these phyllonites developed in sinistral transpression.

A distance 10 m north of the exposure of the Annascaul Formation, blocks protrude from the banks of the Caheracruttia stream that are formed of foliated breccia of the Inch Conglomerate Formation. The breccia is composed of clasts of phyllite and quartz set in red sandy matrix and the foliation is interpreted to be sedimentary (Fig. 15b). The clasts are identical to the lithologies that occur in the cataclasites below. The breccia may be interpreted as a talus deposit, reworked directly from the underlying fault rocks (Capewell, 1951).

8.g. Clasts of basement fault rocks in the Inch Conglomerate Formation

Todd (2000) provided an account of the different metamorphic rocks that form clasts in the Inch Conglomerate Formation, which was deposited as a series of four or more alluvial fans shed off a basement high in the Dingle Bay region during ?Eifellian time after the Acadian deformation of the Dingle Group (Fig. 3). The clasts, which often range up to boulder grade, include gneiss, schist and mylonite. Within the Inch source appears to have been a metamorphic basement hinterland largely composed of medium-grade schists and gneisses, including banded granitoids traversed by extensive mylonitic shear zones (Todd, 2000). The mylonites include coarser-grained blastomylonites with banded segregations of quartz and phyllitic bands of sericite and chlorite (Fig. 15c, d). Finer-grained ultramylonite also occurs as clasts of pebble grade and are typified by a much more even-grained rock of quartz, sericite and chlorite. These clast lithologies provide a window into the uplifted basement core of the Dingle Bay Lineament, exposed as a 500 + m high mountain front above the basin floor during Caherbla Group times (Todd, Boyd & Dodd, 1988; Todd, 2000).

8.h. Dunquin Fault

This fault trends ENE–WSW and dips steeply SSE (Figs 8, 9). The stratigraphic separation of the fault indicates normal displacement in the west at Cooshaun (V 315 998) and reverse displacement in the east, for example at Lateevemore (Q 400 040). However, small-scale structures such pinnate joints and slickenslides indicate reverse motion on the fault. The contrast between the throw indicated by stratigraphic separation in the west and the reverse displacement indicated by small-scale structures suggests that the fault experienced several phases of movement. Moreover, the distinct change in the sedimentary facies of the Dingle Group across the Dunquin Fault indicates that it, or a predecessor, exerted a syn-depositional influence on the development of the Dingle Basin (Todd, Boyd & Dodd, 1988; Boyd & Sloan, 2000).

8.i. Ballyickeen Fault

As with the Dunquin Fault, stratigraphic variation across the NNE-trending Ballyickeen Fault indicates it was probably a growth fault during Dingle Group times (see Section 6). The Ballyickeen Fault crops out inland and is not exposed on the coast, but is mapped by stratigraphic separations and a prominent topographic break (Fig. 9). A number of parallel and probably synthetic faults with lesser displacement are exposed in the region of Clogher Head. One such fault is exposed on the coast south of Clogher Head (V 411 983) and has slickenfibre sheets with gentle plunging lineations and accretionary steps that together indicate a dextral strike-slip sense of motion. These slickenfibres are overprinted by the single, ubiquitous pressure solution cleavage of the area, suggesting a pre-Variscan origin. If this fault is indeed synthetic to the Ballyickeen Fault, the latter can be interpreted as a dextral strike-slip fault forming a lateral ramp between the early Acadian WNW-oriented folds of the Clogher Head region and the main Dingle Basin to the east.

8.j. Dún an Oír Fault

The ENE-trending Dún an Oír Fault (Horne, 1974) replaces the hinge of the Feohanagh Anticline in the area of the Smerwick promontory, in the northwest of the Dingle Peninsula (Figs 8, 9; Todd, Williams & Hancock, 1988). On the western Atlantic coast south of Coosglass, the Dún an Oír Fault is represented as a single, mineralized fault plane that juxtaposes the Coosglass Formation in the northern footwall with younger red beds and lavas of the Foilnamahagh Formation (Holland, 1969; Sloan & Williams, 1991; Higgs & Williams, 2011) in the southern hanging wall. This suggests a normal displacement to the fault, although large quartz veins in the hanging wall are pinnate to the fault and indicate a reverse displacement. On the east side of the Smerwick promontory, the Dún an Oír is represented by a zone of faults that juxtaposes slices of the Clogher Head, Mill Cove and Drom Point formations of the Dunquin Group (Fig. 9). Discrete fault planes observed south of Coosgorrib within the zone bear subhorizontal lineations indicating strike-slip.

As with the Dunquin and Ballyickeen faults, syn-depositional motion on the Dún an Oír Fault apparently induced severe variation in the thickness and facies of the Dingle Group. North of the Dún an Oír Fault, the Dingle Group is represented only by a petromict conglomerate up to 10 m thick. This suggests down-to-the-south displacement on the fault during Dingle Group times, synthetic to the North Kerry Lineament.

9. Nature and timing of deformation episodes

Table 2 summarizes the timing and effects of the deformation episodes that affected the Palaeozoic succession of the Dingle Peninsula. Four aspects of this deduced tectonic sequence merit additional discussion.

9.a. Age and geometry of the main cleavage in the Dingle Group

The single cleavage that is ubiquitously developed in the region has long been held as Variscan in age (Shackleton, 1940; Parkin, 1976; Todd, Williams & Hancock, 1988; Todd, 2000). Meere & Mulchrone (2006) used $R_f/Ø$ strain analysis of clasts in the Dingle and Slieve Mish groups to establish the difference between bulk strain above and below the Acadian unconformity, albeit comparing oligomictic conglomerates to petromict conglomerates and lithic wacke sandstones to quartz arenite sandstones. Having established this difference they argue that the pervasive cleavage in the Dingle Group, manifested by higher bulk strains in Dingle Group mudrocks (no strain indicators from Slieve Mish Group mudrocks were reported), is primarily Acadian in age. While the pre-Late Devonian rocks were undoubtedly deformed prior to the Variscan, the single cleavage discernible in the Dunquin and Dingle Groups is interpreted here to have developed during the Variscan orogeny, intensifying some existing strain (see above and Table 1). Indeed, when mudrocks above and below the unconformity are compared and the effects of volume loss in sandstones included, it appears that bulk strain difference below and above the Acadian unconformity is small (12 %; Table 1). The main difference in shortening appears to be through the folding/faulting, with the Dingle Group shortened by *c.* 40 % compared to *c.* 14 % above the Acadian unconformity.

Similar differences in shortening above and below the Acadian unconformity were recorded by S. D. King (unpub. thesis, 1989) and summarized by Graham (2001). Through bed length balancing and strain indicator assessment, King estimated that the Slieve Mish Group has been shortened by 17 % and by 36 % if volume loss is included (Graham, 2001). King also estimated 24 % pre-Variscan shortening of the Dingle Group (Graham, 2001). Similarly, the interpretation that the cleavage transection of Dingle Group folds is due to Variscan overprint (at an angle of 0–30° in plan view) of the pre-existing Acadian fold structures (Shackleton, 1940; S. D. King, unpub. thesis, 1989; Graham, 2001) is preferred to the interpretation that the cleavage and folds are co-genetic and hence the transection is a product of coeval dextral transpression (Meere & Mulchrone, 2006).

There are a few localities where two cleavages are observed in the Dingle Group, but these are restricted to fault zones and the lowermost Dingle Group where deformation is expected to be most intense. In these cases the later cleavage is Variscan, overprinting an Acadian shear zone fabric. Similarly, the S_2 cleavage in the Annascaul Formation that is easier to discern in the fault zones and in the hinges of F_2 folds outside the fault zones is interpreted to be Acadian in age, cut by a later S_3 fabric which has the characteristic geometry and morphology of the Variscan cleavage (Table 2). The more common occurrence of an Acadian cleavage in the Annascaul Formation

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Table 2. Structural history of the Palaeozoic rocks of the Dingle Peninsula.

Tectonic phase	Broad structural effects	Fault zone effects	Stratigraphic effects	Age
Variscan inversion and shortening of the Munster Basin	Folding of all Palaeozoic rocks and development of pervasive pressure solution cleavage and flattening bulk strain. F_1/S_1 in Slieve Mish and other post-Acadian ORS groups. Main cleavage in Dingle and Dunquin Groups, but locally F_2 and S_2 structures in fault zones. F_3 folds and S_3 cleavage in Annascal Formation.	Reverse reactivation of main lineaments, Dingle Bay and North Kerry, mimicked in outcrop by reverse displacement on MHFZ, Dunquin Fault, Dún an Oír Fault. S_3 overprints S_2 in Annascaul Formation fault rocks.	No post-Westphalian is preserved in SW Ireland except for a minor Cretaceous outlier.	Late Carboniferous
Extension of Munster Basin; Dingle was basin-margin footwall block	In south tilting of Caherbla Group and older rocks; in north progressive development of Feohanagh Anticline as rollover structure.	Normal faulting on e.g. Caherconree Fault, MHFZ mimics major displacements on Dingle Bay and North Kerry Lineaments.	Unconformities between progressively younger ORS groups in the Dingle area and recycling into the Munster Basin.	Middle–Late Devonian
Acadian inversion and sinistral transpression	Folding and faulting of Dingle Group and older rocks. F_1 folds in Dingle-Dunquin groups; local S_1 cleavage in fault zones in Bull's Head Formation. F_2 folds and local S_2 cleavage in Annascaul Formation, particularly in fault zones.	Sinistral transpressive shear in MHFZ and other fault rocks closer to the Dingle Bay Lineament. S_2 in Annascaul Formation intensifies into fault zones. Reverse faulting accompanying folding further to the north.	Acadian unconformity; shedding of Inch Conglomerates from DBL basement ridge.	Late Emsian
Acadian sinistral transtension	Tilting/folding and faulting of Dunquin and early Dingle groups	Subsidence of Dingle Basin adjacent to strike-slip displacement of Dingle Bay and North Kerry lineaments. Syn-depositional displacement on Dunquin, Dún an Oír and Ballyickeen faults.	Variation of Dunquin to Dingle Group transition; shedding of fan aprons from the north and south.	Latest Silurian–Emsian
Caledonian arc volcanism, possibly after cessation of subduction	Tilting	None observed	Basaltic to rhyolitic lavas and pyroclastic rocks interbedded with marine to non-marine sediments	Late Silurian
Caledonian convergence and subduction	Folding (F_1) and cleavage (S_1) of Annascaul Formation	None observed	Unconformity between Silurian and Annascaul Formation	?Llanvirn – ?Early Silurian
Caledonian extension in fore- or back-arc setting	Bedding-parallel cleavage (S*), slumps.	None observed	Deep-water slope deposits including mélange with volcanic debris	Early Ordovician

compared to the Dingle Group is believed to be a result of either: lithological contrasts; different crustal level during Acadian deformation; proximity to the Dingle Bay Lineament; duration of Acadian deformation affecting the older rocks; or some combination of these factors.

9.b. Sinistral and dextral shear patterns in the fault zones

The second point of discussion is the occurrence of both dominant sinistral and less-abundant dextral shear indicators in the Acadian fault zones. Sinistral faults and shears are dominant and are typically oriented from 000 to 090°, with a mode about 060° (Fig. 14). Dextral faults and shears are subordinate, ranging from 000 to 150° with a mode about 080°. Where discernible, the two fabrics may be coeval or the dextral fabric cuts and post-dates the sinistral fabric. One interpretation is that the fabrics are the result of general NNW convergence, with basement structure 'forcing' the structural style and sense of displacement in the transpression zone (cf. Soper, Webb & Woodcock, 1987). In this case, sinistral transpression may not result from through-going strike-slip, but from reactivation of 060°

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structures during NNW–SSE convergence. An alternative model, also supported by field observations, is that the dextral movement on the modal 080° trend was later than the sinistral movement. In this case, sinistral transpression is more likely to involve a through-going, major strike-slip displacement as well as the NNW–SSE convergence. Dextral movement on faults oriented at an anticlockwise angle to the sinistral faults (i.e. 000–040°) could still be accommodated as secondary, coeval Riedel-shear-like faults or lateral ramps such as the Ballyickeen Fault. A third alternative explanation for the observed dextral displacements is that at least some of them are Variscan in age, in agreement with models for the Irish Variscides that infer some dextral shear complementing shortening (e.g. Sanderson, 1984; Price & Todd, 1988; Graham, 2001).

9.c. Partitioned Acadian transpression/transtension

The broad zonation of the Acadian structure of the Dingle Peninsula invites comparison with that of a flower structure (Harding, 1985) or a strain-partitioned transpression zone (Fig. 16; Dewey, Holdsworth & Strachan, 1998; Dewey & Strachan, 2003). During Dingle Basin subsidence structural and stratigraphic evidence indicates coeval strike-slip along the major bounding lineaments and intra-basinal effects such as localized folding, tilting and faulting of the Dunquin Group and older rocks. Similarly, the sinistral transpressive inversion of the Dingle Basin appears to have caused folding and faulting of the basin fill with strain apparently intensifying closer to the Dingle Bay Lineament, manifest by fault rocks in the Annascaul inlier. Although not currently exposed, the Dingle Bay Lineament is considered to represent the main fault segment of the zone. The model predicts simple shear strain increasing towards, or distinctly partitioned in, the main fault segment (Dewey, Holdsworth & Strachan, 1998). Dingle Group and older rocks southeast of the southern limb of the Fahan Syncline are therefore typically subvertical, often overturned and involved in sinistral transpressive fault zones such as the MHFZ. North of the Fahan Syncline the fold belt which shallows northwards, affecting both Dingle and Dunquin groups, is interpreted as the ‘petals’ of the flower; it is predicted to accommodate more compression and reverse slip across thrusts, perhaps with a subordinate amount of strike-slip.

9.d. Late Devonian syn-rift

The observed pattern of Late Devonian normal faulting affecting Caherbla Group and older rocks, then reversed and overprinted by Variscan shortening through folding, faulting and bulk strain, is consistent with the models of Price & Todd (1988) and Todd (1989) invoking the reuse of Caledonian (and Acadian) structures in the development of the Munster Basin framework (Fig. 17). Pre-existing faults such as the Dingle Bay

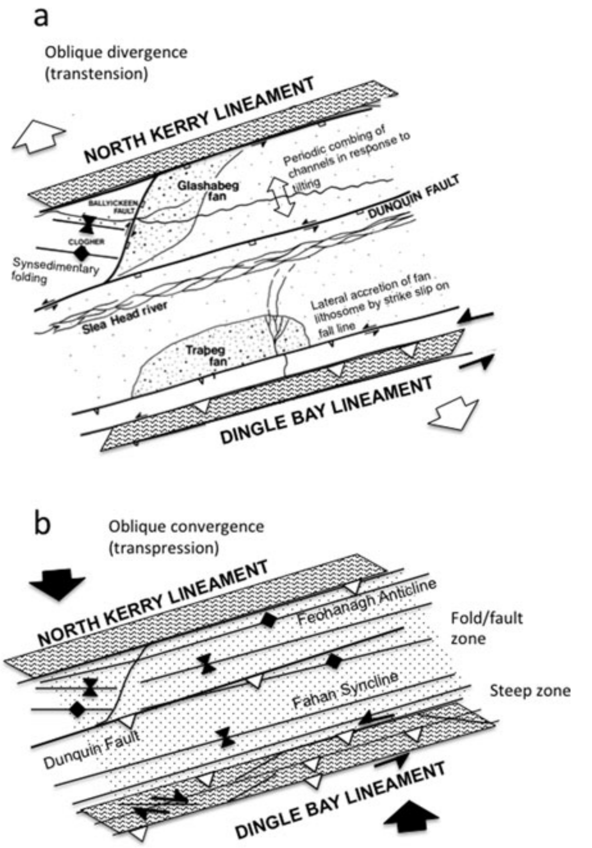

Figure 16. Early Devonian transtensional then transpressive development of the Acadian structure of the Dingle Pensinsula. (a) Schematic plan showing structural and sedimentation patterns in the Dingle Basin during early Emsian time. While the patterns observed could be due to very oblique sinistral transpression, a slightly oblique divergence in sinistral transtension is favoured. (b) Schematic plan showing structural pattern of the Dingle Basin in late Emsian time after inversion by sinistral transpression partitioned across the basin with folding and faulting in the prior basin area and sinistral shear and flattening in the steep zone towards the Dingle Bay Lineament.

Lineament and synthetic faults such as the MHFZ were reactivated as largely normal faults to form the Late Devonian Munster Basin, perhaps with a small amount of dextral strike-slip as is observed on the MHFZ (Todd, 1989).

Williams (2000) presented a detailed modelling analysis of the subsidence history of the Munster Basin. He concluded that extension was accommodated through displacement on steep crustal faults, such as the Dingle Bay Lineament. Deformation of the regional footwall of the Dingle area, north of the Dingle Bay Lineament, was modelled to include the formation of a ‘rim-type’ basin in which the Late Devonian Carrigduff Group was preserved. Williams (2000) only included the Carrigduff Group, preferring the interpretation of Richmond & Williams (2000) that the Smerwick and Pointagare Groups are Early–Middle Devonian in age and ‘pre-rift’ to the Munster Basin. The alternative interpretation that all the post-Dingle Group succession of the Feohanagh Anticline is age-equivalent to the Munster Basin is upheld here as at least equally permissible,

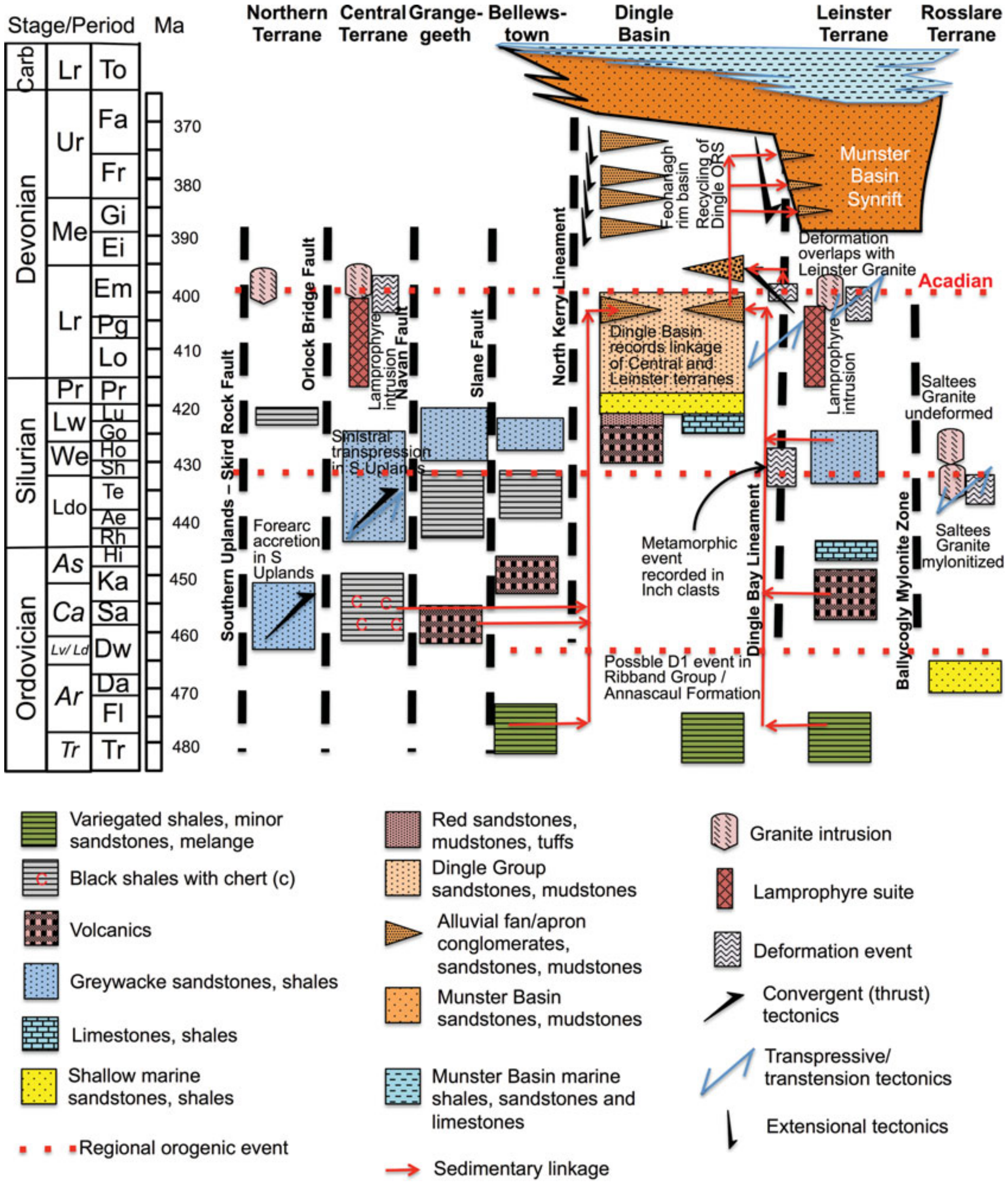

Figure 17. Regional chronostratigraphy of the Iapetus suture zone in Ireland showing the Caledonian and Acadian deformation history and linkages between the main terranes or blocks. Updated from Todd (2000) with correlation with the timing data compiled by Soper & Woodcock (2003) and modification of the time scale to that of Gradstein, Ogg & Schmitz (2012).

if not more consistent, with the available evidence (Price & Todd, 1988; Todd, Boyd & Dodd, 1988; Todd, Williams & Hancock, 1988; Todd, 1989). This model sees the deposition and subsequent tilting of successive unconformity-bounded syn-rift groups in a punctuated fashion through progressive rotations of a half-graben perched on the regional footwall of the Munster Basin (Fig. 17).

10. Tectonic evolution of Dingle Peninsula in a regional context

The oldest rocks exposed in Dingle, the Early Ordovician Annascaul Formation, have been correlated with the Ribband Group in SE Ireland and the Skiddaw Group in the Lake District (Figs 2, 17; Todd *et al.* 2000). As such, a palaeogeographic position on the southern margin of the Iapetus Ocean seems very likely as part of Eastern Avalonia (see Prigmore, Butler & Woodcock, 1997). It is generally agreed that the Iapetus itself was in existence during Cambrian time, although there is some debate on when Avalonia separated from Gondwana. Some favour a late Cambrian or Tremadocian separation, influenced by the Tremadocian onset of arc volcanism. Others, influenced more by palaeomagnetic and faunal data, infer a Llanvirn or early Caradocian separation. In either case, the deep-water facies of the Annascaul Formation, including evidence

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of slope instability to produce slumps and mélange deposits with contemporaneous volcanism, is consistent with both a fore-arc and back-arc ocean margin.

D_1 in the Annascaul Formation, deforming the bedding parallel S* cleavage believed to be associated with deep burial metamorphism, is difficult to date. Correlation with similar early structures in the Ribband Group of east and southeast Ireland (Shannon, 1979; Murphy, 1987) suggests a pre-Caradocian age. If this is correct, then perhaps the stress associated with D_1 is related to subduction either in a fore-arc setting (accretionary processes?) or through inversion of a prior back-arc as subduction trajectories changed. Late Ordovician rocks have not been preserved in Dingle, but in SE Ireland the period is dominated by the Duncannon Group volcanic rocks representing arc volcanism through continued southwards subduction of Iapetus under Avalonia.

It is now generally accepted that Avalonia had accreted to Laurentia by middle Silurian time and that the Silurian succession in the Lake District and central Ireland represents infilling of the vestigial seaway (e.g. Woodcock, Soper & Strachan, 2007). The development of Wenlock subduction-related volcanic rocks (Dunquin Group) in western Dingle has been related to a final burst of Iapetus-related volcanism with a vestigial re-entrant of Iapetus oceanic crust being the preferred explanation for these arc-signature volcanic rocks (Sloan & Bennett, 1990; Boyd & Sloan, 2000). The collision is generally considered to be ‘soft’ within the Iapetus suture zone of Britain and Ireland, with continuous late Silurian subsidence recorded in many places (Soper & Woodcock, 2003). There is evidence of Silurian deformation in South Mayo (Dewey & Strachan, 2003), in Anglesey (Treagus, Treagus & Woodcock, 2011) and in Rosslare (Murphy, 1990). Such an episode might also be recorded in the main metamorphism of the schists and gneisses preserved as clasts in the Inch Conglomerate (Fig. 17; Todd, 2000).

The end of volcanicity in Dingle was followed a transition from marine to non-marine sedimentation and increasingly apparent evidence of syn-depositional tectonics attributed a strike-slip regime (Todd, Boyd & Dodd, 1988; Boyd & Sloan, 2000). The patterns of conformable-unconformable boundary variation, stratigraphic thickness change, sedimentary facies and dispersal patterns of the Dingle Group are strongly evidential of a sinistral transtensional/transpressional regime (Fig. 16a; Todd, 1989). The Dingle Basin was filled by lakes, alluvial fans and rivers sourced from rising hinterlands to the north and south with geological signatures similar to the terranes now north and south of the Iapetus Suture (Fig. 17; Todd, Boyd & Dodd, 1988; Todd, 2000). The Dingle Group was then itself deformed through folds, faults and some penetrative strain. This period of inversion correlates very well with the late Emsian cleavage-forming event in the Acadian slate belts of SE Ireland and Lake District, which was synchronous with granites dated at *c.* 400 Ma (Fig. 17; Soper & Woodcock, 2003). A single Rb–Sr mica-whole-rock age of 398 ± 5 Ma from a clast from the Inch Conglomerate Formation might similarly reflect this event in the basement of the Dingle Bay Lineament (Fig. 17; Todd, 2000).

It is therefore possible to envisage ongoing transtensional/transpressive strain in the region of the Dingle Bay Lineament throughout Dingle Group times (Todd, 1989). The fault zones described above are a window into that major fault zone and display a variety of shortening structures (folds and cleavage), forming together but at different angles with sinistral and dextral strike-slip faults and both normal and reverse faults. It is possible that basin subsidence through pull-apart and/or flexural loading continued synchronously with shortening and uplift of neighbouring blocks, partitioned into or bounded by zones of pre-existing weakness such as the Dingle Bay Lineament (Fig. 16a). Given the local evidence, there is little need to invoke a significant change in plate convergence to accomplish the observed tectonic sequence. For example, the change from a regime of differential subsidence and uplift described for Dingle Group times (Fig. 16a) to general shortening and uplift (Fig. 16b) could be accomplished by a modest rotation of convergence across the 060°-oriented zone from very oblique to less oblique convergence, or by the strike movement of the Dingle block along the Dingle Bay Lineament from releasing to restraining bends (see also Dewey, 2002).

However, in addition to the Acadian orogenic episode being well established as a late Emsian event involving transpression across the slate belts south of the Iapetus Suture, there is now growing evidence of a preceding transtensional period in the latest Silurian – Early Devonian (Dewey & Strachan, 2003; Soper & Woodcock, 2003). This regional transtension reflects large strike-slip displacements inferred for many of the major faults (e.g. Great Glen Fault; Dewey & Strachan, 2003), provides a driver for subsidence and accommodation space for ORS sedimentation (Dewey & Strachan, 2003; Soper & Woodcock, 2003) and enables the intrusion of the lamprophyre suites and ultimately the newer granite suite across the Iapetus suture zone (Brown *et al.* 2008). It is therefore simplest to believe that the Dingle Basin change from differential subsidence/uplift during late Silurian – early Emsian time to general shortening in late Emsian time also reflects the same regional tectonic climate change from largely transtensional to largely transpressional. Woodcock, Soper & Strachan (2007) suggest either a change of subduction trajectory or that the collision of a continental fragment in the Rheic subduction zone induced the stress to generate the slate belts in the fossil Iapetus suture zone (Fig. 17).

A similarly profound change in tectonic style must have occurred to cause the onset of Munster Basin extension that probably began during Givetian time or perhaps earlier during Eifelian time. Leeder (1982) suggested that slab-pull from the Rheic subduction zone caused extension in many upper Palaeozoic basins in Britain. A geotectonic picture therefore emerges for the fossil Iapetus suture zone, in which changes from to transpression to extension are generated by changes in Rheic subduction styles to the south, culminating in the closure of the Rheic with collision of Gondwana and the development of the Irish Variscides.

Acknowledgements. Some of the fieldwork on the Annascaul Formation described here benefited from the involvement of Barry Murphy. Ken Higgs, Pat Meere, Rod Sloan and Nigel Woodcock provided very useful input during discussion of some of the key aspects of the paper prior to submission; two anonymous referees improved the article further with thorough and helpful reviews. The article is dedicated to the memory of Peter Callory and Dr Finbarr O'Shea, two of the many Dingle residents who welcomed and supported the author during his fieldwork in the area, including a memorable day viewing and photographing the dramatic cliffs of the southern coast of the peninsula from the sea.

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