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Structural and stratigraphic relationships across the continuation of the Highland Boundary Fault in western Ireland

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Abstract – The relationship between the Dalradian Supergroup and the Highland Border Complex in Scotland has remained contentious for over a century. In western Ireland, the contact between the Dalradian Supergroup and the Clew Bay Complex (a correlative of the Highland Border Complex) is superbly exposed on the island of Achill Beg on the North Mayo coast. The unfossiliferous South Achill Beg succession has been traditionally assigned to the Clew Bay Complex, and this interpretation is supported by a combination of Sm–Nd model age data, heavy mineral analysis and lithostratigraphic correlation. T_{DM} ages range from 1.99–2.66 Ga (mean = 2.28 Ga, $n = 6$). Detailed structural mapping shows that both the Dalradian and the Clew Bay Complex share the same structural history. A D1 high strain event is common to both units, and is associated with the development of tectonic slides. The D2 event is responsible for the formation of crustal-scale nappes. In both units, beds are consistently downward facing on the S2 foliation. Later dextral shearing (D3) resulted in the tilting of the originally recumbent, S-facing D2 nappes into this downward-facing orientation. Rb–Sr and ^{40}Ar–^{39}Ar radiometric dating of muscovite confirms that both units were deformed contemporaneously as the S2 nappe fabric in each is dated at *c*. 460 Ma. This Middle Ordovician age for deformation of the Clew Bay Complex is highly significant, not least because published microfossil data suggest a Silurian age.

Keywords: Dalradian Supergroup, Highland Boundary Fault, Ireland, structural geology, geochronology, Sm/Nd.

1. Introduction

The Highland Border Complex (Curry *et al*. 1984), and its presumed correlative in Ireland, the Clew Bay Complex (Williams *et al*. 1994), crop out as a series of isolated and lithologically varied exposures along a major NE–SW-trending fault zone. This fault zone, referred to as the Highland Boundary Fault in Scotland, continues into Ireland where it is broadly coincident with a conspicuous magnetic lineation, the Fair Head–Clew Bay line (Fig. 1a; Max & Riddihough, 1975). This major fault zone generally separates the Highland Border Complex and Clew Bay Complex from the Dalradian Supergroup to the northwest. A major phase of sinistral strike-slip movement along the Highland Boundary Fault is believed to be Silurian in age, whereas it is thought that there has been no significant movement along the Fair Head–Clew Bay line since Late Ordovician times (Ryan *et al*. 1995).

However, despite over a century of research in Scotland, the relationship between the Highland Border Complex and the Dalradian is still contentious. It remains uncertain as to whether they share a similar structural history and hence were deformed contemporaneously in the same orogenic event (e.g. Johnson & Harris, 1967; Tanner, 1995) or whether they were originally exotic to each other and juxtaposed at a later stage (e.g. Henderson & Robertson, 1982; Harte *et al*. 1984; Bluck & Ingham, 1997; Dempster *et al*. 2002). In Scotland, such uncertainties are at least partly due to a combination of poor inland exposure and stratigraphic excision along the Highland Boundary Fault. In contrast, the contact between the Dalradian and the Clew Bay Complex is superbly exposed on the western coast of Co. Mayo in Ireland (Fig. 1b), and therefore the structural relationships between the two units can be examined in detail. Consequently, the hypothesis of contemporaneous deformation of the Dalradian and the Clew Bay Complex can be evaluated.

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2. Significance of Dalradian–Highland Border Complex/Clew Bay Complex relationships

The relationship between the Dalradian and the Highland Border Complex/Clew Bay Complex is very important to our understanding of this segment of the Caledonian/Appalachian orogen. The timing of orogenesis within the lower half of the Dalradian Supergroup has proved contentious in recent years (e.g. Tanner & Bluck, 1999), but a detailed discussion on the evidence for Neoproterozoic compressional tectonism is beyond the scope of this paper.

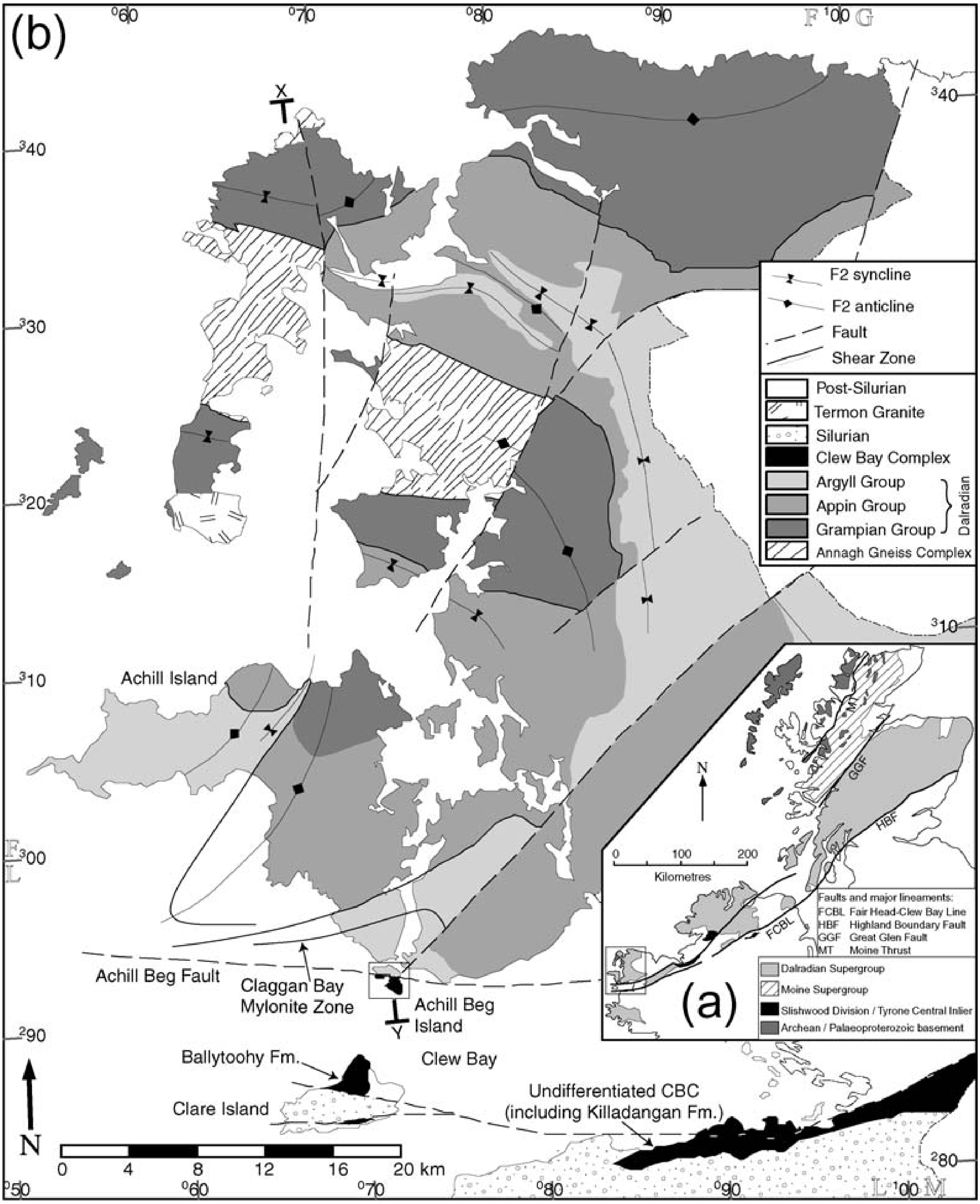

Figure 1. (a) The orthotectonic Caledonides of northwestern Scotland and Ireland with major faults illustrated. Box indicates area shown in (b). (b) Regional geology of the Northwest Mayo inlier. Box indicates area shown in Figure 2.

However, the timing of the orogenic episode which affects the upper half of the Dalradian Supergroup (the Grampian orogeny) has also been contested in the recent past. The Highland Border Complex has yielded a reliable Middle Arenig trilobite and articulate brachiopod fauna (Ingham, Curry & Williams, 1985), and hence contemporaneous deformation of the Dalradian Supergroup and the Highland Border Complex constrains the onset of Grampian deformation to be later than Middle Arenig (*c.* 475 Ma). This Early

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Ordovician minimum age for Grampian deformation contradicts geochronological studies which imply a Middle to Late Cambrian (Dempster, 1985; Dempster, Hudson & Rogers, 1995) minimum age for the Grampian orogeny. Tanner & Pringle (1999) have attributed some of the older (484–520 Ma) Rb–Sr and K–Ar cooling ages of Dempster (1985) to the presence of detrital mica in low-grade schists adjacent to the Highland Boundary Fault, but acknowledge that such a hypothesis cannot account for some of the older (481–514 Ma) Rb–Sr muscovite ages obtained from high-grade schists to the north. An Early Ordovician minimum age for Grampian deformation is compatible with models which attribute Dalradian deformation to a Middle Ordovician arc-collision at the Laurentian margin (Dewey & Shackleton, 1984; Dewey & Ryan, 1990; Van Staal *et al*. 1998; Dewey & Mange, 1999).

In Ireland, the relationship between the Clew Bay Complex and the Dalradian is also contentious, primarily due to the absence of a diagnostic macrofauna. A Middle Cambrian–Tremadoc? sponge (*Protospongia hicksi*, Rushton & Phillips, 1973) and Middle Ordovician coniform euconodonts (Harper, Williams & Armstrong, 1989) have been obtained from the Clew Bay Complex, which is compatible with the involvement of the Clew Bay Complex in a Middle Ordovician Grampian orogeny. However, Williams *et al*. (1994) and Williams, Harkin & Higgs (1996) contend that the rocks containing the faunal elements described above are in fact mega-clasts in a mélange, the matrix of which has yielded trilete miospores interpreted to be of Silurian (Wenlock) age (Williams *et al*. 1994; Williams, Harkin & Higgs, 1996). These microfossil discoveries have proved highly contentious (Graham, 2001) and have been strongly rejected by Johnston & Phillips (1995) and Dewey & Mange (1999).

3. Regional geology of the Northwest Mayo inlier

The Northwest Mayo inlier on the western extremity of the Irish Caledonides (Fig. 1b) preserves an excellently exposed transect from Laurentian basement, the Annagh Gneiss Complex (Daly, 1996), through presumed parautochthonous Neoproterozoic–Cambrian? Laurentian cover (Winchester, 1992; Fitzgerald *et al*. 1994), that is, the Dalradian Supergroup, to outboard oceanic elements, such as the Clew Bay Complex. The majority of the Clew Bay Complex outcrop (as defined by Williams *et al*. 1994) consists of low-grade, deep marine metasediments, and is exposed on the south shore of Clew Bay (Fig. 1b), on the northwest coast of Clare Island (Fig. 1b) and on the island of Achill Beg immediately to the south of Achill Island (Figs 1b, 2, 4b).

The contact between the Dalradian and the Clew Bay Complex is exposed on Achill Beg, where it is defined by the Achill Beg Fault (Figs 1b, 2, 4b). This fault, believed to be equivalent to the Highland Boundary Fault, separates the Dalradian (South Carrowgarve Formation) in the northern part of Achill Beg from the Clew Bay Complex (South Achill Beg Formation) to the south (Figs 2, 4b).

4. Stratigraphic relationships across the Achill Beg Fault

To date, the only detailed study of the geology of Achill Beg is that of C. T. Morley (unpub. Ph.D. thesis, Univ. Dublin, 1966), although a brief account of the geology and lithological descriptions are presented by Max (1989). The stratigraphic nomenclature used closely follows that of C. T. Morley (unpub. Ph.D. thesis, Univ. Dublin, 1966) as it is felt that the boundaries to the three members which comprised the South Achill Beg Formation of Max (1989) need to be revised to take into account the effects of F2 folding (Fig. 2). All quoted stratigraphic thicknesses are estimates, taking into account the effects of tectonism.

4.a. Dalradian Supergroup

4.a.1. South Carrowgarve Formation (2000 m)

The Dalradian succession of the southernmost portion of Achill island (South Achill) and North Achill Beg (Fig. 4b) consists of a sequence of massive psammitic wackes and subsidiary graphitic pelites. The quoted thickness of the formation and the lithological description detailed below include the outcrop of the formation on both South Achill and North Achill Beg. On South Achill, a thin basic metavolcanic horizon is exposed, while serpentinite olistoliths (Fig. 3a; Chew, 2001) and fuchsite (chromian muscovite) are relatively abundant within the graphitic pelite beds. Sparse way-up evidence (occasional grading within pebbly wackes) suggests that the formation youngs southwards towards the Achill Beg Fault. A coarse pebbly wacke horizon with a maximum observed thickness of approximately 100 m occurs on North Achill Beg (Fig. 2) although individual beds are rarely over a metre thick. The dominant clast types are quartz (up to 5 cm long) and altered microcline feldspar, set within a fine-grained psammitic matrix.

The Dalradian affinity of the South Carrowgarve Formation and the immediately underlying formations on South Achill (the Ooghnadarve, North Carrowgarve and Claggan Volcanic formations; Fig. 4a,b) has been questioned (e.g. Harper, Williams & Armstrong, 1989; Dewey & Ryan, 1990; Harris, 1993, 1995). It was felt that the oceanic aspect of this succession as indicated by the occurrence of blueschist-facies metamorphism (Gray & Yardley, 1979) and serpentinite mélanges (Kennedy, 1980) was more characteristic of the Clew

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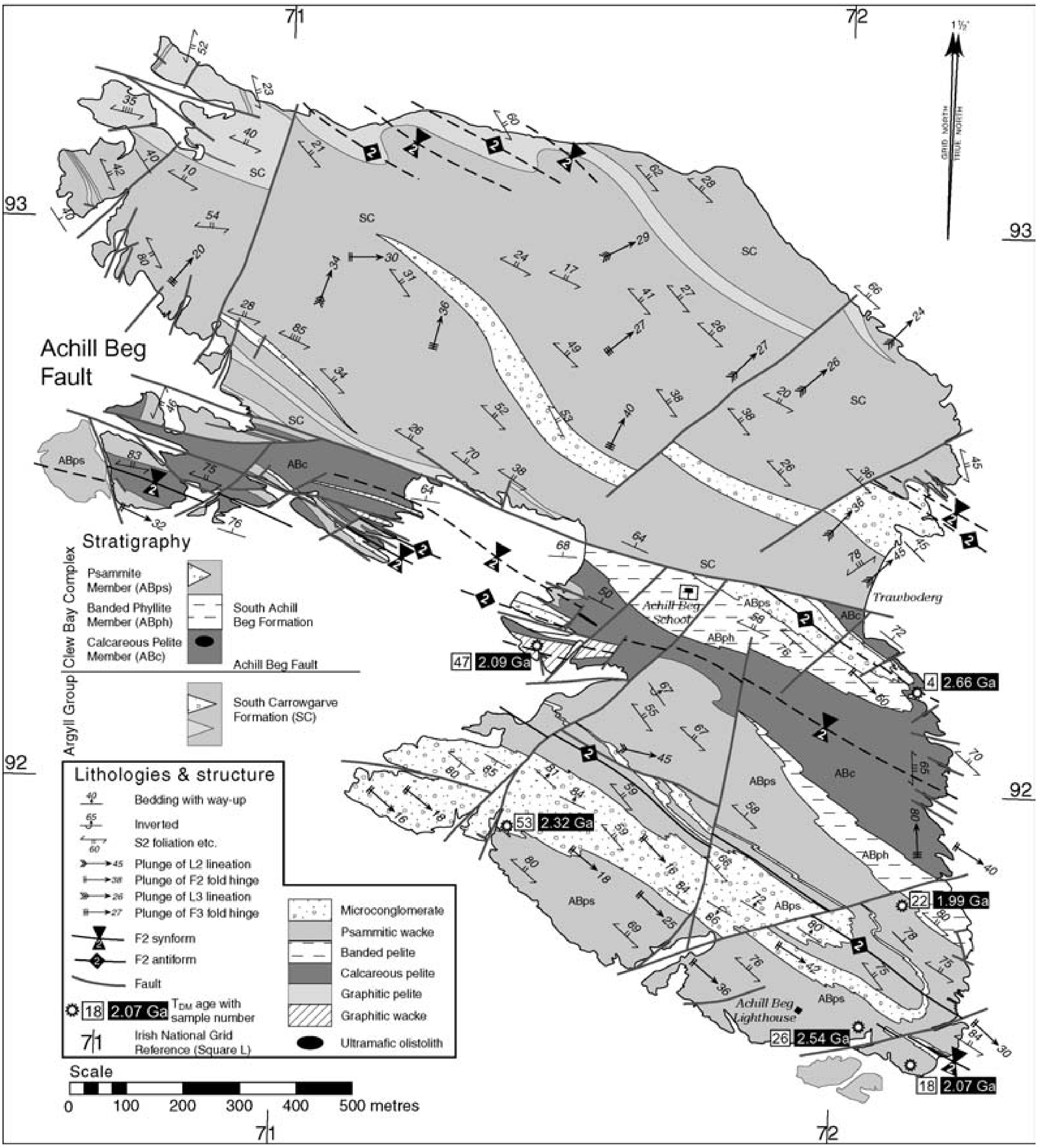

Figure 2. Geological map of Achill Beg based on the author's work.

Bay Complex to the south. However, more recent research detailed below suggests that the South Achill and North Achill Beg succession should be assigned to the upper part of the Argyll Group. An iron- and aluminium-rich pelite horizon which contains abundant chloritoid (North Carrowgarve Formation; Fig. 4a,b) overlying a distinctive basic metavolcanic unit (Claggan Volcanic Formation; Fig. 4a,b) has been recognized within the sequence (Long *et al*. 1992) and this valuable marker horizon can be traced throughout the Easdale Subgroup of the Irish Dalradian (Long, 1997; Fig. 4a). Additionally, serpentinite olistoliths are abundant at this stratigraphic level within the Dalradian of Scotland and Ireland (Chew, 2001; Fig. 4a), and hence the South Achill and North Achill Beg sequence is not regarded as atypical of the Dalradian. These olistoliths are interpreted as protrusions of serpentinized mantle onto the seafloor that were generated in Easdale Subgroup times during a phase of major crustal extension (Chew, 2001).

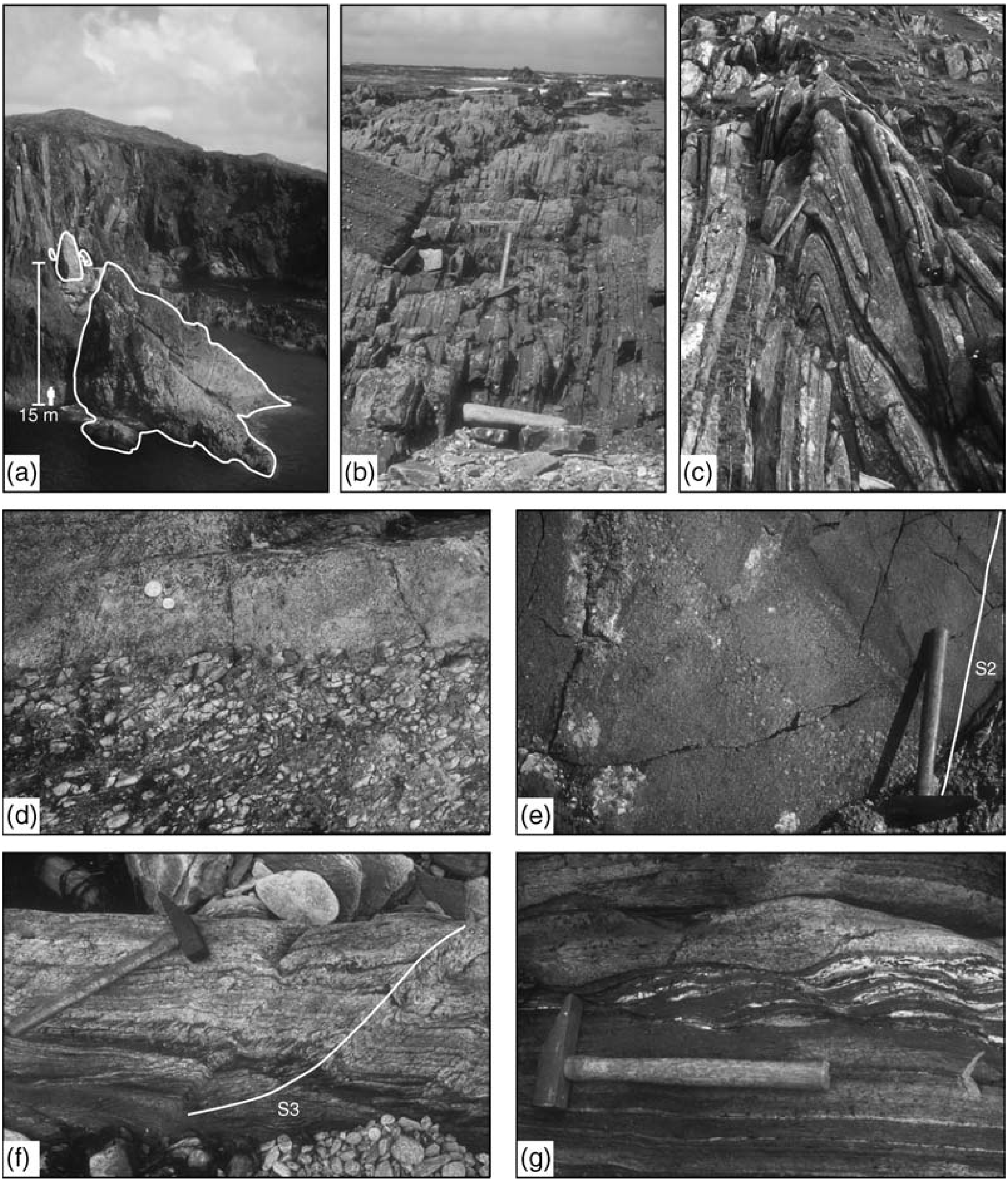

Figure 3. (a) Large serpentinite olistolith enclosed by graphitic pelite. Ooghnadarve Formation, South Achill Dalradian [L69679485]. (b) D1 slide in flaggy quartzite. Cove Schist Formation, South Achill Dalradian [L68949665]. (c) Isoclinal F1 fold refolded by F2 fold. North Carrowgarve Formation, South Achill Dalradian [L68999547]. (d) Prolate L1 pebble stretching in the Keem Conglomerate. Dalradian, North Achill [F58580394]. (e) Grading in coarse pebbly sandstone facing down on the S2 cleavage. Clew Bay Complex, South Achill Beg [L71509194]. (f) Asymmetric buckle folds (F3), interpreted as reverse slip crenulations (RSC). Later rotation of S3 cleavage is due to progressive dextral shear. North Carrowgarve Formation, South Achill Dalradian [L69329495]. (g) D3 dextral shear bands cutting the S2 foliation, interpreted as normal slip crenulations (NSC). North Carrowgarve Formation, South Achill Dalradian [L69029544]. Hammer is 40 cm long, largest coin in Figure 3d is 3 cm in diameter.

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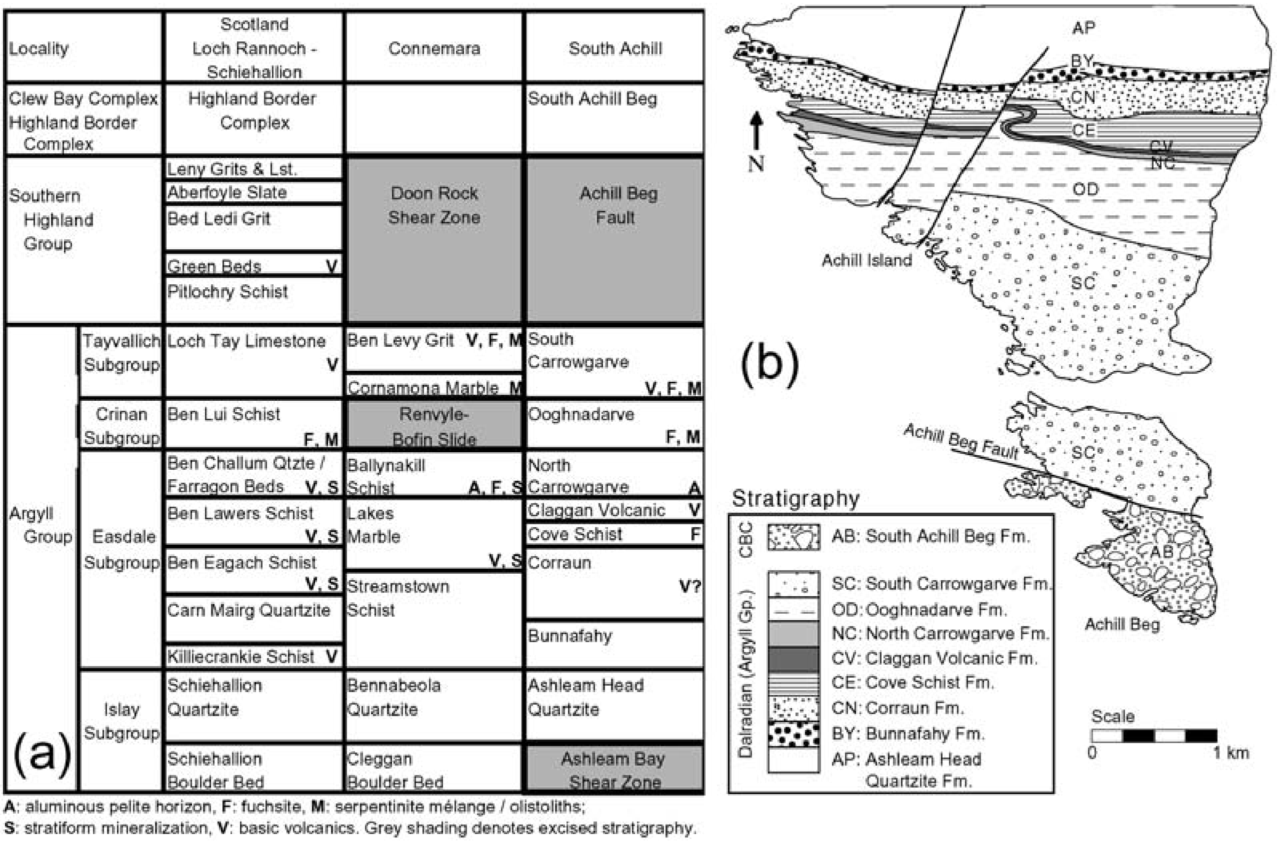

Figure 4. (a) Lithostratigraphic correlation chart for selected Argyll Group and younger successions in Scotland and Ireland. The stratigraphy and occurrences of ultramafic-related detritus are adapted from Chew (2001). (b) Summary geological map of South Achill and Achill Beg.

4.b. Clew Bay Complex

4.b.1. Calcareous Pelite Member (140 m) of the South Achill Beg Formation

This unit is the oldest of the three members that comprise the South Achill Beg Formation and its base is not seen. The dominant lithology is black graphitic slate with thin interbeds of orange/brown-weathered limestone. These thin (0.5–10 cm) limestone beds are laterally continuous at outcrop scale. Microconglomerate beds with clasts up to 2 cm are found locally. Strained polycrystalline quartz is the dominant clast type, with only minor amounts of detrital muscovite and plagioclase feldspar and a matrix composed of graphitic pelite. Slumped horizons are locally developed, with blocks of psammite up to 10 cm long entirely enclosed by the graphitic pelite matrix. A 40 m long talc-rich block occurs within this member. Its contact relationships are poorly exposed, and hence it remains uncertain if it is an altered serpentinite olistolith similar to those encountered on South Achill.

4.b.2. Banded Phyllite Member (110 m) of the South Achill Beg Formation

This unit is in stratigraphic contact with the underlying Calcareous Pelite Member. The dominant lithology is finely laminated grey phyllite with individual laminae ranging from 0.5–5 mm in thickness. The lamination is defined by alternations of quartz-rich laminae and chlorite–muscovite rich bands, and appears to be laterally continuous on the scale of an outcrop. Small quartz grains (up to 0.5 mm) are occasionally visible.

4.b.3. Psammite Member (300 m) of the South Achill Beg Formation

This unit is in stratigraphic contact with the underlying Calcareous Pelite Member and its top is not seen. The majority of the member consists of psammitic wacke and pebbly psammitic wacke, but up to 200 m of microconglomerate is developed within the Psammite Member in the southwest corner of the island. Individual microconglomerate beds are up to 1 m thick, with sharp erosional bases. Pelitic rip-up clasts up to 30 cm long are commonly observed at the base of these coarse pebble beds. Where grading is observed (e.g. Fig. 3e), the tops of the graded microconglomerate beds are usually psammitic, but occasionally laminated pelitic tops are seen. This unit is thought to be turbiditic in origin (Max, 1989; C. T. Morley, unpub. Ph.D. thesis, Univ. Dublin, 1966). Quartz, plagioclase, K-feldspar, muscovite, tourmaline and zircon (in order of abundance) are the dominant clast types. The quartz clasts are up to 5 cm long, and both mono- and polycrystalline quartz are present, commonly containing needles of rutile. Composite quartzofeldspathic clasts are locally encountered, suggesting a granitic or acid gneiss source.

5. Age and provenance of the South Achill Beg Formation (Clew Bay Complex)

This study yielded no conodonts or microfossils from the South Achill Beg Formation. Thirteen samples, each weighing several kilograms, were taken from a variety of pelitic lithologies in the South Achill Beg Formation and processed for microfossils, while five substantial (> 5 kg each) limestone samples from the Calcareous Pelite Member were processed for conodonts. All samples ultimately proved barren and hence the age of the formation remains uncertain. The South Achill Beg Formation has been traditionally assigned to the Clew Bay Complex by various authors (e.g. Williams *et al*. 1994; Williams, Harkin & Higgs, 1996; Harper, Williams & Armstrong, 1989; Max, 1989) based on lithological similarities. The South Achill Beg Formation and the Ballytoohy Formation of the Clew Bay Complex on Clare Island (Fig. 1b) both contain slumped horizons, thinly interbedded limestone and graphitic pelite, and a thick psammitic wacke succession. Heavy mineral analysis and Sm–Nd model age data reported in Section 5.b also support the South Achill Beg Formation being assigned to the Clew Bay Complex.

5.a. Heavy mineral analysis

Eight separate psammitic samples from the South Achill Beg Formation were processed for heavy mineral analysis. The assemblages recovered are in general of low diversity with zircon being by far the most abundant species. It is most commonly present as extremely well-rounded, pale pink to purple varieties, although pale pink subhedral prisms are also observed. Other common species include both euhedral, prismatic tourmaline along with well-rounded varieties, and apatite. Grains of rutile, titanite and euhedral anatase of probable authigenic origin are also occasionally present. There is a marked similarity to heavy minerals recovered from the Killadangan Formation of the Clew Bay Complex on the south shore of Clew Bay (Dewey & Mange, 1999). The provenance of the Killadangan Formation was believed to be a gneissic basement, presumably the Laurentian continent (Dewey & Mange, 1999), and a similar provenance is envisaged for the South Achill Beg Formation. Importantly, neither this study, nor that of Dewey & Mange (1999) have recovered definitive Grampian detritus from the Clew Bay Complex, which might be expected if these parts of the Clew Bay Complex were Silurian in age. A mineral separate containing 30 detrital muscovite grains from the South Achill Beg Formation has yielded a *c*. 2 Ga $^{40}Ar–^{39}Ar$ error plateau (D. M. Chew, unpub. Ph.D. thesis, Univ. College Dublin, 2001). Importantly, although the age is poorly constrained, even the lowest temperature step (1234 ± 21 Ma), shows negligible evidence for a Grampian (*c*. 475–460 Ma) component in the detrital muscovite population (D. M. Chew, unpub. Ph.D. thesis, Univ. College Dublin, 2001).

5.b. Sm–Nd model age data from the South Achill Beg Formation

Several studies have documented that the Sm–Nd isotopic signature of sedimentary rocks is usually unfractionated between the source area and the clastic sediment derived from it (e.g. Mearns *et al*. 1989; McLennan, McCulloch & Taylor, 1989; Goldstein & Jacobsen, 1988; Nelson & DePaolo, 1988). The sediment produced may be of mixed provenance, and as such the Sm–Nd isotopic signature of a whole rock sample is a weighted average of the Sm–Nd isotopic signature of its various protoliths.

Three psammitic samples and one pebbly psammite sample from the Psammite Member of the South Achill Beg Formation, and one psammitic sample and one pebbly psammite sample from the Calcareous Pelite Member, were selected. The $^{147}Sm/^{144}Nd$ ratios (0.099–0.124, Table 1) of all six samples fall within the normal range for clastic sediments. T_{DM} ages range from 1.99–2.66 Ga (mean = 2.28 Ga, n = 6, Table 1). The pebbly psammite samples display the highest Sm and Nd concentrations and the oldest T_{DM} ages. The simplest explanation for this is that the presence of a shaly matrix in the pebbly psammites contains the bulk of the REE in the sample, much higher than the REE abundances of the quartzofeldspathic clasts. The shale matrix in the pebbly psammite samples would therefore be the major contributing factor to the old T_{DM} ages, with the shale matrix possibly sampling a different (older) source area than the psammitic samples. Sm–Nd analysis of the shale matrix is required to test this hypothesis.

Comparing the South Achill Beg Formation with Sm–Nd data from the Dalradian (Daly & Menuge, 1989) and the rest of the Clew Bay Complex (Harkin *et al*. 1996) reveals similarities with Sm–Nd data from the rest of the Clew Bay Complex, rather than with the Argyll or Southern Highland groups of the Dalradian. T_{DM} ages for the Argyll Group range from 1.65–2.39 Ga (mean = 1.99 Ga, n = 5: Daly & Menuge, 1989), while T_{DM} ages for the Southern Highland Group are virtually identical and range from 1.79–2.36 Ga (mean = 2.03, n = 5: Daly & Menuge, 1989). In contrast, T_{DM} ages from the Killadangan and Ballytoohy formations of the Clew Bay Complex (Fig. 1b) are significantly older, ranging from 2.21–2.65 Ga (mean = 2.38 Ga, n = 8: Harkin *et al*. 1996). This compares favourably with the South Achill Beg

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Table 1. Sm-Nd isotopic data for whole rock samples of the South Achill Beg Formation

Sample	Type	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	$\pm 2\sigma$	T_{DM} (Ma)
AB-4	Pebbly psammite	3.82	18.5	0.1249	0.511498	0.000016	2657
AB-18	Psammite	2.60	14.3	0.1103	0.511631	0.000014	2073
AB-22	Psammite	2.95	15.8	0.1127	0.511719	0.000012	1990
AB-26	Pebbly psammite	4.88	24.1	0.1226	0.511526	0.000012	2539
AB-48	Psammite	3.15	16.8	0.1133	0.511659	0.000014	2094
AB-53	Psammite	1.49	9.08	0.0992	0.511287	0.000034	2332

Analyses were performed on a semi-automated single collector VG Micromass 30 mass spectrometer at the Department of Geology, University College Dublin. $^{143}Nd/^{144}Nd$ ratios, Sm and Nd concentrations were determined using a mixed ^{147}Sm-^{150}Nd spike. $^{143}Nd/^{144}Nd$ ratios are normalized to $^{146}Nd/^{144}Nd = 0.7219$. $^{143}Nd/^{144}Nd$ errors are within-run precision and reproducibility is approximately ± 0.00002, while reproducibility of $^{147}Sm/^{144}Nd$ ratios is typically 0.1 %. These analytical errors correspond to typical uncertainties in T_{DM} ages of about 40 Ma. T_{DM} ages were calculated using the depleted mantle curve of DePaolo (1981).

Formation which has T_{DM} ages ranging from 1.99–2.66 Ga (mean = 2.28 Ga, n = 6, Table 1).

6. Structural relationships across the Achill Beg Fault

The major structure in the Northwest Mayo inlier is a S-facing, recumbent F2 nappe which is rotated into a downward-facing orientation approaching the Achill Beg Fault (Fig. 5). The following synthesis focuses on the structural relationships observed on the island of Achill Beg (where the Dalradian/Clew Bay Complex contact is exposed), and also considers the structures observed in the South Achill Dalradian, immediately to the north of Achill Beg. The various structural sub-domains referred to in the text are displayed in Figure 6a.

6.a. D1 deformation event

The D1 deformation event is responsible for the bulk of the high-strain seen in the South Achill and Achill Beg successions. In the Clew Bay Complex of South Achill Beg, pronounced pebble-flattening along S1 cleavage planes is visible when S1 is at an angle to S2. Because bedding and the two primary cleavages (S1 and S2) are usually coplanar in the Dalradian of North Achill Beg and South Achill, it is difficult to establish whether D1 or D2 has imparted the majority of the shape fabric observed in the sporadically developed pebble beds. Many authors have suggested that the pebble-shape fabric in the Northwest Mayo inlier is D2 in age, contemporaneous with the development of a prominent L2 mineral lineation which plunges shallowly east (e.g. Harris, 1993, 1995; Johnston, 1995; Section 6.b). However, pebble-shape fabric data from the Keem Conglomerate in the northern part of Achill Island (Fig. 6a), where the effects of later D2 deformation are minor, show that D1 is associated with prolate strains and that the observed pebble-stretching lineation is clearly L1 in age (Fig. 3d; M. J. Kennedy, unpub. Ph.D. thesis, Univ. Dublin, 1966; Kennedy, 1969). Additional evidence for high D1 strain in the Dalradian adjacent to the Achill Beg Fault is provided by the presence of low-angle ductile shear zones (slides) in the Dalradian of South Achill (e.g. the Claggan Bay Mylonite Zone of Harris, 1993, 1995; Fig. 3b). These slides are clearly refolded by downward-facing F2 folds (D. M. Chew, unpub. Ph.D. thesis, Univ. College Dublin, 2001). F1 folds are locally observed in South Achill (Fig. 3c) and Achill Beg; in both the Clew Bay Complex and Dalradian rocks they are isoclinal, with fold hinges plunging shallowly to the east.

6.b. D2 deformation event

F2 folds are the most commonly observed fold generation in South Achill and North and South Achill Beg. In general they are upright structures that plunge shallowly to the east, similar to the F1 fold plunge. Throughout South Achill, the F2 folds are predominantly asymmetric, with a 'Z' asymmetry viewed down-plunge (Fig. 5). The structure of the F2 fold phase on South Achill Beg is similar to that of South Achill. Way-up evidence is abundant in quartzites of the Ashleam Head Formation (Fig. 4b) of South Achill and the Psammite Member of the South Achill Beg Formation (Fig. 2). Bedding is generally inverted throughout South Achill and Achill Beg and the S2 foliation consistently faces downwards (Fig. 3e). This is consistent with the southward-younging nature of the sequence and an overall 'Z' sense of asymmetry to the eastward-plunging F2 folds. The consistent downward-facing directions on S2 implies that large-scale reversals due to F1 folding must be absent.

F2 folds in general plunge shallowly eastward in South Achill (Fig. 6b), and to the southeast on Achill Beg (Fig. 6c), due to later clockwise rotation of the D2 structures by dextral shear during D3 (Section 6.c). Some F2 folds in South Achill plunge moderately west (Fig. 6b), but there is no evidence of large scale F2 sheath folds. A conspicuous L2 mineral lineation usually defined by muscovite and quartz ribs and consistently parallel to F2 fold axes is well developed in South Achill and Achill Beg and has been assumed to represent the stretching or x-direction in these rocks (e.g. Harris, 1993, 1995; Johnston, 1995).

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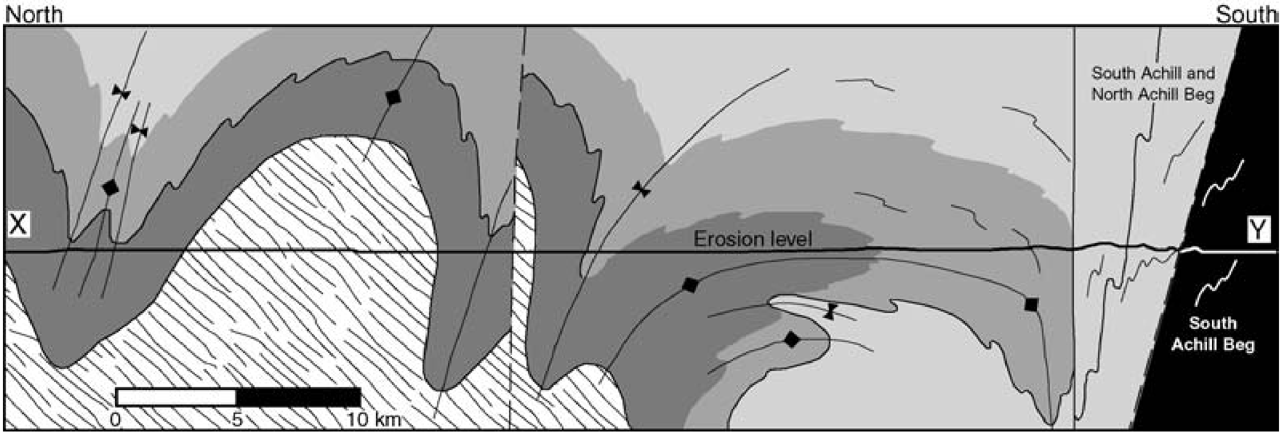

Figure 5. Cross-section across the Northwest Mayo inlier displaying major F2 folds. Legend and line of section (X–Y) as in Figure 1b. Cross-section approximately N–S ([F685416]–[L718897]).

If the L2 mineral lineation does indeed represent a true stretching lineation, then the marked similarity in orientation between the L2 lineation and the F2 fold hinges could be explained by progressive rotation of the F2 fold axes towards the stretching direction under high shear strains. However, no F2 sheath folds have been observed, and strain analysis data (obtained using the R_f/ϕ technique) from six pebble-bearing samples from the South Achill Dalradian are not indicative of high shear strain. The observed elongation in the XY plane is relatively small, with R_{xy} ranging from 1.23–1.56, and this elongation deviates by up to 34° in pitch from the F2 fold axes and the L2 lineation (D. M. Chew, unpub. Ph.D. thesis, Univ. College Dublin, 2001). It is felt more likely that the observed L2 lineation is in fact an S2/S0 intersection lineation, albeit often with an extremely acute angle of intersection. Such a hypothesis is also supported by the fact that the observed L2 lineation is consistently parallel to the S2/S0 intersection lineation, which can be measured where a small angular discordance between S0 and S2 is observed (D. M. Chew, unpub. Ph.D. thesis, Univ. College Dublin, 2001).

6.c. D3 deformation event

The D3 deformation event is represented by an E–W-trending, vertical shear zone with sub-horizontal dextral translation sited close to the former Laurentian margin, similar to the model proposed by Sanderson *et al.* (1980). The northern boundary of the shear zone to the north is not sharply defined, but D3 structures gradually become less abundant to the north of Ashleam Bay in South Achill (Fig. 6a). The D3 deformation event is responsible for much of the S2 strike swing (Figs 1b, 6d) and the production of downward-facing folds in the southern portion of the inlier in the vicinity of the Achill Beg Fault (Fig. 5). Two discrete structural elements are recognized: the reverse-slip crenulations (asymmetric buckle folds; Figs 3f, 6e) and normal-slip crenulations (extensional shear bands; Fig. 3g) of Dennis & Secor (1990). These crenulations are developed in order to compensate for the displacement component of foliation slip normal to the shear zone wall (Dennis & Secor, 1990). A simple shear deformation path is then preserved, with the crenulation morphologies being controlled by the orientation of the pre-existing foliation.

6.d. Geochronology of muscovite defining the S2 nappe fabric

Rb–Sr, ^{40}Ar–^{39}Ar step heating and *in situ* laserprobe dating of muscovite defining the main S2 nappe fabric in South Achill and the Clew Bay Complex (except for samples affected by later deformation) yield *c.* 460 Ma ages, while ^{40}Ar–^{39}Ar *in situ* laserprobe dating of S3 muscovite in South Achill has yielded *c.* 448 Ma ages (D. M. Chew, unpub. Ph.D. thesis, Univ. College Dublin, 2001). Peak metamorphic temperature of the South Achill Dalradian is estimated at *c.* 450 °C by garnet–biotite thermometry while peak metamorphic temperature of the significantly lower grade Clew Bay Complex has been more crudely estimated at 250–350 °C based on thermal alteration values of amorphous organic matter (D. M. Chew, unpub. Ph.D. thesis, Univ. College Dublin, 2001). Significantly, Rb–Sr S2 muscovite dates in both units are almost certainly recording the time of crystallization as the Rb–Sr muscovite closure temperature is usually quoted as *c.* 500 °C (e.g. Cliff, 1985). The geochronology thus helps to confirm the structural observations that the Dalradian and the Clew Bay Complex have shared the same deformation history.

A *c.* 460 Ma age for the development of the main nappe fabric in the Northwest Mayo inlier is marginally younger than the timing of main nappe development on the southern margin of the Dalradian belt elsewhere in Ireland. In Tyrone, the D3 Omagh Thrust has translated inverted Dalradian rocks towards

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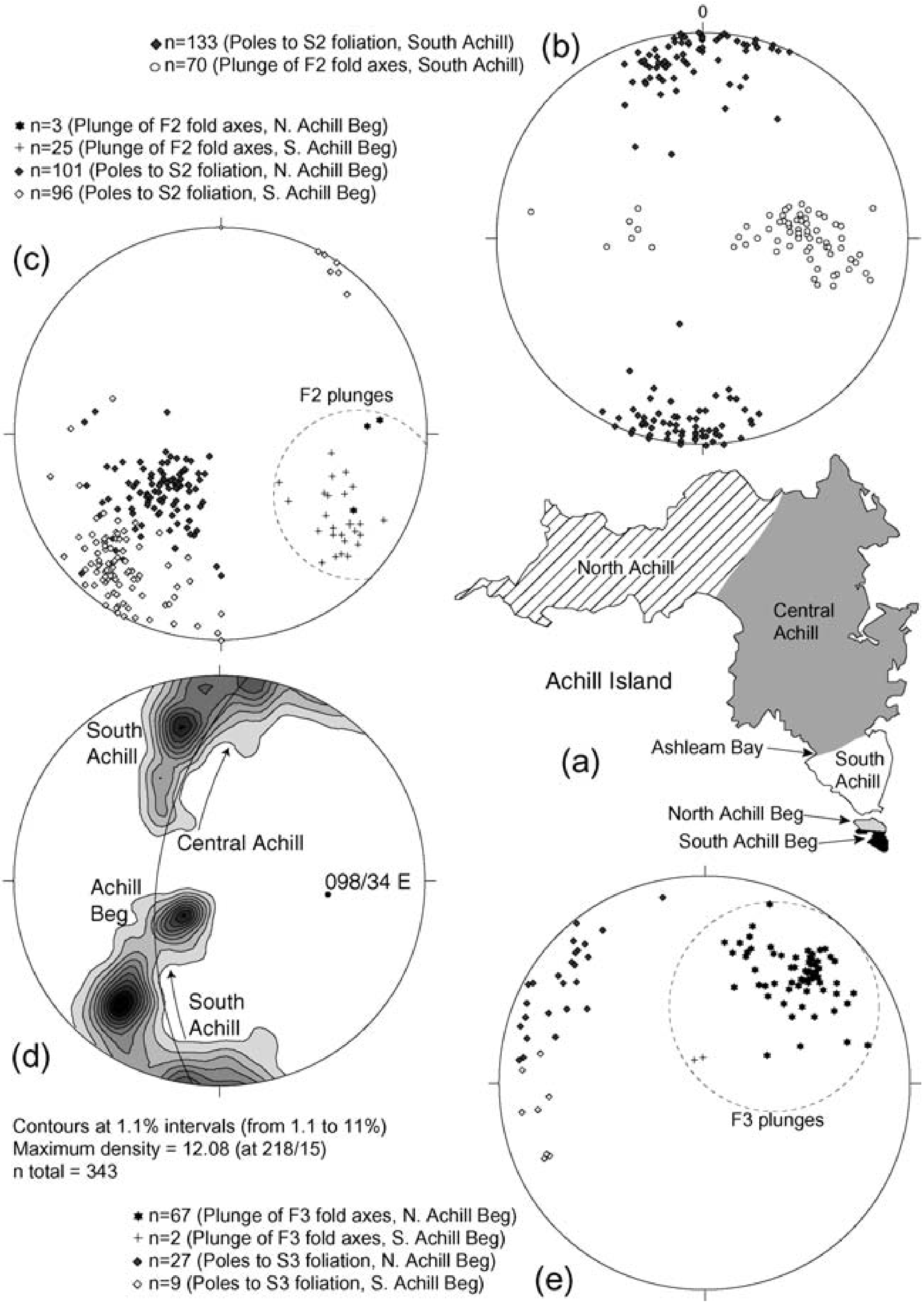

Figure 6. (a) Structural sub-domains referred to in the text. (b) Stereographic plot of the orientation of F2 fold hinges and poles to the S2 foliation. South Achill. (c) Stereographic plot of the orientation of F2 fold hinges and poles to the S2 foliation. North Achill Beg and South Achill Beg. (d) Stereographic plot of the contoured poles to the S2 foliation. Central Achill to South Achill Beg. Arrows indicate the direction of progressive rotation (clockwise) of the S2 poles. This rotation is as a result of D3 dextral shear. (e) Stereographic plot of the orientation of F3 fold hinges and poles to the S3 foliation. North Achill Beg and South Achill Beg.

the ESE (Alsop & Hutton, 1993) over Arenig–Llanvirn shales of the Tyrone volcanic rocks (Hutton & Holland, 1992). Similarly, in southern Donegal and the Northeast Ox Mountains inlier, Dalradian rocks were thrust to the southeast over high-pressure granulite-facies basement (the Slishwood Division) along the Lough Derg Slide (Alsop, 1991) and the North Ox Mountains Slide (Flowerdew, 1998/9) respectively. Tectonic juxtaposition (D3) of the Dalradian and Slishwood Division is likely to have occurred between 470 and 459 Ma based on ^{40}Ar–^{39}Ar, Rb–Sr and Sm–Nd mineral ages (Flowerdew *et al*. 2000).

7. Conclusions

Evidence is presented for an Upper Argyll Group affinity for the South Achill and North Achill Beg Dalradian successions, based on the occurrence of an Easdale Subgroup marker horizon and a new interpretation for the origin of the ultramafic detritus in the sequence. Although no biostratigraphically diagnostic faunas were recovered from the South Achill Beg Formation, a Clew Bay Complex affinity is favoured by lithostratigraphic correlation, Sm–Nd isotopic age data and analysis of heavy mineral species. Thus the Achill Beg Fault is believed to separate Dalradian Supergroup rocks from the Clew Bay Complex. Detailed structural field mapping demonstrates that the Dalradian of South Achill and North Achill Beg has shared the same structural history as the Clew Bay Complex of South Achill Beg. The D1 deformation event is the major high-strain event, and is associated with the development of tectonic slides. The D2 deformation phase is the major nappe-forming event. Beds consistently face down on the S2 foliation in both the Dalradian and the Clew Bay Complex. Muscovite defining the S2 nappe fabric in both units yields *c*. 460 Ma Rb–Sr and ^{40}Ar–^{39}Ar ages and thus provides further evidence for contemporaneous deformation. The D3 deformation event represents an E–W-trending, vertical shear zone with sub-horizontal dextral translation sited close to the former Laurentian margin, and muscovite defining the S3 fabric has yielded *c*. 448 Ma ^{40}Ar–^{39}Ar ages. It is responsible for much of the S2 strike swing and production of downward-facing folds in the southern portion of the inlier. Contemporaneous deformation of the Dalradian and the Clew Bay Complex during a mid-Ordovician Grampian orogenic event casts doubt on the 'exotic status' of the Highland Border Complex of Scotland, and Silurian microfossils retrieved from the Clew Bay Complex.

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