The last glaciers in the Aran and Arenig Mountains, North Wales

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Signed : P. Hughes.

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ABSTRACT

Geomorphological evidence for four former local glaciers has been mapped in the Aran and Arenig Mountains, North Wales. Pollen stratigraphic analysis of infilled lake sediments has enabled an age of the former glaciers to be deduced. Outside of the former glacier limits at Ffridd-y-Fawnog (grid ref. SH 866457) a full suite of Lateglacial and Flandrian deposits exist whilst inside of the former glacier limits at Cwm Gylchedd (grid ref. SH 866457) only Flandrian deposits exist. This implies a Loch Lomond Stadial age of the glacier at Cwm Gylchedd and, by analogy, at the three other glaciers mapped in this study. This finding is also supported by periglacial contrasts between the insides and outsides of the glacier limits. Reconstruction of the four glaciers illustrates a mean ELA of c.504 metres. Variation in ELA between the four glaciers can be primarily attributed to precipitation differences and not variations in local controlling factors. From the reconstructed ELAs and the combination of precipitation and snowblow input in deriving total accumulation, by analogy with Norwegian glaciers, a mean sea-level July temperature was calculated at 8.4°C ± 0.4°C. Also, if the Loch Lomond Stadial is assumed to represent the early stages of ice-sheet build up, then the Arenig Mountains are likely to have been an important centre for earlier Devensian ice-sheet build up.

1. INTRODUCTION

1.1. Aims and Objectives

This study aims to locate sites of former glacier readvance in the Aran and Arenig Mountains, North Wales. The extent of such glaciers can be delimited via detailed geomorphological field mapping. An age of local glaciation, the last at these sites, can be deduced by comparing the pollen stratigraphy of infilled basin sediments within mapped glacier limits with those outside of such limits. Following glacier delimitation, the former glaciers can then be reconstructed by analogy with modern glaciers, from which inferences can be made of former climate and local controlling factors.

1.2. Study Area

The Aran and Arenig Mountains lie in the southern part of the Snowdonia National Park in North Wales (refer to Figure 1.1.). They both provide the western watershed of the River Dee catchment basin and the largest natural lake in Wales, Llyn Tegid (commonly known as Bala Lake). The Arenig Mountains are named after the mountain peaks of Arenig Fawr (854 metres) and Arenig Fach (689 metres) and lie to the north-east of Llyn Tegid (refer to Figure 1.2). The Aran Mountains lie to the southeast of Llyn Tegid and are dominated by the 'Aran Ridge' comprising the peaks of Aran Fawddwy (905 metres) and Aran Benllyn (885 metres) (refer to Figure 1.3).

The Arenig district has been a source of interest to geologists ever since the time when Sedgwick (1843) applied the name of Arenig Ashes and Porphyries to the lower series of North Welsh volcanic rocks. The geology of the Arenig region is dominated by Ordovician volcanic strata comprising ashes, ignimbrites, intercalated andesites, lavas and associated andesitic intrusions, which by differential erosion have given rise to the major peaks of Arenig Fawr and Arenig Fach (Rowlands 1979). North-eastwards of the main Arenig peaks, around Cwm Gylchedd, these volcanic rocks are flanked by Ordovician mudstones, grits and sandstones. The main ridge of the Aran Mountains is also dominated by Ordovician volcanic strata and is flanked to the east by slates and shales, again belonging to the same geological Period. Figure 1.4 provides a simplified illustration of the solid geology in these areas.

Cirques, or cwms in Welsh, are well developed, particularly on the eastern flank of the Aran Ridge, the north-east faces of Arenig Fawr and Arenig Fach and the northern

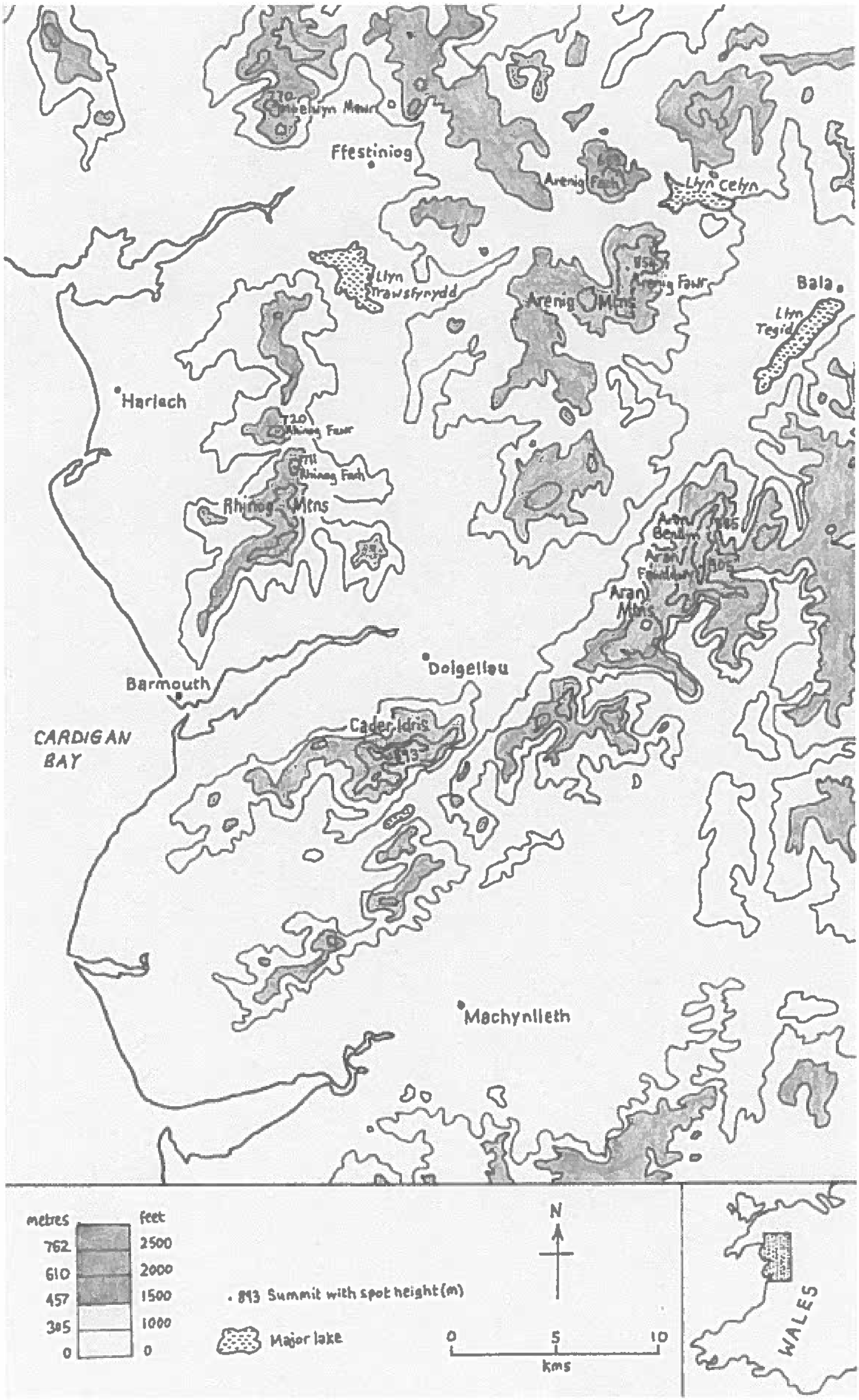


Figure 1.1. Location map of the Aran and Arenig Mountains.

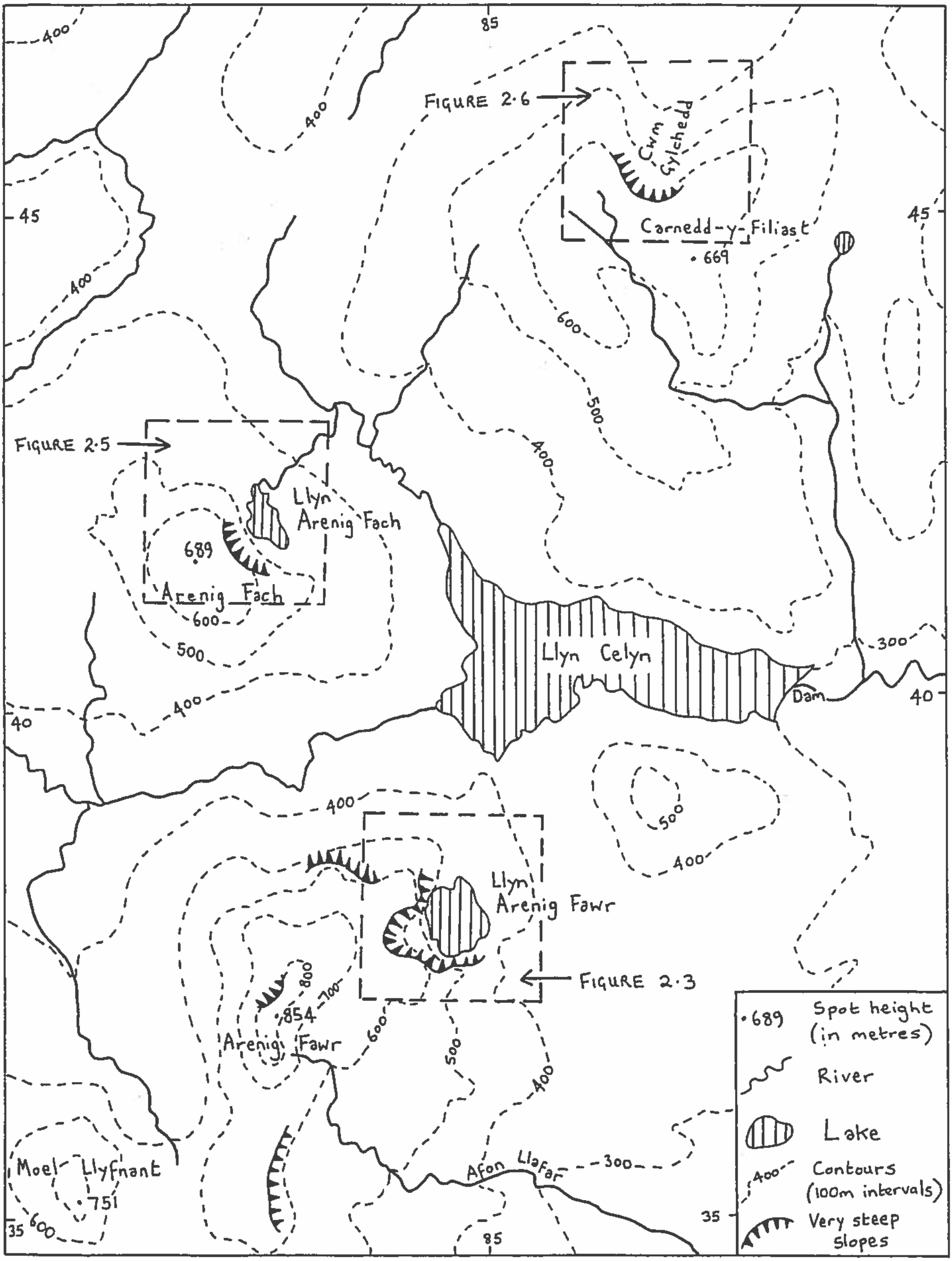


Figure 1.2. Map of the Arenig Mountains at a scale of 1:50,000.

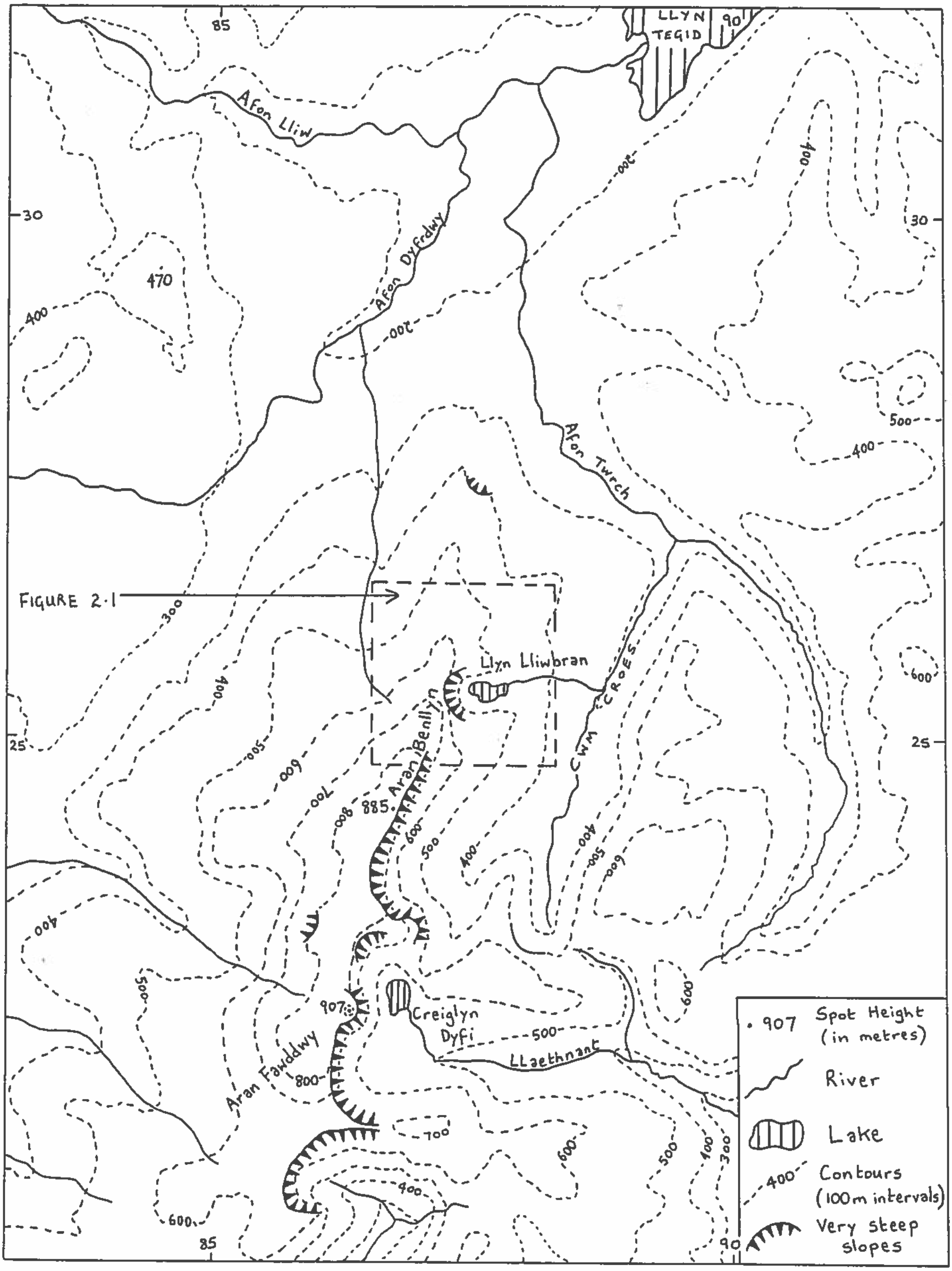


Figure 1.3. Map of the Aran Mountains at a scale of 1:50,000.

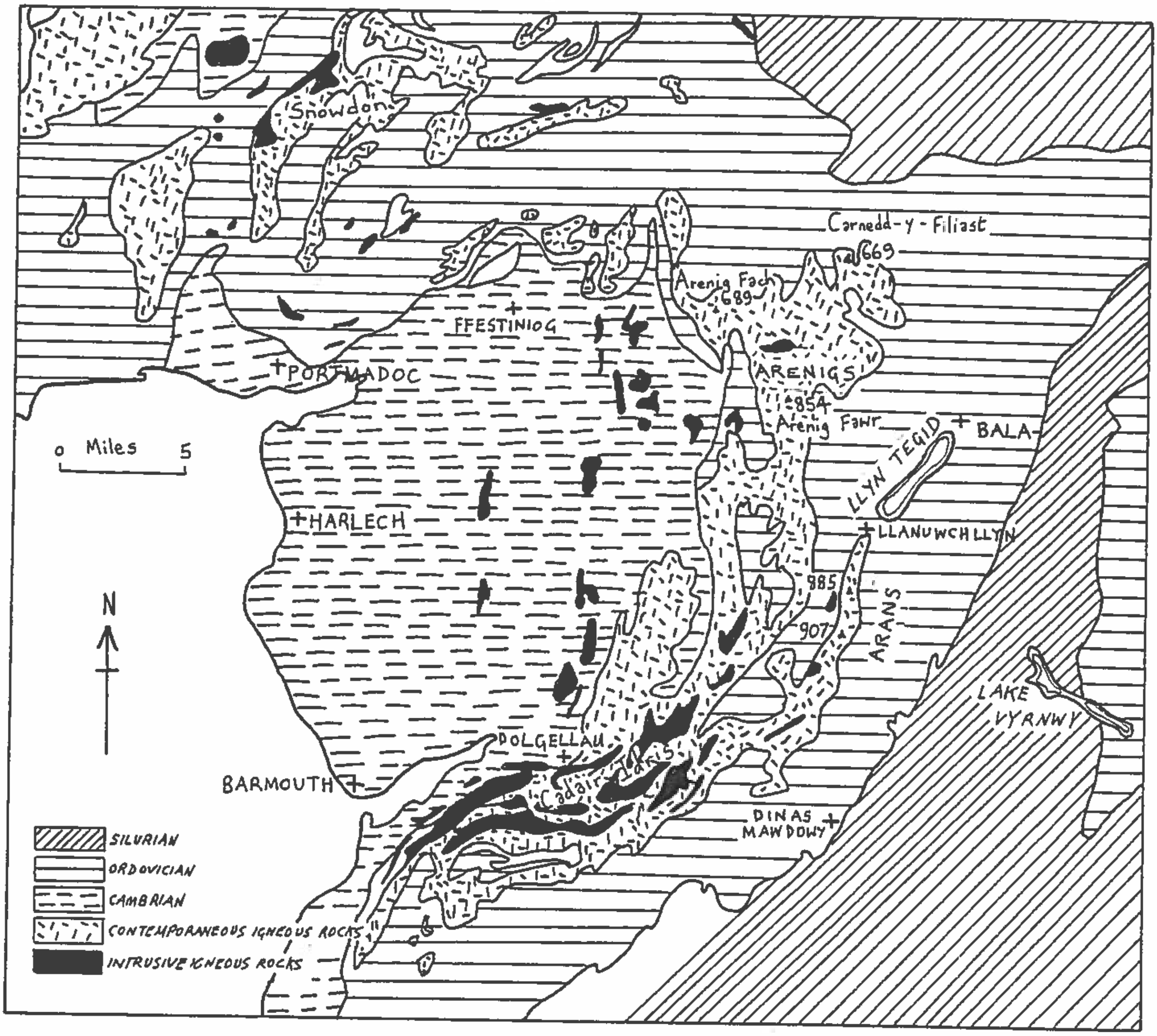


Figure 1.4. Simplified geological map of the southern Snowdonia area.

slopes of Carnedd-y-Filiast. Within these cwms, localised moraines are evident and are accompanied by several geomorphological features suggestive of local glacier occupation. These former glaciers represent the last glaciers of the Arans and Arenigs and are the focus of this study.

1.3. Previous Research

It has long been recognised that cirque moraines in the British Uplands indicate a limited glacial phase after the last ice sheets. Davis (1909) noted the remarkable prevalence of localised moraines in the cwms of Snowdonia and evidence for former local glaciers in various parts of Scotland began to be published as early as the 1840s (Sissons 1980a). It is now widely thought that many, although not all, of these localised moraines are the result of a short reoccupation of the cwms by glacier ice during the Loch Lomond Stadial.

The Loch Lomond Stadial in Britain or Younger Dryas in Europe represents climatic deterioration at the end of the Devensian Lateglacial period following the warm Windermere Interstadial in Britain or Allerød Interstadial in Europe. Some workers propose a temporary shut-down in the ocean conveyor as a primary cause of the Younger Dryas climatic reversal (Broecker and Denton 1990, Broecker et al 1990). The cause of this shutdown was perhaps related to massive freshwater discharge into the North Atlantic , the result of a diversion of meltwater from the modern Mississippi route to the St Lawrence canal (Broecker and Denton 1990). However, Veum et al (1992) found no evidence of a reduction in North Atlantic Deep Water formation during this time and thus no evidence of a shut-down in the ocean conveyor.

Whatever its cause, the Younger Dryas is recorded in the marine, ice and terrestrial records of the northwest Atlantic region. The event appears to have had most impact in this region although has been recorded in various proxies globally (e.g. Denton and Hendy 1994, Mathewes et al 1993, Roberts 1993, Jouzel et al 1992, Linsley and Thunnel 1990). The climatic oscillation during the Devensian Lateglacial is suitably described with reference to the work of Ruddiman and McIntyre (1981) who tracked movements in the Atlantic Polar Front from the marine record. Following the maximum glacial period at c.18,000 BP, the Polar Front moved from northern Portugal to north of Iceland during the warm Allerød/Windermere Interstadial. The Polar Front returned southwards once again between 11,000 and 10,000 BP during the Younger Dryas/Loch Lomond Stadial to a position south of Ireland. Widespread terrestrial palaeoecological evidence such as Coleoptera (Coope and Brophy 1972, Coope and Joachim 1980, Atkinson et al 1987) and pollen also suggest a return to a short period of cold conditions between 11,000 and 10,000 years BP. Lateglacial pollen studies in the British Isles are now very numerous of which some particularly notable North Wales studies include Seddon (1962), Crabtree (1972), Simpkins (1974), Ince (1981), S.Lowe (1981, 1994), J.Lowe et al (1988) and J.Lowe and S.Lowe (1989).

There is now also a large collection of geomorphological data suggesting that glaciers occurred throughout the British uplands during the Loch Lomond Stadial. A large proportion of study has been done in Scotland where Brian Sissons has arguably been the pioneer of Loch Lomond Stadial glacial reconstruction, particularly in his work in the central Grampians (Sissons 1974), southeast Grampians (Sissons and Sutherland 1976), Skye (Sissons 1977a), northwest Scotland (Sissons 1977b) and the Cairngorms (Sissons 1979a). It was work by Sissons and colleagues in Scotland which first quantified Loch Lomond Stadial glacier characteristics and enabled palaeoclimatic inferences from such study. Such early work has been the template for further study although some workers such as Ballantyne (1989) in the Isle of Skye, have questioned the delimitation of Loch Lomond Stadial glaciers by Sissons. Thorp (1986) made a major contribution to the Loch Lomond Stadial glacial record in his delimitation of the West Highland ice-sheet, completing the initial efforts of workers such as Simpson (1933) who first coined the term Loch Lomond Stadial based on field evidence at Loch Lomond itself and at Meinteithin the Upper Forth valley. This major 200 km^2 ice mass stretched from Loch Torridon in the north to Loch Lomond in the south (175 km), and from Loch Shiel in the west to Loch Rannoch in the east (100 km) exceeding 400m in thickness over Rannoch Moor (Gray and Coxon 1991). The Southern Uplands of Scotland also accommodated a significant number of Stadial glaciers with a small ice-cap having existed in the central section (Price 1983) and eleven small glaciers in the western part of this region (Cornish 1981). Elsewhere, localised moraines of possible Loch Lomond Stadial age have been noted on the Isles of Mull (Gray and Brooks 1972), Rhum (Ballantyne and Wain-Hobson 1980), Jura (Dawson 1977), the Outer Hebrides (Sutherland 1993), Orkney (Sutherland 1991) and Shetland (Flinn 1977). Over 200 independent ice masses, separate from that of the main Highland ice-sheet, have been mapped in the Scottish Highlands and Inner Hebrides alone (Sutherland 1984a), illustrating the significance of the Loch Lomond Stadial as a glacial event.

In England, 64 former glaciers were identified in the mountains of the Lake District, Cumbria, by Sissons (1980a). A further 3 sites containing moraine ridges and mounds of Loch Lomond Stadial age were mapped in the eastern Lake District by Wilson and Clark (1998). In the north Pennines, Rowell and Turner (1952) mapped 8 small glaciers between Mallerstang and Ravenstonedale commons, though Mitchell (1996) argued that the morphology at some of these sites was more easily explained as the result of complex mass movements, particularly deep-seated rotational failures and mudslides. Mitchell (1996) has identified the former existence of five cirque glaciers of Loch Lomond Stadial age in the western Pennines and Sissons (1979b) mentions an isolated glacier on the Cheviot in Northeast of England. Elsewhere in the British Isles, in southern Ireland, Anderson et al (1998) found evidence of six small high level glaciers in the Macgillycuddy's Reeks. Also, Colhoun and Synge (1980) showed, via radiocarbon dating of organic silts from ice-pushed silts, that a series of cirque moraines deposited on the floor of Lough Nahanagan in eastern Ireland were formed between c.11,000 and 10,500 BP.

In northern Snowdonia Seddon (1957) mapped the overall distribution of cwm moraines. The mapping is however, presented at a very small scale (c.1: 225,000) and 33 end-moraines are marked only as simple crescents. Pollen analysis of sediment cores both inside and outside of the morainic arcs showed that the end-moraines were formed during both the retreat of the last ice sheet and during the Loch Lomond Stadial. The results of such investigations are published in Godwin (1955) and Seddon (1962). Unwin (1975) provided a more detailed analysis of local moraines at a larger scale albeit still at a scale of c.1:150,000 with detailed maps only for 4 cwms. Gray (1982) provided the first detailed examination and mapping of the former local glacial limits in the area. Gray (1982) mapped evidence for 35 former glaciers in northern Snowdonia covering a total area of c.17.5 km^2 although there is still some debate over the actual limits of the former glaciers such as at the famous Cwm Idwal site (Gray 1990).

Elsewhere in Wales, the Brecon Beacons and surrounding hills have been studied by Ellis-Gruffydd (1977), Walker (1980), Robertson (1989) and Shakesby and Matthews (1993). Both Robertson (1989) and Shakesby and Matthews (1993) reconstructed glacier form and characteristics in the mould of Sissons and co-workers in Scotland, as did Gray (1982) in North Wales. Walker (1980) dated 2 sites with local moraines using pollen stratigraphic evidence from both inside the glacier limits and outside at the Traeth Mawr Lateglacial site. Walker combined this with radiocarbon dating and showed that the local moraines were formed during the Loch Lomond Stadial.

The Aran and Arenig Mountains have received scant published attention relating to glacial readvance during the Loch Lomond Stadial. Gray, Ince and S.Lowe (1981) do mention briefly the moraines at Llyn Arenig Fach in a report on a short field meeting in North Wales. Rowlands (1970, 1979) noted locally derived end moraines in the cwms of Llyn Arenig Fawr and Llyn Arenig Fach although they do not receive much attention. Rowlands' study was concerned with other matters, namely the Arenig region as a centre of the former Welsh ice-cap during the last main glacial c.18,000 BP. Evans (1999) makes note of the "cirques northwest of Arenig Fawr and very clear cirques south east of Bala in the Aran range" and includes some of these in his map showing the distribution of Welsh cirques containing moraines of possible Lateglacial age. Even so, apart from the literature described here there has been no detailed published work regarding Loch Lomond Stadial glaciers in the Aran and the Arenig Mountains (J.M.Gray personal communication). S.Lowe (1994) has however produced a PhD thesis dealing with the Lateglacial and Flandrian stratigraphy of sites in southern Snowdonia. In his study 7 glaciers were mapped in the Cadair Idris, Moelwyn, Aran and Arenig mountain areas. Of the latter two areas only two local glaciers were mapped, at Llyn Lliwbran in the Arans and Llyn Arenig Fach in the Arenigs. The following study not only independently maps, reviews and reconstructs these sites using newly devised analytical techniques, it also provides evidence for at least a further two local glacier sites and another which is open to debate. This, as well as pollen-stratigraphic analysis at two new sites, will hopefully further our knowledge and understanding of the Loch Lomond Stadial in Wales.

2. GLACIAL GEOMORPHOLOGY AND GLACIER DELIMITATION

2. 1. Methods of investigation

2.1.1. Mapping

The maximal limits of 4 former local glaciers were mapped in the field onto Ordnance Survey maps at a scale of 1:10,000. These were at Llyn Lliwbran (grid ref: SH 875255) in the Aran Mountains and at Llyn Arenig Fawr (grid ref: SH 845380), Llyn Arenig Fach (grid ref: SH 827417) and Cwm Gylchedd (grid ref: SH 866455) in the Arenig Mountains. Field mapping was aided using evidence derived from 1:15,000 scale aerial photographs which were consulted beforehand in the Cambridge University Aerial Photographic Unit. The field evidence was mapped onto acetate overlays allowing accurate re-drawing for presentation and reconstruction purposes. Features mapped in the field included end moraines, hummocky moraine, drift and boulder limits, fluted moraine, periglacial features and periglacial trimlines. The use of such features in delimiting former Loch Lomond Stadial glaciers is discussed in the following text, Section 2.1.2.

2.1.2. Glacier delimitation

i) *End and lateral moraines*. These are often particularly valuable in deducing the maximal extent of former glaciers. In northern Snowdonia, Gray (1982) found that only 5 of 35 former local glaciers failed to produce an end moraine. Similarly, in the English Lake District Sissons (1980a) found that only 10 of 64 former local glaciers failed to produce an end moraine along part of their limit whereas in the Gaick area of the Scottish Highlands Sissons (1974) found that end moraines were very poorly developed. It is important to recognise the most distal moraine ridge as many localities display numerous recessional moraines such as at Lough Nahanagan in Ireland (Colqhoun and Synge 1980) (See Figure 2.1b).

ii) *Hummocky moraine*. This is the term given to numerous mounds of drift within glacier limits. Boulders frequently rest on the mounds which are often composed of till containing many large clasts or sometimes entirely of boulders. In most cases hummocky moraine lies immediately behind end moraines and thus where end moraines are absent,

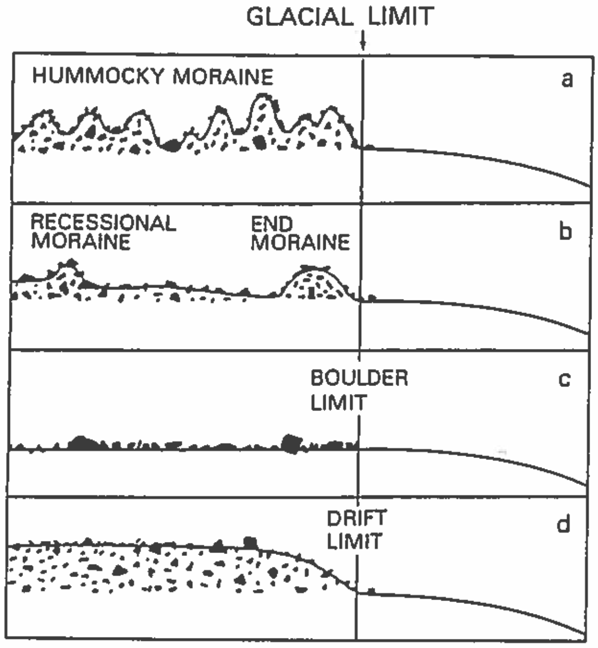


Figure 2.1. Some types of geomorphological evidence used to delimit the extent of Loch Lomond Stadial glaciers; a) limit of hummocky moraine, b) end moraine, c) boulder limit, d) drift limit. (From from Gray and Coxon 1991, Fig.37)

the down valley limit of hummocky moraine is often indicative of the approximate limit of the former glacier (See Figure 2.1a).

'Hummocky moraine' can only be regarded as a morphological term as there is debate about its genesis at different sites. Most of the literature in the 1970s and early 1980s argued that many Loch Lomond Stadial glaciers decayed by mass or 'areal' stagnation and that Scottish 'hummocky moraine' was the chaotic product of this stagnation (Sissons 1967, Sugden 1970, Gray 1982). Through the 1980s and 1990s there became unease with this interpretation and workers such as Eyles (1983), Bennett and Glasser (1991), Bennett and Boulton (1993) and Bennett (1994) have argued that hummocky moraine topography can be formed via active glacier retreat. Bennett and Glasser (1991) for example, re-interpreted hummocky moraine mapped by Sissons (1979a) as closely spaced recessional moraines. It is also possible that in some cases hummocky moraine is the product of the decay of detached ice-blocks from an actively retreating glacier (Benn and Evans 1998). It is therefore likely that so-called 'hummocky' moraines encompass a number of different sediment-landform associations, some of which formed at active ice margins while others record ice stagnation.

Only via detailed morphological and sedimentological analysis can the genesis of 'hummocky moraine' be ascertained. Moraine formed via ice melt-out in stagnation would be characterized by diamicton lenses interbedded with water sorted sands and gravels. These moraines are also likely to appear morphologically chaotic (Benn 1991) although this can only be deduced from aerial photographs or from a well elevated position in the field. Moraines formed at the glacier margin during active retreat are likely to consist of 3 broad types of sediment: (1) sediments showing shear-folding; (2) heterogeneous sands, bouldery gravels and diamictons; or (3) homogenous diamictons (Bennett and Boulton 1993). The morphological arrangement of the moraine is likely to consist of ridges and hummocks arranged in closely-spaced chains oriented obliquely downvalley. Individual ridges commonly have a beaded long profile, which enhances the hummocky appearance (Benn 1991) and as such must be differentiated from chaotic stagnant ice forms by consultation of aerial photographs as well as via sedimentological analysis.

It is evident that if the dynamics of deglaciation during the Stadial are to be understood, careful considerations of landform genesis are needed. Whilst acknowledging this point, this study is primarily concerned with reconstructing the maximal extent of the former local glaciers. As such the genesis of 'hummocky moraine' is not a key issue. Whatever the genetic origin of hummocky moraine, its presence will usually indicate a position within the maximal limits of a former local glacier. It is possible however that pre-Stadial hummocky moraine will lie outside of the Stadial limits although should be easily differentiated based on differences in preservation (freshness) and its position within the whole assemblage of features.

iii) *Boulder limits*. These mark the transition between areas of glacially deposited boulders within glacier limits and areas boulder-free or with very different boulder lithologies outside of these limits (See Figure 2.1c). Sissons (1980a) noted that glacially deposited boulders are often very abundant in ground formerly occupied by localised cirque glaciers. Some of the best published descriptions are from northwest Scotland where Sissons (1977b) reported a boulder limit of white Cambrian Quartzite contrasting with red Torridonian Sandstone bedrock outside of the former glacier limit.

iv) *Drift limits*. These are similar to iii) but correspond to a sharp change in drift thickness at the former glacier margins (see Figure 2.1d). In this study drift limits have been mapped in areas where drift margins are obvious and in delimiting hummocky moraine along a former glacier margin.

v) *Fluted moraine*. Flutes are typically low (<3m), narrow (<3m), regularly spaced ridges which are usually less than 100m long and are aligned parallel to the direction of ice flow. As such they can can be used to indicate the flow direction of former local glaciers. They usually begin up-valley at a boulder or bedrock obstacle and are typically composed of lodgement till, although they may also contain fluvial sands and gravels (Bennett and Glasser 1996). Fluted moraines can be particularly vague on the ground and are often best observed using aerial photographs or alternatively from an elevated position in the field.

vi) *Periglacial trimlines*. These are sometimes useful in delimiting the upper extent of former local glaciers. During the Loch Lomond Stadial, the last major cold period in Britain, areas outside the limits of former local glaciers would have been subject to severe periglacial weathering. Within Stadial glacier limits no periglacial activity would have occurred, only glacial action. A major exception here is on cwm backwalls where frost shattering could have occurred well below the the upper glacier limit due to the presence of bergschrunds. Thus, apart from on some cwm backwalls, trimlines can act as accurate delimiters of upper lateral glacier limits (Thorp 1981).

The basis of this method assumes little periglacial activity during the whole of the Holocene. This seems a fair assumption, as short cold phases such as the Little Ice Age were not sufficiently cold to have had significant effect at the altitudes of the sites concerned. Indeed, Campbell and Bowen (1989) consider most of the periglacial features in Wales, previously covered by the Devensian ice sheet, to be relics of the Loch Lomond Stadial.

Collectively, the methods of glacier delimitation discussed here allow the limits of former local glaciers to be mapped particularly accurately. Using similar methods as in this study, Sissons (1977b) estimated that boulder limits identified the former ice margins to c.1-2 metres whilst end moraines were probably accurate to c.10 metres and overall interpolated ice margins accurate to within 50-100 metres. Where trimlines are unclear or non-existant the upper limits of former glaciers are likely to the main source of error, although this can eliminated somewhat with well judged interpolation based on cwm morphometry, the overall assemblage of features and glaciological considerations.

2.1.3 Ice-directional analysis

i) *Small-scale erosional forms*. These include friction cracks and glacial striae which can be used to interpret former ice-flow directions of the former local glaciers. The term 'friction crack' was introduced by Harris (1943), and is now used to cover a variety of erosional forms including crescentic gouges, lunate fractures and crescentic fractures. Striae are lines or scratches on a rock surface produced by glacial abrasion.

In this study, only striae were measured in any great number although friction cracks were used as a check on striae direction in the field. Striae have probably been used more in determining ice movement (Gray and Lowe 1982) although there are several problems in in their interpretation, including (a) possible changes in ice movement over one glacial phase; (b) influence of more than one phase of glacier activity; (c) influence of local factors such as relief and rock structure and (d) possible occurrence of striae produced by agencies other than ice (Flint 1971, pp. 90-93). However, a general picture of former ice movement within the limits of former local glaciers can be achieved provided enough samples are measured. In this study, at each glacier site 200 striae were measured. However, no striae measurements were taken from Cwm Gylchedd due to the lack of suitable bedrock exposures.

ii) *Till fabric analysis*. The orientation and dip of pebbles in Quaternary glacial sediments have been used by many workers to indicate ice-flow directions of past glaciers (e.g. Holmes 1941, Andrews and Smith 1970, Boulton 1971, Rose 1974). Elongated pebbles have a tendency to lodge in till with their long-axis parallel to the direction of flow and thus the preferred orientation of the long-axis of pebbles sampled from till will give the direction of movement of the ice which deposited the till (Harrison 1957, West and Donner 1956). Also, elongated pebbles in basal tills are usually inclined in the direction of approach (Hirvas and Nenonen 1990 pp. 228).

In this study only direction and not dip were measured because up and down ice directions are easily deduced given the small scale of the glaciers under consideration. Also, dip is not a conclusive indicator of flow direction. Saarnisto and Peltoniemi (1984) found that on the basis of over a hundred till fabric analyses in eastern Finland about 60% of the stones in the till were inclined in the direction of approach and 40% were inclined in the down-glacier direction or not inclined at all.

Till fabric analysis was performed at a site which could be assumed to represent regional flow conditions i.e. a flat area of basal till, well away from obstacles which may have locally altered flow. Also, sites well away from the ice margins were preferred because of the directional distortion which occurs at the margins, particularly near lateral and end moraines. At least 50 clasts were measured at each measuring station. In the field, till fabric was analysed following the methods described in Hirvas and Nehonen (1990, pp.226 - 230) and the directional data subsequently plotted onto mirror image rose diagrams.

2.2. Results

2.2.1. Llyn Lliwbran (SH 875255)

(Refer to the geomorphological map at 1:10,000 scale given in Figure 2.2.)

The maximal extent of the glacier occuping the basin containing Llyn Lliwbran in the Aran Mountains is marked by an end moraine. This, well developed, clear end moraine bounds the easternmost part of the lake (see Plate 1.). The moraine is 100 metres long and 4 metres high. The northern limit of the former glacier is marked by a well defined lateral moraine. The actual glacier limit is mapped slightly north of this limit in parts as it is likely that the moraine has moved downslope from its original position upon glacier retreat. This is inferred because of the steepness of the slope and the shape of the moraine with relation to cwm morphometry. Also, a less distinct line of smaller boulders, marked as a boulder limit in Figure 2.2, exists c.30m beyond the distal slope of the moraine. This lateral moraine extends some 400 metres upslope. The southern limits are unclear on the ground although can be confidently interpolated given the tight closure of the cwm morphometry. With little evidence for ice extent up the headwall, the common practice of assuming glacier extension to c.30m from the headwall summit has been followed (Gray 1982). This is somewhat arbitary although can be given confidence if related realistically to the overall glacier shape. Although a clearly defined trimline is not evident on the cwm headwall, there is a marked difference between the very smooth striated bedrock on the south-western shores of the lake and the frost shattered rock on the ridge above to the west.

The striae data is displayed in Table 2.1.

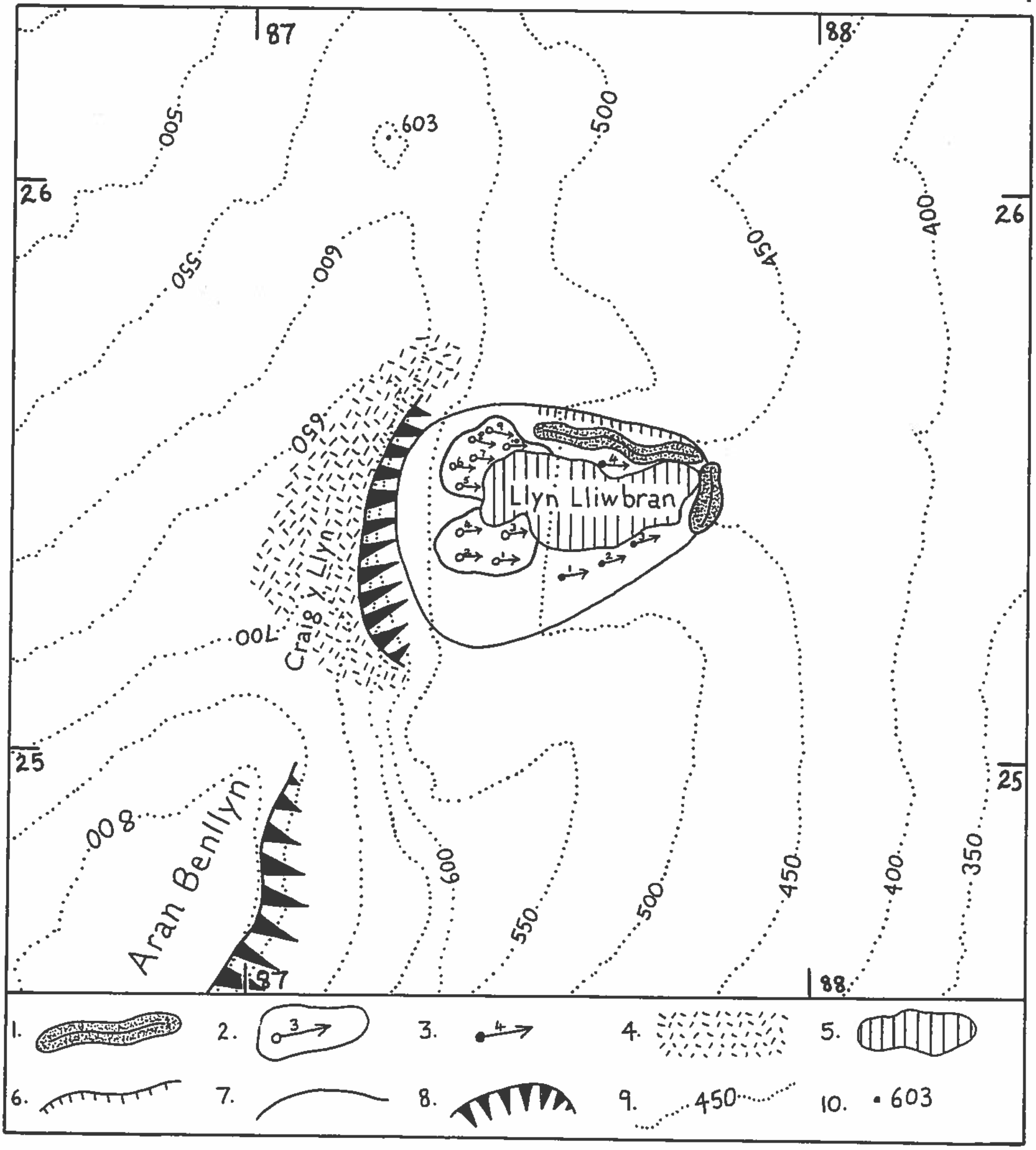


Figure 2.2. Geomorphological map of the Llyn Lliwbran glacier. Key: *1. Moraine with lines showing ridge crests; 2. Ice-smoothed bedrock showing striae stations (showing general direction); 3. Till fabric analysis stations (showing general direction); 4. Periglacially disturbed ground; 5. Lake; 6. Boulder limit; 7. Interpolated glacier limit; 8. Steep cliffs; 9. Contours at 50 metre intervals (incorporating former ice surfaces where glaciers were present); 10. Spot height in metres.*

Plate 1. Llyn Lliwbran viewed from the north-west. Note the clear end moraine in the centre of this picture.

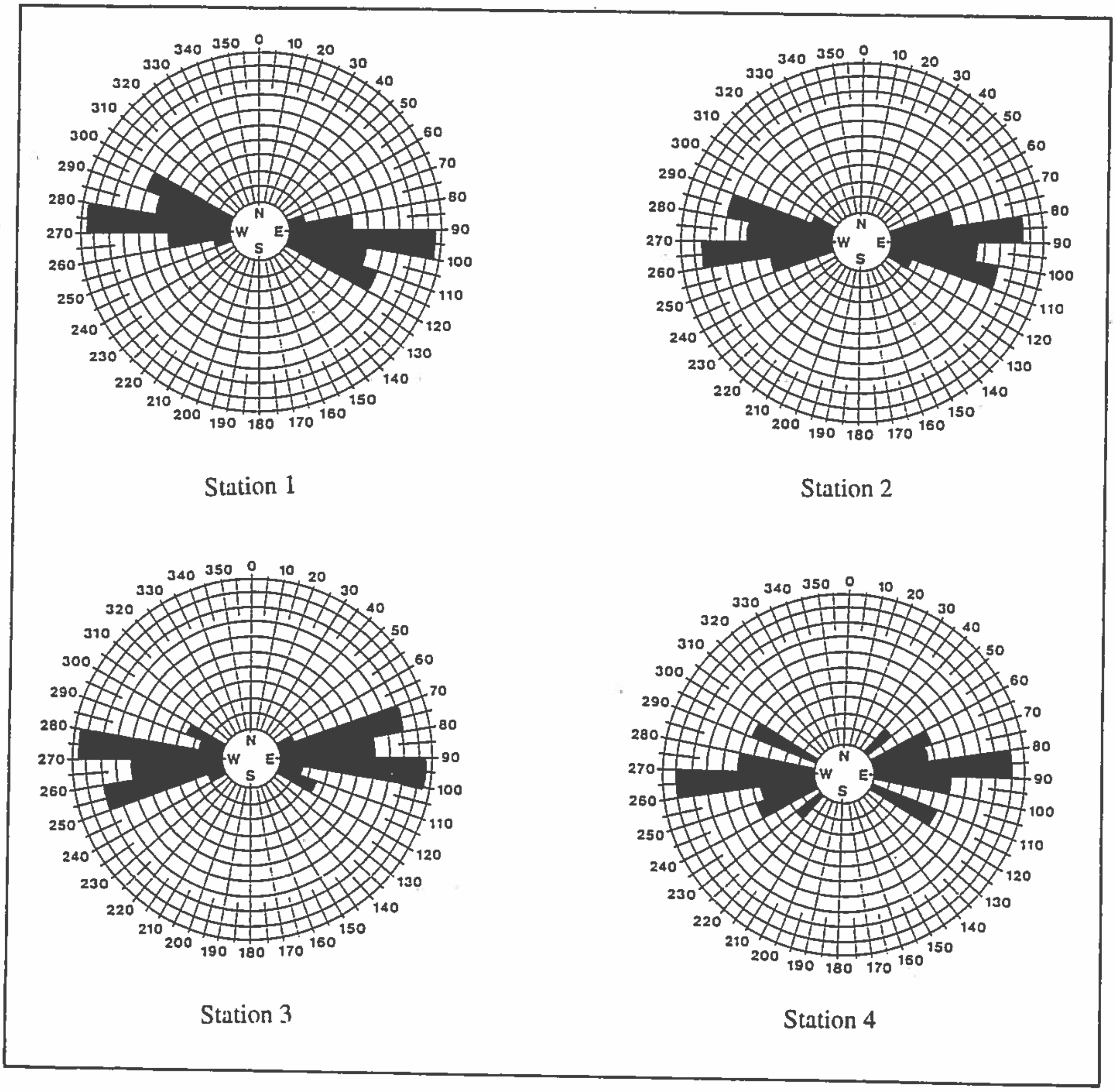


Figure 2.3. Till fabric analyses from sections at the Llyn Lliwbran site.

Table 2.1. Striae measurements at the Llyn Lliwbran site.

Station	Range of values	Mean	Number of measurements
S1	87-93°	90.6°	20
S2	86-92°	88.5°	20
S3	89-95°	93.4°	20
S4	88-94°	92.3°	20
S5	86-91°	88.5°	20
S6	86-90°	87.9°	20
S7	84-90°	86.3°	20
S8	85-89°	87.7°	20
S9	89-97°	94°	20
S10	88-92°	90°	20

The striae measurements show values in the range 84° - 97° from ten sample stations with an overall mean of 89.9°. The results of till fabric analyses from 4 sections are shown in Figure 2.3. The till fabric data supports the striae data with ice-flow direction being along a west-east plane. It is clear from the field evidence that ice-flow was from west to east. The locations of the sample points for both striae and till fabric analyses are shown in Figure 2.2.

2.2.2. Llyn Arenig Fawr (SH 845380)
(Refer to the geomorphological map at 1:10,000 scale given in Figure 2.4.)

The lake of Arenig Fawr is retained by a broad ridge which arcs from 0°N to 120°SE. There has been some debate in previous literature as to whether this ridge is an end moraine or largely solid rock (Rowlands 1970). There has even been suggestion that the lake is dammed by a lateral moraine of a valley glacier below the cwm (Fearnsides 1905). A similar suggestion was made by Fearnsides (1905) in explanation of the Llyn Arenig Fach moraines. It is clear in the field that the lake is situated in an over-deepened rock basin and not dammed entirely by an end moraine. There are several solid rock outcrops along much of the length of the retaining ridge such as at grid reference SH 848385 and SH 851380. Also, local planning documents record that the small dam with which the

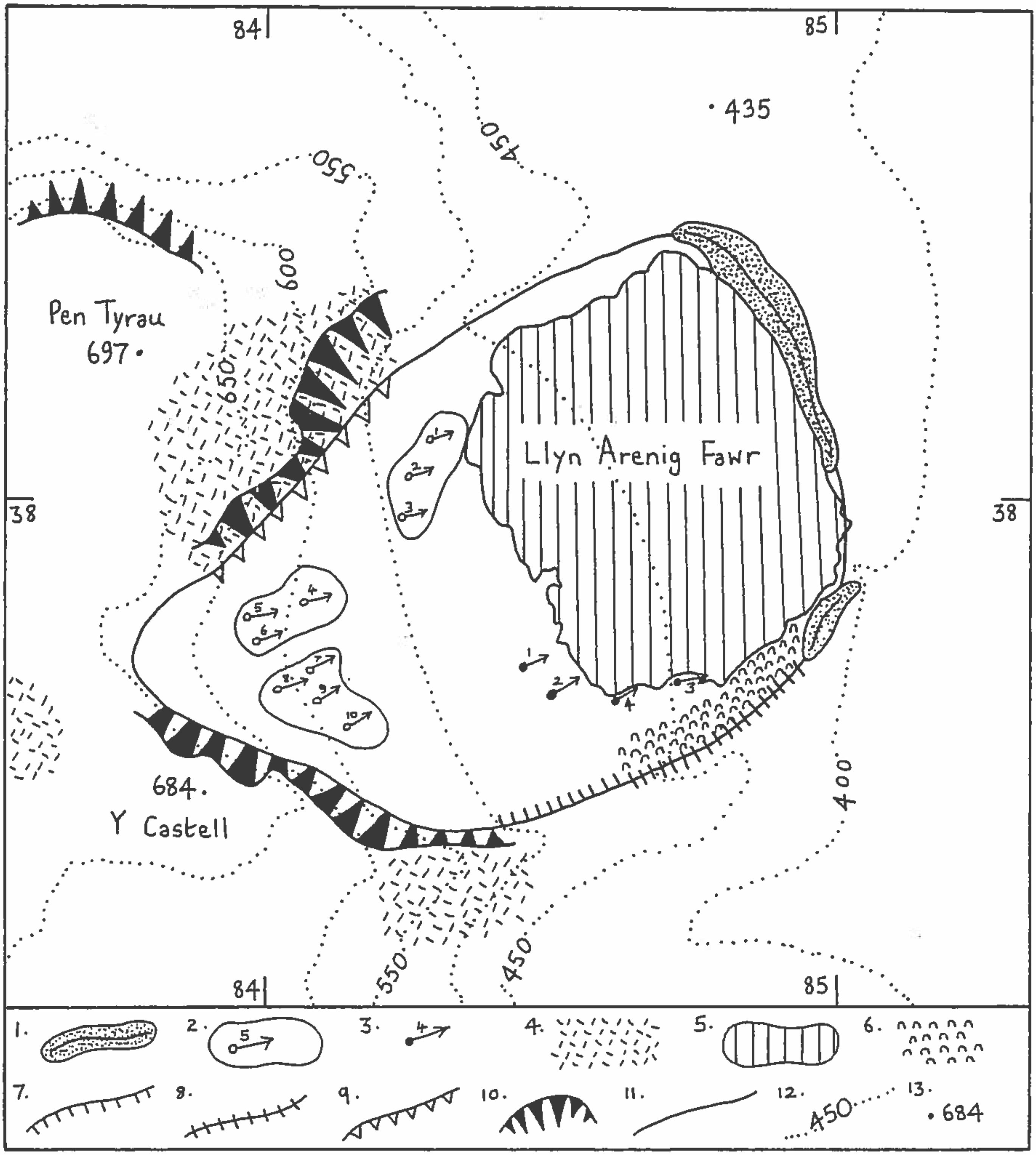


Figure 2.4. Geomorphological map of the Llyn Arenig Fawr glacier. Key: *1. Moraine with lines showing ridge crests; 2. Ice-smoothed bedrock showing striae stations (showing general direction); 3. Till fabric analysis stations (showing general direction); 4. Periglacially disturbed ground; 5. Lake; 6. Hummocky moraine; 7. Boulder limit; 8. Drift limit; 9. Periglacial trimline; 10. Steep cliffs; 11. Interpolated glacier limit; 12. Contours at 50 metre intervals (incorporating former ice surfaces where glaciers were present); 13. Spot height in metres.*

lake was artificially raised in the 19th century was founded on solid rock (Rowlands 1970). Despite this fact, there is clear evidence of localised moraines along the north-eastern shores of the lake as well as to the south of the small dam. Whilst Rowlands (1970) discounts the possibility that the ridge is a lateral moraine and maintains that the ridge is predominantly a solid feature he does not discount the possibility of some locally derived moraines along part of its length. Indeed in a later paper Rowlands (1979, pg. 21) notes "a series of purely locally-derived block moraines within the cirque of Arenig Fawr". The investigations of this study confirm this observation. The moraines are not as clear as at other sites in this study as they are broader and less pronounced. At the north end of the lake the moraines rise 2-3 metres above the surrounding ground on both sides. No differentiation could be made of individual morainal crests even though the moraine was 80 metres wide in parts, and thus the moraine was mapped as a singular moraine. The moraines to the south of the dam are more pronounced, similar to those at Llyn Lliwbran, and c.4 metres in height. It is likely that the moraines continued over the present position of the dam and reservoir workings but were excavated to the lake level and to bedrock when construction took place. It would also be interesting to know what features are submerged beneath the lake raised from its natural level by 14 feet (4.3 metres). The moraines which are evident are likely to mark the maximal extent of the former local glacier as till and morainal features are absent beyond these features.

Hummocky moraine is evident immediately behind the southern end moraine mentioned above. The moraines are 1-2 metres in height, fresh in appearance and appear entirely composed of boulders. The moraines extend up-valley for c.400 metres after which the mounds become less distinct. The overall extent of hummocky moraine is concealed by the artificially raised waters of Llyn Arenig Fawr. The distinct distal limit of hummocky moraine represents a clear lateral limit, which extends westwards from the southern end moraine.

A boulder limit marks the southern limit of the the former glacier. Where hummocky moraine fades out there is a distinct boundary between dense glacially deposited boulders and smooth peat-covered slopes ouside the glacier limit.

The north-western limit of the former glacier is marked by a periglacial trimline. Beginning from the southwest, the trimline is defined by a transition between glacially moulded bedrock with little sign of periglacial action to a periglacially weathered upper section. The upper rock walls show clear signs of frost-shattering with the rock displaying deep jointing. The whole plateau above the trimline displays evidence of widespread periglacial activity, particularly frost-weathered bedrock and deeply jointed rock pavements. Further east, the trimline is less distinct, although evidence still remains. Here, the upper glacier limit is mapped as the approximate line of transition from glacial drift cover to frost-shattered scree. In contrast to the frost-shattered rocks outsides of the former glacier limits, the rocks of the upper cwm basin to the south-west of the lake are glacially moulded showing little sign of subsequent alteration by frost action. Several large scale roche moutonee occur in this area, displaying very smooth stoss side surfaces with abundant striae.

The striae results are shown in Table 2.2.

Table 2.2. Striae measurements from the Llyn Arenig Fawr site

Station	Range of values	Mean	Number of measurements
S1	57-68°	63.4°	20
S2	60-65°	63.2°	20
S3	64-70°	69.4°	20
S4	68-76°	72.8°	20
S5	67-78°	73.4°	20
S6	66-74°	72°	20
S7	65-73°	71.3°	20
S8	69-74°	72°	20
S9	75-81°	77°	20
S10	78-85°	80°	20

The striae measurements show values in the range 57°- 85° from ten sample stations with an overall mean of 71.5°. The greater range of directional values compared with the previous glacier Llyn Lliwbran reflects the far greater size of the Llyn Arenig Fawr glacier and subsequent greater variation in flow direction across the cwm. The results of till fabric analyses from 4 sections are shown in Figure 2.5. Two of these were taken at cliff exposures at the lake edges. The rose diagrams clearly illustrate an east to north-easterly ice movement (50 - 120°). The locations of the sample points for both striae and till fabric analyses are shown in Figure 2.4.

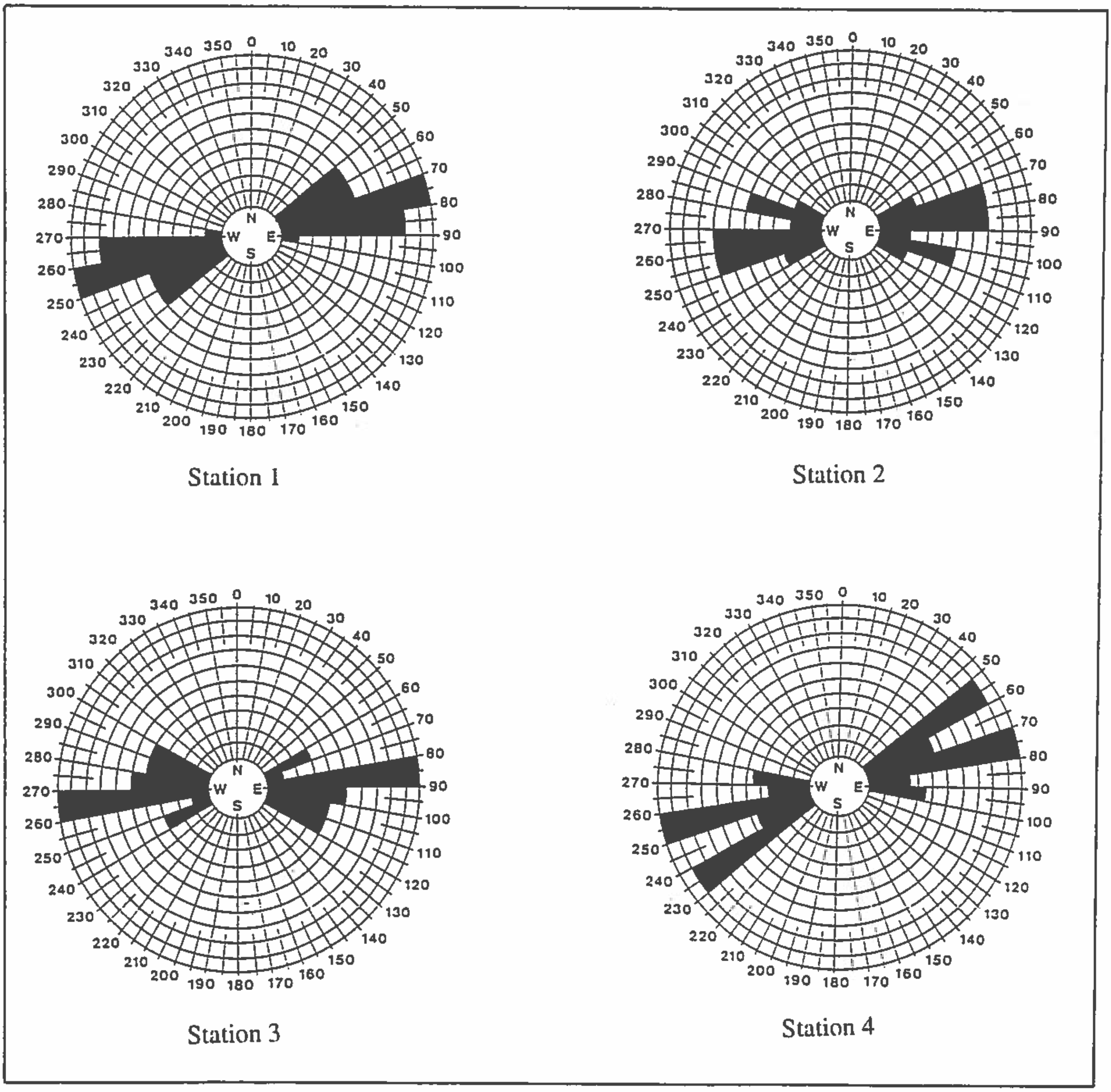


Figure 2.5. Till fabric analyses from sections at the Llyn Arenig Fawr site.

2.2.3. Llyn Arenig Fach (SH 827417)

(Refer to the geomorphological map at 1:10,000 scale given in Figure 2.6. An aerial photograph of this site is given in Plate 2.)

This site displays very impressive readvance features, in particular long the eastern shores of Llyn Arenig Fach. Here, a series of very clear morainal ridges bound the lake for around 500 metres from north to south (see Plate 3.). There are four major ridge crests covering a width of upto 100 metres and reaching greater than 10 metres in height. Although the moraine complex is very prominent there is strong field evidence suggesting that the former local glacier, at its maximal extent, extended beyond the outermost moraine of the complex. The moraines therefore represent a series of closely spaced recessional moraines.

A boulder limit, in parts nearly 100 metres beyond the end moraine complex, would appear to mark the maximal extent of the former glacier. The bouldery drift is relatively flat and appears morphologically separate to the distinctive end/recessional moraine complex, although may be just a continuation of the end moraine form. In Figure 2.6 it has been marked as a boulder limit in preference to a continuation outwards of the end moraine complex purely on morphological grounds. What is clear in the field is that the former glacier did extend further than the position of the outermost crest in the moraine complex, thus suggesting that these crests were formed after the glacier had reached its maximal extent. Once having reached the maximal extent marked by the clear boulder limit, the glacier probably retreated c.100 metres to the position of the end/recessional moraine complex where the glacier terminus must have existed for some time to create such large morainal crests. There is a well defined contrast between the boulder drift inside the limits of the former glacier and the smooth, boulder-free, slopes outside of these limits. This well defined boulder limit extends over 1 km from the south-east of the lake to the far north-west. To the south of the lake a boulder limt also marks the former glacier limit for around 250 metres.

Hummocky moraine extends up-valley behind the southern end of the moraine complex. The extent of the moraine marks a lateral drift limit extending over 250 metres before changing to a boulder limit, which bounds flatter bouldery drift. Hummocky moraine is also evident to the north-west of Llyn Arenig Fach. Here, the moraines are particularly pronounced, 1-2 metres in height and covered in large boulders. The

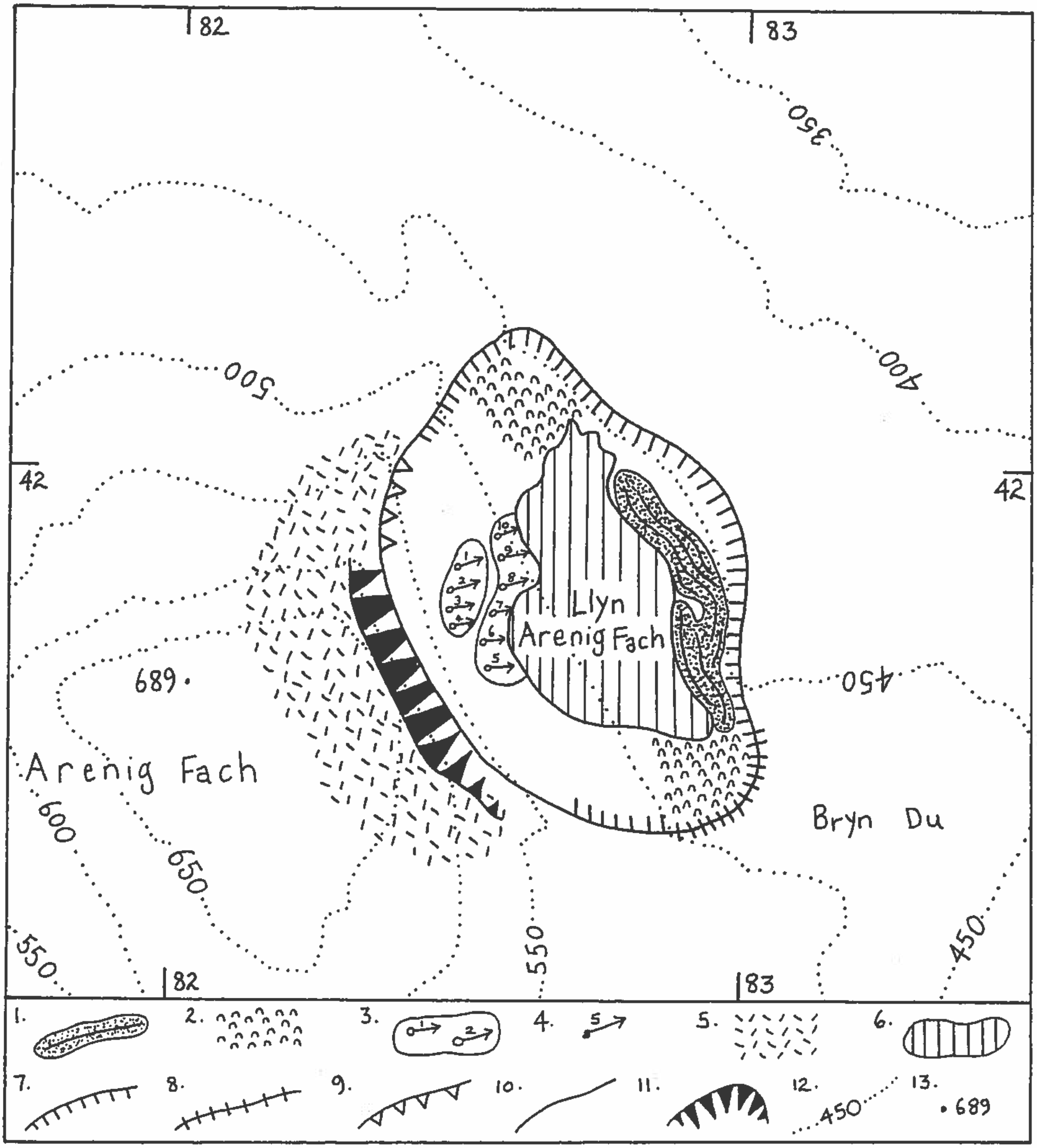


Figure 2.6. Geomorphological map of the Llyn Arenig Fach glacier. *Key: 1. Moraine with lines showing ridge crests; 2. Hummocky moraine; 3. Ice-smoothed bedrock showing striae stations (showing general direction); 4. Till fabric analysis stations (showing general direction); 5. Periglacially disturbed ground; 6. Lake; 7. Boulder limit; 8. Drift limit; 9. Periglacial trimline; 10. Interpolated glacier limit; 11. Steep cliffs; 12. Contours at 50 metre intervals (incorporating former ice surfaces where glaciers were present); 13. Spot height in metres.*

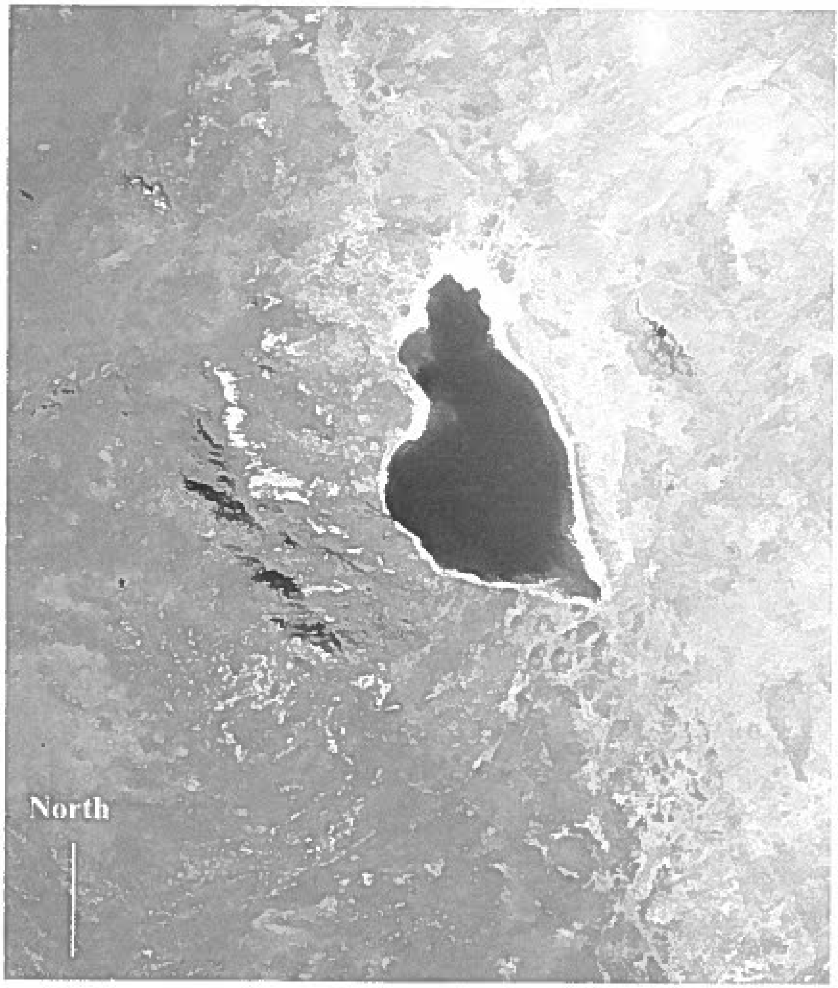


Plate 2. Aerial photograph of Llyn Arenig Fach. Note the backwall and the moraine complex on the eastern shores of the lake (Scale: approx. 1:15,000).

Plate 3. The Llyn Arenig Fach moraines.

moraines fade out towards the glacier limit where flatter bouldery drift dominates. The sedimentology is exposed in several places, consisting of diamicton lenses interbedded with water sorted sands and gravels, typical of ice melt-out in stagnation. Aerial photographs do not appear to show any order within these moraines supporting this deduction. The moraines also consist of intermittent large bouldery clasts which are a particularly prominent feature of the moraine surface morphology.

The upper limit of the former glacier cannot be easily deduced. The cwm backwall does not display a clear periglacial trimline, as is to be expected on backwalls. There is widespread periglacial evidence, largely in the form of frost shattered bedrock, on the ground surrounding the backwall and beyond the lateral limits of the former glacier. In contrast, the base of the cwm to the west of the lake is free of such features and is ice-moulded, displaying well preserved striae and friction cracks. The only clear trimline between glacial and periglacial features is to be found north of the backwall cliffs and is mapped as a short trimline in Figure 2.6.

The striae results, taken from near the western shores of the lake, are shown in Table 2.3.

Table 2.3. Striae measurements from the Llyn Arenig Fach site.

Station	Range of values	Mean	Number of measurements
S1	48-56°	53.2°	20
S2	49-55°	53.7°	20
S3	52-60°	56°	20
S4	54-66°	59.4°	20
S5	59-71°	66.5°	20
S6	58-68°	64.3°	20
S7	51-58°	55°	20
S8	50-60°	56.1°	20
S9	51-58°	55.7°	20
S10	47-54°	51.9°	20

The striae measurements show values in the range 47° - 66° from ten sample stations with an overall mean of 57.2°. The measurement stations are shown in Figure 2.6. Till fabric analysis was not carried out at this site because of the lack of till sections representative of ice-flow direction. The large areas of hummocky moraine were unsuitable because the sedimentology suggests that the till would have been reworked during melt-out.

2.2.4. Cwm Gylchedd (SH 866455)

(Refer to the geomorphological map at 1:10,000 scale given in Figure 2.7.)

The maximal extent of the former Cwm Gylchedd glacier is marked by a well defined drift limit. The well defined drift limit arcs from c.320°NW to 82°E and is dissected at its approximate mid point by a clear meltwater channel. The drift limits continue upslope revealing the lateral limits for some distance, more particularly on the north-eastern side of the cwm. The distal slope is c.20 metres high at its maximum and slopes steeply at around 30°-40°. The drift limit suggests local glacier occupation and the deposition of locally eroded material on the floor of the cwm, and not beyond the drift limit. Inside of the drift limits there is evidence of an infilled lake basin, indicated at the surface by a level area of boggy ground, slightly lower in height than the lip of the drift limit. The stratigraphy of this basin is investigated in Chapter 3.

Within the drift limits and interpolated glacier limit, hummocky moraine is evident. The moraines extend 300 - 400 metres back from the drift limit. The moraines die out quickly upslope and enclose the small infilled lake basin towards the drift limits. The sedimentologies of several exposed sections consist of a matrix-supported diamicton interbedded in parts with water-sorted sands and gravels with a notable lack of bouldery clasts. The sedimentology of the moraine is similar to those found at Llyn Arenig Fach in that it suggests formation by ice melt-out. However, the lack of bouldery clasts is in marked contrast to the clast dominated bouldery moraines at both Llyn Arenig Fawr and Llyn Arenig Fach. This difference can be attributed to the differing geologies in the erosional zones, in Cwm Gylchedd grits and sandstones whilst in the cwms of Llyn Arenig Fawr and Llyn Arenig Fach, Ordovician volcanics of a more resistant nature. This geological difference could account for the absence of boulder drift in Cwm Gylchedd compared with the dense boulder fields found at the other sites in this study.

Linear ridges are evident on the steeper slopes up-valley of the hummocky moraine. These are 0.5 - 2 metres in height and trend in a north to north-easterly direction

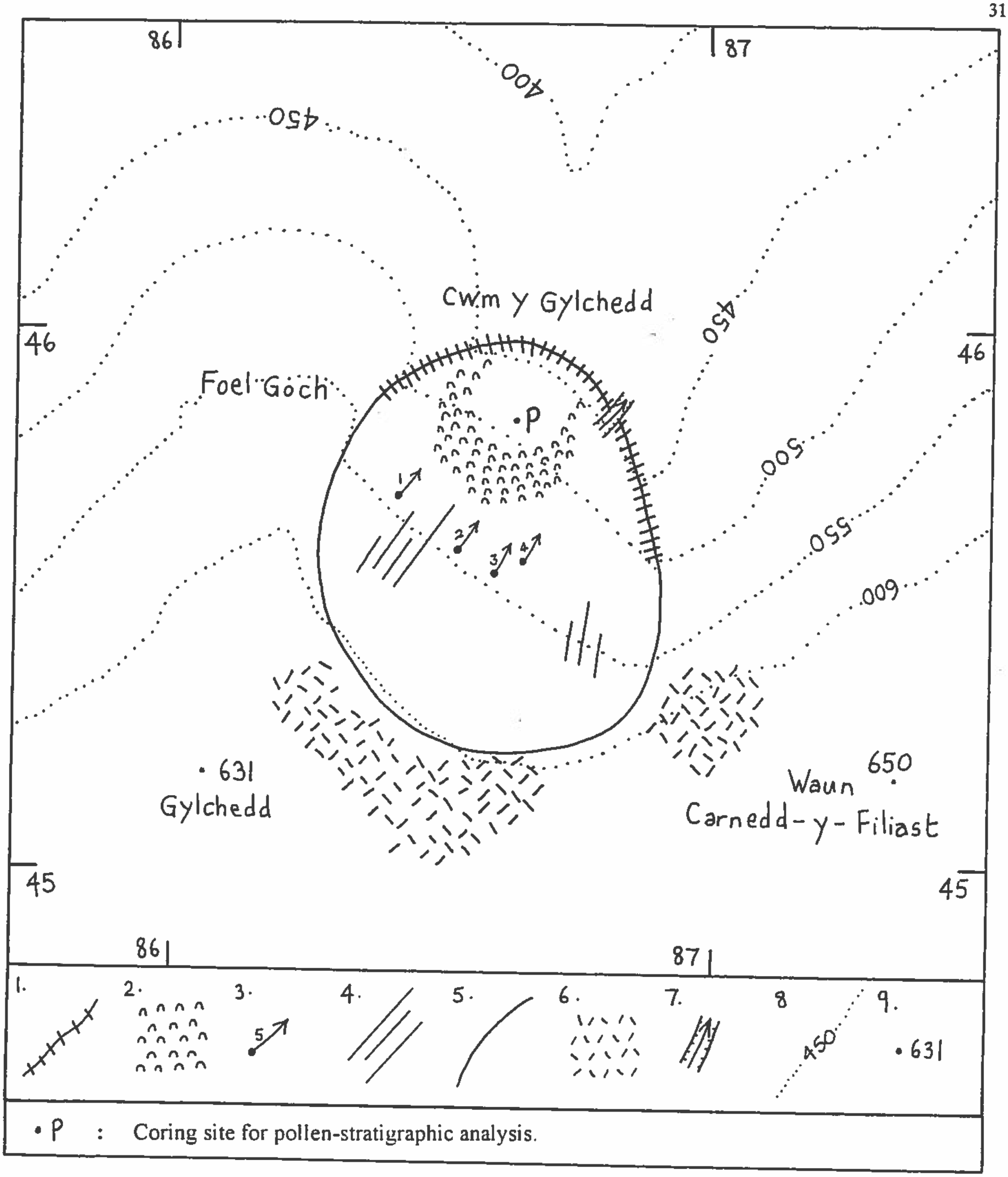


Figure 2.7. Geomorphological map of the Cwm Gylchedd glacier. Key: *1. Drift limit; 2. Hummocky moraine; 3. Till fabric analysis stations (showing general direction); 4. Fluted moraine; 5. Interpolated glacier limit; 6. Periglacially disturbed ground; 7. Meltwater channel; 8. Contours at 50 metre intervals (incorporating former ice surfaces where glaciers were present); 9. Spot height in metres.*

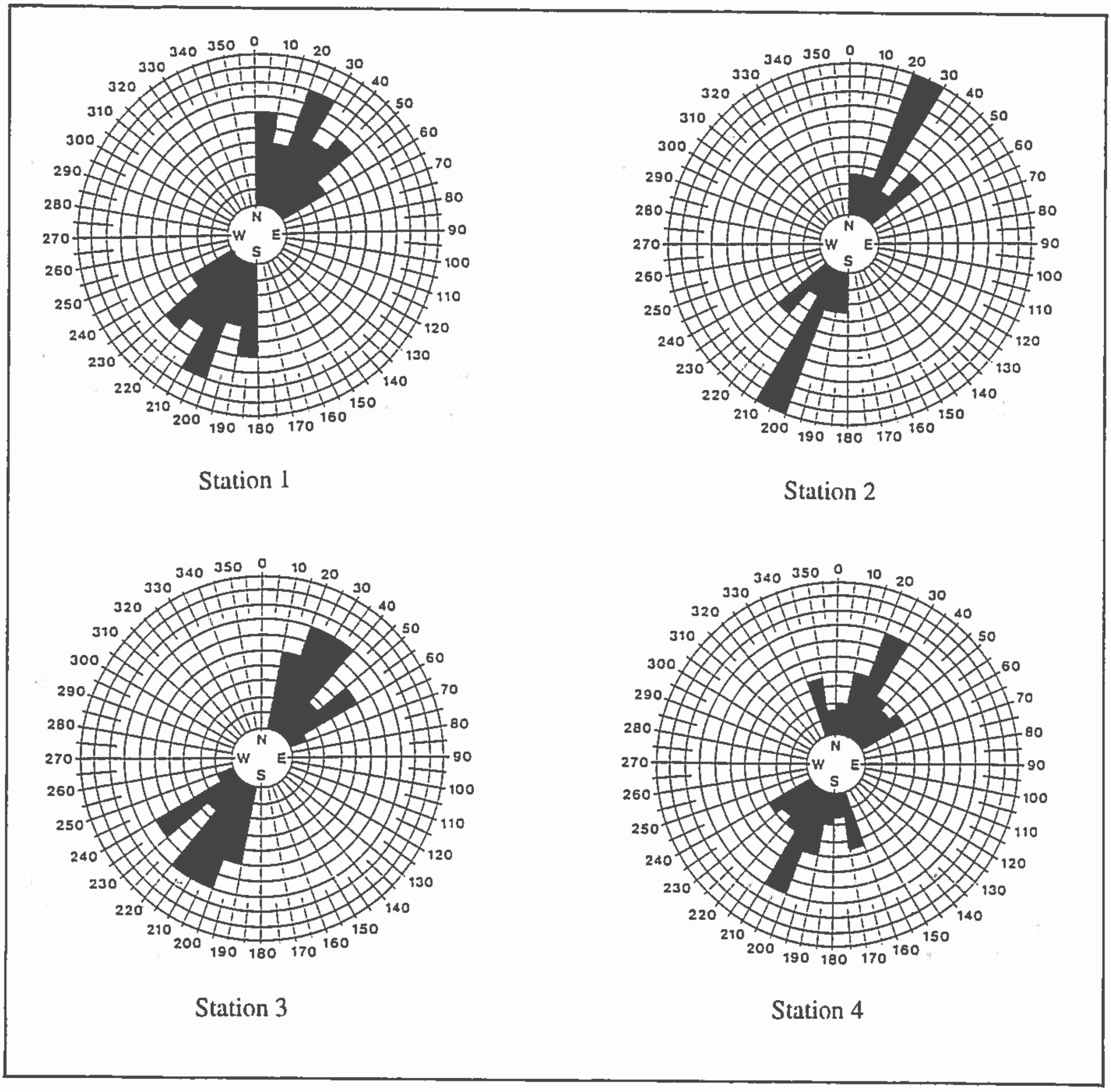


Figure 2.8. Till fabric analyses from sections at the Cwm Gylchedd site.

for upto 300 metres. At close inspection, although evident, the flutes are vague. However, when viewed from above the lineations appear much clearer.

Gelifluction sheets and gelifluction lobes are in evidence on the surrounding slopes outside of the mapped glacier limits. These are evident morphologically and sedimentologically in the form of highly disturbed, angular, layered sediment. Isolated rock outcrops also display evidence of intense frost shattering around the rim of Cwm Gylchedd. Within the mapped glacier limits periglacial features are absent. The nature of the periglacial evidence and the lack of an obvious trimline result in the glacier limits being interpolated onto the map in Figure 2.7. The interpolated limit, being shaped to the cwm morphometry and c.30 metres below the lip of the headwall (cf. Gray 1982) as well as separating glacial and periglacial features, gives a reasonably accurate estimate of the upper glacier limit.

No striae analysis was undertaken at this site due to the lack of suitable bedrock exposures within the mapped glacier limits. Till fabric analysis was done at four sites. The results, illustrated in Figure 2.8, show that ice flow was in the north-east quadrant of the compass. Although the till fabric is illustrated as a mirror image rose diagram, the overall assemblage of geomorphological features firmly supports this conclusion. The locations of these measurement stations are shown in Figure 2.7.

2.2.5. Other possible sites

The clearest evidence of former localised glaciation in the Aran and Arenig Mountains has been discussed above. There is however reason to believe that local glaciers could have existed at other sites also, an important example being at Creiglyn Dyfi in the Aran Mountains (grid reference: SH867226. See Figure 1.3. for its location). At this site the lake of Creiglyn Dyfi occupies a well-defined south-easterly facing cirque on the eastern side of Aran Fawddwy (905 m). The lake occupies the cwm floor and sits at an altitude of 580m. The lake is dammed by a rock bar at its southern shores and not by moraines. The bedrock is clearly exposed by the outflow stream which is incised deeply into rock bar. Although no moraines appear to mark a maximal limit to a former local glacier there are remarkably well-preserved erosional features in the erosional zone, possibly the best observed in the Aran and Arenig Mountains (see Plate 4.). Here, striae and friction cracks are very well preserved and there is no evidence of periglacial modification in contrast to the slopes outside of the cwm. There is also evidence of a boulder limit to the east of the

Plate 4. The 'fresh' glacially smoothed bedrock in the cwm of Creiglyn Dyfi.

lake although this is rather intermittent and ill-defined. However, the lack of a clear moraine in the cwm, particularly at the lip does not allow any suitable conclusion. As such, a former glacier was not mapped in this locality.

S.Lowe (1994) also declined to recognise a glacier at this site on the basis of the lack of moraines. However, a lack of moraines need not imply that the site was not occupied by a glacier during the Loch Lomond Stadial as these may have been eroded away since. The very impressive erosional features favour this view. Some support would be possible via pollen stratigraphic analysis of the deepest sediments accumulated in the basin but unfortunately stratigraphic analysis was not undertaken at this site due to its remoteness and the logistical problems involved. In any case, it is doubtful whether the deepest sediments could be reached because they are likely to exist beneath the lake waters. Without more substantial evidence no justifiable argument can be given, as yet, to further support a theory of localised glacial occupation at this site.

3. POLLEN-STRATIGRAPHIC ANALYSIS AND THE AGE OF THE FORMER GLACIERS

3.1. Providing an age for the former glaciers

The former glaciers mapped in Chapter 2 represent a phase of local glaciation after the last great ice-sheets had retreated. The timing of this localised glaciation is open to debate. The glaciers could have readvanced prior to the Windermere Interstadial during a brief reversal of deglaciation. For example, Benn (1996) has suggested a link between ice-sheet readvances in northwest Scotland and the massive discharge of ice into the North Atlantic during the Heinrich event H1. Alternatively, the glaciers built up from afresh and readvanced during the Loch Lomond Stadial. This problem can be partially resolved with reference to the periglacial evidence at the sites concerned. If the glaciers existed during the Loch Lomond Stadial, periglacial features would be poorly developed inside of the former glacier limits in comparison with the land outside of these limits. This is because of glacial erosion under ice rather than sub-aerial periglacial weathering at the sites of the former glaciers. Such contrasts in periglacial weathering points to a Loch Lomond Stadial age as comparatively little severe periglacial action has occurred through the Flandrian. Events such as the Little Ice Age, although significant, are unlikely to have affected the 'freshness' of Loch Lomond Stadial features.

Periglacial contrasts are evident at all of the sites in this study although this is a rather subjective line of argument. A more solid technique is that of pollen-stratigraphic analysis of infilled lake basins inside and outside of the former glacier limits. Outside of the glacier limits the full Lateglacial stratigraphy would exist. This is typically represented by a tripartite sequence of minerogenic-organic-minerogenic sediments beneath Flandrian lake muds and peats. Inside of the former glacier limits, the sedimentary sequence of an infilled basin would comprise of only Flandrian deposits although bottomed sometimes by clays deposited during the cold, vegetation-less period immediately following deglaciation from the Loch Lomond Readvance (Gray 1975). Provided the deepest sediment and full sedimentary sequence can be found within these infilled basins then an age of last glacier occupation can be inferred. Figure 3.1. provides a generalized summary of this theory and the inferred ages of the individual stratigraphic units.

Sediment stratigraphy as a proxy dating method is usually complemented by pollen-stratigraphic analysis. The Lateglacial period and corresponding pollen zones is represented in

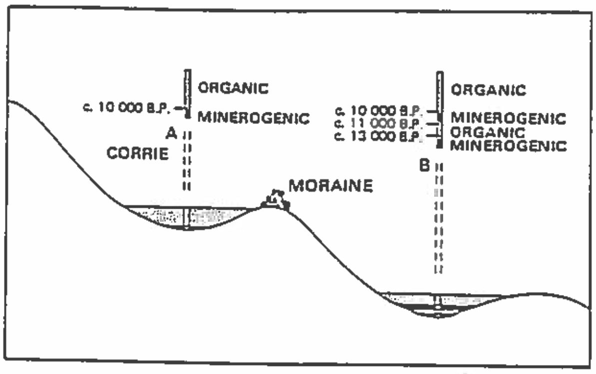


Figure 3.1. Schematic diagram to illustrate the different bog stratigraphies inside and outside of Loch Lomond Stadial limits. (From Gray and Coxon 1991)

Figure 3.2. Donner (1957) and (Sissons et al 1973) employed pollen stratigraphic analysis in establishing the extent of Loch Lomond Stadial glaciers in Scotland. In this work, the absence of sediments of pre-Zone IV age was regarded as evidence for ice occupation during Zone III (Loch Lomond Stadial). The importance of sediment and pollen-stratigraphic studies in studies of former local glaciers was highlighted by Seddon (1962) at Cwm Dwythwch in north Snowdonia. Here, a sedimentary sequence behind a set of local moraines, inside of former local glacier limits, was found to display the full Lateglacial sequence therefore implying a pre-Loch Lomond Stadial age for the former local glacier.

However, this technique of inferring an age of the last glacial occupation is not without its problems. Using the logic whereby a lack of Lateglacial sediments implies occupation by Loch Lomond Stadial ice is arguing on the basis of negative evidence. The absence of Lateglacial deposits may be due to factors other than occupation of the area by ice during the Loch Lomond Stadial (Gray 1975). These factors include (a) the persistance of 'dead-ice' through much of the Lateglacial period and; (b) the failure to locate the deepest and earliest sediments in the basin. It is implicit in previous investigations that the earliest sediments in a basin lie in the deepest part. In studies where workers have extruded cores from lake margins (e.g. Godwin 1955, Pennington 1978, MacPherson 1978), these methodologies cannot be relied upon to locate the deepest sediments. Lowe and Walker (1976, 1980, 1981) and Walker and Lowe (1977, 1979, 1980, 1985) chose to examine completely infilled basins, and developed detailed methodologies for depth-probing sites to ensure the location of the deepest part. S.Lowe (1994) employed the skills of sub-aqua divers to extrude sediment from beneath lake waters at sites in southern Snowdonia. The problem of dead-ice persisting through the Lateglacial is difficult to resolve. Although Porter and

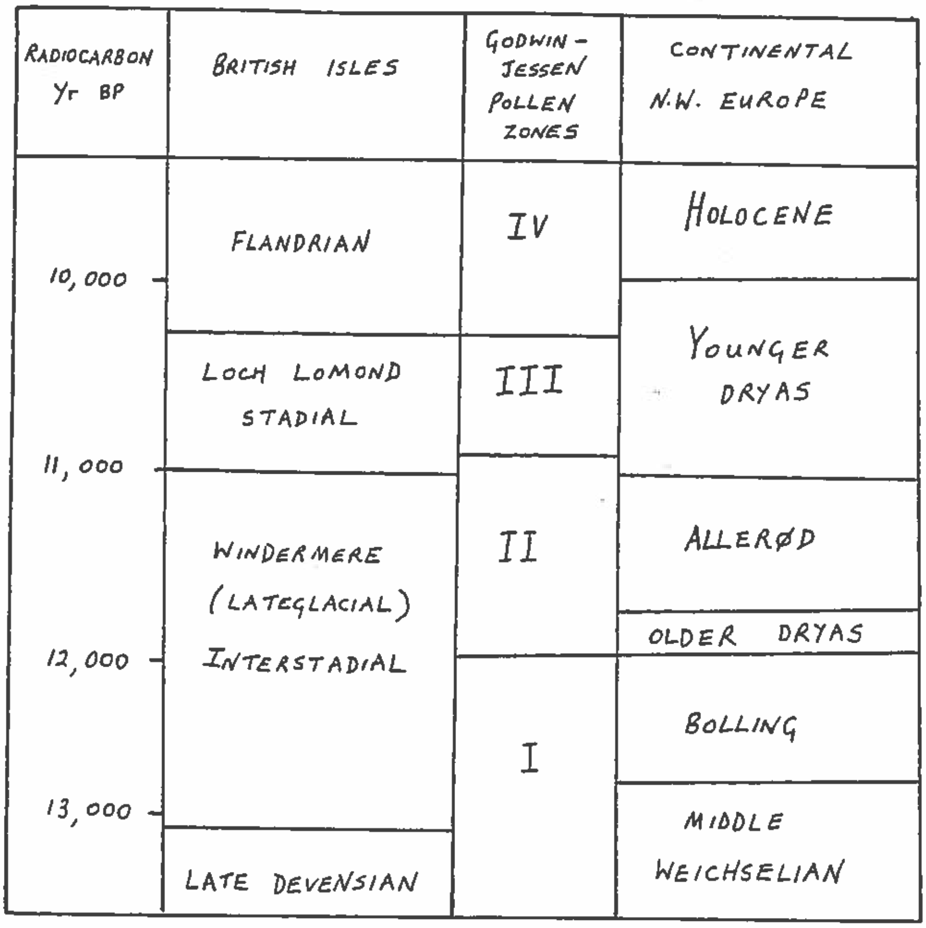


Figure 3.2. The Lateglacial period and associated pollen zones. (Adapted from Walker 1980)

Carson (1971) have described dead-ice features in North America and Østrem (1965) in Scandanavia, it is probable that the climates pertaining to these studies were not comparable to those experienced in the British Isles during the earliest postglacial (Tipping 1988). It likely that 'dead-ice' would have melted in Britain by the close of the Windermere Interstadial. This assumption is supported by the length of the Interstadial (c.2,000 years) and the fact that temperatures rose by 1°C per decade with an overall warming of 7°C per century (Coope and Brophy 1972) while the temperature of the coldest month may have risen by up to 20°C (Atkinson et al 1987, Walker et al 1993).

3.2. Methods

3.2.1. Field methods

Cores were extruded from infilled lake basins at Cwm Gylchedd (grid ref. SH 866457) and Ffridd-y-Fawnog (grid ref. SH 860385). The latter site lies outside of former local glacier limits whilst the Cwm Gylchedd site lies inside the limits of a former local glacier. Both sites represent infilled lake basins and are situated within six miles of each other (see Figure 3.3. for a map of the site locations). Cwm Gylchedd was used to investigate the stratigraphic age of the former glaciers because the former lake within the glacier limits is now infilled, unlike at Llyn Lliwbran, Llyn Arenig Fawr and Llyn Arenig Fach. As noted in the previous section (3.1.) there are practical problems in locating the deepest and fullest sedimentary sequence at lake-side sites. The late-glacial site at Ffridd-y-Fawnog is not immediately outside the limits of the Cwm Gylchedd glacier because of a lack of suitable sites in the area. Even so, the site is still sufficiently near for stratigraphic comparison and is no different from other studies such as that of Walker (1980), in the Brecon Beacons, whose Lateglacial site at Traeth Mawr lay over 3 miles away from the glacial sites.

At Cwm Gylchedd, a flat and enclosed area of boggy ground was systematically traversed with the sediment being sampled with a gouge auger. Once limnic sediments were discovered beneath blanket peat, the extent of the former lake basin was measured by means of a series of test bores with a small bore gouge auger. These bores were taken along a rectilinear grid set out over the bog surface using a level mounted at the side of the site to maintain a common datum for individual test bores (cf. Walker and Lowe 1977). Borings were taken every 5 metres through a grid measuring 30 metres by 30 metres. Once the deepest sediment was located, the sediment was sampled using a Livingstone piston corer (cf. Wright 1967). Core segments, each 1 metre long, were wrapped in cling film and then in foil and placed in plastic guttering to prevent disturbance. The upper 1 metre of sediments were sampled using a large bore (5cm) gouge auger. The Livingstone corer reached over 3.5 metres before hitting bedrock.

At Ffridd-y-Fawnog the above procedure was not necessary. Once a full suite of Lateglacial sediments were located, following exploratory borings with a gouge auger, the sediment was extracted using a Livingstone piston corer. A number of bores were made with the gouge auger around the Livingstone site to ensure that the tripartite sequence was not merely a localised distortion of the stratigraphy. Once the cores reached the laboratory they were stored in a dark room at 4°C prior to subsampling. The Ffridd-y-Fawnog site is shown in Plate 5.

3.2.2. Laboratory methods

(i) **Lithostratigraphy**. The main lithostratigraphic features of the two cores were described using the system of Troels-Smith (1955). The sediment colour was described using Munsell colour charts (Munsell© 1975).

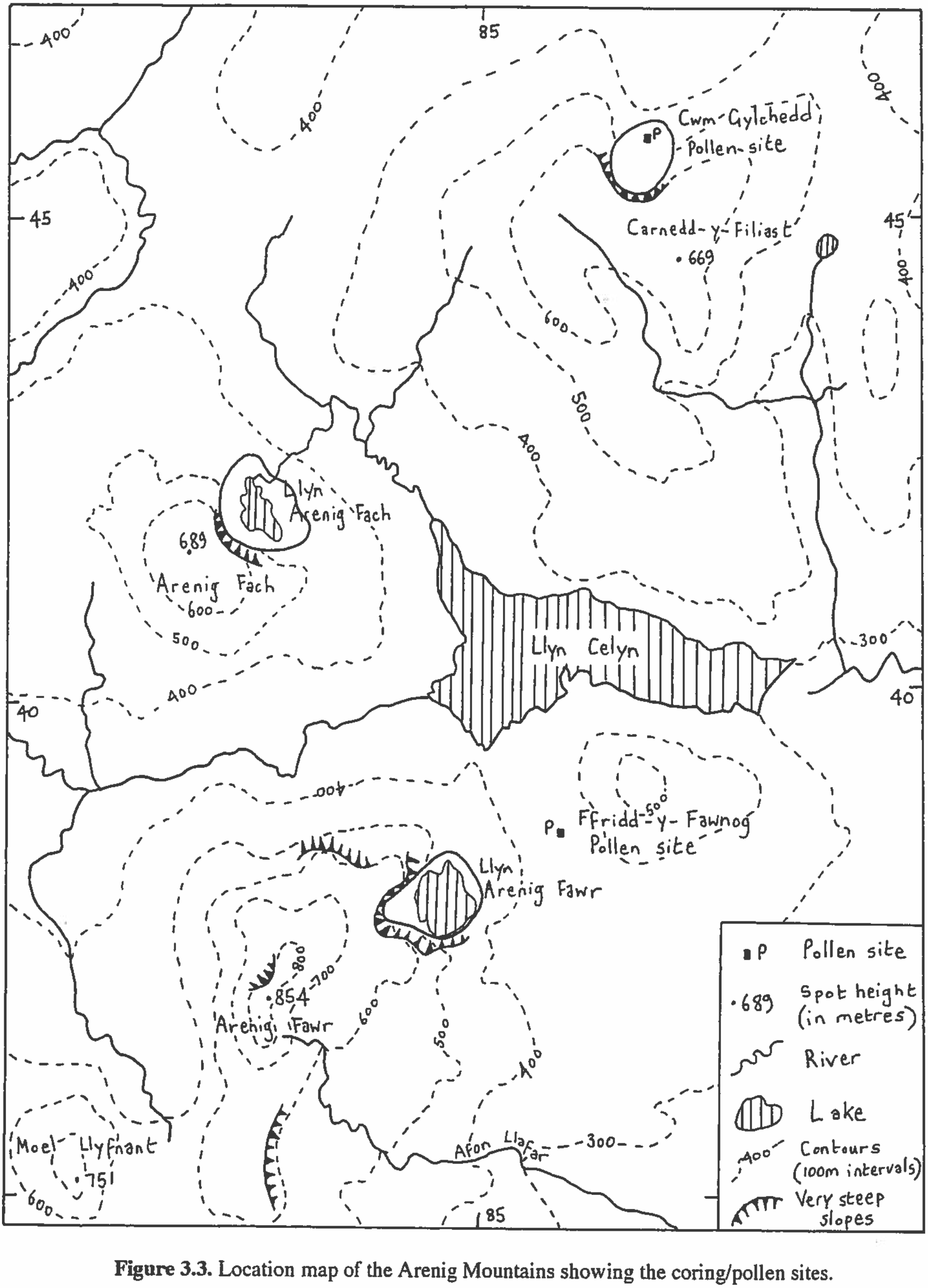


Figure 3.3. Location map of the Arenig Mountains showing the coring/pollen sites.

Plate 5. The Ffridd-y-Fawnog coring site. The sampling position is marked by an 'X' in the picture. The cwm behind is that containing Llyn Arenig Fawr.

(ii) **Physical properties.** 1 cm^3 samples for loss-on-ignition were taken at regular intervals (every 5 - 10 cm) from the bottom of cores. Both the Cwm Gylchedd and Ffridd-y-Fawnog cores were sampled through the bottom 1.5 metres of sediment. Sediment dry weight was determined after oven-drying the sediment at 105°C for twenty-four hours. Percentage organic matter and percentage calcium carbonate were calculated as weight loss after ignition at 550°C and 950°C, respectively, and expressed as percentages of sediment dry weight. The weight loss between 550°C and 950°C was multplied by 2.27 in order to convert CO_2 to $CaCO_3$. Percentage mineral matter expresses the relative abundance of material other than organic matter and carbonate in the dried sediment.

(iii) **Magnetic susceptibility.** Measurements of magnetic susceptibilty were taken every 2cm along core sections using a Bartington Instruments single-sample magnetic susceptibility meter. These measurements covered the same basal length of core as that for loss-on-ignition. Measurements were taken up and down the core respectively and the values of each traverse averaged and corrected taking into account variations in the Earth's magnetic field. This parameter measures the ease with which a sample can be magnetized and is proportional to the concentration of ferrimagnetic minerals in a sample. Since ferrimagnetic minerals are usually associated with inputs of allochthonous mineral material to a catchment basin, this parameter serves as a useful measure of the relative importance of inorganic materials in the sediment.

(iv) **Pollen analysis.** Samples for pollen analysis were prepared by standard procedures (Faegri and Iversen 1975, Berglund and Ralska-Jasiwiczowa 1986) and mounted in silicone oil. At least 300 grains were counted on each slide. The pollen data was then synthesized using Psimpoll© (Bennett 1993). As with loss-on-ignition and magnetic susceptibility, pollen analysis was done for the basal sections of core with the principal objective being the recognition of Lateglacial and/or early Flandrian vegetational change. The pollen data acts as a biostratigraphic classification of the sediments.

3.3. Results

3.3.1. Ffridd-y-Fawnog

(i) **Lithostratigraphy.**

The basal stratigraphy at this site displays the classic tripartite sequence, typical of the Lateglacial period. At the base there are grey lake clays (436 - 448 cm). In the field these appeared distinctly blue-grey and appear to have changed colour slightly upon exposure to the air. Above this there is a distinct black organic unit 10 cm thick (426 - 436 cm). This is then overlain by a grey clay unit with a significant proportion of coarse sand-sized particles (416 - 426 cm). Above this tripartite sequence the sediment consists of over 400 cm of organic lake muds and peats. Only the organic lake muds below 393 cm were analysed in this study. The Troel-Smith notation of this sequence is displayed in Table 3.1.

(ii) **Sediment properties**.

The results of the loss-on-ignition analyses are shown in Figure 3.4. Water content fluctuates over the tripartite sequence with low percentages (< 30%) in the basal clay unit (436 - 448 cm), high percentages (> 80%) in the overlying organic unit (426 - 436 cm) and low percentages (< 30 %) in the succeeding clay unit (416 - 426 cm). Water content then remains high throughout the sampled section of core (393 - 416 cm). Organic content fluctuates between values of < 10% in the two clay units and >80% in the organic unit in-between. The remainder of the core is dominanted by organic material with values consistently above 90%. Residue oscillates inversely to that of organic

Table 3.1. Sediment description of the Ffridd-y-Fawnog core

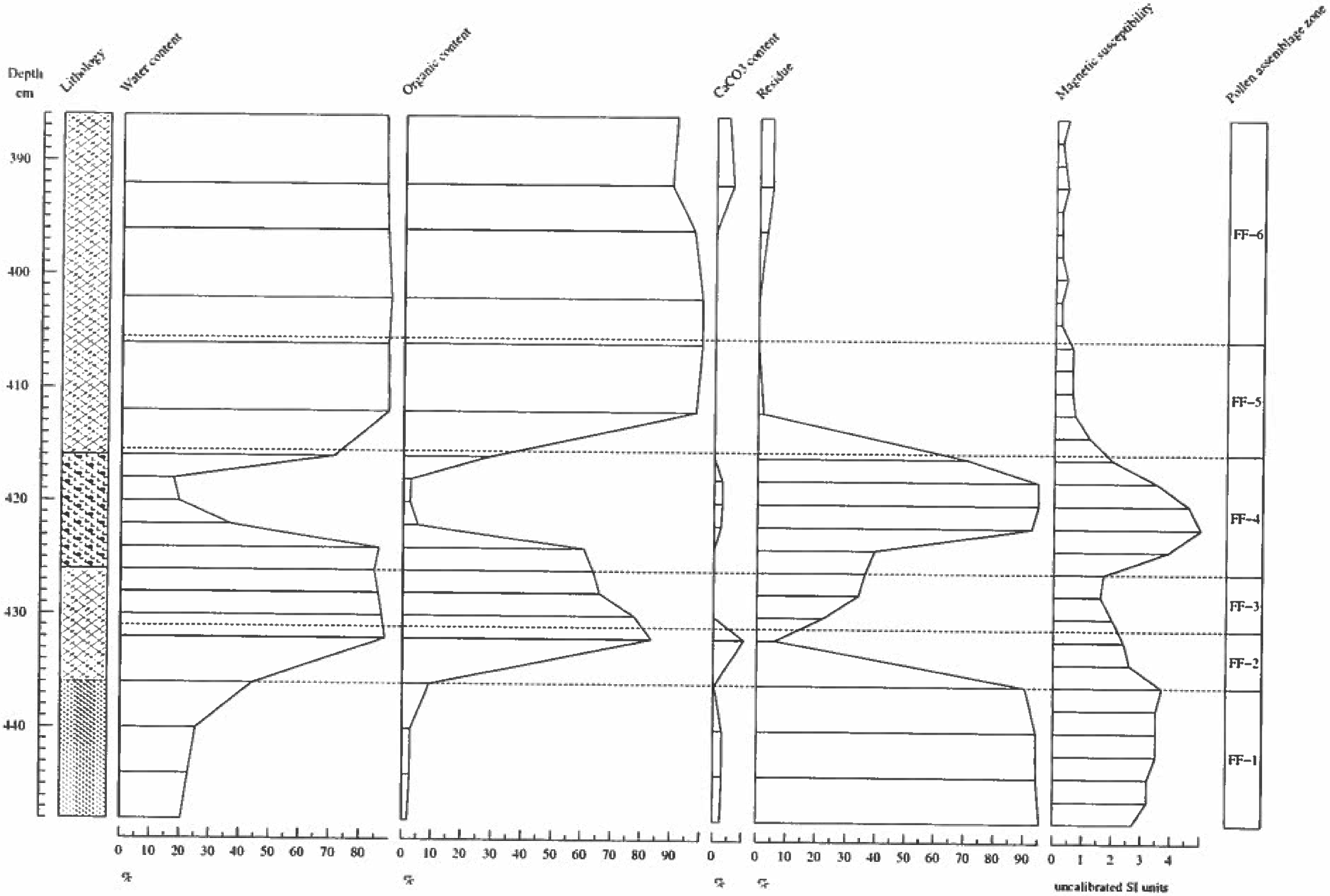
Depth (cm)	Sediment description
393 - 416	**Ld2 Dg2** Munsell chart colour: 10YR 2/2 (very dark brown) Organic lake mud. A coarse detritus nekron mud.
416 - 426	**As3 Gs1** Munsell chart colour: 7.5YR 6/0 (Grey) Predominantly a grey clay with some sand sized clasts.
426 - 436	**Ld3 Dg1** Munsell chart colour: 5YR 2.5/1 (Black) Organic lake mud. A fine detritus organic mud.
436 - 448	**As4** Munsell chart colour: 7.5YR 6/0 (Grey) Grey clay.

material with high values in the clay units and low values in the intervening organic lake mud. The remainder of the core is dominated by low residue content (<10%). $CaCO_3$ appears to be of little significance and never reaches over 10% of the sediment content although a prominent peak corresponds with high organic levels and low residue levels at 434 cm.

(iii) **Magnetic susceptibility**.

The results of the magnetic susceptibility are shown alongside the loss-on-ignition data in Figure 3.4. Prominent positive peaks occur at 436 cm and 422 cm. As in the loss-on-ignition results, the fluctuations in magnetic susceptibility closely mirror that of the tripartite lithostratigraphic

Figure 3.4. Ffridd–y–Fawnog Sediment Analysis



sequence. At depths above 415 cm magnetic susceptibility is low, consistently reaching less than 0.5 uncalibrated SI units..

(iv) **Pollen stratigraphy**

The pollen diagram for the Ffridd-y-Fawnog core is presented in Figures 3.5a and 3.5b. and the total pollen concentrations are displayed separately in Figure 3.6. The assemblage zones are described below.

Zone FF1 (446 - 436 cm)

Poaceae - Cyperaceae pollen assemblage zone

This zone is characterized by abundant herb pollen (>80%). Poaceae and Cyperaceae dominate reaching values of 40% and 35% respectively. *Rumex* occurs at levels above 5% whilst other herbs such as *Artemisia*, Carophyllaceae, Chenopodiaceae and *Helianthemum* occur intermittently at low levels (<5%). Tree pollen is around 10% and comprises *Betula* and *Pinus* at around 5% each. Shrub pollen is low at the base of the zone comprising 5% of total land pollen although this increases to 10% towards the top of the zone. The shrub pollen comprises primarily *Juniperus* and lesser amounts of *Salix*. Pollen concentrations are low, reaching only 50,000 grains/cm^3 through much of the zone.

Zone FF2 (436 - 431 cm)

Juniperus - Betula pollen assemblage zone

This zone is characterised by a peak in *Juniperus* pollen with values reaching 15%. *Betula* also rises through this zone from 5% to 20% as does *Salix*, to 5% of total land pollen. Both Poaceae and Cyperaceae decline by nearly 20%.The rise in tree pollen is clear from the total land pollen diagram where tree pollen rises from <10% to >25%. Caryophyllaceae and Chenopodiaceae also peak in this zone and *Rumex* declines. Total pollen concentrations increase slightly through this zone reaching nearly 100,000 grains/ cm^3.

Zone FF3 (431 - 426 cm)

Betula - Poaceae - Salix pollen assemblage zone

Betula peaks at over 30% in this zone and dominates the pollen assemblage along with Poaceae at 30% also. Whilst *Juniperus* shows a sharp decline in the bottom half of the zone, *Salix* remains relatively stable at around 5%. *Pinus* also peaks at the base of the zone (5 - 10%) although declines to near zero at the top of the zone.

Figure 3.5a. Percentage pollen diagram from Ffridd–y–Fawnog. Trees, shrubs and dominant herbs

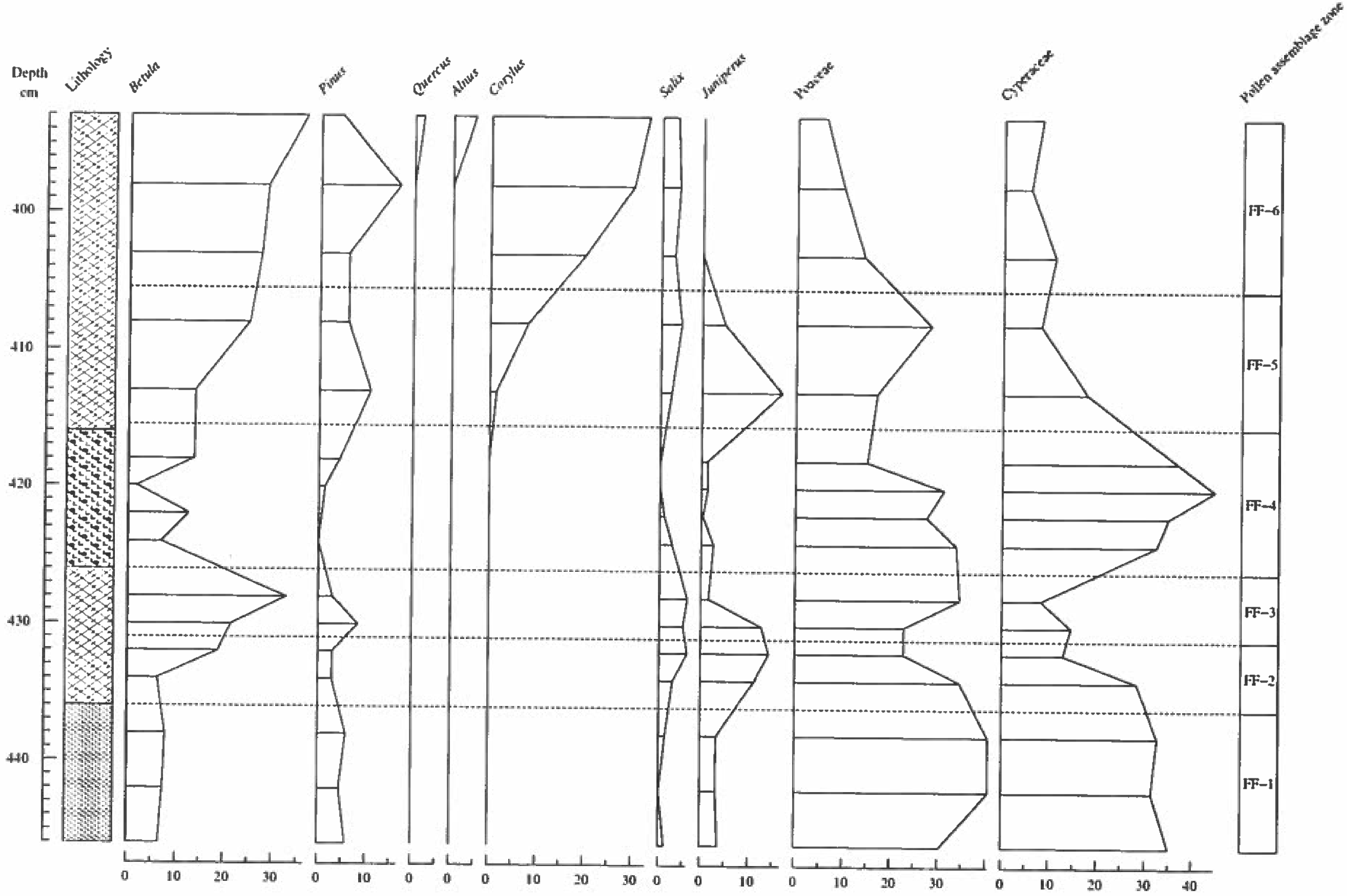
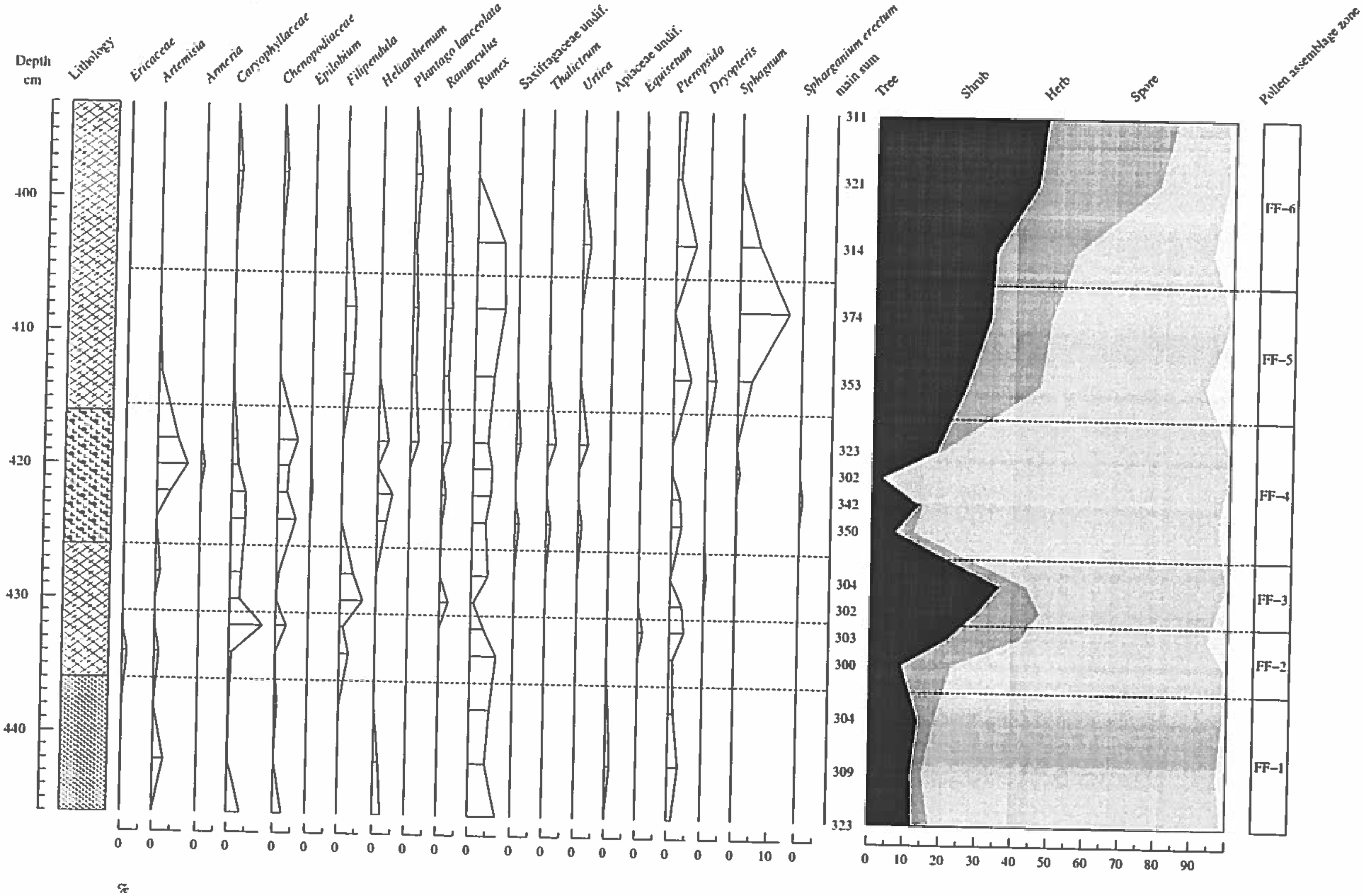


Figure 3.5b. Percentage pollen diagram from Ffridd–y–Fawnog Herbs, spores, aquatics and summary diagram



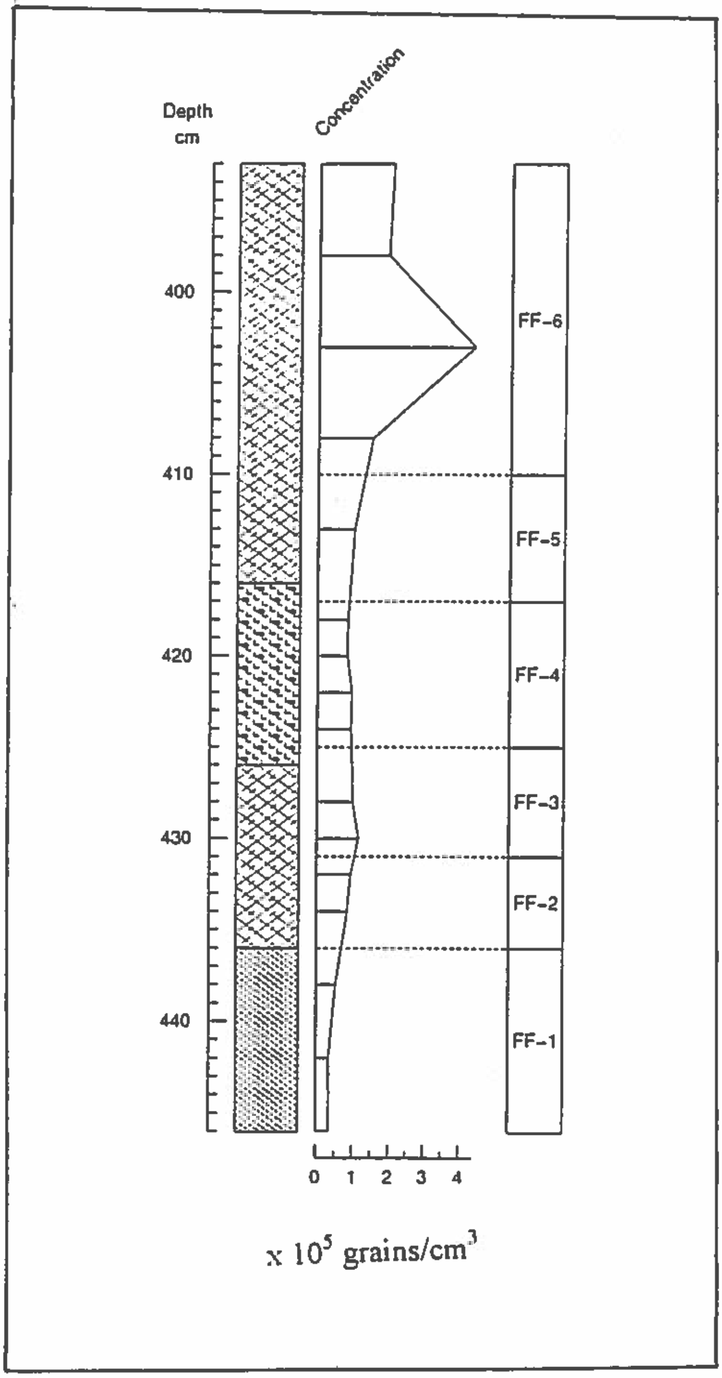


Figure 3.6. Total land pollen concentration from Ffridd-y-Fawnog.

Zone FF4 (426 - 415.5 cm)

Cyperaceae - Artemisia pollen assemblage zone

This zone is dominated by a peak in Cyperaceae at 40% and a lesser but significant peak of *Artemisia* at over 5%. Herb pollen overall reaches over 95% of the total land pollen and herbs such as Caryophyllaceae, Chenopodiaceae, *Helianthemum* and *Rumex* reach 5% or more. *Epilobium* and *Armeria* also make an appearance although only at values less than 2%.

Zone FF5 (415.5 - 405.5 cm)

Juniperus - Betula - Poaceae pollen assemblage zone.

This zone is characterized by a peak in *Juniperus* at over 15%. *Betula* rises throughout this zone from 10 - 15% to 20 - 25% and Poaceae peaks at around 25% of total land pollen. *Corylus* rises from near zero at the base of the zone to over 10% at the top of the zone. Differentiation between *Corylus* and *Myrica* pollen grains proved difficult and most pollen grains of this type were counted as *Corylus*. If this causes any discrepancies in the interpretation of the pollen data it will be taken into account in the discussion. Cyperaceae values decrease from over 25% to less than 10% through this zone. Overall, herb pollen declines from 80% to less than 50% although *Rumex* remains stable at around 5% and *Filipendula* makes an appearance reaching nearly 5%. *Sphagnum* pollen rises sharply through this zone reaching 10% of total sum aquatic and land pollen.

Zone FF6 (405.5 - 393 cm)

Betula - Corylus - Pinus pollen assemblage zone.

This zone is dominated by *Betula* which reaches 35% of total land pollen. *Corylus* pollen is also dominant and rises from around 15% to over 30% of total land pollen. *Pinus* also peaks in this zone at nearly 20% and significantly, towards the top of this zone, *Quercus* and *Alnus* appear for the first time. Poaceae declines throughout this zone as do herbs in general. *Sphagnum* declines through this zone from 10% to 0% of total aquatic and land pollen. Pollen concentrations increase dramatically in this zone, peaking at 400,000 grains/ cm^3 at 403cm and stabilizing at around 200,000 grains/ cm^3 towards the top of the zone.

3.3.2. Cwm Gylchedd

(i) Lithostratigraphy.

The lowest sediments (347 - 358 cm) consist entirely of grey lake clays. These grade into a fine grained lake mud consisting largely of organic material but with a significant clay component (337

- 347 cm). Overlying these lake deposits the sediment contains abundant organic fragments and can be described as a coarse-detritus nekron mud (258 - 337 cm). This grades into telmatic humified peats between 210 and 258 cm. The sediment above this was not analysed in this study but was a continuation of the latter deposit grading upwards into terrestrial peats. The Troel-Smith notation of this sequence is displayed in Table 3.2.

Table 3.2. Sediment description of the Cwm Gylchedd core

Depth	Sediment description
210 - 258	**Th4** Munsell chart colour: 10YR 2/2 (Very dark brown) Humified peat containing roots and rhizomes of herbaceous plants.
258 - 337	**Ld2 Dg2** Munsell chart colour: 10YR 2/2 (Very dark brown) Organic lake mud. A coarse detritus nekron mud.
337 - 347	**Ld3 As1** Munsell chart colour: 10YR 4/2 (Dark greyish brown) Fine detritus organic mud but with a significant proportion of clay.
347 - 358	**As4** Munsell colour chart: 7.5YR 6/0 (Grey) Grey clay.

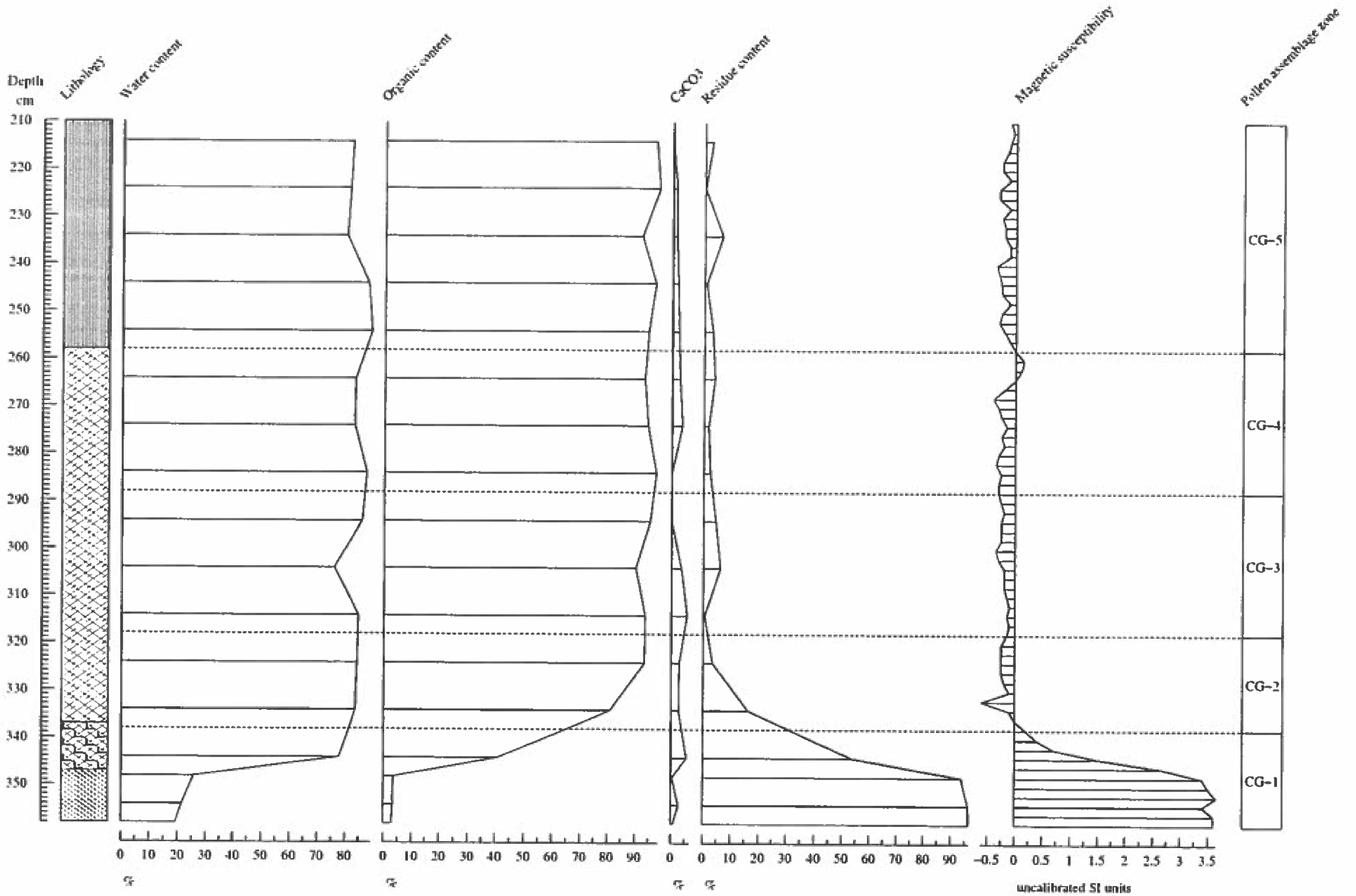
(ii) **Sediment properties.**

The results of the loss-on-ignition analyses are shown in Figure 3.7. Water content is low (< 30%) in the basal clay deposits (347 - 358 cm) and increases rapidly upwards to a consistently high value (> 75%) through the rest of the analysed core section. Organic content is very low (< 5%) in the basal clays and increases rapidly upwards through the core reaching values of > 90% in the rest of the sequence. $CaCO_3$ never reaches more than 5% of the sediment content and does not display any discernible trends although does reach zero between 294 cm and 284 cm. Residue content displays a pattern inverse to that of organic content, as would be expected. Values reach 95% in the basal clay deposits. Residue content then decreases markedly upwards being less than 10% of the content upwards of 324 cm.

(iii) **Magnetic susceptibility.**

The results of the magnetic susceptibility measurements are shown in the Figure 3.7 alongside the loss-on-ignition data. The basal sediments correspond with pronounced positive peaks in magnetic susceptibility. The magnetic susceptibility values decrease rapidly upwards and become negative at around 335 cm and remain slightly negative for much of the rest of the analysed core.

Figure 3.7. Cwm Gylchedd Sediment Analysis



(iv) Pollen stratigraphy

The pollen diagram for the Cwm Gylchedd core is presented in Figures 3.8a and 3.8b. and the total pollen concentrations are displayed separately in Figure 3.9. The assemblage zones are described in the following text.

Zone CG1 (353 - 338 cm)

Poaceae - Cyperaceae pollen assemblage zone.

This zone is characterised by abundant herb pollen (>80%). Principle among these are Poaceae (20 - 30%) and Cyperaceae (15 - 20%). Other significant herbs include *Artemisia* (10%) and *Rumex* (10%). Pollen of *Plantago lanceolata*, *Helianthemum*, Caryophyllaceae, Chenopodiaceae and *Urtica* occur also, although in low amounts (<5%). Tree pollen is low (10 - 20%) as is shrub pollen (<5%) with dominant species being *Betula* and *Pinus* in the former category and *Juniperus* in the latter. Aquatic pollen such as *Sphagnum*, *Isoetes*, *Myriophyllum verticillatum*, *Potamogeton* and *Typha angustifolia* occur in small amounts (<5%).

Zone CG2 (338 - 318cm)

Betula - Juniperus assemblage zone.

This zone is characterized by a peak in *Juniperus* (18%). *Betula* is also dominant at 20% throughout this zone. The spore *Equisetum* is also of significance in this assemblage zone peaking at over 5%. This zone marks the last appearance of aquatic pollen such as *Isoetes*, *Myriophyllum verticillatum*, *Potamogeton* and *Typha angustifolia* in the core sequence.

Zone CG3 (318 - 288cm)

Betula - Salix- Corylus assemblage zone.

In this zone *Betula* peaks at 35% of total land pollen. *Salix* levels reach 10% consistently in this zone whilst *Corylus* rises steadily through the zone from <5% at the base to nearly 20% towards the top. *Pinus* is also present at values below 5%. Total tree pollen remains steady at 35 - 40% of total land pollen and no major changes occur in the overall percentages of tree, shrub, herb and spore pollen.

Zone CG4 (288 - 258cm)

Corylus - Betula - Pinus - Salix assemblage zone.

Figure 3.8a. Percentage pollen diagram from Cwm Gylchedd. Trees, shrubs and dominant herbs

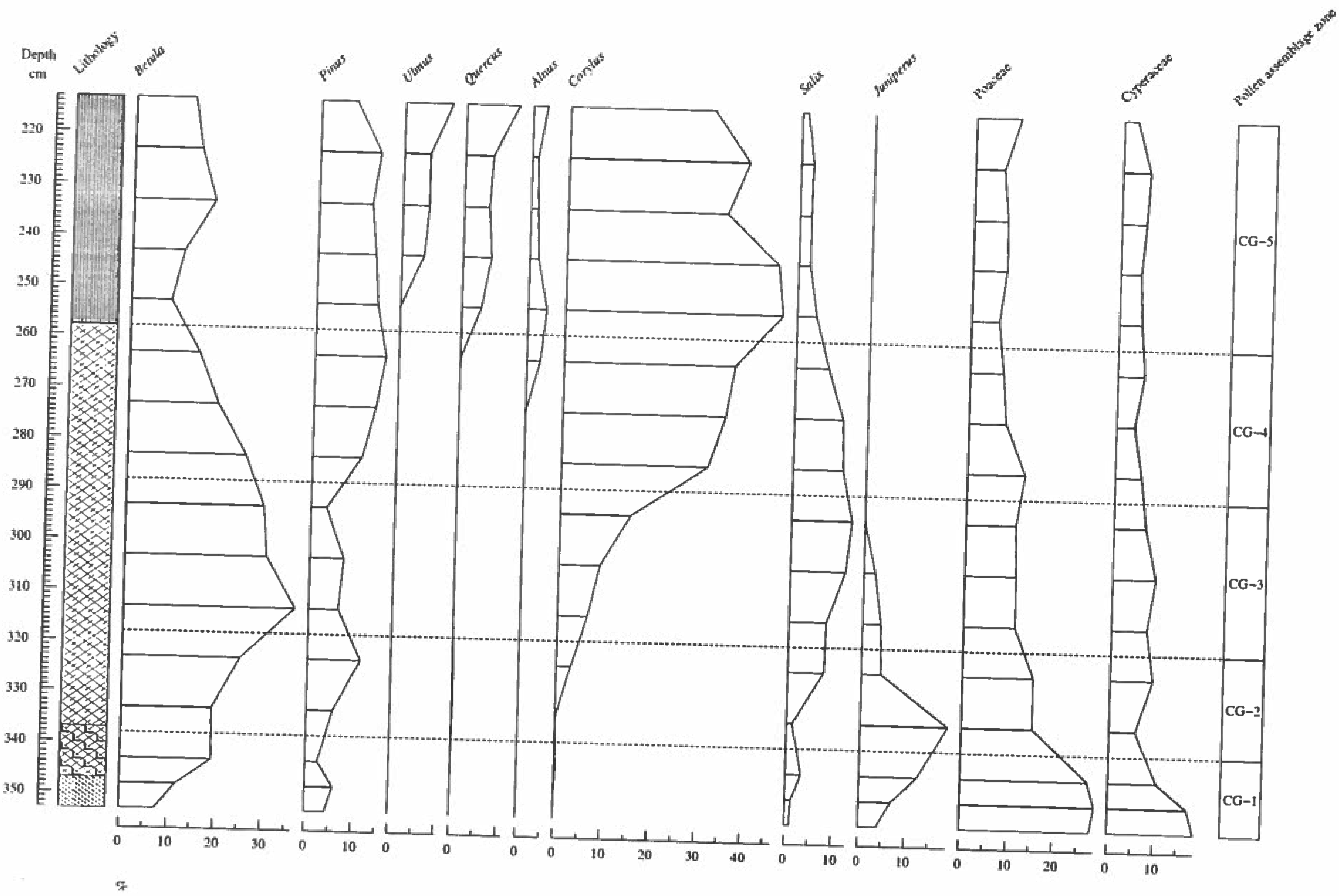
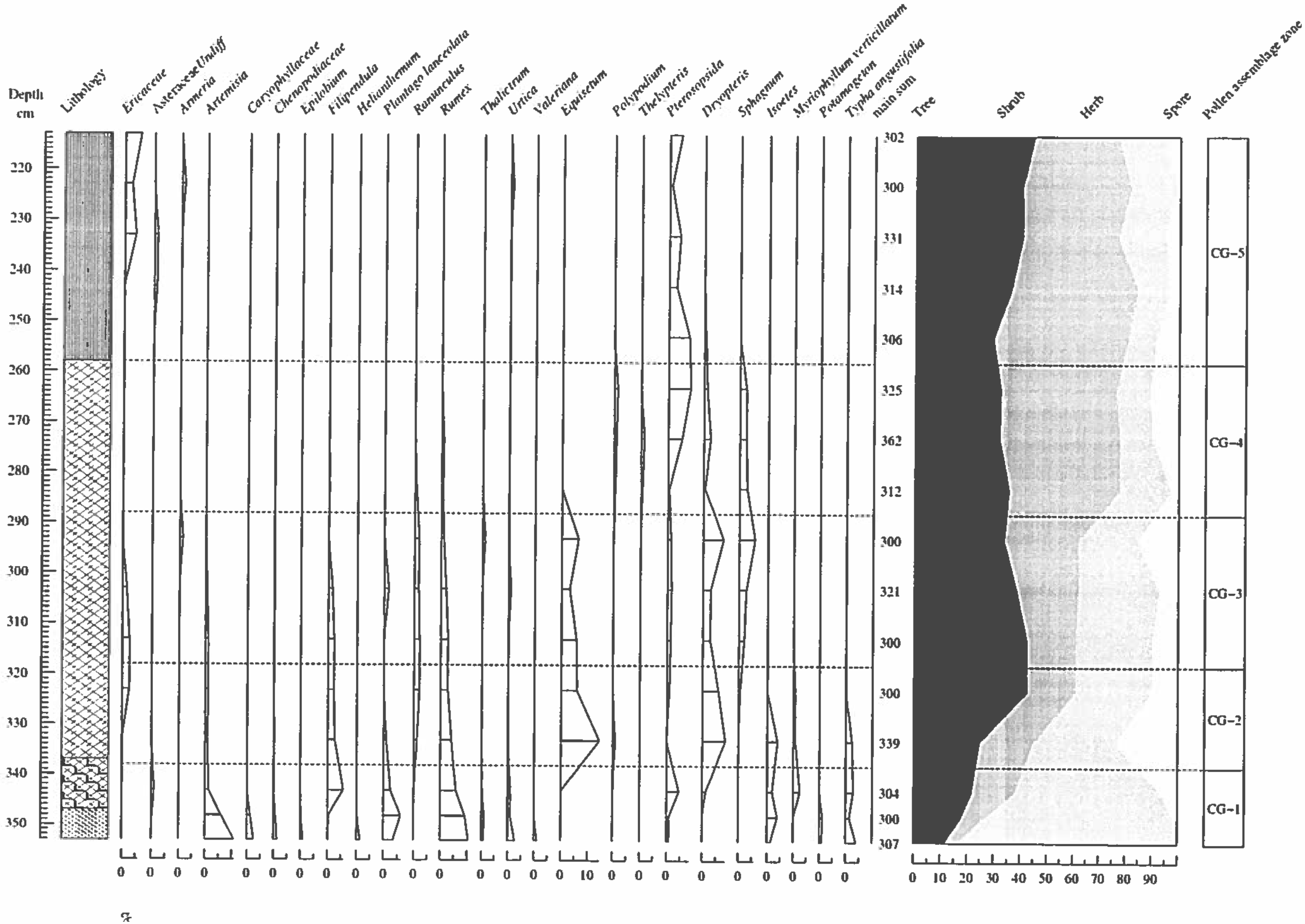


Figure 3.8b. Percentage pollen diagram from Cwm Gylchedd. Herbs, spores, aquatics and summary diagram



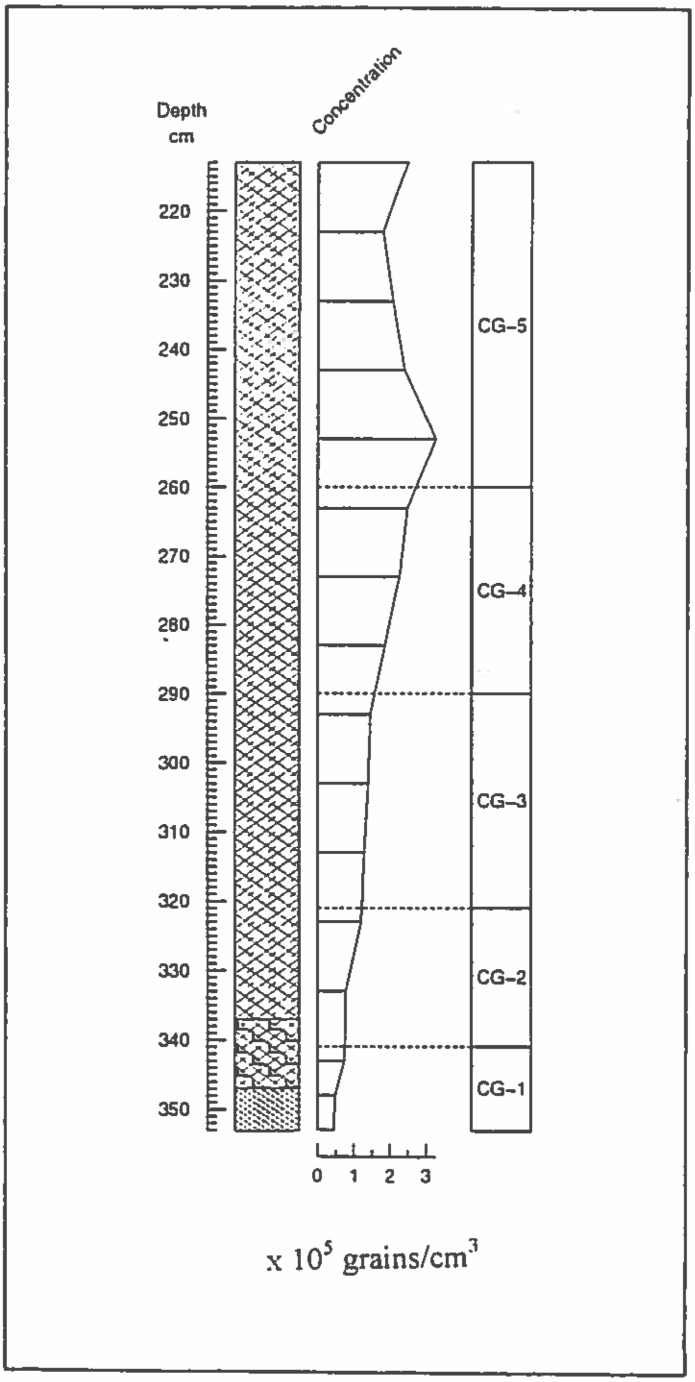


Figure 3.9. Total land pollen concentration diagram from Cwm Gylchedd.

This zone is dominated by *Corylus* which rises to 40% at the top of this zone. *Betula* is still prominent although declines steadily through the zone from 30% to <15%. *Pinus* reaches 15% of total land pollen in this zone. *Salix* remains important although declines towards the top of the zone to less than 5%. *Juniperus* disappears in this zone whilst Alnus appears for the first time at the top of the zone. Pterosopsida (monolete) Undif. rises significantly towards the top of zone albeit only to values just under 5%.

Zone CG5 (258 - 213 cm)

Corylus-Betula-Pinus-Quercus-Ulmus assemblage zone.

In this assemblage zone a mixed deciduous forest is evident, dominated by *Corylus* at 35 - 40%. *Betula* and *Pinus* remain the dominant trees at 15 - 20% and 5-10% respectively although *Ulmus* and *Quercus* both appear and rise through this zone to levels just over 5%. *Alnus* is also present although at consistently low levels (<5%). The tree - shrub forest is illustrated in the overall percentage diagram with the combined percentages of these two pollen types consistently reaching 60%. *Pteropsida* (monolete) Undif. remains present although at low values less than 5%.

Total pollen concentrations rise steadily through zones CG1 to CG4 from 50,000 grains/cm^3 to over 200,000 grains/ cm^3. Total concentrations peak early in zone CG-5 after which concentrations never fall beneath 150,000 grains/ cm^3.

3.4. Discussion

3.4.1. Ffridd-y-Fawnog

The following discussion provides an analysis of the data described in Section 3.3.1.

The basal pollen zone, **FF-1 (436 - 446 cm)**, is characterized by open habitat, herb dominated communities consisting largely of grasses and sedges. The presence of *Rumex*, Chenopodiaceae, Caryophyllaceae, *Helianthemum* and *Artemisia* as well as Poaceae and Cyperaceae, in the lower part of this zone reflects local vegetational developments during an early stage of primary succession following the decay of the Late Devensian ice sheet. Birch and pine must also have been present in small amounts during this phase. *Rumex* appears to be the most significant herb after Poaceae and Cyperaceae and its role in early Lateglacial grassland succession is recognised at many other sites in Snowdonia (Seddon 1962, Crabtree 1972, Simpkins 1974 and Ince 1981). Modern day analogues for this vegetation phase have been drawn from pioneer herb communities on recently deglaciated terrain from Jostadalsbrae in Southern Norway (Pennington 1977). High residue and low organic sediment content as well as strongly positive magnetic susceptibility in this pollen zone indicates high levels of inwash. The recently deglaciated terrain would have been vulnerable to erosion especially given the sparsity of vegetation cover and the absence of expansive woodland cover. The low pollen concentrations in the basal zone are likely to indicate the sparsity of vegetation biomass as well as rapid sedimentation rates at a time of high inwash.

The peak in Juniperus pollen in zone **FF-2 (436 - 431 cm)** reflects an amelioration in climate. Sediment organic content and subsequently sediment water content also peak in this zone illustrating a shift to a temperate interstadial climate. The onset of organic sedimentation following Devensian deglaciation is dated at 13,200 ± 120 yr BP at Llyn Gwernan in southern Snowdonia (S.Lowe 1981) and the *Juniperus* rise has been dated at around 12,500 yr BP at Llanilid in South Wales (Walker and Harkness 1990). The *Juniperus* peak is considered to reflect the warmest part of the Lateglacial (Lowe and Walker 1986) and the occurrance of thermophilous herbs such as *Filipendula* also indicate a warming climate.

In zone **FF-3 (431 - 426 cm)** Juniper is replaced by Birch, probably because of increased shading. Willow is likely to have occurred at damp sites such as at the former lake margin and as an understorey shrub within birch woodland. This interstadial temperate phase is illustrated clearly in the sediment stratigraphy, which in this zone comprises organic lake muds, as well as in the total pollen summary diagram where tree pollen peaks sharply. The *Betula* peak during the Lateglacial interstadial is thought to correspond with a radiocarbon date of 11,500 yr BP in South Wales (Walker and Harkness 1990).

The base of zone **FF-4 (426 - 415.5 cm)** can be taken as the approximate boundary between the Windermere Interstadial and the Loch Lomond Stadial. This is marked lithologically by an abrupt change from organic lake muds to clay-rich minerogenic sediment. The onset of the Younger Dryas is radiocarbon dated to 11,160 ± 90 yr BP at Llyn Gwernan in southern Snowdonia by S.Lowe (1981) and to around 11,000 yr BP in South Wales by Walker and Harkness (1990). Open herb habitats are evident, dominated by Cyperaceae. Open ground and soliflucted soils of the Stadial environment are reflected by a sharp increase in the frequency of *Artemisia* and the presence of Caryophyllaceae, Chenopodiaceae, *Epilobium*, *Helianthemum* and *Thalictrum* together with the continued presence of *Rumex*. Inwash to the former lake basin is highlighted in the peaks of sediment residue content and magnetic susceptibility.

Zone **FF-5 (415.5 - 405.5 cm)** represents the Early Flandrian and its basal boundary the Loch Lomond Stadial/Flandrian transition. This zone corresponds with organic lake muds in contrast to the minerogenic clays of the previous zone, FF-4. Vegetation development and substrate stabilization is reflected in the low magnetic susceptibility values and markedly reduced residue levels. *Juniperus*, whose pollen peaks in this zone, represents the pioneer shrub of an ameliorating climate. The Early Flandrian *Juniperus* peak in Wales spans the time interval from 9,750 to 10,200 yr BP (Hibbert and Switsur 1976, Ince 1983, Walker and Harkness 1990). A warming climate is also indicated by the reappearance of *Filipendula*, a thermophilous herb, in this zone. The top of the zone can be correlated with the Early Flandrian rise in *Corylus* and corresponds to a date of around 9,000 yr BP (Hibbert and Switsur 1976, S.Lowe 1981, Walker and Harkness 1990).

Zone **FF-6 (393 - 410 cm)** is characterized by the development of full temperate woodland. The woodland was dominated by birch, hazel and pine with oak and alder making an appearance at the top of the zone. The top of the zone, on the basis of the appearance of *Alnus*, is likely to date to around 6,900 yr BP (Hibbert and Switsur 1976). The development of a richly vegetated woodland is also reflected in the dramatic rise in pollen concentrations in this zone as well as the low values of magnetic susceptibility and residue. The rise in pollen concentrations is also likely to be due to reduced sedimentation rates at a time of reduced inwash.

3.4.2. Cwm Gylchedd

The following discussion provides an analysis of the data described in Section 3.3.2.

The basal pollen zone, **CG-1 (353 - 338 cm)**, is indicative of open habitat, herb dominated communities. Peaks in magnetic susceptibility and residue within this zone indicate high levels of inwash and reflect the minerogenic nature of the lake clays. It is likely that this zone represents the end of the Loch Lomond Stadial and the onset of deposition upon glacier retreat from the basin. The presence of disturbed ground taxa (especially *Artemisia*) attest to the relatively unstable ground conditions at this time. It is apparent that *Rumex* played a significant role in the colonization and stabilization of the recently deglaciated landscape and climatic amelioration towards the end of this zone is marked by the presence of thermophilous herbs such as *Filipendula*.

The beginning of zone **CG-2 (338 - 318 cm)** marks the Loch Lomond Stadial/Flandrian transition. The basal clay unit is replaced by organic lake muds. The peak in *Juniperus* pollen indicates a rapid rise in temperature leading to increased flowering of this shrub (Iversen 1954). The biostratigraphic horizon is a characteristic feature of early Flandrian profiles. In Wales, the age of the *Juniperus* peak spans the time interval from 9,750 to 10,200 yr BP (Hibbert and Switsur 1976, Ince 1981, Walker and Harkness 1990). The peak in Dryopteris probably represents the development of a fern-rich understory to the developing *Betula* woodland, the pollen of which rises through this zone. The peak in *Equisetum* spores, a plant characteristic of wet and marshy ground

(Stace 1991), perhaps implies rapid expansion in the absence of competition on the lake edge (Tipping 1993).

A characteristic of the two basal pollen zones, CG-1 and CG-2, is the presence of the aquatics *Isoetes*, *Myriophyllum alterniflorum*, *Potamogeton* and *Typha angustifolia*. *Isoetes* is indicative, with *Myriophyllum alterniflorum*, of moderately acid water of very low nutrient status (Godwin 1975). The complete absence of aquatic pollen and spores in the rest of the core perhaps implies early infilling, possibly as soon as the base of zone CG-3.

In zone **CG-3 (318 - 288 cm)** the peak in *Betula* and the decline in *Juniperus* pollen suggests that birch was able to rapidly establish some woodland tracts. The light demanding juniper shrubs were unable to effectively compete with birch at this time, being ultimately shaded out from their previous niches (S.Lowe 1994). The relatively high levels of *Salix* pollen in this zone (10%) suggest that willow was able to successfully compete with juniper and exist within birch woodland as an understorey shrub or in isolated damp localities such as at the site or margins of the former lake. The rise of *Corylus* through this zone implies an age of c.9,000 yr BP (Hibbert and Switsur 1976, S.Lowe 1981, Walker and Harkness 1990) for the base of zone CG-3. The ground substrate is likely to have stabilized upon vegetation development resulting in lower levels of minerogenic inwash to the basin. This is illustrated in the high organic and low residue content of the sediment in this zone and indeed through the rest of the analysed section of core. The same applies to magnetic susceptibility data which is slightly negative in this section and for much of the rest of the core.

In zone **CG-4 (288 - 258 cm)** the consistent rise in *Corylus* and decline in *Betula* pollen can be interpreted as the invasion of the upland birch woodland by hazel. The rising values of *Pinus* pollen (reaching 15%) suggest that pine may have been locally present on the drier mire surfaces surrounding the site. These levels of *Pinus* may be the result of windblown pollen from higher sites around the tree-line as S.Lowe (1994) suggests that pine probably formed the regional tree line along with scattered copses of birch. *Pteropsida* (monolete) Undif. appears to replace Dryopteris as an understorey fern whilst *Equisetum* disappears from the pollen record, perhaps implying complete infilling of the former lake by this time as suggested earlier. The appearance of *Alnus* in this zone perhaps represents the development of localised Alder carr on the surface of the infilled lake basin. The arrival of *Alnus* in the uplands of Wales is dated at around 6,900 yr BP from sediments within Nant Ffrancon in north Snowdonia (Hibbert and Switsur 1976).

In zone **CG5 (258 - 213 cm)** *Corylus* declines slightly allowing *Ulmus* and *Quercus* pollen percentages to increase. In this zone, full temperate deciduous woodland has developed. The appearance of Ericaceae in the top half of this zone perhaps implies localised clearance and the development of open heath.

On the basis of the discussion above it is likely that the rising total pollen concentration through zones CG1 to CG5 reflects increasing vegetation development and reduced sedimentation rates through the Early Flandrian.

3.4.3. Significance of the pollen-stratigraphy

The evidence from the Cwm Gylchedd site clearly implies a post Loch Lomond Stadial age of the sediments. No Lateglacial sequence has been been found unlike at the Ffridd-y-Fawnog site outside of former local glacier limits. Therefore, given that the deepest sediments within an infilled lake basin have been sampled then the evidence supports ice occupation at this site during the Loch Lomond Stadial. On the basis of this evidence, in conjunction with the geomorphological evidence described in Chapter 2, it would appear feasible to correlate all 4 of the Aran and Arenig glaciers to the Loch Lomond Stadial. The pollen stratigraphy of both the Cwm Gylchedd and Ffridd-y-Fawnog cores are correlated in Figure 3.10.

Chronozone	Age (yr BP)	Ffridd-y-Fawnog	Cwm Gylchedd
Flandrian			CG-5 (*Corylus-Betula-Pinus-Quercus-Ulmus*)
	6,700[#]	FF-6 (*Betula-Corylus-Pinus*)	CG-4 (*Corylus-Betula-Pinus-Salix*) CG-3 (*Betula-Salix-Corylus*)
	9,800[†]	FF-5 (*Juniperus-Betula*-Poaceae)	CG-2 (*Betula-Juniperus*)
Loch Lomond Stadial	11,000[†Δ]	FF-4 (Cyperaceae-*Artemisia*)	CG-1 (Poaceae-Cyperaceae)
Windermere Interstadial	13,200[Δ]	FF-3 (*Betula*-Poaceae-*Salix*) FF-2 (*Juniperus-Betula*)	
Devensian Deglaciation	Pre - 13,200	FF-1 (*Poaceae-Cyperaceae*)	

Hibbert and Switsur (1976), deduced from the appearance of *Alnus*.
† Walker and Harkness (1990).
Δ S. Lowe (1981).

Figure 3.10. Correlation diagram between the Ffridd-y-Fawnog and Cwm Gylchedd pollen-stratigraphy.

4. GLACIER RECONSTRUCTION

4.1. Methods of reconstruction

The former glaciers were reconstructed using the evidence mapped in the field. Using such data, the three-dimensional form and glaciological characteristics of the glaciers were calculated, the details of which are elaborated upon below.

i) **Glacier surface contours**. The former glacier surfaces were contoured at 10 metre intervals on maps of scale 1:10,000 (presented in Figure 4.1 at a scale of 1:25,000). This allowed particularly accurate reconstruction of other glacier features. Contours were plotted from the mapped ice margin and follow the general pattern as seen on modern glaciers whereby the contour curvature closely corresponds with snout curvature in the snout area but diminishes up-glacier to become reversed in the upper parts (Sissons 1974, 1977b, Lowe and Walker 1997, pp.43). In addition to the overall glacier shape, contour plotting was guided by the striae and till fabric measurements measured in the field.

ii) **Glacier area**. Former glacier areas were derived by superimposing a grid of squares with sides equivalent to 25 metres on to 1:10,000 maps of former glacier extent. The number of squares that lay predominantly within the glacier limits were counted and the glacier area then calculated.

iii) **Glacier volume**. Former glacier volumes were derived by measuring ice thickness at 30-80 regularly spaced points within each glacier (the number obviously a function of glacier size). Glacier thickness equals glacier surface altitude (obtained from the surface contours) minus land surface altitude. The average ice thickness was then multiplied by the glacier area to give glacier volume. In the case of the glaciers with the cwms now occupied by lakes (Llyn Lliwbran, Llyn Arenig Fawr and Llyn Arenig Fach) there is the problem of unknown depths over the lakes themselves. The problem is eliminated in the case of Llyn Arenig Fawr and Llyn Arenig Fach as the volume of the lakes are known (Roberts 1995). This was not the case for Llyn Lliwbran and thus only a minimum volume could be deduced in this case.

iv) **Ice surface gradient and glacier aspect**. The mean ice surface gradients and mean glacier aspects were derived from the same 30 - 80 regularly spaced points as in the glacier volume calculations. Ice surface gradients were calculated from the horizontal distance between contours

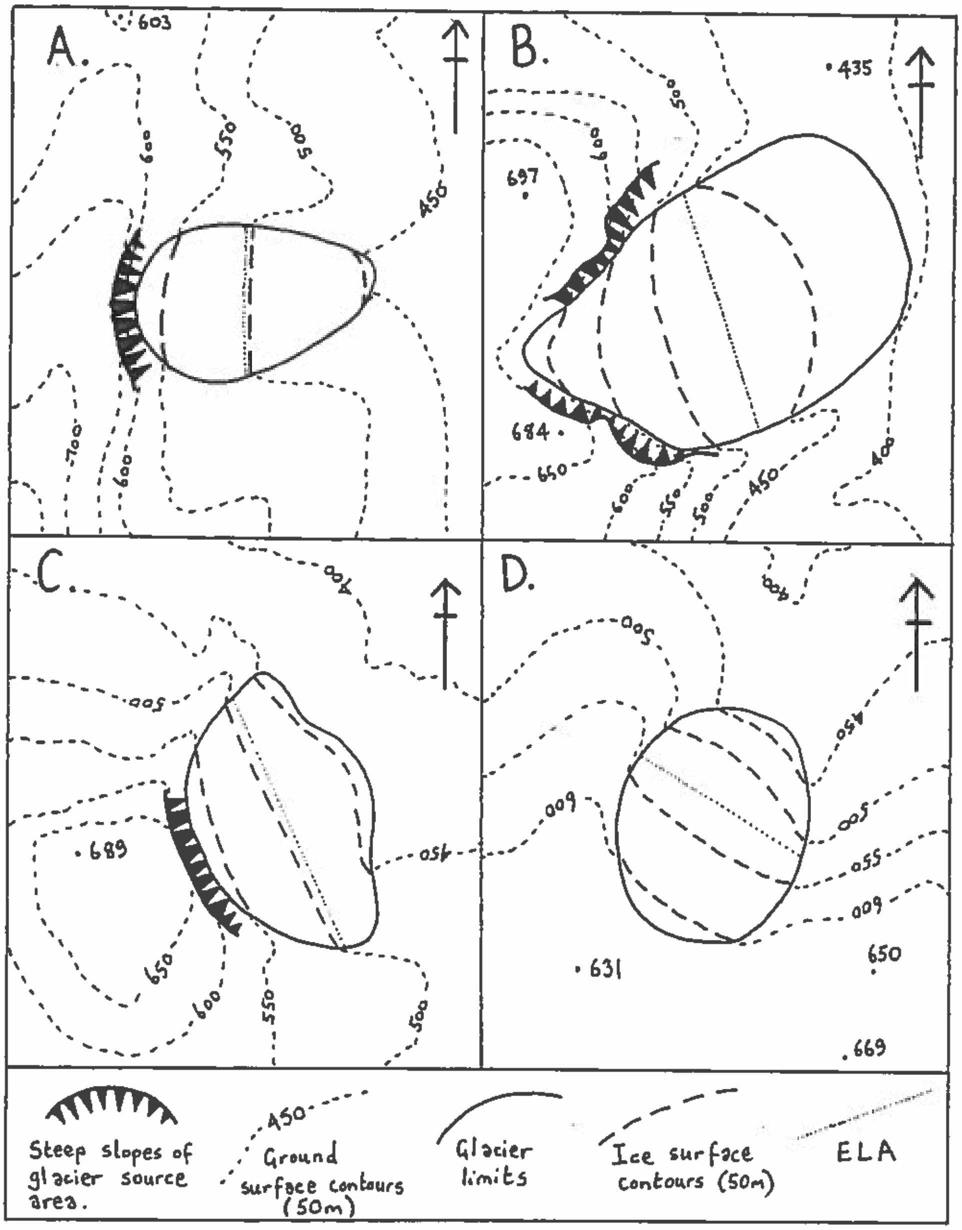


Figure 4.1. Reconstructed glacier surface contours at a scale of 1:25,000. A= Llyn Lliwbran glacier, B= Llyn Arenig Fawr glacier, C= Llyn Arenig Fach glacier, D= Cwm Gylchedd glacier.

using simple trigonometry to establish ice surface angles. Glacier aspects were measured at each regularly spaced point from flow direction lines drawn perpendicular to the ice surface contours.

v) **Insolation factor**. The combined measurements of ice-surface gradient and aspect at each regularly spaced sample point were used to calculate an insolation factor. This was calculated using the methods given in Sissons and Sutherland (1976) and represents the percentage of clear-sky solar radiation that would be absorbed, on average, across the whole glacier during the assumed May-September ablation season. The insolation factor for a horizontal glacier is 14.7 at latitude 53°N. This value for a horizontal glacier decreases northwards as would be expected with values such as 14.1 for the English Lake District and 11.8 for northern Scotland (Sutherland 1984). A high insolation factor indicates an adverse location since the glacier would have absorbed a large amount of direct insolation and therefore suffered greater potential ablation.

The insolation factor cannot be used to give absolute values of direct radiation absorbed by the glaciers since the former cloud cover is not known. However, it does give a measure of the relative importance of direct insolation on different glaciers. Another problem with this technique is that of shading by higher ground as this is not taken into account. However, small cirque glaciers tend to slope away from the rock walls that provide the main cause of shadow and also fill a large part of their basins, unlike valley glaciers. As such this was not perceived to be a significant issue and was ignored.

vi) **Snowblow and avalanche factors**. The importance of snowblow and avalanche to accumulation can be assessed using one a number of similar methods. Sissons and Sutherland (1976) developed the concept of a snowblow and avalanche factor to represent the relative importance of each input factor on glacier development. Potential avalanche areas are defined as those which slope at more than 20° directly on to the accumulation area of a former glacier and potential snow-blowing areas as those slopes, within the catchment, that sloped directly onto the accumulation area or potential avalanche area. Snowblow and avalanche factors represent the ratios between the snowblow or avalanche areas and the total glacier area. Although this method was applied by Sissons and colleagues in much of their work (Sissons and Sutherland 1976, Sissons 1979c, 1980b and in Cornish 1981) the writer feels that the more recent methods developed by Dahl et al (1997) are more suitable for a number of reasons.

Dahl et al (1997) propose that the ratio between the drainage area (D) and the reconstructed accumulation area (A) can be used as a rough measure of the potential for additional accumulation by snowblow (methods in deriving the accumulation area and the ELA are described later). The D/A ratio is preferable to the snowblow factor of Sissons and Sutherland (1976) for two principal reasons. Firstly, the D/A ratio relates extra input from snowblow to the accumulation area unlike the snowblow factor. Although the snowblow factor considers the potential snowblow onto the accumulation area, the ratio is determined by the whole glacier area. The D/A ratio would seem more logical as the importance of snowblow input would be calculated with reference only to the area of the glacier where it has any effect on mass balance and subsequent development. Secondly, and of more significance, the D/A ratio can be combined with estimated precipitation values to provide a reasonable estimate of total accumulation (Dahl et al 1997). This allows calculations of ablation season temperatures at the ELA to be made for a small number of glaciers, even for single glaciers (See Section 5.2).

While Dahl et al (1997) only consider using the entire drainage basin above the accumulation area, a consideration of prevailing winds and the recognition of areas of greatest snowblow potential would improve on the accuracy of the method. Based on work in the Lake District, Sissons (1980b) provided a strong argument suggesting that snowfalls were associated with southerly winds. It is probable that the synoptic situation was one where occluded fronts approached from the west or south-west. The main snowfalls were therefore occurring with south to south-easterly winds that preceded the fronts, with additional snowfall occurring with south to south-westerly winds after the fronts had passed. Considering this synoptic situation, only that part of the upper drainage area susceptible to winds in the south-west and southeast quadrants of the compass need to be assessed. Other possible directions are not important as most snow transfer by wind occurs during and immediately after the snowfall event (Sissons 1980a, pg.24). The Llyn Arenig Fawr glacier is given as an illustrating example of this technique in Figure 4.2.

The ratio between the total area susceptible to avalanche (slope >20°) leading directly on to the accumulation area (V) and the glacier accumulation area (A) represents the avalanche ratio (V/A) in this study. Note that the avalanche ratio here differs from the avalanche factor used in earlier work (e.g. Sissons and Sutherland 1976, Sissons 1979c, 1980a, 1980b Cornish 1981), where ratios are over total glacier area and not accumulation area. The reasons for this are the same as given for snowblow in the previous paragraph.

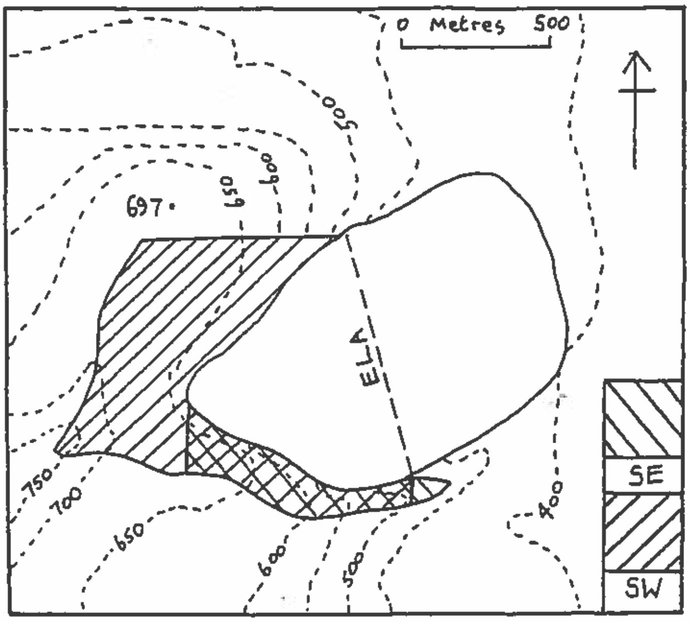


Figure 4.2. Derivation of D/A snowblow ratios. The importance of snowblow input was calculated using the ratio between the potential catchment area (shaded areas - showing the area of the SW and SE quadrants susceptible to snowblow) and the accumulation area of the glacier (area above the ELA). This example is based on the Llyn Arenig Fawr glacier.

vii) **Equilibrium Line Altitude (ELA).** The firn line or ELA for each glacier was calculated using the method of area-weighted mean altitude introduced by Sissons (1974). This method employs the following equation (see overleaf):

$$\textbf{ELA} = \sum Ai \cdot hi / \sum Ai \qquad \text{(Equation 4.1.)}$$

where the **Ai** are the areas between particular contour intervals and the **hi** are the corresponding mid-point altitudes for those contour intervals. This procedure is based on two assumptions: (a) during the glacial maxima the glacier was in equilibrium; and (b) both the accumulation and ablation gradients have a linear relationship with altitude. Sutherland (1984b) noted that a linear mass balance curve is unlikely in practice, although went on to show that the method can still be relied upon. In order to establish the extent to which the assumption of a linear mass balance curve biases the results, Sutherland (1984b) calculated the steady-state ELAs for 25 glaciers from a wide variety of climatic regimes using original mass balance data, thus allowing for the non-linearity of mass balance curves. These results were compared with the results gained using Equation 4.1. and a close relationship was found, although the ELA values calculated using the equation of Sissons (1974) tended to overestimate the real ELA. This slight overestimation is also noted by Rune (1996) and Nesje and Dahl (2000) but is thought to be the best method of ELA reconstruction for small cirque glaciers with even area/altitude distributions. Other methods such as toe-to-headwall ratios (THAR) and accumulation area ratios (AAR) have been ignored, the former being rather crude and the latter being better applied for larger valley glaciers longer than 3km (Porter 1977). Also, these methods and others such as the Balance Ratio method (Furbish and Andrews 1984) are ignored because the method used here has been most widely used to calculate equilibrium line altitudes of reconstructed Loch Lomond Stadial glaciers in upland Britain (e.g Sissons 1980a, Cornish 1981, Gray 1982, Lawson 1986, Ballantyne 1989, Shakesby and Matthews 1993, Wilson and Clark 1995, Mitchell 1996, Wilson and Clark 1998, Anderson et al 1998). Thus meaningful comparison can be made with other British sites using this method.

4.2 Glacier characteristics

The results of the measurements and calculations outlined in the previous section are summarized in Tables 4.1 and 4.2. Table 4.1 displays the physical characteristics of the former glaciers such as area, volume, depth, slope, aspect and ELA whilst Table 4.2 displays the local controlling factors of snowblow, avalanche and insolation.

The total area covered by the four glaciers was c.1.85 km^2 with a total volume of at least 0.0635 km^3. The glacier at Llyn Lliwbran in the Aran Mountains had the smallest area of 0.181 km^2 whilst the glacier at Llyn Arenig Fawr in the Arenig Mountains was the largest with an area of 0.92km^2. These areas compare with an average of 0.5km^2 for 35 north Snowdonia glaciers (Gray 1982) and 0.85 km^2 for the Lake District (Sissons 1980a). The largest glacier, at Llyn Arenig Fawr was therefore particularly large with an area greater than both the average glacier areas of north Snowdonia and the Lake District. However it was of no comparison to the largest glaciers in these areas which reached 4km^2 at Llyn Llydaw in Snowdonia and 4.02 km^2 at a site southeast of Ullscarf in the Central Fells of the Lake District. The Llyn Arenig Fawr glacier accounted for nearly 50% of the glacier-covered area and 54% of the total glacier volume of all the four glaciers in the Aran and Arenig Mountains. The Arenigs held a great deal more ice during the Loch Lomond Stadial than the Arans with over 90% of the total ice area and 87% of the calculable glacier volume. These figures would be much reduced if a glacier had occupied the Creiglyn Dyfi site east of Aran Fawddwy, although as noted earlier the evidence for such a glacier is weak and inconclusive.

Table 4.1. Physical characteristics of the former glaciers

Glacier	Area (km^2)	Volume (km^3)	Maximum depth (m)	Gradient	Aspect	ELA (metres)
Llyn Lliwbran	0.181	0.008§	68#	15°	88°	503
Arenig Fawr	0.92	0.034	65#	9.8°	72°	470
Arenig Fach	0.394	0.0115	57#	13.3°	56°	511
Cwm Gylchedd	0.363	0.01	55	15°	25°	531

§ The volume value for the Llyn Lliwbran glacier is a minimum value because the volume of the lake is not known.

These depths are minimum values as there are unknown depths over the lake.

Table 4.2. Local controlling factors of the former glaciers

Glacier	Insolation factor	Snowblow ratio D/A	Avalanche ratio (V/A)
Llyn Lliwbran	13.7	2.9	0.5
Arenig Fawr	13.5	1.85	0.26
Arenig Fach	12.1	2.62	0.42
Cwm Gylchedd	10.9	3.4	0.11

The Llyn Lliwbran glacier had the thickest ice with a maximum depth of at least 68 metres. This is ironic considering it was the smallest glacier but is probably the result of the considerable closure of the cwm in comparison to the Arenig sites. The glaciers at Llyn Arenig Fawr, Llyn Arenig Fach and Cwm Gylchedd follow with depths of 65 metres, 57 metres and 55 metres respectively. The depths of all the glacier sites now occupied by lake basins (Llyn Lliwbran, Llyn Arenig Fawr and Llyn Arenig Fach) are all minimum values as the absence of bathymetric charts did not allow ice-depths to be calculated over the lakes. Depths over these lakes represent the difference between the ice surface height and the lake surface height. Nevertheless, these values are significant as they represent depths at which firn can be transformed into ice, with 30 metres being generally accepted as the critical thickness at about which firn becomes glacier ice (Embleton and King 1975, pp.143).

The ELAs of the glaciers were 503 metres, 470 metres, 511 metres and 531 metres at Llyn Lliwbran, Llyn Arenig Fawr, Llyn Arenig Fach and Cwm Gylchedd respectively, averaging at about 504 metres. This compares with a mean of 600 metres for north Snowdonia (Gray 1982), 540 metres for the English Lake District (Sissons 1980a) and 700 metres for the Macgillycuddy's Reeks at a similar latitude in southern Ireland (Anderson et al 1998). The Aran and Arenig glaciers therefore had low ELAs in comparison although in north Snowdonia the Loch Lomond Stadial ELAs ranged from 410-815 metres (Gray 1982).

5. PALAEOCLIMATIC INFERENCES

5.1. Local controls on glacier development

During the Loch Lomond Stadial the climate in North Wales was one which was marginal to glaciation (Gray and Coxon 1991). In such conditions localised glacier development is often the result of certain local controls, which combine with the prevailing climate to enable glacier development. Although local controls were probably numerous and complex, dominant local controls would have included wind-blown snow, avalanche and direct insolation. Wind-blown snow and avalanche representing accumulative controls and insolation an ablative control. The reasons why the Aran and Arenig glaciers formed at the sites that they did can be explored with reference to the calculated local controls shown earlier in Tables 4.1. and 4.2. Also, the influence of local controls on ELA variation between the four glaciers can be examined.

All four glaciers had aspects within the northeastern quadrant of the compass with the Cwm Gylchedd glacier having the most northerly aspect of 25° and the Llyn Lliwbran glacier having the most easterly aspect of 88°. Insolation factors, combining glacier aspect and slope, illustrate that the Cwm Gylchedd glacier absorbed the least solar radiation during the ablation season (insolation factor 10.9) followed by the Llyn Arenig Fach glacier (insolation factor 12.1) and the Llyn Arenig Fawr glacier (insolation factor 13.5). The Llyn Lliwbran glacier absorbed the most solar radiation of the four glaciers (insolation factor 13.7). Collectively, when one considers that the insolation factor for a horizontal glacier at this latitude is 14.7, all four glaciers occupied sites favourable in terms of insolation. However, insolation would appear to have little influence in controlling the ELAs of the former glaciers as the glacier with the highest ELA, at Cwm Gylchedd, is in the most favourable location with respect to solar radiation absorption. Also, the two lowest glaciers, at Llyn Arenig Fawr and Llyn Lliwbran, are in the least favourable locations with respect to incoming solar radiation. As an increase in insolation and a decrease in altitude implies greater ablation it must follow that the Llyn Arenig Fawr and Llyn Lliwbran glaciers had greater accumulative input than the Cwm Gylchedd glacier in order to offset such increased ablation.

Wind blown snow can significantly add to the accumulative input to a glacier. Wind would have been a particularly important transportive agent during the Loch Lomond Stadial due to the vigourous interaction of polar and maritime air masses resulting in generally more stormy conditions (Sissons 1980b). The transport of falling and fallen snow and its subsequent accumulation is supported by numerous workers (e.g. Manley 1959, Sissons 1980b, Cornish 1981, Dahl et al 1997).The relative importance of snowblow input in the Aran and Arenigs can be assessed with reference to the D/A ratios calculated for the SW and SE catchments displayed in Table 4.2.

At 3.4 the Cwm Gylchedd glacier had the largest D/A ratio for the SW and SE catchments. This value indicates that under the assumption of the dominant synoptic conditions described in Section 4.1., the potential snowblow catchment was over three times the accumulation area of the glacier it supplied. It is therefore highly likely that in addition to low insolation, the Cwm Gylchedd glacier owed its existence largely to the influence of snowblow. The Llyn Lliwbran glacier had the next highest D/A ratio of 2.9, followed by the Llyn Arenig Fach (2.62) and the Llyn Arenig Fawr (1.85) glaciers. All of the glaciers therefore had snowblow catchments significantly larger than their accumulation areas. This is largely a result of topography. The three Arenig glaciers (Arenig Fawr, Arenig Fach and Cwm Gylchedd) all lie in the lee of plateau surfaces and the Aran glacier at Llyn Lliwbran lies in the lee of a broad rounded ridge.

Avalanche ratios are low for all four glaciers. The highest value was for the Llyn Lliwbran glacier, although at 0.5 indicates that the potential avalanche area was only half that of the accumulation area it supplied. These low values are the result of the topography noted in the previous paragraph. The glaciers lay in the lee of plateau surfaces or broad ridges and tended to largely fill their cwms. In any case, snow available for avalanche is often the result of wind accumulation on lee side slopes (Male 1980, pp. 233) and is therefore more or less accounted for in the snowblow D/A ratios. It can therefore be concluded that avalanching was not a major influence on accumulation and not a significant local control.

The discussion above implies that variations in the equilibrium line altitudes of the four glaciers cannot be explained with reference to local factors. The highest glacier at Cwm Gylchedd (531 m) occupied the most favourable location with relation to insolation and snowblow. Theory would suggest that this glacier should have had the lowest altitude assuming that snowblow and insolation were the dominant local accumulative and ablative factors. The lowest glaciers at Llyn Arenig Fawr (470 m) and Llyn Lliwbran (503 m) occupied the least favourable sites with relation to insolation and in the case of the Llyn Arenig Fawr glacier, snowblow input also. These lower glaciers must have had greater accumulative input than the Cwm Gylchedd glacier in order to offset the increased ablation via insolation and lower altitudes. At Llyn Lliwbran the high snowblow ratio of 2.9 suggests that such ablation could have been offset by snowblow input although this would have still been less than that at Cwm Gylchedd where the snowblow ratio was 3.4. Assuming ablation is well accounted for by insolation and a linear altitudinal temperature gradient existed, greater accumulation must have occurred via mechanisms other than snowblow and avalanche.

This spatial varaition in accumulation is likely to have been due to precipitation differences between the sites of the former glaciers and also differences in cloud cover.

The Llyn Lliwbran and Llyn Arenig Fawr glaciers existed on the slopes of the highest mountains in the area, Aran Benllyn (885 m) and Arenig Fawr (854 m). The Cwm Gylchedd and Llyn Arenig Fach glaciers existed on the slopes of lesser mountains, Carnedd-y-Filiast (669 m) and Arenig Fach (689 m). Precipitation is likely to have been higher at the sites of the Llyn Lliwbran and Llyn Arenig Fawr glaciers than at Llyn Arenig Fach and Cwm Gylchedd. Indeed, modern meteorological maps support this assumption with annual precipitation reaching 2500-3000mm around Aran Benllyn and Arenig Fawr and only 1500mm at Cwm Gylchedd and 2000mm at Arenig Fach (Meteorological Office 1977). Similarly it is very likely that localised cloud cover would have been greater over the large mountain masses of Aran Benllyn and Arenig Fawr than over Arenig Fach and Carnedd-y-Filiast. The lowest precipitation at Cwm Gylchedd was also likely to be due to an inland precipitation gradient as this was the most easterly and inland glacier.

Greater precipitation and more prevalent cloud cover at Llyn Lliwbran and Llyn Arenig Fawr allowed the glaciers here to develop at lower altitudes than at Llyn Arenig Fach and Cwm Gylchedd even though insolation was less favourable. The far greater snowblow input and low insolation at the highest glacier, Cwm Gylchedd, was probably offset by lower precipitation than the other glaciers including Llyn Arenig Fach which was lower but with lesser snowblow and less favourable insolation. The reasons for this difference in precipitation is related to the altitude and mass of the mountain masses as well as an inland precipitation gradient.

5.2. Ablation Season Temperatures

Former ablation season (May - September) temperatures can be derived from reconstructed glaciers using recent glacier-climate relationships as an analogy for past glacier-climate conditions. Sutherland (1984b) noted that mean summer (ablation season) temperatures (**t**) on modern glaciers is closely related to accumulation at the equilibrium line, which in turn approximates average accumulation (**A**) over the whole glacier. The form of this relationship has been established for ten Norwegian glaciers (Liestøl in Sissons 1979b, Sutherland 1984b) and has been shown by Ballantyne (1989)to correspond to the following regression equation:

$$\mathbf{A = 0.915_e^{0.339t}\ (r^2 = 0.989, P < 0.0001)} \qquad \text{(Equation 5.1)}$$

where **A** is in metres water equivalent and **t** is in °C (r is a measure of correlation, r^2 is the coefficient of determination and P is a measure of probability). This relationship is shown in Figure 5.1. The positive correlation between these variables for different glaciers reflects the fact that higher levels of mass turnover at the ELA require higher ablation and thus higher summer temperatures to balance the annual mass budget. This relationship has also been demonstrated by Loewe (1971) and Ohmura et al (1992) and is of global application (Nesje and Dahl 2000).

It is important to note that when accumulation (A) in Equation 5.1. is taken as entirely the winter precipitation, this is only of use on plateau glaciers unaffected by local accumulative inputs such as windblow and avalanche (Ballantyne 1989, Dahl et al 1997). The temperature-precipitation ELA (TP-ELA) can therefore be used synonymously with the lowest altitude for 'instantaneous glacierisation' on a plateau, as defined by Ives et al (1975). Cirque glaciers, like those in the Aran and Arenigs, receive far more winter accumulation than is suggested by merely precipitation input,

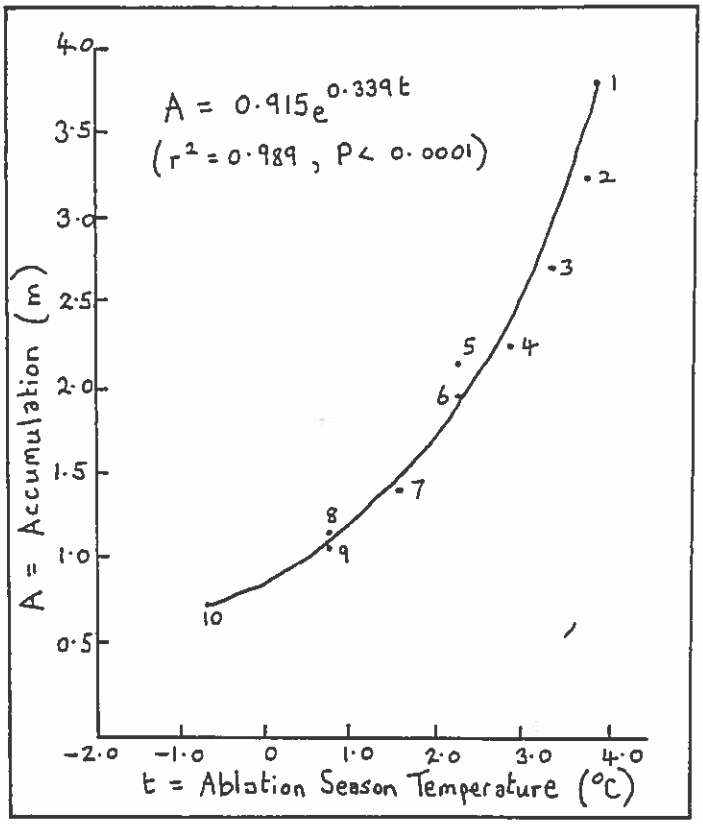


Figure 5.1. Non-linear relationship between accumulation at the ELA and ablation season temperatures for ten Norwegian glaciers (1. Alfotbreen; 2. Engabreen; 3. Folgefonni; 4. Nigardsbreen; 5. Tunsbergdalsbreen; 6. Hardangerjokulen; 7. Storbreen; 8. Austre Memurubreen; 9. Hellstugubreen; 10. Grasubreen). (The curve is modified from Liestøl in Sissons (1979b, Fig. 3) and Sutherland (1984, Fig. 4))

due to the leeward accumulation of windblown snow and avalanching (the former of most importance in the Arans and Arenigs). The temperature-precipitation-wind ELA (TPW-ELA) corresponds to the mean ELA on cirque glaciers (see Figure 5.2.) and is regarded as a measure of the combined influence of regional precipitation and local leeward accumulation of windblown snow and avalanche at the ELA (Dahl and Nesje 1992, Dahl et al 1997). The incorporation of windblown snow and avalanche input into the accumulation variable (A), in Equation 5.1. allows a more accurate estimation of temperature than if only winter precipitation were accounted for. Quantification of local accumulative inputs also allows temperature calculations to be made for individual glaciers, important where the sample of reconstructed glaciers is small.

Dahl et al (1997) propose that a rough estimate of total accumulation can be derived using the D/A ratio in conjunction with precipitation values. As noted in Chapter 4 the D/A ratios for the reconstructed Aran and Arenig glaciers are 2.9 at Llyn Lliwbran, 1.85 at Llyn Arenig Fawr, 2.62 at Llyn Arenig Fach and 3.4 at Cwm Gylchedd. This suggests that the TPW-ELA at Llyn Arenig Fawr was closest to the contemporaneous TP-ELA and thus the most representative for palaeoclimatic reconstruction. Before the D/A ratio for the Llyn Arenig Fawr glacier can be used to calculate total accumulation, values of precipitation during the Stadial need to be established.

The present precipitation at the site of the Llyn Arenig Fawr glacier is around 2500mm.

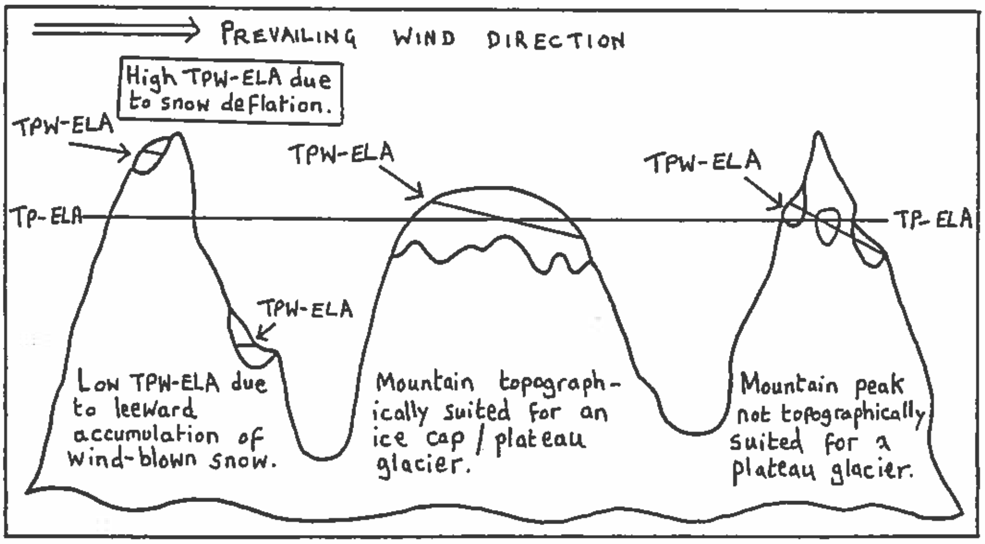


Figure 5.2. Idealized diagram illustrating the effect of leeward snowblow accumulation on the TP-ELA. Note that the TPW-ELA occurs at cirque glaciers while the TP-ELA may be regarded as synonymous with the altitude for 'instantaneous glacierization' at a plateau. (From Dahl and Nesje 1992, Fig. 2)

During the Loch Lomond Stadial, lower temperatures would have caused saturated air to hold less moisture than at present thus reducing the precipitative potential of the air. However, vigourous interaction of competing air masses would have tended to increase precipitation. On the assumption that these tendencies roughly cancelled each other out, Sissons (1980b) argued that precipitation amounts during the Stadial were similar to that of today. This is however open to debate as several workers have advocated a drier Stadial. In Britain this has been argued on the basis of high *Artemisia* percentages in Stadial pollen sequences. Birks and Mathewes (1978) interprete high *Artemisia* pollen assemblages as a reflection of an arid climate. However, although a number of *Artemisia* species are considered to be characteristic of xeric habitats, Moore (1980) argues that identification to genus level alone is insufficient evidence for arid conditions. Palaeohydrological data from The Netherlands and Belgium indicate that the later part of the Younger Dryas was significantly more arid than the early period prior to 10500 yr BP (Bohnke et al 1988, Cleveringa et al 1988) supporting palynological of xeric environments due to the presence of *Artemisia* (e.g. Tipping 1991). Given this evidence, it is likely that the Loch Lomond Stadial glaciers in the Aran and Arenigs developed and reached their maximum by 10500 yr BP after which time greater aridity led to their retreat. This would not necessitate any adjustment of Sissons' (1980b) theory regarding precipitation and thus levels can be assumed to have been similar to that of today at the time of maximum glacier extent.

Assuming that 20% of total precipitation was as summer rain (Manley 1959, Sissons and Sutherland 1976) a value of c.2000mm can be used as the winter precipitation input for the Llyn Arenig Fawr glacier. By multiplying this value by the D/A ratio for this glacier of 1.85, a reasonable estimation of the combined winter precipitation and local snowblow can be made. For the Llyn Arenig Fawr glacier this gives a total accumulation value of c.3700mm. As this is a largely estimated value, a reasonable range of total accumulation could therefore cover 3700mm ± 500mm (3200mm - 4200mm). This likely range of values can now be substituted into Equation 5.1. allowing an approximation of mean ablation season temperatures to be calculated.

For the Llyn Arenig Fawr glacier, the temperature at the ELA calculates at 3.7 - 4.5°C or 4.1 ± 0.4°C. Using a lapse rate of 0.6°C/100 metres this range of values equates to 6.9 ± 0.4°C and represents the mean sea-level ablation season temperature. The equivalent July temperature would be 8.4 ± 0.4°C. As this estimated range relates to the ELA of glaciers in equilibrium, when temperatures are likely to have ameliorated slightly, mean July temperatures may have been at the lower end of this range. Even so, the generous range of estimation given here would make any adjustment superfluous (Details of these temperature calculations can be found in Appendix 1).

The reconstructed temperature range for July of 8.4 ± 0.4°C compares well with that calculated for north Snowdonia of 8.5 - 9°C by Gray (1982) and the figure of 9°C for central England and North Wales derived from Coleoptera studies (Coope 1977). The temperature value is also substantially higher than that derived for sites further north such as 7.5°C for the Lake District (Sissons 1980a), 6°C for the southeast Grampians (Sissons and Sutherland 1976) and 6.3°C for the Isle of Skye (Ballantyne 1989), as would be expected at the lower latitude of the Aran and Arenig Mountains.

5.3. The Loch Lomond Stadial as an analogy for earlier Devensian ice-sheet build up

It is likely that the Arenig Mountains, and to a lesser extent the Aran Mountains, played a major role in earlier Devensian ice-sheet build up in Wales. The development of cirque glaciers represents the embryonic phase of major ice-sheet build up. Ice-caps would have built up rapidly once the snowline fell below that producing cirque glaciation (Evans 1999). Ice-caps would have developed over north Snowdonia, the Arenigs and the Brecon Beacons and eventually coalesced to form the Welsh ice-sheet. Subsidiary ice-caps may have also formed over isolated peaks like Cadair Idris.

It is possible that the ice-sheet was spawned from a bigger ice-cap over the Arenigs than over the mountains of north Snowdonia or the Brecon Beacons. In north Snowdonia, cirque glaciers would have reached lower ground, and thus suffered greater ablation, rapidly due to the sharp topography of the region. The glaciers would have also been susceptible to calving in the nearby waters of Tremadoc Bay, Caernarvon Bay and Conwy Bay. In the Arenigs, cirque glaciers would have expanded onto high-level tablelands without great loss in altitude and thus little increase in ablation. As such, plateau and piedmont glaciers would have developed here before they did in north Snowdonia from which an ice-cap would have developed rapidly. The conditions in this scenario are similar to those for ice-sheet growth suggested by Barry et al (1975). In the Brecon Beacons, the southerly latitude and high altitude of Loch Lomond Stadial glaciers effectively removes the area as an initial source of the Welsh ice-sheet. Here a smaller ice-cap is likely to have developed later and eventually coalesced with the southward advancing Arenig ice-cap which had by now coalesced with the north Snowdonian ice-cap.

The possibility of the Arenig region as the initial source of the Devensian Welsh ice-sheet is further supported by the low ELAs of the Stadial glaciers. The mean ELA of 35 north Snowdonian glaciers was 600 metres (Gray 1982) compared with just over 504 in the Arans and Arenigs. Of the north Snowdonian glaciers, only 11 (30%) had ELAs lower than that of the highest Arenig glacier, at Cwm Gylchedd (531 m) and only 4 (11%) had ELAs lower than that of the lowest Arenig glacier at Llyn Arenig Fawr (470 m). This is significant because if the Arenig glaciers could develop at a lower altitude than those in north Snowdonia during earlier Devensian ice-sheet build-up then the Arenig ice could rapidly expand out of their cwms onto the surrounding high-level plateaux. This scenario is interesting although is unfortunately rather speculative without further evidence as well as detailed modelling. It does however provide a thought provoking conclusion to the thesis.

6. CONCLUSIONS

1. Detailed field survey and geomorphological mapping in the Aran and Arenig Mountains, north Wales, has revealed evidence for at least four sites of former localised glacier occupation. Three former glaciers were mapped in the Arenig Mountains and one in the Aran Mountains. Collectively, the glaciers occupied an area of 1.86 km^2 and had a mean equilibrium line altitude of 503 metres.

2. Pollen stratigraphic analysis and periglacial evidence supports a Loch Lomond Stadial age for the former glaciers. The pollen stratigraphy of infilled lake basins inside and outside of mapped glacier limits shows that sedimentation inside of the former glacier limits is entirely Flandrian whilst outside the typical tripartite Lateglacial sequence exists. The 'freshness' of the local moraines and the absence of Lateglacial periglacial activity within the mapped glacier limits also suggests that the local glaciers existed during the Loch Lomond Stadial of c.11 - 10,000 yr BP.

3. The Aran and Arenig glaciers are likely to have grown and developed as a result of local factors such as low insolation and snowblow input in addition to the prevailing climate. However, when comparing the four glaciers, while local factors were important they were not the dominant control on ELA variation between the glaciers. This was primarily controlled by precipitation differences between the sites of the glaciers as well as probable differences in localised cloud cover during the Stadial.

4. By analogy with modern glacier-climate relationships in Norway, a mean Stadial sea-level July temperature was calculated at approximately 8.4 ± 0.4°C. This compares well with other temperature reconstructions for North Wales based on similar palaeoglaciological reconstructions (Gray 1982) and also Coleopteran studies (Coope 1977, Atkinson et al 1987).

5. It is proposed that the Arenig Mountains were an important centre for ice-sheet initiation in earlier Devensian glaciations, assuming that the Loch Lomond Stadial provides an analogy to the early stages of ice-sheet build up. The low ELAs of the Stadial glaciers relative to north Snowdonia and the nature of the surrounding topography support a tentative model of rapid ice-sheet development in the area.

APPENDIX 1

Ablation Season (Summer) Temperature Calculations.

Glacier One was used in these calculations as it had the lowest D/A ratio suggesting that the TPW-ELA was closest to the contemporaneous TP-ELA and thus most representative for palaeoclimatic reconstruction.

Precipitation: 2,500mm

Winter precipitation: (80%) (**P**): 2,000mm

D/A ratio (**D**): 1.85

TOTAL ACCUMULATION = D x P = 1.85 x 2,000 = 3,700mm

Estimated range of possible values: **3,700 ± 500mm**

Substitute this range of values into the following regression equation (Ballantyne 1989):

$\mathbf{A = 0.915_e^{0.339t}}$ ($r^2 = 0.989$, $P < 0.0001$)

Where A = total accumulation in metres water equivalent
t = temperature ° C

r is a measure of correlation
r^2 is the coefficient of determination
P is a measure of probability

For **3,200mm**:

$3.2 = 0.915_e^{0.339t}$

$t = 1/0.339\ \mathit{ln}\ (3.2/0.915)$

$t = 3.7°C$

For **4,200mm**:

$4.2 = 0.915_e^{0.339t}$

$t = 1/0.339\ \mathit{ln}\ (4.2/0.915)$

$t = 4.5°C$

Likely range of ablation season temperatures at the ELA of Glacier One (470m a.s.l.): **4.1 ± 0.4°C**

Sea-level equivalent (using a lapse rate of 0.6°C/100m): **6.9 ± 0.4°C**

Sea-level July equivalent (± 1.5°C): **8.4°C ± 0.4°C**

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