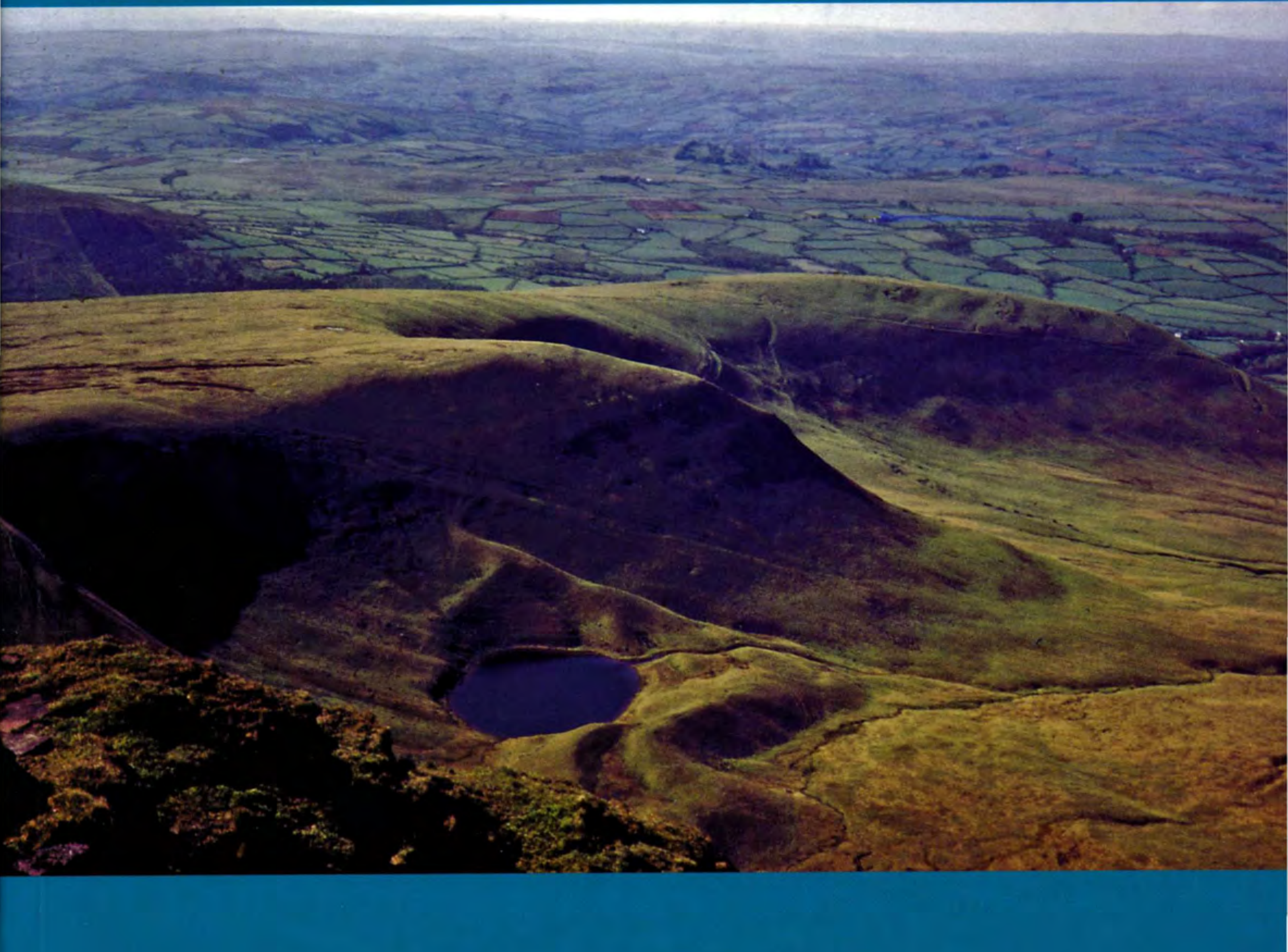
The glaciations of Wales and adjacent areas



Edited by Colin A. Lewis and Andrew E. Richards

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Front cover illustration: Llyn Cwm Llwch from the summit of Pen y Fan, Brecon Beacons. The lake is enclosed by a moraine formed when a small cirque glacier occupied the shaded area below the backwall of Cwm Llwch and extended as far as the outer moraine, little more than 11,000 years ago. Glaciation has left many imprints on the landscapes of Wales and adjacent areas, from features associated with small cirque glaciers as at Cwm Llwch, to those caused by major ice-sheets, as in North Wales and the Cheshire-Shropshire Plain. (Photo: Colin A. Lewis)

Contents

Preface

The landscapes of Wales and adjacent areas have been profoundly influenced by glaciation. The rugged mountains of Snowdonia, for example, owe much of their beauty to the effects of glacial erosion. The lowlands of Herefordshire, with their gentle landscapes and almost garden-like appearance, owe much to the effects of glacial deposition, as ice sheets from mid-Wales melted and deposited rich fertile sediments on the underlying bedrock. On the western side of the Irish Sea, in County Wexford, the hummocky kettled topography of the Screen Hills results from deposition due to melting of an ice sheet that occupied the Irish Sea basin during the Late Quaternary. Further north, in County Wicklow, impressive gorges cut through rocky uplands, such as The Scalp (near Enniskerry) and The Glen of the Downs (near Delgany), are the remains of glacial melt-water channels. Similar channels exist on the eastern side of the Irish Sea, as near Talybont, between Aberystwyth and Machynlleth.

Much attention has been paid to the origins of the Welsh landscape and, especially, to its geomorphology. In 1960, for example, Eric Brown published *The relief and drainage of Wales. A study in geomorphological development*. A century earlier, in 1860, Andrew Ramsay published a booklet on *The old glaciers of Switzerland and North Wales*, in which he remarked on the significance of glacial erosion: 'all glaciers must deepen their beds by erosion'.

In 1970 a group of eleven Geographers combined, under the leadership of a young Welshman then resident in Ireland, to publish *The glaciations of Wales and adjoining regions*. 'This book depicts the known glaciations not only of Wales but also of the Severn valley, South West England and the south and east coasts of Ireland', proclaimed the dust-jacket blurb. Three and a half decades have passed since that book appeared and there have been many advances in the knowledge of glacial events. Consequently a new group of fifteen scientists, under the leadership of the same but older Welshman, now resident in South Africa, and of a much younger geomorphologist from Herefordshire, have combined to produce the present book. Unlike the previous text, little attention is paid to the south coast of Ireland, since that area is discussed in *The Quaternary history of Ireland*, edited by Kevin J. Edwards and William P. Warren in 1985.

The authors thank all who have made this book possible: those who encouraged their interests in landscapes and, in particular, in glacial geomorphology; the funding agencies that have supported their research; the cartographers who prepared the final Figures; the Quaternary Research Association; and, of course, Logaston Press for publishing this text. The editors especially thank all the contributors for preparing their scripts and for their tolerance during the arduous editorial processes; the referees who read and commented on each chapter; and Mrs D. Brody, Miss B. Tweedie and Mrs J. Naidoo of Rhodes University for immense help with cartographic and secretarial matters.

The preface to the 1970 book stated that: 'Many details of the glaciations of Wales are still uncertain, whilst others are probably completely unknown.' That is still the case! In 1970 the editor wrote that:

'we have tried to present a synopsis of the knowledge available to us ... in the hope that it will encourage, and possibly guide, further research in the years ahead.' The aims of the present book are exactly the same.

Three of the authors of the 1970 book have also written for the present text (Bowen, Lewis, Worsley). They remember with affection and gratitude those who contributed to the 1970 book but who have passed to the life eternal, including Clifford Embleton, Edward Watson and Francis Synge. They also remember Frank [G. F.] Mitchell, Fred [F. W.] Shotton and many others with whom they had stimulating discussions about the Quaternary history of the English Midlands, Wales, South West England and the Irish Sea Basin. They remain fascinated with the study of the glacial history of Wales and its borderlands, the many remaining uncertainties of that history and the need to integrate the evolution of the area into a global overview. Like the other contributors to the present book, they hope that it stimulates a new generation of geoscientists and guides their researches in the years ahead.

Colin A. Lewis (Rhodes University) and Andrew E. Richards (University College, Worcester) (Editors), St David's Day, 2005.

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1 Introduction

by Colin A. Lewis

The glacial theory

On the sixteenth of October, 1841, William Buckland wrote in the visitors' book of the Goat Hotel, Beddgelert:

> Notice to Geologists.- At Pont-aber-glass-llyn, 100 yards below the bridge, on the right bank of the river, and 20 feet above the road, see a good example of the furrows, flutings, and striae on rounded and polished surfaces of rock, which Agassiz refers to the action of glaciers. See many similar effects on the left, or south-west, side of the pass of Llanberis (Davies, 1969).

Buckland (1784–1856) was, in 1841, Rector of the living of Stoke Charity in Hampshire and a Canon of Christ Church Cathedral, Oxford, having resigned his Readership in Geology at Oxford University in 1825 (Hunt, 1886). In the late 1830s he came under the influence of a young Swiss scientist named Louis Agassiz (1807–73).

In 1836 Agassiz had accompanied Ignace Venetz (an engineer), and Jean de Charpentier (a graduate of the Freiberg Mining Academy and director of the salt mines at Bex, in Switzerland), on a tour of the area around Bex and the Valois to examine evidence for the former existence of larger glaciers than those that then existed in the region. Agassiz accepted the ideas of his companions: that the glaciers had formerly been larger; and thereby accepted the similar hypotheses of Bernard Kuhn (a Swiss cleric, 1787); James Hutton the Edinburgh geomorphologist (1795); John Playfair the geologist (1802); and Jean-Pierre Perraudin, a professional mountaineer from Lourtier in southern Switzerland (1815). Agassiz, however, extended the existing hypotheses by stating, in his paper to the *Societè Helvètique* in July 1837, '... that the northern hemisphere from the North Pole down to the latitude of the Mediterranean and Caspian Seas had until recently been shrouded beneath a thick mass of glacier ice'. In other words, Agassiz conceived of '... a full-scale ice-sheet glaciation of continental proportions' (Davies, 1969).

Buckland had been aware of Agassiz' non-glacial researches since at least the early 1830s, when the two men corresponded with each other. In 1834, during Agassiz' first visit to Britain, he and Buckland became friends. Thus it was not surprising that in 1838 Buckland and his wife travelled to Switzerland to stay with Agassiz and inspect the evidence for his glacial theory. The evidence: of erratics, and of polished and striated rocks analogous to those at the margins of existing glaciers; did not initially impress Buckland. [Erratics are sediments that differ from the underlying bedrock. Examples of striations in Wales are shown on Fig. 7.2E.] After taking their leave of Agassiz the couple toured the Oberland and looked at real glaciers, and Buckland became a convert to the glacial theory. Furthermore, Buckland realised that similar features to those accepted as evidence of former glaciation in Switzerland also existed in Britain. He therefore realised that Britain had formerly been glaciated, although it was Agassiz, in a paper at the 1840 meeting of the British Association for the Advancement of Science, who formally made this statement (Davies, 1969).

Initially British scientists were sceptical of the new theory. When Agassiz read a paper to the Geological Society of London in the autumn of 1840 he was asked whether he really thought that Lake Geneva (which is very deep) had formerly been occupied by ice some 3,000 feet thick? The depth of that lake had been explained as due to catastrophism, to some earthquake or other event that had cleft the rocks, forming a deep trench that the lake subsequently occupied. Catastrophism had theological implications. William Shakespeare, for example, had written in the seventeenth century about the perfect sphere, and it was understood in clerical circles that God had made the world without blemish: a perfect sphere. The sins of humans, or so it was believed, had caused the perfect sphere to lose its symmetry and to develop hills and mountains, valleys and chasms, even deep hollows such as that occupied by Lake Geneva.

Agassiz replied that 3,000 feet '... must be regarded as a minimum thickness for the ice in the area' (quoted from Davies, 1969. Quotes from this source are shown hereafter as Davies, 1969). His questioner responded that the glacial theory was the 'climax of absurdity in geological opinions'. Another sceptic asked whether, if scratches were explicable as the work of former glaciers, all scratches should be explained as caused similarly?

Buckland, like Agassiz, was not deterred by scepticism. In 1841 he undertook field work in Snowdonia, finding ample evidence of former glaciation. On the coastlands around Snowdonia, as at Dinas Dinlle near Caernarfon, Buckland noted deposits that had been 'disturbed'. In the following year Charles Darwin examined parts of North Wales and became thoroughly converted to the glacial theory. 'The valley[s] about here', he wrote, '... must once have been covered by at least 800 or 1,000 feet in thickness of solid ice!' (Davies, 1969).

Glacial submergence

During the 1840 meeting of the British Association Agassiz had suggested that the melting of glaciers that had formerly covered large areas of the globe was responsible for rises in sea levels. Consequently, large areas were inundated and '... marine currents had imported boulders and other debris and deposited them widely over the continental surfaces' (Davies, 1969). The concept of a late glacial submergence soon became part of the general understanding of the glacial theory and was used to explain the origin of sediments that we now consider glaciofluvial. Buckland believed that the 'disturbed' deposits at Dinas Dinlle (Chapter Three), and elsewhere, had been caused by ice-bergs floating in the glacial submergence that had thrust into superficial sediments (*drift*) as they grounded, thereby disturbing them.

In 1831 Trimmer had described '... the diluvial deposits of Caernarvonshire' and noted that sands with marine shells exist at an altitude of 396 m at the Alexandra Slate Quarry on Moel Tryfan on the Lleyn peninsula(520560; Davies,1969). Buckland (1841) explained these, and other shell-bearing deposits in the Vale of Clwyd, as due to deposition under the waters of the glacial submergence.

The concept of 'glacial submergence', the submergence of large areas by rising sea levels due to glacier melting, was long lasting. In the 1850s Sir Andrew Ramsay (1814–91) claimed, on the evidence of deposits on Carnedd Dafydd and Carnedd Llywelyn, that North Wales had been inundated to a depth of 2,300 feet. (Ramsay became Director General of the Geological Survey in 1871). In 1863, when Sir Charles Lyell (1779–1875) visited the Moel Tryfan deposits, he believed that 'These shells show that Snowdon and all the highest hills which are in the neighbourhood of Moel Tryfan were mere islands in the sea at a comparatively late period' (Davies,1969). Even in 1910 Edward Hull, a former Director of the Geological Survey of Ireland, believed that the Moel Tryfan deposits were evidence of marine transgression.

The deposits at Moel Tryfan are now interpreted (Chapter Three) as sediments dredged from the floor of the Irish Sea basin by ice moving southwards down that basin and impinging on the North Wales uplands. The shells have been radiocarbon dated to about 33,700 years BP, indicating 'that the Lleyn

Peninsula was glaciated, at least in part, by extraneous ice after this date' (Foster,1970). They have amino acid ratios that suggest that they predate the Last Glacial Maximum (Bowen *et al.*, 2002).

Chronology

By the 1850s a number of scientists had suggested chronologies of the British Pleistocene. Among the first of these was Ramsay's (1852) division of the glacial period in North Wales into three stages:

i) the advance of valley glaciers from the mountains, forming striations and roches moutonnées on the valley floors.

ii) marine submergence as glaciers melted back into the uplands and sea level rose, as evidenced by the Moel Tryfan and other deposits. During this submergence Ramsay postulated that 'marine currents swept a stream of ice-bergs south-westwards across Anglesey, and the frequent grounding of these ice-masses caused the island's rocks to become heavily striated in a north-east to south-west direction. Further to the east a few rogue icebergs left the main stream and floated into the Welsh valleys to deposit Scottish and other northern erratics' (Davies, 1969).

iii) readvance of valley glaciers.

Other chronologies were produced for other areas of Britain. By the 1870s a tripartite stratigraphic division of the British drifts was generally accepted. At the base of the succession was the Lower Boulder Clay, supposedly deposited by an ice-sheet or by large valley glaciers; on top of this lay the Middle Sands and Gravels, which were thought to date to the supposed glacial submergence; finally, at the top of the sequence lay the Upper Boulder Clay, left by relatively small valley glaciers. Although it was believed that southern England had lain beyond the limits of glaciation nobody seems to have wondered how North Wales could be drowned by high sea levels of up to 2,300 feet while there was no such evidence of submergence in the former area!

The tripartite glacial chronology dominated thinking in the British Isles until the twentieth century, when new ideas resulting from the researches of Penck and Brückner (1909) in the Alps and surrounding areas suggested that the glacial sequence may have been more complicated.

Glacial erosion

Buckland, Darwin and many other scientists in the early days of the glacial theory concentrated on the recognition of striae and glaciated rock faces in order to establish that glaciation had formerly taken place, often overlooking the magnificent cirques, moraines, glacial breaches and other spectacular features that are the unmistakable evidence of the former work of ice. In 1859 Ramsay published an essay on 'The Old Glaciers of Switzerland and North Wales', which was published as a booklet in 1860. Ramsay stated '... that all glaciers must deepen their beds by erosion' and that they excavated rock basins. Over twenty years earlier an Irish engineer, Robert Mallet (1838), had claimed that '... the bed of a glacier is in continual process of degradation, or deepening by the resistless passage of these vast masses of ice and rocks over it'. It was Ramsay's booklet, however, that drew widespread attention to glacial erosion.

In 1860 Edward Hull suggested that many of the water-filled depressions in the English Lake District had been formed 'by the scooping action of glacier ice' and acknowledged his indebtedness to Ramsay for introducing him to the concept of glacial erosion. Nevertheless there was opposition to Ramsay's erosion concept.

In 1870 Sir Roderick Murchison (1792–1871), Director General of the Geological Survey and President of the Royal Geographical Society, wrote: '... where in any icy tract is there evidence that any glacier has by its advance excavated a single foot of solid rock? In their advance, glaciers striate and polish, but never excavate rocks'. Murchison's opposition was too late to elicit much support and, six years later when Judd questioned whether rock basins could have been caused by glacial erosion, many geologists wrote in support of the concept.

Subsequently a handful of geologists, led by Bonney and Garwood, argued that rivers have much greater erosive potential than glaciers and maintained that a covering of glacier ice actually protects the sub-glacial topography from experiencing as much erosion as it would have been subjected to if exposed to fluvial processes. Although this protectionist school was later scorned its proponents were not entirely incorrect, for *cold-based ice* (in which temperatures are below freezing from the surface to the base of the ice, Paterson, 2001) probably does have protective properties.

Wales played a major role in the development of the glacial erosion hypothesis, for it was research in the Principality, and especially in North Wales, that introduced the concept to the general scientific community and integrated it into the glacial theory.

Demise of glacial submergence

In 1862 Ramsay reconsidered his idea of a decade earlier: that ice-bergs had swept across Anglesey and, when grounding, striated the island's rocks. He noted that striations exist on the floors of deep hollows as well as elsewhere and now wrote that: 'An ice-berg that could float over the margin of a deep hollow would not touch the deeper recesses of the bottom'. In other words, he realised that a medium other than floating ice must have been responsible for the initiation of many striations. This medium, he now believed, had been '... sheets of true glacier-ice in motion, which moulded the whole surface of the country, and in favourable places scooped out depressions that subsequently became lakes.'

Ramsay was not the only, nor possibly the first, scientist to reject the theory of glacial submergence. Robert Jamieson, the Edinburgh scholar, argued in 1862 that 'striations, glacial polish, and roches moutonnées ... could only be the work of land-ice such as exists today in Greenland and Antarctica' (Davies, 1969). Geike (1863), Croll (1875) and other distinguished scientists also maintained that it was land-ice, and not glacial submergence, that had been responsible for the landforms and sediments that were so widespread in highland Britain and that were increasingly recognised as glacial. When Croll suggested in 1870 that shelly till in eastern Scotland, Orkney and Shetland was material dredged from the floor of the North Sea by glacier ice flowing westward from Scandinavia, the problem of shelly drifts was essentially resolved. Nevertheless Ramsay continued, into the 1880s, to believe that the shelly drifts of Moel Tryfan were due to glacial submergence. Others, however, were less blinkered and were increasingly aware of the findings that were beginning to be made in the Alps and in the upper reaches of the Danube valley and the valleys of its right bank tributaries.

Alpine glacial chronologies

In 1847 Collomb recognised evidence for two former glaciations of the Vosges Mountains, north of Switzerland (Bowen, 1978), then in 1856 Morlot discovered evidence of two former glaciations of Switzerland (Bowen, 1978). Thereafter increasing interest focussed on Switzerland and its surrounds as scientists sought to unravel the glacial chronology of that region. Their work peaked in 1909, when Penck and Brückner published *Die Alpen im Eiszeitalter*. In this book the authors presented evidence for what they argued were four different glaciations of the Alps, the main 'evidence' for which, they suggested, was a series of outwash terraces ('schotter') that could be traced upstream to end moraines in valleys of rivers south of Munich.

The basis for Penck and Brückner's argument came from mapping in Bavaria, especially in the valleys of the rivers Iller, Günz, Mindel and Lech, that Penck had first published in 1885.Subsequently mapping was extended to the valley of the River Würm and to other areas. This mapping showed that four outwash terraces could be traced up-valley to merge with end moraines. Each terrace and its associated end moraine, Penck and Brückner thought, evidenced one glaciation. The earliest glaciation, they reasoned, was evidenced by the outermost end moraine and the uppermost outwash terrace, which was associated with the outermost end moraine. Subsequent glaciations were evidenced by consecutively lower outwash terraces and by moraines. In other words their scheme was based mainly on morphology: the existence of end moraines and of terraces that could be traced back to those moraines.

Penck and Brückner's model of Alpine glaciations was essentially based on *morphostratigraphy*, which is the classification of bodies of sediment mainly on the basis of their surface morphology. They also examined the nature of the terrace deposits, concluding that the oldest (and highest) terraces were more weathered than the others. They maintained that the morphological evidence indicated that there had been four glaciations of the Alps, and they named these glaciations after river valleys in which they had studied the glacial and fluvioglacial landforms. The oldest glaciation, they thought, was the Günz, succeeded in turn by the Mindel, Riss and Würm. They also argued that the dissection of outwash plains down stream of end moraines, which transformed those outwash plains into outwash terraces, took place during interglacials. Their scheme therefore supposed that deposition occurred during glacials and erosion during interglacials and failed to consider that both processes could exist within glacials and/or interglacials.

The extent of weathering of terrace deposits led Penck and Brückner to suggest that the interglacial between the Mindel glaciation and that of the Riss was by far the longest of all interglacials that they recognised in Alpine Europe. They also suggested that the Riss glaciation lasted longer than the Würm because more sediments accumulated during the former glaciation. In some cases organic remains, that appear to be of interglacial character, are mixed with colluvial and scree deposits (such as the *Hottinger Breccia* near Innsbruck) and rest on terraces. These organic remains appeared to support the concept that outwash deposition was glacial and that erosion was interglacial.

In 1924 the German climatologist, Wladimir Köppen, and his son-in-law, Alfred Wegener (who introduced the concept of continental drift), published a book entitled *Die Klimate der Geologischen Vorzeit.* In it they included a graph by a great Serbian mathematician, Milutin Milankovitch, that showed how the intensity of summer sunlight (solar radiation) varied over the past 600,000 years (Fig. 1.1). Milankovitch correlated some of the low points on the graph with the four glaciations of the Alps as identified by Penck and Brückner. There was no geological proof for this correlation, but it became widely believed. Four years later, in 1930, Milankovitch published *Mathematical Climatology and the Astronomical Theory of Climatic Changes*. In this book he showed how the amount of solar radiation reaching the surface of the earth is influenced by the tilt of the earth on its axis, which takes place in a regular 41,000 year oscillation, and by the 22,000 year oscillation of the distance between the earth and

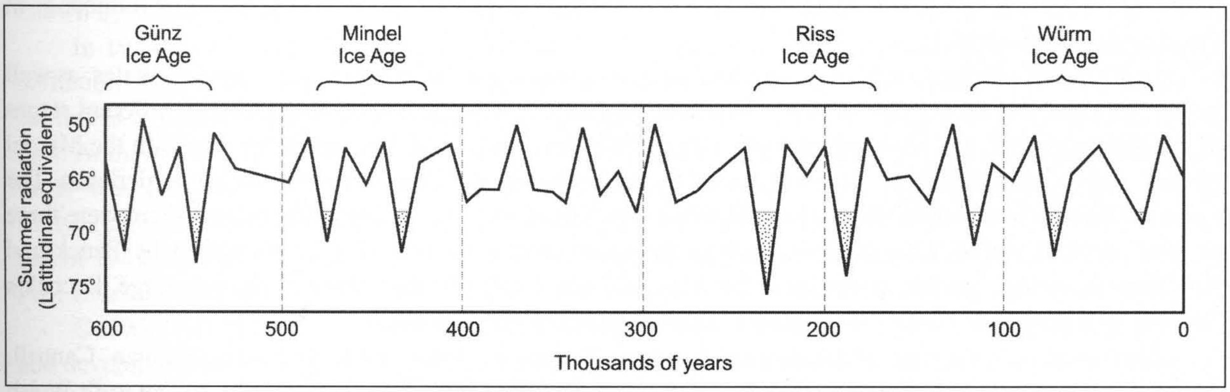

Fig. 1.1 Milankovitch's radiation curve for latitude 65°N was used by Köppen and Wegner to 'date' the classic glaciations of Alpine Europe. The radiation received 400,000 years ago at 65°N, for example, was equivalent to that now received at 67°N. Low points on the radiation curve were 'correlated' with glaciations. (Redrawn and simplified after Köppen and Wegener,1924)

the sun, as described in Chapter Two. Milankovitch thereby provided an astronomical explanation for the existence of glacials and interglacials and is honoured scientifically by the almost universal recognition of *Milankovitch cycles*.

The Alpine model, as Penck and Brückner's classification came to be called, was extended by the work of Eberl (1930) and others, who believed that there had been at least one pre-Günz glaciation: the Donau. The Günz, they suggested, had been divided into substages, as had some of the other glaciations. Eberl (1930), like Penck and Brückner, mapped the morphology of landforms, but he also examined the *lithostratigraphy* of terrace gravels. *Lithostratigraphy* is the organisation of strata into units based on their lithological characteristics, as discussed in Chapter Two.

By the beginning of the Second World War it was widely believed that the four or five fold model of Alpine glaciations must have been repeated elsewhere. There was little, if any, appreciation that the model itself, which was by then an established classic, might be seriously flawed and might even be incorrect and misleading.

Bowen (1978) wrote that 'Pleistocene time, as conceived by Penck and Bruckner [*sic*] ... is represented in the type area by glacial and fluvioglacial (outwash) deposits that only represent part of that time. A good deal of Pleistocene time is represented by erosional breaks in the stratigraphic record. ... interglacial time is not represented by sediments, but merely by inferred erosion during which the outwash schotter were dissected into terraces. ... not only do the sediments not represent the full time-span ... but the unconformities between each successive terrace probably conceal events lost to the record ... As such ... because it is primarily a morphostratigraphical model, it is inherently deficient.'

Wales and the classic Alpine glacial chronology

By the 1880s it was widely accepted that large areas of Wales, and not just North Wales, had been glaciated. In 1883, for example, Edgeworth David published 'On the evidence of glacial action in south Brecknockshire and east Glamorganshire'. David showed, using erratic and other evidence, that ice from the Brecon Beacons/Fforest Fawr uplands had passed down the valleys of the adjacent South Wales coalfield to debouch onto the Vale of Glamorgan around Cardiff. Twenty years later, in 1903/4, Howard noted the distribution of erratic material and of striae and concluded, correctly, that ice had flowed from northerly sources (which were the plateaux of Pumlumon and its southerly extensions) down the valleys of the Wye and Usk. Howard and David had written in terms of one period of glaciation, but after the publication in 1909 of Penck and Brückner's findings in the Alps some scientists presented evidence that indicated, at least to them, that there had been four periods of glaciation in Wales (the same number as in the classic Alpine model).

In 1925 Pocock suggested that there had been two major glaciations of at least parts of Wales, as well as 'two limited and recent glaciations'. 'It seems probable that these last correspond with different stages of the Würm period, the more general glaciation with the Riss period and the maximum with the Mindel period'. He thought that these were evidenced by terraces and drift deposits in the Welsh borderlands. The basis for Pocock's dating of the various deposits was flimsy in the extreme: just because there were three or four terraces in the Wye or other valleys in Wales, and a similar number identified by Penck and Brückner in valleys leading north from the Alps, did not mean that they were of the same age. Pocock's 'dating' was therefore based on supposition rather than scientific proof.

The work of a number of Geological Survey Officers in Wales, including Jehu, Strahan, Cantrill, Thomas, Dixon and O. T. Jones, was of greater scientific value than that of those who strove to fit Welsh evidence into the (flawed) Alpine model. They mapped deposits and phenomena carefully throughout much of Wales. Another geological surveyor, albeit of the Geological Survey, Ireland, W. B. Wright (1914), formalised the division of the glacial deposits of these islands into an Older Drift and a Newer

Drift. In other words, careful research in Wales and adjacent areas (including Ireland) had, by 1914, produced evidence for an older and a newer glaciation rather than for four glaciations as proposed by the Alpine model.

The delineation of glacial limits in and adjacent to Wales

A major advance in glacial studies came in 1929, when Charlesworth delineated 'The South Wales end-moraine', which he thought was the outermost limit of Welsh ice during Newer Drift times (*vide* Chapters Seven and Ten). Further delineations of the supposed limits of Welsh and Irish Sea ice were made subsequently, especially in the 1950s and thereafter (Bowen, 1981; Chapter Two). These were essentially based on a combination of morphological and lithological mapping. Mitchell (1960) suggested that there had been three glaciations of Wales and adjacent regions, although his map of the outermost limits of the most recent ice-sheet suggested that west Wales, from the uplands of the Lleyn peninsula south to Gower (apart from the fringes of Swansea Bay) lay beyond the ice-margin.

Detailed studies of glacial deposits and supposed glacial limits in various limited parts of Wales and its borderlands also appeared during the half-century following Charlesworth's seminal 1929 paper. Dwerryhouse and Miller (1930), for example, recognised a depositional feature in the northern area of the Hereford basin that they termed *The Kington-Orleton kettle moraine*. This feature is discussed in Chapter Nine. Elsewhere, as in Snowdonia, emphasis was placed on the identification of cirque moraines, as by Seddon (1957). Further south, in the upper Usk valley, moraines were linked to readvance and retreat stages of the last glacier to occupy that valley (Ellis-Gruffydd, 1977, *vide* Chapter Eight). Nevertheless new directions in Quaternary research were taking place in Wales, as elsewhere, and were to become increasingly influential.

Palynological studies

One of the most important forms of Quaternary research used in Wales from the late 1930s onwards has been *palynology*. This is the study of pollen and other spores, which had been pioneered in Scandinavia by Von Post in the early years of the twentieth century. Pollen grains from different plants have characteristics that enable scientists to identify the plants from which the grains are derived: oak (*Quercus*), elm (*Ulmus*) and so on. By identifying and counting the number of grains from each species, layer by layer stratigraphically, it is possible to gain an idea of what past vegetation used to be like. This enables scientists to reconstruct palaeoenvironments and to divide the sequential pollen record into zones of similarity, known as *pollen zones*.

In 1934 Knud Jessen, Professor of Botany at the University of Copenhagen, visited Ireland at the invitation of the Committee for Quaternary Research in that country in order to study pollen deposits. He continued to visit Ireland up to and including 1949 and worked there in collaboration with two Irish scientists: Anthony Farrington and Frank [G. F.] Mitchell (Lewis, 1984). The first paper resulting from these visits was published in 1938 and was written jointly by Jessen and Farrington, on 'The bogs at Ballybetagh, near Dublin, with remarks on late-glacial conditions in Ireland'. Further papers followed and that of 1949, followed by Mitchell's paper of 1956, established the pollen zonation of the Late Glacial and Post Glacial in Ireland.

Godwin and Mitchell pioneered pollen analysis in Wales with their 1938 paper on the stratigraphy and development of two raised bogs near Tregaron. Subsequently Godwin published two more important papers based on the analysis of pollen from various parts of Wales (1938b (with Newton), 1955). Professor Sir Harry Godwin (1901–85) was the founder of the Godwin Institute for Quaternary Research at the University of Cambridge, at which fundamental research into Quaternary history, and especially the dating of Quaternary events, has been undertaken.

After the end of the Second World War, Seddon (1957, 1962) published on the palynology of Late Glacial deposits in some of the cirque basins in Snowdonia, Bartley on a site in Radnorshire (1960), Trotman (1963) on evidence from parts of South Wales, P. D. Moore (*e.g.* 1966, 1968, 1970, 1972, 1978) on the vegetational history of mid-Wales, (in which he shed much light on Late Quaternary environments), J. J. Moore (1970) on the Late Glacial pollen sequence from Mynydd Illtyd, on the northern margins of the Brecon Beacons (a site that was later reworked by Walker, 1980, 1982), and Crabtree (1972) on Cors Geuallt in Snowdonia. These and other studies helped to show the complexity of Late Quaternary environmental changes and indicated the dynamic and ever-changing nature of climate. Some of them demonstrated that interstadial conditions (known as the Bølling–Allerød in northern mainland Europe; *e.g.* Bos *et al.*, 2001) existed after ice-sheets melted and before renewed climatic deterioration caused cirque glaciation to take place. Many palynological studies have been produced in recent decades, such as that by Robertson (1988) on the Brecon Beacons National Park region that is discussed in Chapter Eight.

The study of Coleoptera

Coleoptera are beetles and weevils. Beetles are very sensitive to climatic changes and different species of beetles exist at present under specific environmental conditions that may be related to temperatures and to precipitation. By identifying the remains of beetles in particular stratigraphic layers in Quaternary sediments, and then establishing the mutual climatic range of the beetles in individual layers, it is possible to establish the climatic conditions that existed when those animals lived. The *mutual climatic range* is that portion of the temperature range under which all the beetles in a particular stratum could have lived, and the portion of the precipitation range that is common to all the species in that stratum. Coleopteran remains have been discovered at a number of sites in North Wales and the Borderlands (*e.g.* Coope, 1972 (with Brophy), 1977: Chapter Three), Mathon in Herefordshire (Coope *et al.*, 2002; Chapter Nine), Llanilid in Glamorgan (Walker *et al.*, 2003; Chapter Ten) and from the Midlands (*e.g.* Coope and Sands,1966; Morgan, 1973; Chapter Five).

Subdivision of the glacial deposits

The lithology of many of the superficial deposits of Wales and adjoining regions was mapped, in the late nineteenth and early twentieth centuries, by such Surveyors of the Geological Survey as Strahan, Cantrill, Thomas, Dixon, O. T. Jones and Jehu. In other words, they studied the physical and chemical characteristics of the rocks and finer sediments that are incorporated within superficial deposits. In the Wicklow Mountains, on the western side of the Irish Sea, glacial and associated deposits were also examined lithologically by a geologist of the Geological Survey of Ireland: Anthony Farrington (1893–1973), who was later employed by the Royal Irish Academy. In a series of papers Farrington (1934, 1942, 1944, 1949, 1957, 1966) established the lithostratigraphy of those deposits (Davies, 1963).

In 1973 a special report of the Geological Society of London attempted to correlate different regions with a standard classification based on inferred climate change (*climatostratigraphy*) in East Anglia (Mitchell *et al.*, 1973). The second edition of this report adopted a formal *lithostratigraphical* classification (Bowen, 1999) which, in Wales and the Borderland, closely followed an earlier systematic review of the stratigraphy of Pleistocene deposits (Bowen, 1973, 1974). Chapter Ten, with its subdivision of superficial deposits in south Wales into lithostratigraphic formations, bodies of sediment that are mappable and that may be subdivided into members and beds, exemplifies the use of a formal Quaternary lithostratigraphy.

Periglacial studies

Deposits of angular stony rubble held in a finer matrix, which were sometimes referred to as 'rubble drift' or *head* (de la Beche, 1839; Prestwich, 1892; George, 1933) have long been known in southern England and Wales. These deposits commonly infill valley bottoms and mantle the lower part of hill slopes and are semi-stratified. Spurrell (1886) thought that head accumulated by 'the intermittent flowing, under its own weight, of a soil undergoing thaw, that is, in a viscous state'. By the end of the nineteenth century head was considered, by most geologists, to be a periglacial deposit. *Periglacial* environments are those that are non-glacial but are dominated by cold climatic conditions in which frost is common but in which ground is seasonally snow free (Washburn, 1973). Head deposits have been mapped, particularly in coastal areas of Wales and in south-west England, by employees of the Geological Survey since the latter part of the nineteenth century, but little interest was otherwise shown in periglacial deposits and landforms in Wales until the second half of the twentieth century.

In 1961 Albert Pissart, a Belgian geomorphologist, visited Wales and identified the remains of pingos in the Llangurig area of mid-Wales. *Pingo* is 'An Eskimo [*sic*., Inuit] word for a domed, perennial ice-cored mound of earth formed ... under permafrost conditions' (Whittow, 1984). *Permafrost* is a condition in which sediments below the ground surface are frozen for two or more consecutive years and commonly exists in periglacial environments.

Pingos melt as temperatures rise. This allows the inorganic sediments within the domed mounds, which were cemented by and incorporated within ice as the pingos formed, to sludge down the sides of the domes to form circular or sub-circular ramparts around the margins of the formerly active landforms. The remains of pingos are described in Chapters Seven and Eight.

Pissart published his findings in 1963, by which time he had interested Edward Watson, a geomorphologist in the Department of Geography at the University College of Wales in Aberystwyth, in periglacial studies. Pissart also identified fossil solifluction terraces in Wales (1963b). In the years that followed Watson provided firm foundations for periglacial studies in the Principality, writing on such periglacial sediments as *grèzes litées* (1965a), ice wedge casts (1981) and other periglacial structures (1965b), pingos (*e.g.* 1971), and nivation cirques with their associated landforms and sediments (1966, 1969). By the end of the 1960s periglacial studies were well established in Wales although it was not until the following decade that they were emphasised in Ireland (*e.g.* Mitchell, 1971; Lewis, 1979).

Sand and ice wedges form under specific environmental conditions and the casts of such wedges can be used to indicate former climatic conditions, as has been done for the Lower Severn Valley in Chapter Six. On a more extensive scale Huijzer and Vandenberghe (1998) have used ice wedge casts and other periglacial features to indicate climatic conditions in north-western and central Europe between ~72–13 ka, during the glacial stage known in the Netherlands, Scandinavia and the southern borderlands of the Baltic Sea, as the Weichselian.

The use of numerical dating

i) Radiocarbon dating

Nuclear research, especially during the Second World War, led to the appreciation that sediments could be dated by measuring the proportion of carbon atoms in an organic deposit. A radioactive form of carbon (radiocarbon) is produced in the atmosphere by cosmic rays. Radiocarbon is then absorbed into the bodies of living plants and animals but decays at a known and measurable rate once the plants/animals die.

Willard Libby, working at the University of Chicago, developed *radiocarbon dating* in the late 1940s (Imbrie and Imbrie, 1979). In the years that followed radiocarbon dating was applied to organic deposits in many parts of the world, including Wales and adjoining areas. As a result it was possible to construct a numerical timescale, in radiocarbon years, for geological events.

Radiocarbon dating was applicable to suitable deposits up to about 40,000 years old and now, with accelerator mass spectrometry (AMS), may be used to date suitable samples that are up to 50,000 years old. In 1970, when *The glaciations of Wales* was first published, some of the regional chapters did not contain a single date. Others, such as that on Pembrokeshire, contained 'radiocarbon age determinations' that some scientists regarded as questionable. The present book contains numerous radiocarbon dates, as in Chapters Three, Five and Eight. Many other numerical dating techniques now exist, such as uranium-series ages on stalagmites, thermoluminescence (TL), optically stimulated luminescence (OSL), electron spin resonance (ESR), amino acid dating of bivalves and gastropods and cosmogenic chlorine-36 rock exposure dating.

ii) Amino-acid dating

Aminostratigraphy is the correlation of stratigraphic units based on the ratios of particular amino acids preserved in the fossil protein of gastropods and bivalves. *Amino acid geochronology* relies on the age calibration of such ratios by, for example, radiocarbon, Uranium-series or other independent means. It has been applied to raised beach, terrestrial and shelly glacial deposits in Wales (Chapter Ten) and to raised beach deposits in south-west England (Chapter Eleven).

iii) Cosmogenic nuclide surface-exposure dates

Cosmogenic nuclide surface-exposure dates are quoted in Chapter Ten. Cosmogenic nuclides (^{3}He, ^{10}Be, ^{21}Ne, ^{26}Al, ^{36}Cl) are produced in rocks once they are exposed to cosmogenic ray bombardment at the surface of the Earth (*e.g.* Phillips *et al.*, 1994; Bowen *et al.*,2002). Their accumulation is a measure of rock exposure. Arthur's Stone, on Cefn Bryn in Gower, is a glacial erratic with a Chlorine-36 age of about 23,000 years, indicating that deglaciation in that part of South Wales took place at about that time (Chapter Ten). At Cwm Idwal in Snowdonia, Chlorine-36 ages of boulders on the surface of the outer Late Glacial moraine (Phillips *et al.*, 1994) indicate that ice build-up at that site occurred during the interstadial preceding the Younger Dryas (the Bølling–Allerød), when precipitation took place, (a considerable proportion of which must have been of snow), derived from a relatively warm North Atlantic Ocean (Bowen, 1999).

Oxygen isotope stratigraphy ($\delta^{18}O$)

The sedimentary record of the continents is incomplete because of erosion, whereas sediments accumulated more or less continuously on the floor of the deep open (pelagic) ocean and contain benthic and planktonic micro-organisms (mainly the Foraminifera). The geochemistry of these organisms, such as their oxygen isotope, carbon isotope, cadmium-calcium and magnesium-calcium content, contain valuable information about the chemistry of the oceans from which they secreted chemicals for their shells. Oxygen isotope stratigraphy (Emiliani, 1966), expressed as δ^{18}O, the ratio of ^{18}O to ^{16}O in fossil shells, has been used as a monitor of the changing isotopic composition of the oceans. During ice ages the evaporation of the lighter (^{16}O) isotope enriches continental ice-sheets, while residually enriching the isotopic composition of the ocean in the heavier (^{18}O) isotope. When deglaciation occurs, the ocean is enriched in the lighter isotope as the δ^{16}O enriched ice-sheets melt. An oxygen isotope stratigraphy from deep sea cores thus provides a record of the changing ice volume (glaciation) of the continents and, thereby, a record of Pleistocene glaciations, of which there were about 50 as well as a comparable number of interglacials back to some 2.5 million years ago (Bowen, 2004). These *oxygen isotope stages* are numbered backwards in time, with odd numbers denoting interglacials and even numbers denoting ice ages.

Correlation of oxygen isotope stages established from studies of ocean floor sediments with terrestrial events, such as the glaciations of Wales, is challenging but necessary if the global climatic system is to be understood. The deep sea isotopic record is tracked by astronomical calculations of the earth's orbital variations (the *Milankovitch cycles* that have already been discussed). Exactly how astronomical forcing of insolation changes is translated into climatic changes, and how millennial changes are superimposed on those of orbital origin, has yet to be established in causal terms. The estimation of sea surface temperatures is complicated and is usually done using a combination of oxygen isotopes, ecologic water masses characterised by combinations of plankton species and, recently, alkenone thermometry.

The ages of ice ages and interglacials, as established by oxygen isotope analyses, are calibrated by magnetic polarity through reference to the ages of major reversals of the Earth's magnetic field, as have been established on terrestrial lava flows (*e.g*. Shackleton and Opdyke, 1973).

Oxygen isotope stages are referred to in many of the following chapters, as in Chapters Three, Four, Six, Ten and Eleven, and their utility as palaeoclimatic indicators has helped scientists to realise that events in Wales and surrounding areas reflect developments in the global system. Such recognition has coincided with the development of a palaeoglaciological approach to glacial studies in and adjacent to the Principality.

Palaeoglaciology

In 1977, after publication of the first edition of this book, Boulton *et al.* attempted a reconstruction of the ice-sheet that covered much of Britain and Ireland during 'the last glaciation'. This reconstruction was based on 'flowlines derived from inspection of the glacial geology, to establish the ice surface topography' (Siegert, 2001) and proposed that the ice surface was some 1,200 m above modern sea level in North Wales, declining to under 400 m south of the Brecon Beacons, with large areas of south-west and south-east Wales being ice free. In 1985 Boulton *et al.* produced a revised reconstruction of the ice-sheet surface, indicating a thickness of little more than 250 m of ice in North Wales.

Lambeck (1993a, b) produced further models of the British and Irish ice-sheet, for 22,000 BP, 18,000 BP, 16,000 BP, 14,000 BP, 13,000 BP and 12,750 BP (BP means Before the Present, and the Present is regarded as 1950 AD). At 22,000 BP, according to the model, the ice-sheet had a surface of some 600 m in north central Wales but less elsewhere, thinning to 400 m by 18,000 BP. By 16,000 BP, or so the model indicated, only a small residual ice-cap existed in central Wales, remaining areas of the Principality being essentially ice free. Lambeck's model considered isostatic effects on the ice-sheet and concluded that by 18,000 years ago the Irish Sea was free of grounded ice, although geological and geomorphological evidence in north-east Ireland indicates otherwise (Clark *et al*., 2004). Isostatic effects include the depression of the land due to the weight of the overlying ice, and the consequent rebound when the ice melts and that weight, or part thereof, is removed.

In contrast to the large-scale glaciological approach of ice-sheet modelling, reconstructions were published for central south Wales (Bowen, 1980), while Shakesby and Matthews (1993) published a reconstruction of a small glacier in the Mynydd Du uplands west of the Brecon Beacons in South Wales, showing the interpolated contours on the glacier surface (Chapter Eight). They also discussed the size of the most recent cirque glaciers in the Brecon Beacons region and computed the Equilibrium Line Altitudes (ELA) of some of them. The *Equilibrium Line Altitude* is the altitude at which accumulation on the higher portion of a glacier is exactly balanced by ablation (melting) on lower areas. Once the ELA is known it is possible to indicate former temperatures and/or precipitation.

Carr (2001) presented a more developed glaciological approach in his investigation of the most recent glaciers in the central area of the Brecon Beacons. From the geomorphological evidence Carr calculated ELAs, ablation gradients, glacier mass balance and ice discharge (mass flux), velocity, ice deformation, basal slippage/subglacial deformation and glacier flow through basal slippage. From these calculations he proposed that four of the five supposedly most recent glaciers were 'relatively slow moving...with a significant component of basal slip ... and compare well with modern [small glaciers] in southern Norway'. Carr concluded that 'the geomorphological evidence ... reflect[s] the activity of small glaciers'.

In 2004 Jansson and Glasser used satellite imagery to map glacial lineations in northern Wales, which enabled them to reconstruct former palaeo-ice-flow systems. They were then able to calculate former ice-sheet surface profiles and indicate the corresponding subglacial thermal regimes. They concluded that 'Landform creation and reorganization beneath the [last] Welsh Ice Cap appears to have been limited ... possibly because the ice cap was predominantly cold-based', which harks back to the protectionist ideas of Bonney and Garwood a century earlier. They also concluded that 'northern Wales was covered by glacier ice reaching an altitude of 1000–1200 m a. s. l.'

Glaciological studies are at an early stage of development in Wales and surrounding areas and it is important that they are consistent with the established facts of glacial geomorphology, lithostratigraphy and age estimates for the events they purport to model.

Wales and the global system

Introduction

Although there was general appreciation, when *The glaciations of Wales* appeared in 1970, that events in Wales and adjacent regions correlated with those elsewhere, there was no discussion of the role of global systems in causing those events. There was, for example, no discussion of the origins of ice ages, or of the role of atmospheric gasses in climate change, nor of the importance of ocean currents or even of glaciological events (such as the melting or surging of North American and Greenland ice-sheets and the consequent discharge of armadas of ice-bergs into the North Atlantic, with a concomitant lowering of sea temperatures with associated atmospheric consequences). That was mainly because research into those topics had scarcely commenced. Results from the first deep-polar ice core, in Greenland, were published in 1969 (Dansgaard *et al.*, 1969, 1971). They heralded a new scientific field of study in which millennial fluctuations of climate in both Greenland and Antarctica, and their interconnections, are investigated (*e.g.* EPICA,2004).

i) The origin of ice ages

Changes in insolation caused by changes in the orbit of the Earth seem to be the primary pacemakers of climate change although the role of sub-orbital processes, especially at millennial timescales, is superimposed on these larger changes. Exactly how orbital variations are translated into climate change is still unknown, although it seems likely that variability in the irradiances of the Sun, as well as the circulations of the oceans and atmosphere, all interact with various feedback effects. The glaciations of Wales were influenced by modulation of the obliquity (41,000 years), by precessional (23,000 and 19,000 years) pacings and by eccentricity (100,000 years; Chapter Two). Variability on sub-orbital scales has also been important.

The role of plate tectonics in opening and closing seaways , as well as in the uplift of the Himalayas and Colorado Plateau (which influence global climates through the effects of high altitude areas on global atmospheric circulation), has also been important in preparing the global boundary conditions in which the glaciations of Wales have taken place (Raymo and Ruddiman, 1992; Ruddiman, 2000; Wilson *et al.*, 2000).

ii) Ocean Circulation

In 1987 Broecker suggested that there is a mass transfer of water globally as a result of salinity differences, resulting in the *Global Conveyor or thermohaline circulation.* Dense, saline water from the North Atlantic flows south and then east at depth into the Indian and southern Pacific Oceans. It then flows northward, in the Pacific, becoming warmer as it does so and becoming less saline. In the northern Pacific, around the latitude of Japan, the now warmed and less saline water starts to flow back near the surface, north of Australia, to the Indian Ocean and thence, around the Cape of Good Hope, into the Atlantic (Fig. 1.2). As the Global Conveyor flows northward through the Atlantic so it carries warmth into that Ocean. With progress northwards the Atlantic arm of the Global Conveyor becomes cooler and increases in salinity so that, between Scotland and Iceland, the current sinks below the surface and starts its return journey south and then east into the Pacific.

The Global Conveyor has many effects on climate, not only transferring heat around the globe but also, through releasing heat to the atmosphere in the North Atlantic, nurturing atmospheric water absorp-

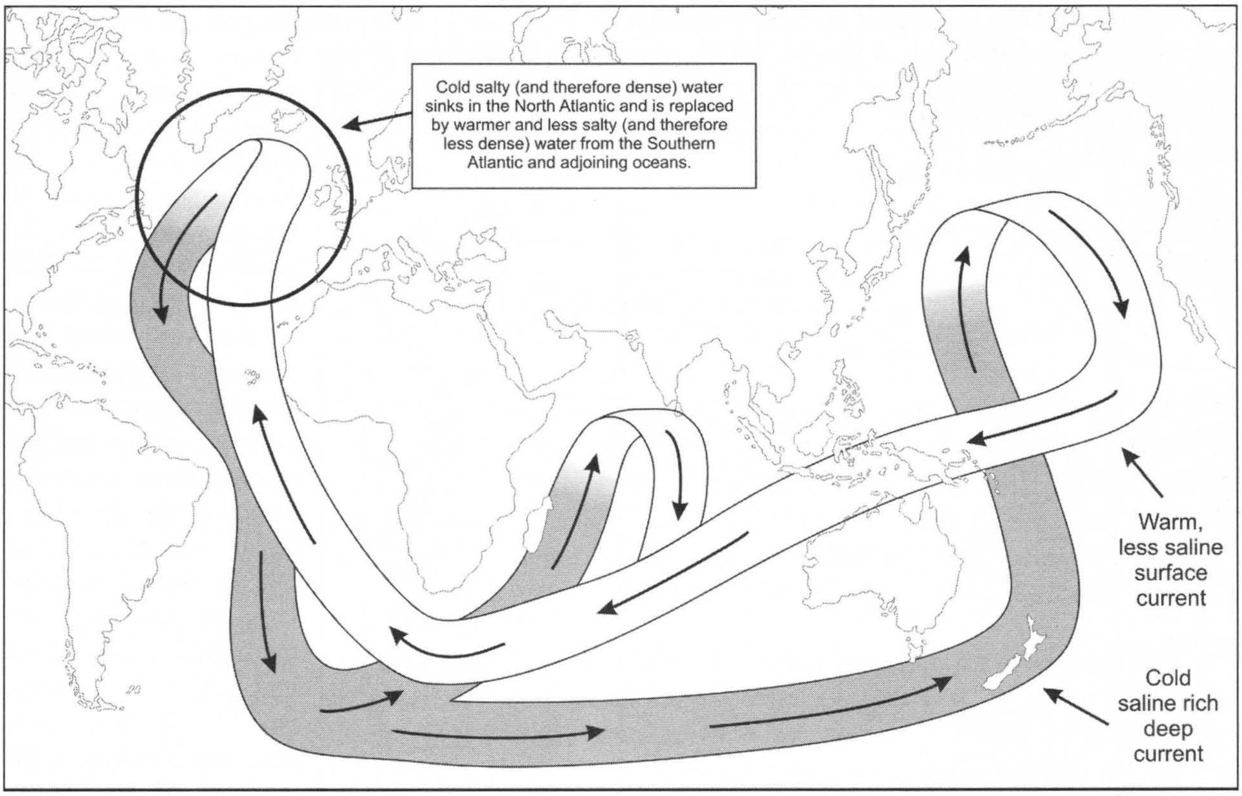

Fig. 1.2 The Global Conveyor, or thermohaline circulation, showing the main ocean currents that distribute heat from the Pacific and Indian Oceans northwards into the Atlantic Ocean. They return as cooler and more saline (and therefore denser and deeper) currents to flow south and then east into the Pacific Ocean where they flow northwards, warming and then returning to the Atlantic as less saline (and therefore less dense) surface currents

tion, cloud formation and, eventually, precipitation on adjacent continental areas influenced by the prevailing south-westerly winds that blow over that portion of the North Atlantic. The effectiveness of the Global Conveyor compared with atmospheric flux is debatable because, during transfer of heat poleward in the North Atlantic there is an atmospheric flux of 5 PW (one Peta Watt is 10^{15} watts/m^{-3}) compared with 1 PW through the ocean. There is also growing evidence that the Global Conveyor did not switch off during ice ages, as was previous thought (Raymo *et al.*, 2004). Atmospheric pulses of warmth derived from lower latitudes might have been the cause of at least some climatic changes that have been superimposed on larger changes of orbital origin (Cane and Clement, 1993; Bowen, 2000).

iii) Heinrich Events

There appear to have been a number of occasions in the past when the Global Conveyor was affected by the release of pulses of cold and fresh melt-water from decaying, or surging, ice-sheets in North America, Greenland and the islands of the arctic. Other cold freshwater pulses came from the (rapid) drainage of proglacial lakes as ice-barriers disappeared. Some pulses were associated with the release of ice-bergs into the Atlantic from the Laurentide (North American) and Greenland ice-sheets, forming *Heinrich Events* in which sediments from the melting ice-bergs were deposited over a broad swathe of the floor of the North Atlantic from Hudson Strait and Davis Strait to a zone west of the Celtic Sea/Bay of Biscay (Fig. 1.3). Another swathe of deposition from ice-bergs extended southwards from the vicinity of Spitsbergen towards Iceland and the Norwegian Sea.

Heinrich Events are named after Helmut Heinrich, the scientist who first reported their former existence (Heinrich, 1988; *vide* Andrews, 1998). The causes of Heinrich Events remain unclear, but may have included climatic amelioration, possibly associated with the release of pulses of warm water from the Caribbean and Gulf of Mexico, that caused ice-sheet decay. They may also have been the result of glaciological events, such as *surging*. Surging is when an ice-mass advances rapidly (MacAyeal, 1993; Clarke *et al.*, 1999). During a surge large volumes of ice are transferred to the terminal part of the ice-mass, possibly leading to a reduction in altitude of the ice-mass. Surging is the result of instability within the glacier system, possibly due to a build-up of stress as ice thickens in the accumulation area, leading to a loss of cohesion within the ice or between it and its bed (Patterson, 2001). Surging may therefore be glaciologically controlled in at least some cases, rather than a reflection of climatic change.

The freshwater pulses released into the Atlantic during Heinrich Events reduced the intensity of the Global Conveyor and, in some cases, caused it to turn southward considerably further south than it does

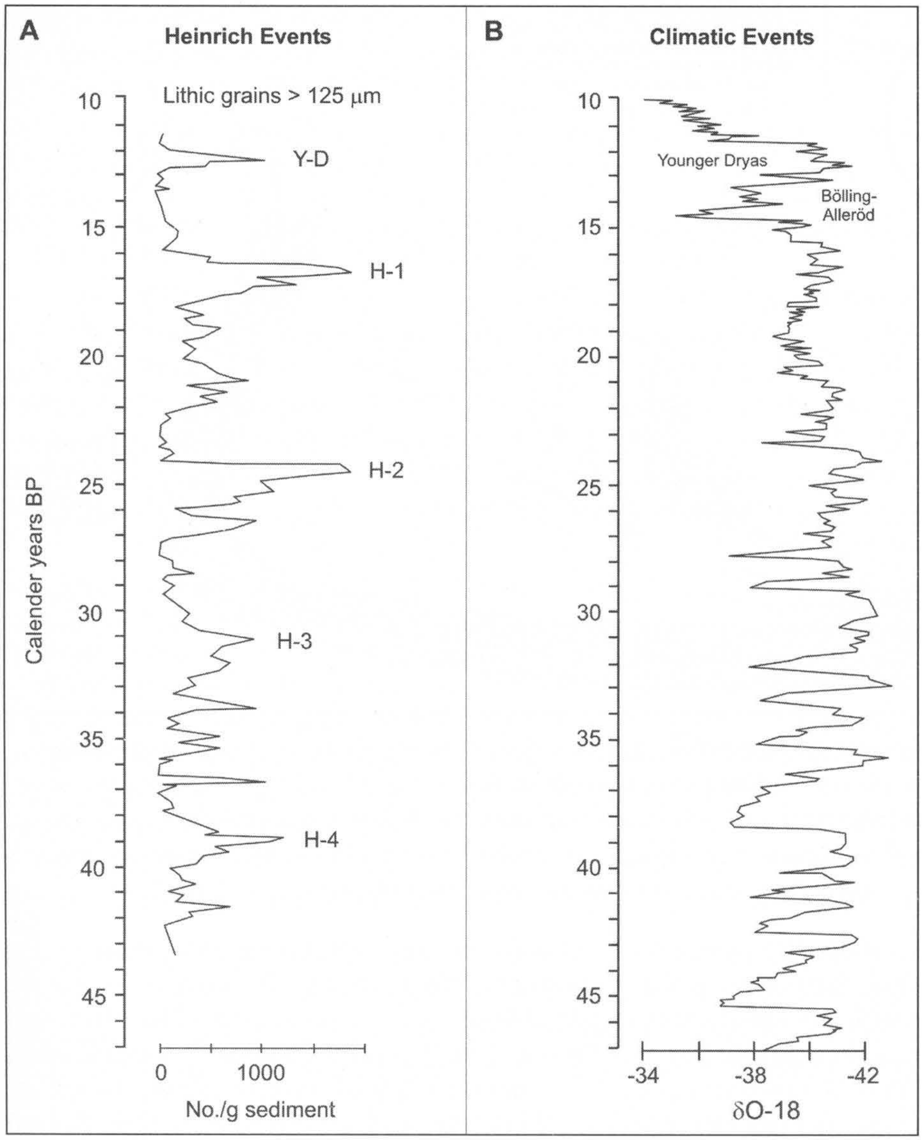

Fig. 1.3 a) Heinrich Events, as evidenced by the number of lithic grains larger than 125 μm per gram of sediment recovered from a core taken from the floor of the Atlantic Ocean west of Scotland (Barra Fan core MD95-2006). The peak numbers of these grains were deposited as a result of major iceberg discharges (Heinrich Events) into the Atlantic Ocean. H-1 to H-4 were Heinrich Events. Y-D: Younger Dryas stadial, sometimes denoted as H-0

b) Climatic events during the last 47,000 years as evidenced by analysis of an ice core from a site in Greenland (GISP2). Fluctuations in δO-18 reflect temperature oscillations, many of which occurred rapidly and are grouped into Dansgaard-Oeschger cycles. (Source: Bowen et al., 2002) at present in the North Atlantic. Temperatures consequently fell in the northern North Atlantic and adjacent areas, causing renewed glacial activity in some regions. Since the oceans form a coupled system there was an associated rise in sea temperatures in the tropical Atlantic, Southern Ocean and in the tropical Pacific, albeit after a time-lag of several hundred years (Clarke *et al.*, 2004). This tropical warming may have then influenced higher-latitude deglaciation.

Heinrich Events are referred to in Chapters Ten and Twelve. Based on cosmogenic rock exposure ages on glacial boulders it has been suggested that Heinrich Event 2 corresponded with the Last Glacial Maximum in Ireland and Wales (Bowen *et al.*, 2002). A ^{36}Cl rock exposure age on an erratic deposited at the ice-margin in Gower (Chapter Ten) apparently confirms this suggestion.

iv) Dansgaard-Oeschger and Bond cycles

During the 1960s the first cores were extracted from the world's major ice-sheets: from Camp Century in Greenland in 1966 and from Byrd Station in the Antarctic in 1968. Cores have since been recovered from ice-masses in many parts of the world, including Dye 3 (1981), Renland (1988), GRIP (1992) and GISP (1993) in Greenland; Devon Island in arctic Canada; Dome C (1979) and Vostok (1985) in the Antarctic, and from much smaller ice-masses including the Quelccaya ice-cap in the Peruvian Andes; Lewis Glacier on Mount Kenya and Kilimanjaro, both in Africa; the Inilchek Glacier in Kyrgystan; and many other locations (NOAA, 2004).

In 1969 Dansgaard *et al.*, from their study of evidence from the Camp Century core from north-west Greenland, showed that there have been rapid temperature changes over the past 100,000 years. Further work showed that these changes form cycles, now known as *Dansgaard–Oeschger cycles*, that have pronounced 1,450–1,500 year pacings. The warmer events are *interstadials* and the colder are *stadials*. There appear to have been rapid changes of temperature between stadials and interstadials, sometimes of as much as 10°C in less than a decade (Fig. 1.3). In some cases, particularly between 20,000 and 80,000 years ago, Dansgaard–Oeschger events were grouped into cooling cycles of some 3-6,000 years. These are know as *Bond Cycles* (Bond and Lotti, 1995). Chapter Twelve suggests that ice in the Irish Sea basin retreated from still-stand positions after the Last Glacial Maximum with the same pacing as Heinrich Events, Bond Cycles and Dansgaard-Oeschger cycles, indicating that glaciation in Wales and adjacent areas was related to much larger, probably global, systems.

v) Greenhouse gasses

Bubbles of air within ice-sheets provide information on greenhouse gas concentrations in the atmosphere at the time when those bubbles were sealed: lower concentrations of greenhouse gasses (carbon dioxide and methane) existed during glacial ages and higher concentrations during interglacials. Greenland and Antarctic ice cores show that greenhouse gas variability pulsed on orbital and millennial time scales. Whether greenhouse gasses changed climate, or whether they amplified changes in temperature caused by insolation variability, is as yet uncertain. Air temperature variability in Antarctica has now been estimated for the last 740,000 (EPICA, 2004) and 800,000 years (Jouzel *et al.*, 2004), with the analysis of greenhouse gas variability to follow soon (EPICA, 2004). Such variables, at orbital and millennial timescales, will have influenced climatic and other events in Wales (Bowen, 2000, 2004, 2005a, b), as is hinted at in Chapter Twelve.

Conclusion

A thorough understanding of glacial events, of glacials and interglacials, stadials and interstadials, requires interdisciplinary collaboration, an appreciation of the evolution of the global climate system and of its modelling.

The aim of the present book is to provide a compendium of existing knowledge of the glaciations of Wales and surrounding districts so as to provide a benchmark for further studies. The book consists essentially of a series of regional chapters that describe glacial and associated landforms and deposits in the different regions of Wales and surrounding areas.

Chapter Two provides an introduction to glacial deposits, Quaternary stratigraphy and Welsh climate history (Bowen, 1999). The Chapter also lists the major lithostratigraphical formations of glacial origin in and adjacent to the Principality. The latter part of Chapter Eight pays particular attention to events towards the end of the most recent (Devensian) glacial stage, when an interstadial began before 14,000 BP, following melting of the last regional ice-sheet and before reversion to cold, stadial conditions at about 12,450 BP, when cold conditions returned during the Younger Dryas Stadial. A small ice-sheet developed in the Western Highlands of Scotland at approximately the same time, reaching its maximum extent at the southern end of Loch Lomond, after which it has been named the *Loch Lomond Glaciation* (Sissons, 1976). By contrast, Chapters Six, Nine, Ten and Eleven draw attention to glacial deposits that may be more than 400,000 years old, possibly correlating with Oxygen Isotope Stage 16, 'the first major glaciation of the hundred thousand eccentricity world' (Bowen, this volume) or with Oxygen Isotope Stage 12. No evidence of earlier glacial deposits has yet been found in Wales and adjacent areas so that events during the earlier stages of the Quaternary in the region remain unknown.

Much of the initial work in Britain that was related to the establishment of the glacial theory was undertaken in Wales, and especially in North Wales. Recent research in South Wales has concentrated on the establishment of a geochronology that should enable scientists to correlate events in Wales and adjoining districts with the evolution of the global climate system. Wales therefore holds an honourable place in the continuing development of ice age investigations.

8 The upper Wye and Usk regions

by Colin A. Lewis *and* G. S. P. Thomas

Introduction

Although it is impossible to prove the precise morphology of the pre-glacial landscape (Brown, 1960; Clarke, 1936–7; Thomas, 1959) it is apparent that the basic outline of the Usk and Wye valleys, and of their tributaries, as well as of the uplands of the region, originated prior to the onset of Quaternary glaciation. The ice-sheets that developed over the region, the valley glaciers, the cirque glaciers and the periglacial features (such as protalus ramparts) thus formed on a landscape whose outlines already existed and which influenced glacial developments.

The mid-Wales ice-sheet

The maximum known extent of Quaternary ice in the Wye-Usk region is evidenced by the altitude to which erratics from areas further north have been deposited on the Black Mountains and the lack of comparable erratics on the Brecon Beacons. Presumably such erratics do not occur on the Brecon Beacons because those uplands nurtured their own ice cover, which prevented their invasion by mid-Wales ice.

Howard (1903–4) recorded Silurian slates, sandstones, quartz and dark basic igneous clasts at altitudes up to 400 m beneath the Black Mountains escarpment. Such erratics may be seen south of Hay-on-Wye on the interfluve between the Dulas and Cilonw Brooks in the vicinity of Twyn y Beddau (SO 245386), at altitudes around 380 m (Fig. 8.1).

The erratics discovered by Howard were derived from the Builth Wells-Aberedw area of the Wye valley and indicate that ice flowed generally southwards down the Wye valley to impinge upon the northern escarpment of the Black Mountains. The gathering grounds for this ice were presumably the uplands of Pumlumon and their southerly extensions.

Howard noted that Silurian erratics occur in the Rhiangoll valley, indicating that mid-Wales ice passed through the Pen-y-genffordd col at the head of that valley. The floor of the col is at little more than 320 m, some 240 m above the floor of the nearby Wye valley. Howard also considered that ice from the Wye valley entered the Usk valley by flowing up the Llynfi valley and passing over the low ground in the vicinity of Bwlch and near Tal-y-llyn station. On the basis of the evidence available to him Howard concluded that 'none of the Wye drift passed over the...escarpment between the Golden Valley and the Rhiangoll.' Dwerryhouse and Miller (1930) suggested that Wye ice entered the head of the Golden Valley.

North of the Wye valley, near Llanbella in the vicinity of Huntington (SO 240530), on the interfluve between the Arrow and Gladestry Brook valleys, there are hummocky glacial deposits. These contain Silurian clasts derived from the region to the north and west of Llanbella. The deposits exist at altitudes up to about 340 m and, like the erratics recorded by Howard beneath the escarpment of the Black Mountains, evidence the former existence of an ice-sheet over the region.

Erratics of Builth Wells olivine dolerite have been recorded in the Tywi valley (Bowen, 1970, this volume). Although it is not known when these erratics were deposited, they are believed to be of glacial origin (Griffiths, 1940), indicating that ice from the Builth region spilled south-westwards across the

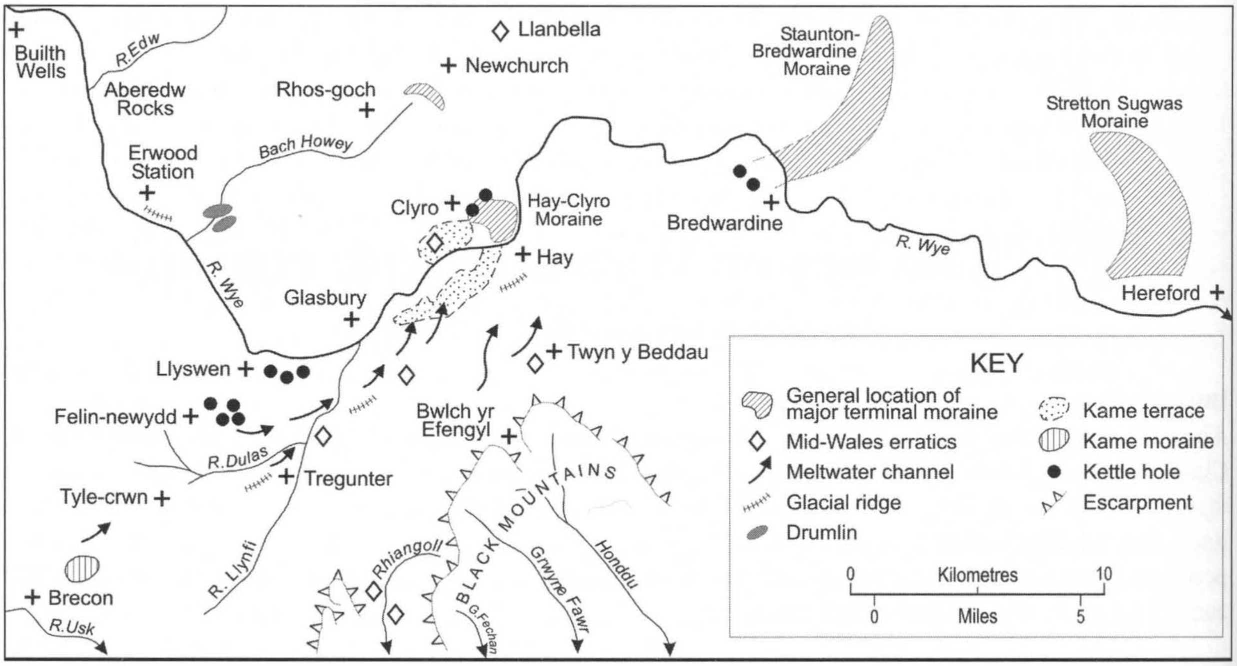

Fig. 8.1 The location of major terminal moraines in the Wye valley downstream of Glasbury and of erratics, meltwater channels and other indicators of glaciation north of the Black Mountains and south of Builth Wells

watershed between the Tywi and rivers that flow towards the Wye. The watershed is at an altitude of 290 m.

Mid-Wales, north of the Black Mountains-Brecon Beacons uplands, was therefore inundated by an ice-sheet that was at least 240 m thick in the Glasbury-Talgarth area when ice passed over the col at the head of the Rhiangoll valley, between the Wye and Usk drainage basins. This ice-sheet must have been continuous with ice over the Brecon Beacons-Fforest Fawr uplands, otherwise it would have impinged upon those uplands and deposited indicator erratics. The main escarpment of the Black Mountains apparently projected above the ice-sheet. Glaciers existed in the dip slope valleys of the Black Mountains, as is evidenced by glacial deposits and landforms in the Honddu, Grwyne Fechan and other valleys, although the correlation between them and the mid-Wales ice sheet has not been established.

M'Caw (1936) suggested that mid-Wales ice breached the Black Mountains escarpment at Bwlch yr Efengyl (the Gospel Pass; SO 235351; Fig. 8.1), at the head of the Honddu valley. This valley displays features of glaciation, being trough-shaped and having a remarkably well-developed terminal moraine at its southern end, at Llanfihangel Crucorney (SO 325208; Fig. 8.4). The floor of Bwlch yr Efengyl is at an altitude of 542 m so that, if the interpretation of M'Caw (1936) is correct, the mid-Wales ice-sheet on the northern side of the Black Mountains must have been at least 220 m thicker in the Glasbury-Talgarth area than the evidence presented by Howard (1903–4) suggests, attaining a depth of at least 460 m. Until mid-Wales erratics have been found in the upper Hondddu valley, however, it cannot be proved that mid-Wales ice flowed through the Gospel Pass.

The age and eastern limits of the mid-Wales ice-sheet as evidenced by erratics deposited beneath the Black Mountains escarpment and in the Rhiangoll valley, remains uncertain. In the Usk valley, downstream of the general vicinity of Crickhowell, ice of assumed Late Devensian age extended as a valley glacier almost to Usk. This may have been an outlet glacier of the mid-Wales ice-sheet that Howard (1903–4) had discovered.

In the Hereford basin mid-Wales ice of presumably Late Devensian age (Richards, this volume) spread out as a piedmont lobe, leaving well developed features at Stretton Sugwas, just west of Hereford, and in the Kington-Orleton region (Luckman, 1970). Older, probably Anglian Cold Stage (Middle Pleistocene, Oxygen Isotope Stage 12) glacial deposits exist north-east of Hereford (Richards, this volume) and contain Welsh erratics. Whether the mid-Wales ice-sheet, identified by the erratics recorded by Howard (1903–4) in the Black Mountains region, correlates with the older or younger glacial deposits in the Hereford basin is unknown. If, however, the hummocky glacial deposits near Llanbella (SO 240530) are of the same age as the deposition of mid-Wales erratics on the flanks of the Black Mountains, it is likely that the mid-Wales ice-sheet that they indicate was younger than the Anglian Stage, since glacial deposits of that antiquity are unlikely to have retained their morphology.

The maximum ice extent, as recorded by Howard (1903–4), was therefore probably not that responsible for the extension of glacier ice north-eastwards of Hereford. Consequently it is likely that there was an earlier and larger ice-sheet in mid-Wales, the evidence for which, within mid-Wales, has yet to be identified.

High-level meltwater channels

A series of now-dry channels occur at altitudes around 380 m on the commons near Twyn y Beddau (SO 240387) and cut across the summit of the interfluve between the Digedi and Cilonw valleys to the south of Hay-on-Wye, beneath the escarpment of the Black Mountains (Fig. 8.1). These channels lead north-eastward and probably originated as glacial meltwater channels utilised by meltwater flowing essentially parallel to the escarpment of the Black Mountains as it drained towards the lower ground of the Hereford basin.

The Wye glacier

Major terminal moraines

At least three major terminal moraines exist in the Wye valley downstream of Glasbury: at Stretton Sugwas (SO 465426) near Hereford (Luckman,1970; Richards, this volume); Staunton-on Wye/Bredwardine (SO 330445; Luckman,1970; Richards, this volume); and between Hay-on-Wye and Clyro (SO 220435; Fig. 8.1). All three features contain mid-Wales erratics and were formed by ice from that area.

The Stretton Sugwas moraine

This moraine formed at the edge of a piedmont lobe as the mid-Wales ice-sheet extended onto the lower ground of the Hereford basin. Both Luckman (1970) and Richards (this volume) have described the moraine in detail.

The Staunton-on-Wye/Bredwardine moraine

The Staunton-on-Wye/Bredwardine moraine formed at the end of a valley glacier that extended eastwards down the Wye valley from accumulation areas in Powys. Richards (this volume) has described how the moraine extends from Brobury Scar (SO 353444), where a section is exposed, through Staunton and Norton Canon to Hyatt Sarnesfield (SO 380500). Up-valley of Bredwardine, especially in the vicinity of Weston Farms (SO 455320), well developed kame and kettle holes exist in the glacial deposits and some of the kettle holes contain minor lakes. There was thus a dead ice area upvalley of the terminal moraine, where the glacier stagnated and melted in situ.

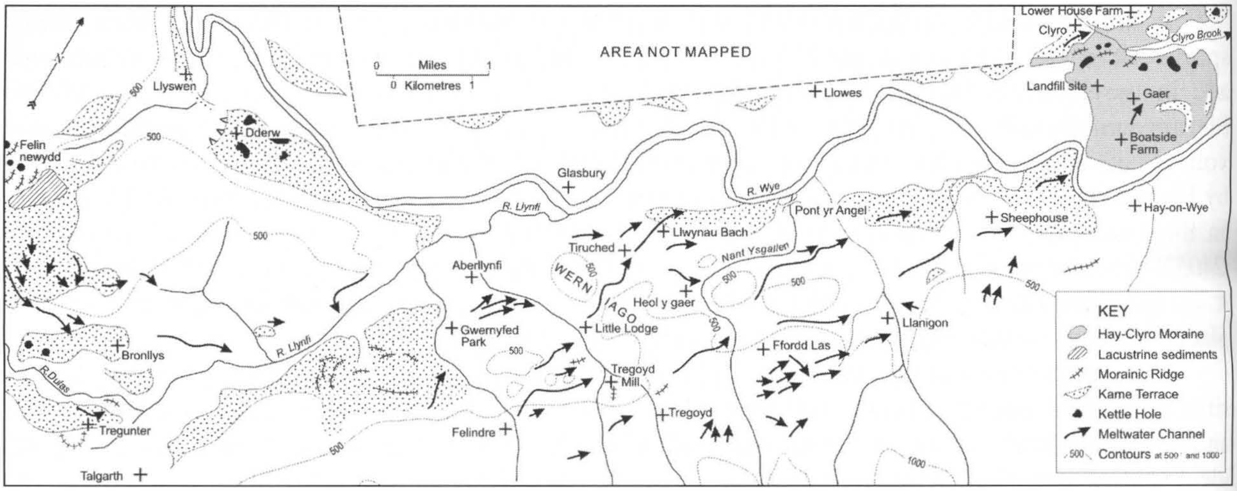

Fig. 8.2 Meltwater channels and other glacial features in the Talgarth- Hay region. (Redrawn, with minor additions and alterations, after Williams, 1968a)

The Hay-Clyro moraine

The Hay-Clyro moraine (Williams, 1968a; Fig. 8.2) forms a ridge, partly plastered on bedrock, that is followed by the B4351 road. In 1990 a landfill site adjacent to this road (SO 219434) exposed 4 m of diamicton containing lenses of sand and gravel. Some of the clasts within the deposit were of fossiliferous Silurian rock, presumably derived from outcrops in the Aberedw area, further up the Wye valley, near Builth Wells. Similar clasts, some of which bear shallow striations, exist in the lateral moraine/kame terrace between Llowes and Clyro that leads into the terminal feature. They have been exposed by road works beside the A438 opposite the point where a meander of the River Wye lies immediately below the road. The striae on the clasts indicate that the clasts were transported within the glacier, where clasts rubbed against and scratched each other.

The minor road that leads from the B4351 past Boatside Farm (SO 432225) and a house named "Gaer" to the A438 at Lower House Farm (SO 222443) traverses the moraine. Kame and kettle holes exist between Gaer and Lower House Farm as well as what appear to be isolated perforation kames. Morainic ridges, aligned approximately parallel to the axis of the Wye valley, occur between Clyro village and the minor road upstream of the point where that road crosses the Clyro Brook that flows past Lower House Farm. Kame and kettle topography also exists adjacent to the right bank of the Clyro Brook below the point where it is crossed by the minor road. Occasional meltwater channels run across the moraine, being most prominent in the vicinity of Clyro. An outwash terrace extends downstream of the moraine, towards Clifford. Symonds (1872) reported that the remains of boar, deer and ox had been discovered in this terrace, although these may have been part of the post-glacial, possibly Holocene, fauna of the valley.

Kame terraces and meltwater channels associated with the Hay-Clyro moraine

Lateral deposits, mapped by Williams (1968a) as kame terraces, exist on both sides of the Wye valley downstream of Glasbury and terminate at the Hay-Clyro moraine (Fig. 8.2). They are best exhibited on the right bank of the valley, erosion having obliterated much of the kame terrace/lateral moraine material on the Radnor side of the valley. Cuttings in these features, as already noted for the lateral moraine/kame terrace immediately downstream of Llowes, expose occasional Silurian clasts as well as the remains of the local country rock, indicating that the glacier responsible for these features flowed down the Wye valley from the vicinity of Builth Wells. Shallow meltwater channels are cut into the terraces, especially in the vicinity of Sheephouse (SO 210414), about 1.4 km upstream of Hay-on-Wye. These channels are aligned essentially parallel to the axis of the valley.

A complicated series of meltwater channels exists higher up the right hand side of the valley, especially between Felindre and the vicinity of Llanigon (Fig. 8.2; Williams, 1968a; Lewis, 1970a). The Wye glacier apparently spread out as a piedmont lobe downstream of the funnel-like valley above Llyswen to form a mer-de-glace between Hay-on-Wye and an unknown area in the Llynfi valley upstream of Talgarth, where the Wye ice might have been in contact with ice from the Brecon Beacons and adjacent uplands. These channels allowed meltwater from the decaying ice to drain north-eastward into the Wye valley beyond the limits of the Hay-Clyro moraine.

Some of the meltwater channels, such as that which cuts across the Wern Iago (Tregoyd Wood) ridge between Little Lodge Farm (SO 187382) and Tiruched (SO 185389), appear to have been superimposed upon and thus cut through ridges. The Little Lodge-Tiruched channel starts beside Little Lodge Farm, cuts through the Wern Iago ridge, crosses the Tiruched valley that opens into the Wye at Glasbury, and proceeds across the ridge between the Tiruched valley and Llwynau Bach (SO 395187) to emerge on the kame terrace on the right bank of the Wye valley. The intake of the channel at Little Lodge lies down-valley of a morainic ridge, containing clasts from the Builth/Aberedw area, on the southern side of the minor valley occupied by the Tregoyd Mill Brook some 200 m down-valley of the hamlet of Tregoyd Mill (SO 188378). The ridge is best seen in the fields west of the minor road from Felindre to Tregoyd Mill. The channel was therefore not of ice marginal origin. Instead, it was probably initiated subglacially, as glacial meltwater drained north-eastwards to escape along the kame terrace at the edge of the main Wye glacier downstream of Glasbury.

A different origin apparently applies to the deeply incised valley of Nant Ysgallen, which forms a gorge between Heol-y-gaer (SO 195391) and the point where the stream debouches onto the kame terrace near Pont yr Angel (SO 199403). Upstream of Heol-y-gaer the stream drains a flat area between Tregoyd and Ffordd Las (SO 198380–207393) in which there are sediments that appear to be of lacustrine origin (Williams, 1968a). An ice margin ending immediately west of this flat area would have blocked drainage off the Black Mountains and could have ponded a pro-glacial lake. As the ice retreated towards the lower land of the Wye valley floor the lake apparently drained, via Nant Ysgallen, cutting the gorge near Heol-y-gaer as it did so.

Other, and much less impressive channels, exist on the southern flank of the Wye valley between Gwernyfed Park (SO 175373) and Glasbury. Some of these are semi-circular, as if one side of the channel is missing, as is the case with a number of small channels north of Gwernyfed Park school that apparently drained towards the Tregoyd Mill Brook. These may have developed as ice-marginal channels, allowing meltwater to drain down-valley along the ice-front, with one side of the channel being the ice-front itself.

The age of the Hay-Clyro moraine

This moraine is a spectacular and obvious feature in the landscape and on morphological grounds it is likely that it dates to the Late Devensian, although its age has not been established numerically.

The Llyswen-Dderw kame moraine

An area of kettled sediments extends downstream from a steep slope, that is up to 8 m high, located on the eastern side of the stream that flows past the former Nonconformist chapel (SO 133378) on the eastern side of Llyswen (Fig. 8.2). Sand and finer sediments exist in the flattish area on the Llyswen side of the steep slope. Kettle holes, some of which are occupied by ponds, exist on the Dderw Farm (SO 140376) and extend down-valley to the vicinity of Boxbush (SO 154377). The steep slope near the former chapel at

Llyswen may well be a former ice-contact slope, while the sands and gravels that form the kettled area were probably deposited by outwash streams that flowed out of the melting ice. The kettle holes presumably occupy areas where blocks of ice, carried onto the outwash plain by meltwater streams and covered or partly covered in fluvioglacial sediments, subsequently melted. The flattish area up-valley of the steep slope may have been occupied by a pro-glacial lake, in which fine sediments were deposited, as the ice-front melted back from the steep slope but still blocked drainage down the main Wye valley.

The Wye valley between Llyswen and Builth Wells

The River Wye flows within an U-shaped gorge between Llyswen and Builth Wells (Fig. 8.1), the valley floor being some 100 m or more below the level of the uplands on either side of the valley. Silurian bedrock is particularly well exposed down-valley of the confluence of the Edw and the Wye, at Aberedw Rocks (SO 080460), where some of the outcrops resemble roches moutonnee and may owe their shape to glacial plucking. Well developed terraces, some of which may be of glacial outwash origin, occur on the valley floor and lower valley sides between Llyswen and the vicinity of Builth Wells, as between Erwood and the site of the former Erwood railway station (SO 089439). Linear ridges downstream of the road bridge across the River Wye near Erwood station may be the remnants of a former ice-margin.

Drumlins exist at altitudes around 210 m in the Bach Howey valley, which is a left bank tributary of the Wye, between Erwood and Painscastle, in the vicinity of the Castle Mound (SO 125450). Terminal moraines cross the Bach Howey valley east of Painscastle, as between Rhosgoch Common (SO 195485) and Newchurch. Sections in glacial till at altitudes around 300 m in the vicinity of Llanbella (SO 240530) expose striated Silurian clasts and evidence the former movement of ice eastward over the uplands of the region from an ice centre further west.

An area of moundy terrain on the southern flank of the Duhonw valley west of Glanwye (SO 066498), and the till exposed on the wooded east bank of the Wye adjacent to a house and garage upstream of Llanfaredd (SO 064512), may be the remains of a terminal moraine of the Wye glacier.

The Vale of Irfon

The Wye valley changes its nature upstream of Builth Wells as it crosses a strike vale north of the north-facing escarpment of the Eppynt-Carneddau uplands. The River Irfon, which with its tributaries drains the mid-Wales uplands between the Elan-Claerwen and Tywi valleys, flows southwards off those uplands before turning east near Llanwrtyd Wells and flowing as a strike stream along the Vale of Irfon to join the Wye at Builth Wells.

North of Llanwrtyd Wells, particularly upstream of Abergwesyn, the Irfon flows in a trough-shaped valley. Deposits of glacial till occur at various sites within the valley, a section on the east side of the valley downstream of the site of the former church of Abergwesyn, exposes some 8 m of till derived from the local rocks of the upper Irfon valley (SN 864525). Ice from the mid-Wales uplands therefore drained down the Irfon valley, which it imprinted with the shape of a glacial trough in the region upstream of Llanwrtyd, before debouching onto the lowlands of the Vale of Irfon between Llanwrtyd Wells and Builth Wells.

The Vale of Irfon contains many features of glacial deposition. Landforms resembling drumlins are well developed in the region between Cilmery (SO 005515) and Builth Wells, with their long axes predominantly aligned parallel to the long axis of the Vale. Till is exposed on the flanks of one such feature between Cilmery railway station and Glan-Irfon (SO 003509). Upstream of Cilmery, between there and Llanfechan House (SN 979504), kames occur on the floor of the Vale, as in the field (SN 989506) beside the A483 between the road and a house adjacent to the track that leads north to Cilmery Farm. Interpretation of many landforms in the Vale, including those that resemble drumlins, is hindered by the paucity of exposures.

Further west in the Vale, as at Aber-Dulas-uchaf (SN 918468), there is blue-coloured till, containing Ordovician clasts that indicate that the ice responsible for its deposition flowed southwards from the central Welsh uplands and into the Vale. The till at Aber-Dulas-uchaf, which is exposed to a depth of 2 m, exists under outwash and younger (possibly Holocene) gravels.

Between Llanwrtyd Wells and Llawr dre-fawr (SN 858450) exist kames, with the boggy area of Waen Rydd occupying a basin with steep faces that resemble ice-contact slopes to north, south and west. Immediately further west, at Erwbeili (SN 853447), beside the A483 road, there is a delta deposit with well-marked foreset bedding dipping south-westward. This indicates that a pro-glacial lake was formerly trapped between an ice-front at Erwbeili and the present watershed between the Vale of Irfon and the Tywi catchment to the west. Presumably the lake drained over that watershed.

The complicated glacial features in the Vale of Irfon and elsewhere imply that the Vale was formerly inundated under an ice-sheet derived from mid-Wales but that ice subsequently only occupied the lower portions of the Vale. Whether the mid-Wales ice-sheet, that covered the Vale and abutted against the Black Mountains, melted entirely but was succeeded by readvance of ice from the mid-Wales mountains into the Vale, is unknown. The drumlin-like landforms between Cilmery and Builth indicate subglacial drainage towards the Wye valley during deglaciation. Tills in the western Vale, extending across the lower part of the Eppynt and into the south flowing Cilieni valley, indicate that mid-Wales ice also flowed into the upper Usk catchment. The pro-glacial delta at Erwbeili indicates that although ice formerly spilled into the upper Tywi catchment, across the Sugar Loaf col (SN 845440), its western margin was later restricted to the Irfon Vale. Whether the Erwbeili deposits relate to a withdrawal stage of the main ice-sheet, or a readvance of a glacier down the Irfon valley, is presently unknown.

The Wye valley upstream of the Vale of Irfon

Hummocky topography exists upstream of Builth Wells in the vicinity of Builth Road railway station, where there are kame-like landforms, suggestive of the former presence of stagnant ice. Thicknesses of up to 6 m of till are exposed in stream side sections and in cuttings beside the abandoned railway line some 200 m north-west of Builth Road. Kame-like mounds occur on the floor of the Wye valley between Builth Road and Penmincae (SO 004543), where there is a wooded esker-like ridge. Unfortunately the lack of sections renders interpretation of these features uncertain.

North of Newbridge-on-Wye, where the Wye emerges from its Doldowlod gorge section, a gravel pit near the junction of the A470 with the A4081 exposes bedded sands and gravels that were deposited by meltwater from a glacier issuing out of the gorge. Similar deposits near Llanyre (SO 037614) indicate that ice formerly covered the Ithon lowlands. At Caerfagu abattoir (SO 045657), in the Dulas valley, an esker, or possibly delta deposit, is exposed, indicating that meltwater from an ice-mass in the Nantmel-Rhayader area drained eastwards as the ice melted.

On the western side of the Wye valley, opposite Doldowlod House where Nant Cymrun joins the main valley, a well marked lateral moraine exists at altitudes around 300 m (SN 985623). The right bank tributary further north, that joins the Wye at Llanwrthwl, is trough-shaped and carried diffluent ice from the Elan valley as it spilled towards the Wye. The valley sides are markedly plucked while boulders, probably derived from the Elan valley uplands, are in evidence on the valley floor.

A lateral moraine exists on the south side of the Elan valley above the confluence of that valley with the Wye near Rhayader. Terminal moraines exist on the floors of valleys draining off the uplands between Rhayader and the Elan valley reservoirs, as in the valley of Glanllyn (SN 948690), where a small lake nestles below a trough's end over which a stream tumbles in waterfalls. Similar trough's ends exist in other valleys that carry drainage off the central Welsh uplands, indicating the importance of those uplands in nurturing the mid-Wales ice-masses that formerly developed into ice-sheets that extended into the Welsh border-lands and the Hereford basin.

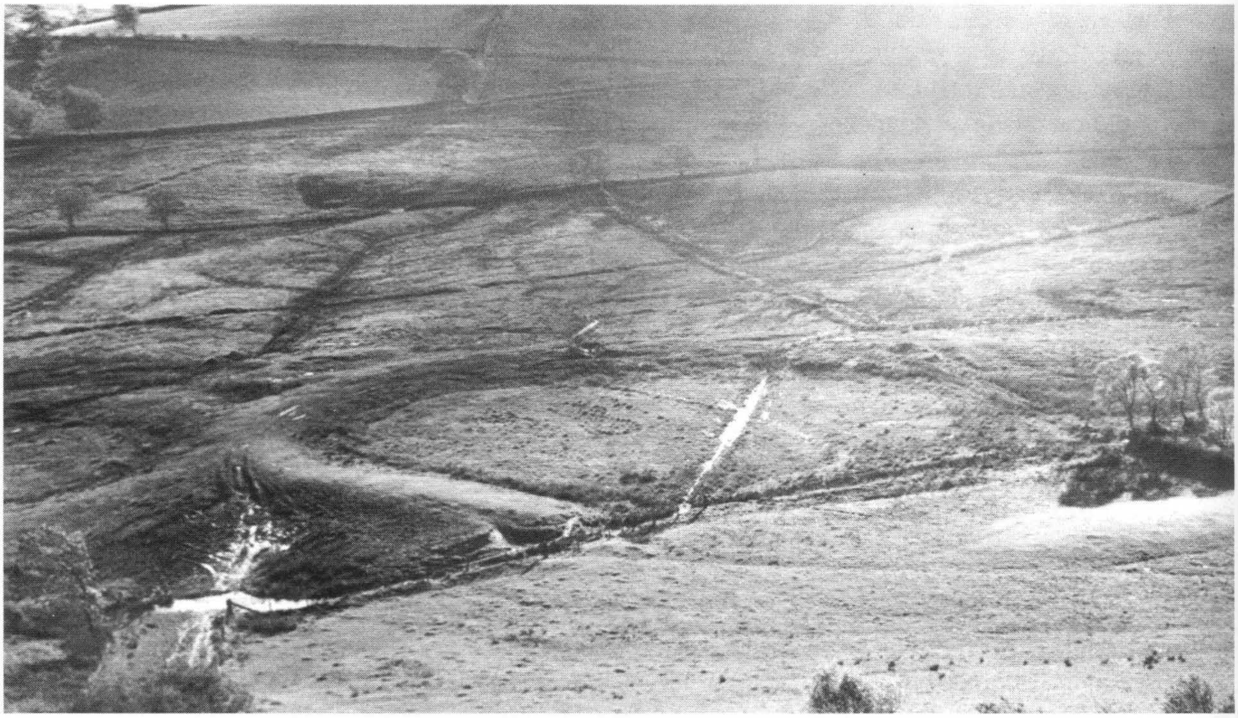

Fig. 8.3 Pingo remnants in the valley of the Nant Cae-garw, near Llanidloes, photographed when they were being examined by Pissart in 1961. The circular ramparts formed as the ice of the active pingos melted, allowing sediments incorporated within that ice to sludge down the sides of the pingos. The ill-drained basins within the ramparts represent the central portions of the formerly ice-cored landforms. The formation of pingos is described by Ballantyne and Harris (1994). (Photo: C. A. Lewis)

Periglacial features, in the form of solifluction deposits, ice-wedge casts and pingo remnants, exist in the isolated areas north of Rhayader. Pingos near Llangurig, in the Nant Cae-garw valley (SN 950798; Fig. 8.3), which is a tributary of the Dulas that flows northward towards Llanidloes, have been described by Pissart (1963) and are thought to be Late Glacial in age. They evidence the existence of at least sporadic permafrost in the area following its deglaciation. Ice-wedge casts in the Claerwen and other valleys of the central Welsh uplands (Potts,1971) are also indicative of the existence of permafrost in those areas following ice-sheet deglaciation.

The Usk valley

The glacial features of various parts of the Usk valley have been mapped by Williams (1968a), Lewis (1970a, b), Ellis-Gruffydd (1972, 1977) and Thomas (1997). South of Abergavenny a large morainic structure runs in a broad arcuate band across the valley of the middle Usk. This moraine has long been recognised as marking the limit of the last glaciation in the area and has been repeatedly incorporated into maps of the Late Devensian glaciation (Charlesworth, 1929; Bowen, 1973, 1981; Campbell and Bowen,1989). Except for the work of Thomas (1997) and the excellent but unpublished geomorphological maps of Williams (1968a), however, very little work has been undertaken in the area.

The Usk valley may be divided into three regions: the middle Usk between Talybont and Usk; the lower Usk valley, downstream of Usk; and the upper Usk and Brecon Beacons-Fforest Fawr uplands and the Eppynt. Each of these regions, as well as the adjacent Black Mountains and Grwyne Fawr/Cwm Coed-y-Cerrig trough, will be described following presentation of a glacial sediment-landform assemblage system and a general statement on drift characteristics.

Sediment-landform assemblages

The middle and upper Usk valley is a type example of a glacial valley depositional system that includes at least four major sediment-landform assemblages. These assemblages are mappable units in which relatively homogenous morphologic, stratigraphic and lithologic characteristics occur, as shown on Fig. 8.4. Similar assemblages have already been described from the Wye valley, although they were not then related to a sediment-landform assemblage model.

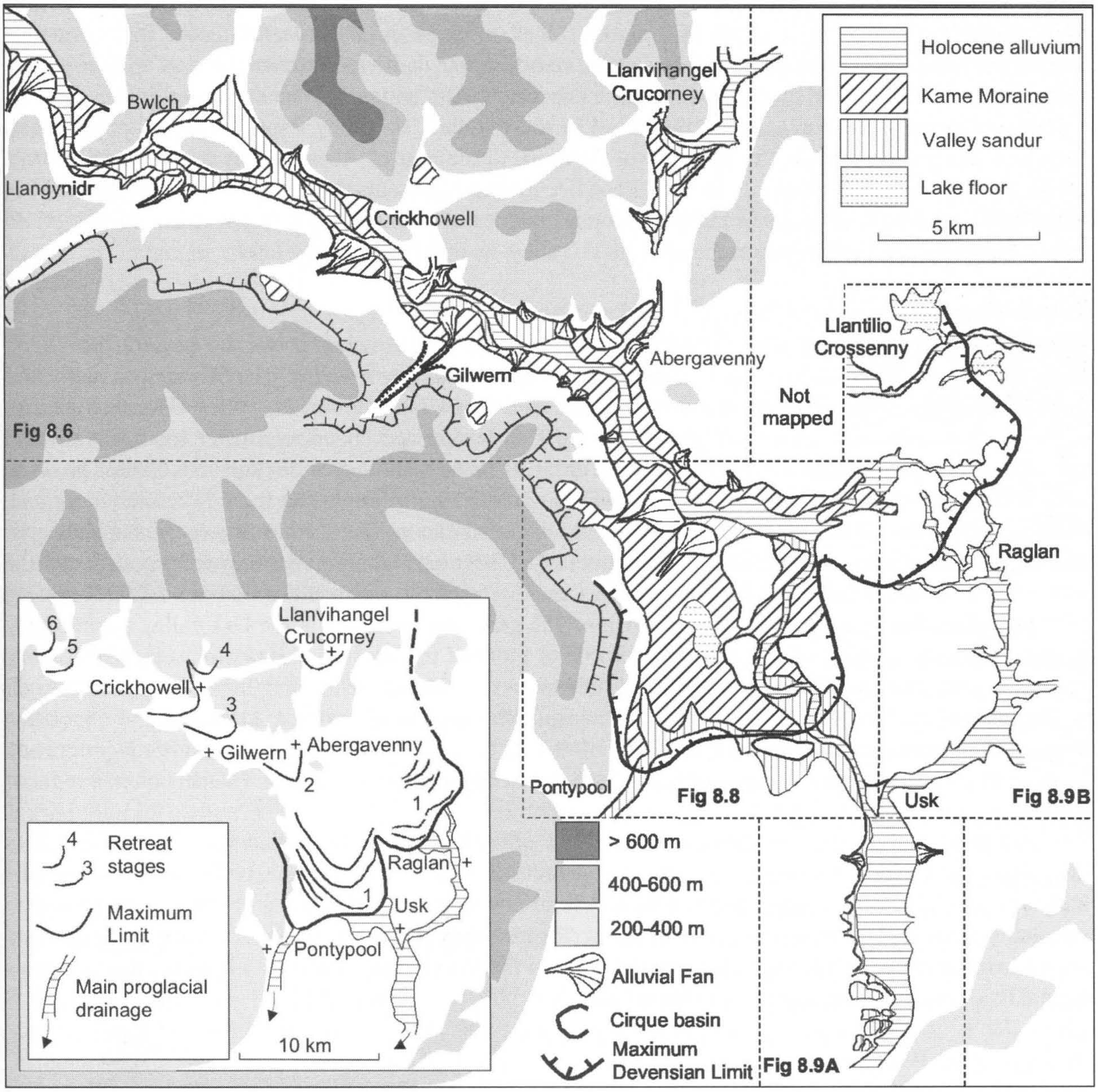

Fig. 8.4 Outline geomorphological map of the middle and lower Usk showing distribution of sediment-landform assemblages, maximum of the Late Devensian glaciation and location of geomorphological maps in Figs 8.6, 8.8 and 8.9. Inset shows major moraine systems and distribution of main proglacial drainage

i) Kame moraine assemblage

This assemblage consists of extensive tracts of hummocky topography, up to 20 m in height, comprising ridges, mounds, basins and intervening channel systems. Individual tracts occur either as arcuate bands running across valley or as irregular borders to the valley floor. They are composed of sand and gravel overlying basal diamict and bedrock. The hummocky topography was formed at, or immediately in front of an ice-margin during temporary retreat, still-stand or minor readvance, by deposition of ice-marginal sedimentation on top of dead and decaying ice. When the glacier retreated and the buried ice melted, the resultant surface was left as a series of irregular ridges, mounds and basins marking the former arcuate ice-margin. Much of this topography was later removed by meltwater stream erosion as the glacier retreated, or was buried by subsequent meltwater sedimentation and post-glacial alluviation.

Kame moraine assemblages are preserved in the middle Usk valley, especially west of Talybont, west of Llangynidr, opposite Crickhowell, at Gilwern, to the east of Abergavenny and in a major tract between Abergavenny, Pontypool and Usk. Each occurrence represents a major still-stand position during retreat of the Usk valley glacier from its maximum limit and at least six major episodes can be recognised (inset, Fig. 8.4). They also exist in the upper Usk valley, as will be described later.

ii) Valley sandur assemblage

This assemblage consists of gently sloping depositional surfaces, elevated above the modern river flood plain, that front tracts of cross-valley kame-moraine and extend down-valley. They formed as meltwater from the melting ice carried sediments down valley and deposited them on the valley floor downstream of the ice-front to form an outwash plain, or sandur. This feature was abandoned as the ice-front retreated up-valley and a new sandur was deposited. The deposits underlying the sandur surfaces consist predominantly of gravel. The older sandur deposits were at least partially eroded as the newer sandur formed. Consequently, through a succession of retreat still-stands, sandur surfaces occur as a set of stacked depositional features grading into one another down-valley. Under normal circumstances the proportion of the valley floor occupied by sandur surfaces diminishes downstream as a result of increased lateral erosion by fluvio-glacial and post-glacial stream incision. The gradients of sandur in the Usk valley are normally steeper than either that of the existing river profile or those of recent river terraces.

Valley sandur form from a single-exit meltwater stream system discharging from an ice-margin defined by the cross-valley moraine ridge on the up-valley side of the sandur. At the time of deposition outwash sediments accumulated across the whole width of the Usk valley floor, but subsequent erosion associated with the deposition of younger sandur plains and with Holocene fluvial incision has reduced the extent of the original deposits.

A typical sandur section occurs in the west bank of the Usk north-east of Gilwern (SO 262148). This displays 8 m of clast-supported, sub-rounded, pebble to cobble gravel in stacked, sub-horizontal sets indicative of high-energy, upper fan braid-bar sandur environments. The clasts are mainly of Devonian conglomerates, sandstones and mudstones derived from outcrops upstream. Apart from tuff, which may derive from Ordovician volcanic rocks outcropping in the Wye basin around Builth Wells, no rock types from outside the Usk catchment have been found.

iii) Alluvial fan assemblage

Alluvial fans occur frequently on the margins of the Usk valley. Over twenty fans, ranging in size up to more than a kilometre in area, occur between Bwlch and Usk (Fig. 8.4). These fans extend from tributary valleys out onto the floor of the main valley. They are mostly gently sloping and either rectilinear or gently concave upwards. They consist of coarse gravel that is often little rounded, intercalated with debris flow deposits.

The alluvial fans were probably initiated as deglaciation occurred. Water and debris flows transported sediment down the valley sides and from tributary valleys to form the alluvial fans on the floor of the main Usk valley. By analogy with contemporary valley glacier systems it is likely that in their early stages they were deposited partially upon the lateral margins of the glacial ice. As glacier retreat continued water and debris flows transported sediment downslope from parts of the declining ice-sheet that still occupied the upland flanks of the Usk, or, as in the case of the two large fans around Llanover and similar fans opposite Abergavenny and Crickhowell, from cirque glaciers occupying large hollows along the rim of the Carboniferous Limestone escarpment.

Some of the alluvial fans at least partially overlie and bury kame moraine and sandur deposits formed during earlier stages of valley glacier retreat. Consequently they form a diachronous series, younging up-valley.

iv) River terrace assemblage

River terraces border the Usk throughout most of its length. In general terms the horizontal and vertical extent of terracing markedly increases down-valley. Two main terraces occur, though intermediate terraces exist in some areas.

The Upper Terrace occurs at a height of approximately 4-5 m above the modern flood plain while the Lower Terrace is at a height of 2-3 m. Abandoned meander scars and cut-offs, some of the latter being still filled with water, frequently exist on the Lower Terrace.

A river bank cut on the outside of a large meander bend east of The Bryn (SO 335099), between Abergavenny and Usk, exposes details of both terraces (Fig. 8.5). The Upper Terrace, 4 m above river level, truncates a small kame mound and consists of laminated red silts and fine sands. These pass down into grey, partially laminated organic clay, with tree trunks and scattered plant debris, and then into peat. To the east the Upper Terrace is truncated by the Lower Terrace at a height of 2 m above river level. The age of these terraces is unknown but is believed to be Holocene. They probably formed as the River Usk adjusted to changes in sediment supply, climatic fluctuation and to anthropogenic influences.

Drift distribution, thickness and type

Drift deposits in the middle Usk are largely confined to the valley bottom, either as a series of flat or gently sloping terraces across the valley floor or as narrow, hummocky strips along the lower flanks of the valley. These deposits are up to 50 m thick. South of Abergavenny the valley widens, drift distribution is thinner, (usually less than 30 m), and more diffuse, forming an irregular veneer across partially buried Silurian bedrock escarpments. Drift is sparse south of the town of Usk, but the over-deepened rock floor of the valley towards Newport is overlain by sands and gravels succeeded by thick Holocene river alluvium bordered by narrow gravel terraces (Williams, 1968b).

Analysis of borehole records and mapping of bedrock outcrops (Crimes *et al.*, 1992) suggests that the middle Usk forms a series of shallow, over-deepened rock basins separated by rock-bars at or near the surface. Each basin consists of accumulations of sand and gravel overlying diamict. Rock bars occur at Llangynidr, Crickhowell and Abergavenny (Fig. 8.4) with an approximately 5-6 km spacing. This is a pattern similar to other radial valley systems investigated elsewhere in Wales (Thomas *et al.*, 1982). Although few boreholes reach bedrock, average over-deepening is a minimum of 50–60 m.

Although drift outcrop is poor, boreholes indicate that four major sediment types can be identified. Diamict is rarely exposed but occurs extensively at depth in thicknesses of up to 10 m across the bedrock floor. It comprises a massive, matrix-supported, sandy clay crowded with well-rounded, often striated clasts of predominantly Devonian and Carboniferous rock types, but with a sprinkling of more far-travelled Silurian rocks from the Wye basin (Lewis, 1970a). By its lithological characteristics and

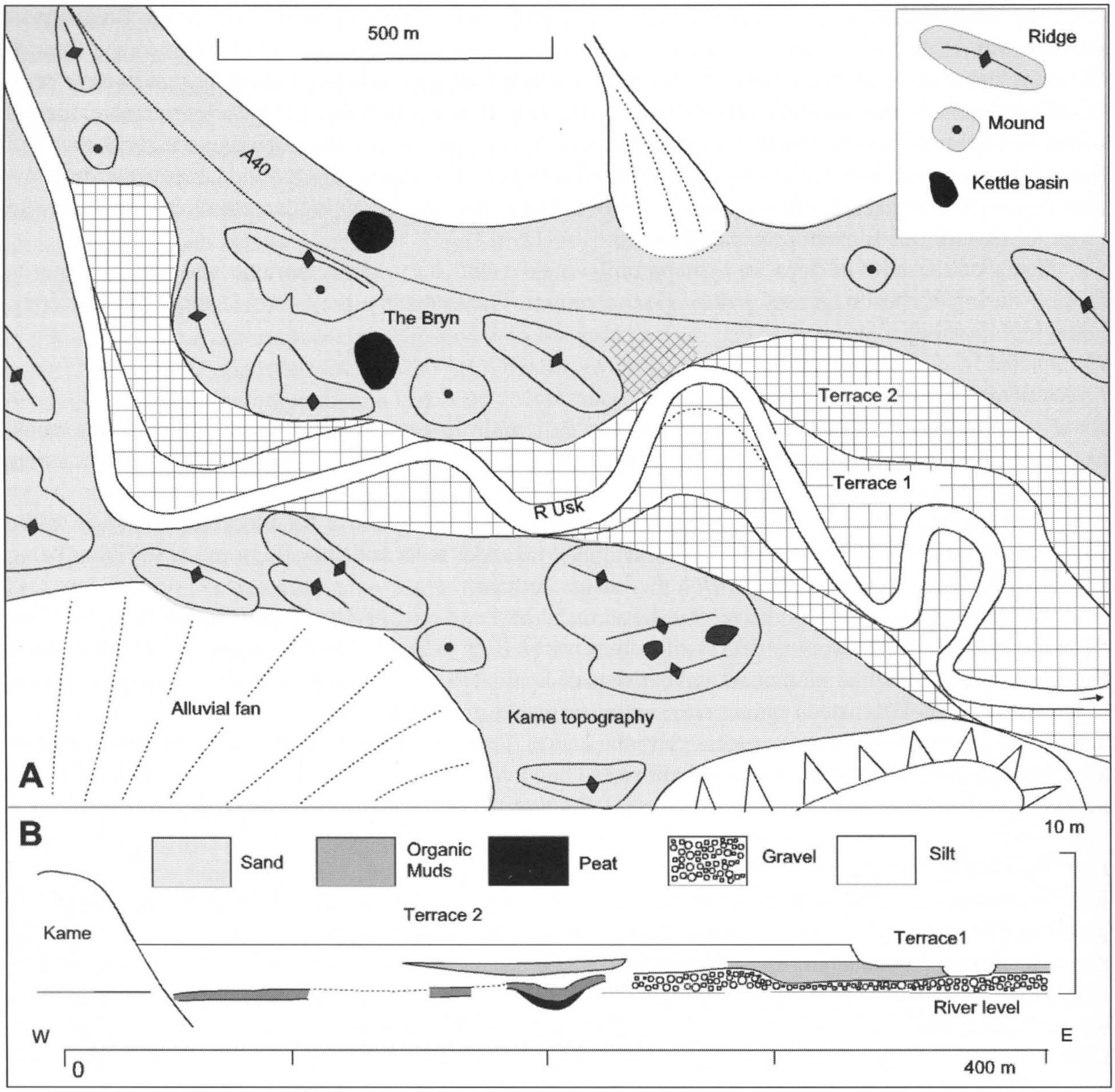

Fig. 8.5. A: geomorphological map of the area around The Bryn, showing distribution of Holocene river terraces. B: sketch section showing relationships between Devensian fluvioglacial and Holocene river terraces in section east of The Bryn

its stratigraphic position above the bedrock, the diamict is interpreted as the product of sub-glacial deposition. Sand and gravel ranges from coarse, sometimes boulder gravel, to fine sand deposited in a number of different depositional environments including ice-front alluvial fans, ice-marginal sandur and valley-sandur. Finer gravel and sand also comprises much of the alluvial flood plain. Mud includes laminated or thinly bedded brown or reddish clay, silt and fine sand occurring locally within the succession but also as extensive surface outcrop in flat-floored basins east of Llanover and in the upper headwaters of the River Trothy. These mud deposits are interpreted as glacio-lacustrine sediments.

Regional descriptions

Introduction

The underlying geology plays a significant role in the morphological development of the Usk basin. North of Abergavenny the middle Usk valley floor is narrow and developed along the strike of the mudstones of the Lower Devonian St Maughans Formation (Barclay, 1989). It is bordered on its western side (Fig. 8.4) by a prominent escarpment formed by the more resistant Upper Devonian Quartz Conglomerate Group and the Lower Carboniferous Castell Coch Limestone. This escarpment is broken north of Abergavenny by the deep entrenchment of the Clydach gorge. South of Abergavenny the valley widens along the axial trend of the Silurian Usk anticlinal inlier and some strong local relief occurs in the form of minor escarpments across the valley floor.

The middle Usk: Talybont to Abergavenny

The salient features of the geomorphology of the Usk valley between Talybont and Abergavenny are shown on Fig. 8.6. At Talybont a large alluvial fan drains north into the margin of the Usk valley and is flanked to its west by a strip of kame-moraine topography diversified by a number of arcuate cross-valley moraine ridges and marginal drainage channels. To the west of Bwlch a further area of kame-moraine topography is preserved on both margins of the Usk and displays a complex sequence of arcuate cross-valley moraine ridges, small channels running parallel to the ridges, and small kettle holes (Lewis, 1970b). Together these cross-valley moraines represent the final ice-marginal retreat stage of the Usk glacier in the middle Usk valley (Stage 6) although further ice-marginal stages occur in the upper Usk region. Further east the valley narrows abruptly into a shallow gorge below Buckland Hill and bedrock is exposed in the river bed. This marks the position of a shallow rock bar separating overdeepened rock basins upstream and downstream.

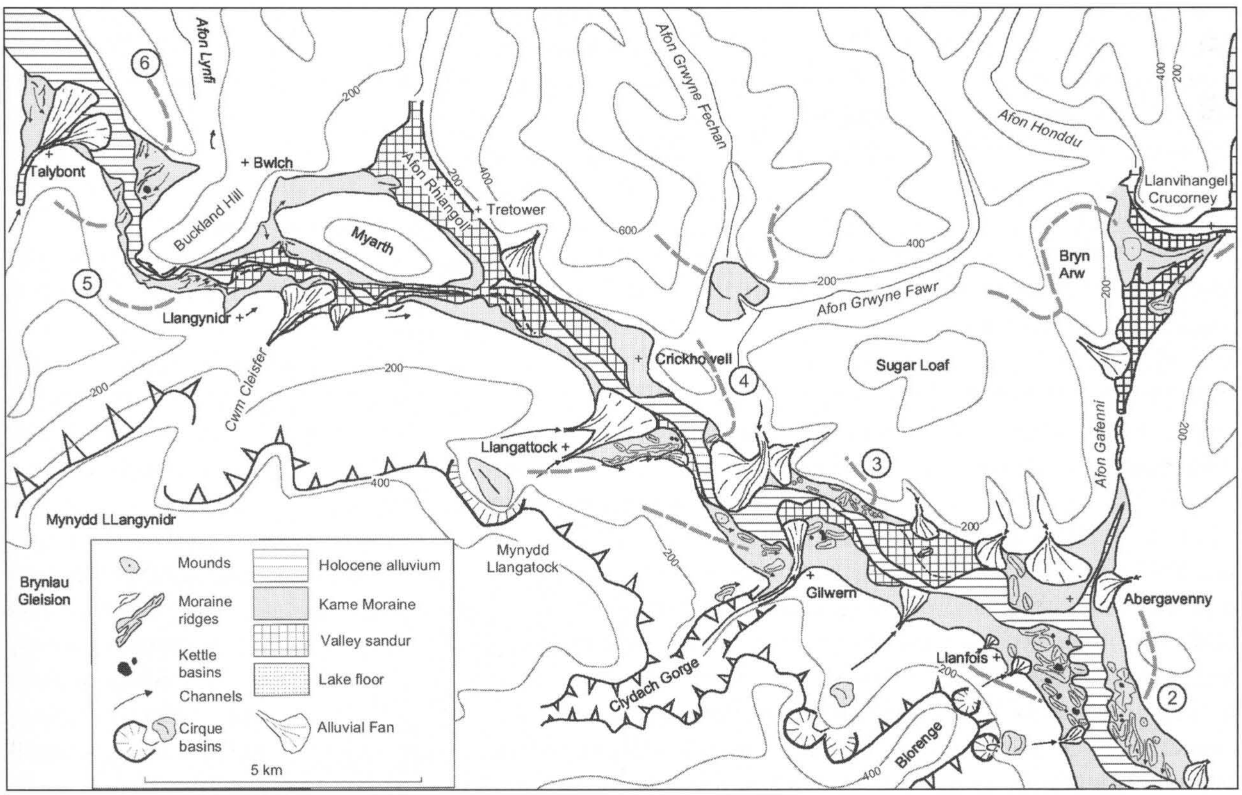

Fig. 8.6 Geomorphological map of the area between Talybont and Abergavenny. For location, see Fig.8.4

Downstream of the gorge the valley floor widens and extensive kame-moraine topography, accompanied by short arcuate moraine ridge segments and associated channels, occurs on the southern valley margin west of Llangynidr, marking an earlier stage in ice-marginal retreat (Stage 5). On the opposite side of the valley a large area of subdued kame-moraine topography extends eastwards below the slopes of Myarth and passes through the shallow col between Buckland Hill and Myarth to extend into the open valley to the east of Bwlch. The kame moraine topography on both sides of the Usk around Llangynidr is fronted by a large valley sandur (Williams, 1968a) that extends east as far as Crickhowell and is joined by an extensive sandur emerging from the Rhiangoll valley, south of Tretower. At the confluence of the Usk and Rhiangoll the extensive sandur surface on the southern side of the Usk forms a flight of terraces. Immediately to the east of Llangynidr a large alluvial fan, draining north from the Carboniferous Limestone escarpment fronting Mynydd Llangynidr and Mynydd Llangattock, via Cwm Claisfer, extends onto the floor of the Usk valley and partly buries the kame-moraine topography and sandur surface on its southern margin.

The town of Crickhowell is built on an extensive area of kame-moraine topography bordering the northern margin of the Usk. Opposite the town, two coalescing alluvial fans drain north from a large cirque basin on the rim of the Carboniferous Limestone escarpment below Mynydd Llangattock. They are bordered on their eastern side by a strip of kame-moraine topography that displays a major set of arcuate cross-valley moraines, separated by parallel channels that define a further ice-marginal retreat position (Stage 4). This stage may correlate with a prominent moraine at the confluence of the Afon Grwyne Fechan with the Afon Grwyne Fawr to the north of Crickhowell.

The geomorphology around Gilwern (Fig. 8.7 A) characterises the relationships between the various sediment-landform assemblages identified throughout the middle Usk. East and west of the town a wide strip of kame-moraine topography flanks the southern side of the valley and carries arcuate moraines that loop across the valley to further moraine fragments on a similar strip of kame-moraine topography bordering the northern side of the river. These mark another ice-marginal retreat position (Stage 3; Fig. 8.6). To the south, the kame-moraine topography is cut through by a large alluvial fan draining from the Clydach gorge, which is a major entrenchment into the face of the Carboniferous Limestone escarpment. A similar fan descends from the mouth of the Grwyne Fawr on the north side of the Usk valley and buries much of the kame-moraine sediment.

East of Gilwern extensive sandur surfaces front the cross-valley moraine and extend towards Abergavenny where they occupy much of the valley floor. The kame-moraine topography is truncated by two separate sandur surfaces that, in turn, are entrenched by the alluvial floor of the modern river. The highest sandur occurs only to the east of the cross-valley moraine, off the down-ice flank of which it grades. A matching, but more extensive sandur, exists at the same height on the northern bank of the Usk and extends downstream towards Abergavenny. The lower sandur has eroded through the cross-valley moraine and is entrenched into the upper sandur. The upper and lower sandur deposits relate to two separate outwash systems: the upper sandur formed in front of the cross-valley moraine and the lower sandur originated from an ice-margin further upstream.

Fig. 8.7 B depicts a down-valley section through the sediment-landform assemblages in the vicinity of Gilwern. None of the boreholes shown on Fig. 8.7 B reached bedrock but two penetrate a stony, red diamict that appears to underlie most of the area. This diamict is overlain by thick red sand that coarsens upwards into pebble and cobble gravel. Upstream of the moraine ridges these gravels are overlain by a stony red diamict that terminates against the outermost moraine ridge and is replaced on the down-ice side by thickening gravels. The upward coarsening signature suggests that the cross-valley moraine was built by a small-scale, localised readvance in which basal diamict was emplaced across a floor of former valley-sandur sediment. At the readvance maximum the melting of dead-ice created irregular kettle basins, which are now occupied by organic sediments in the water-filled basin of 'The Swamp'. Sediment-laden meltwater also deposited a coarse apron of sandur material immediately beyond the ice-margin.

Abergavenny is located on a large area of kame-moraine topography that extends along the north bank of the Usk and into the valley of the Afon Gafenni to the north of the town. Part of this area is buried by large alluvial fans issuing from steep valleys on the south-east flank of the Sugar Loaf. Although much topographic detail is obscured by urban development, a series of arcuate moraine features on the western

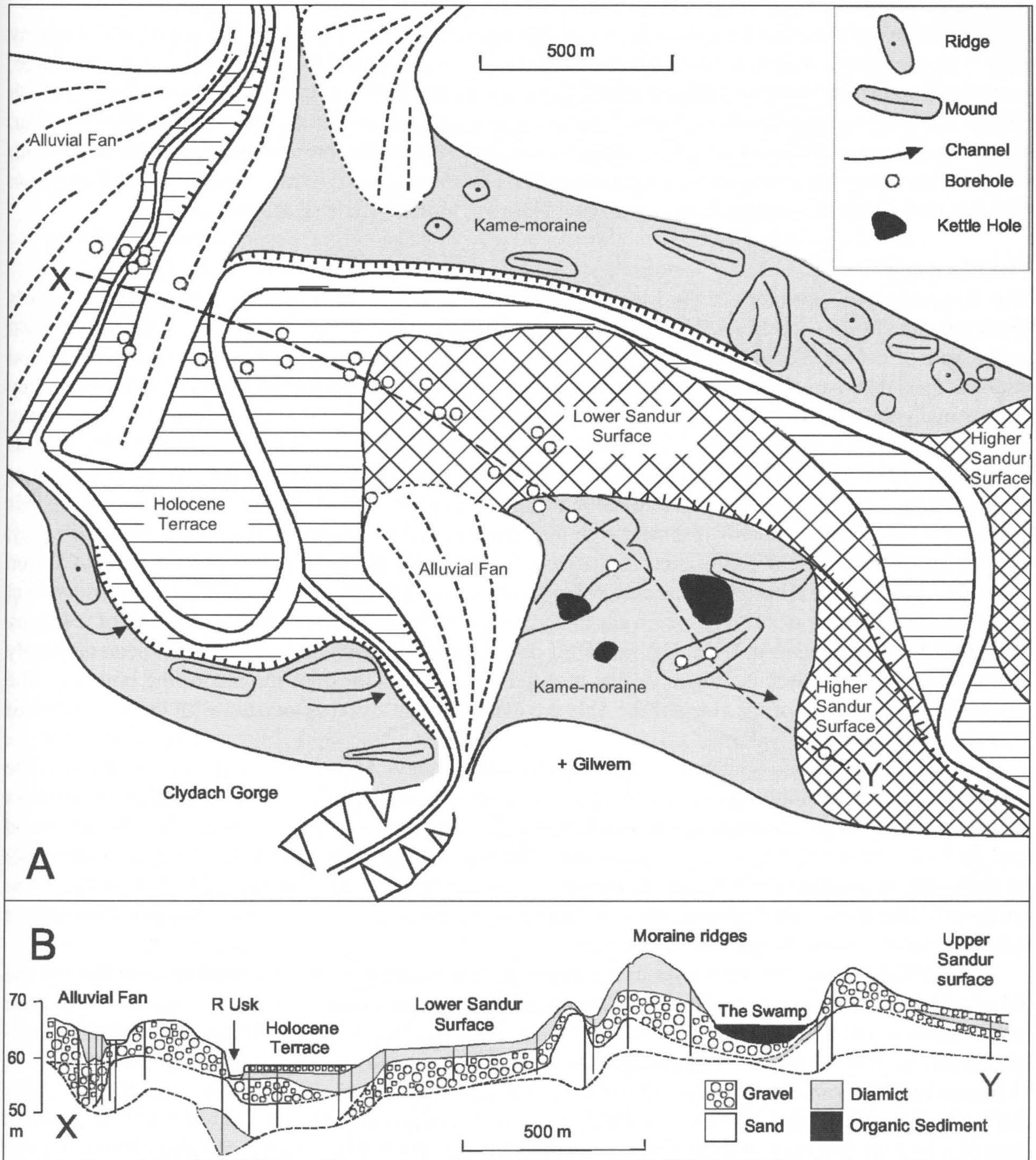

Fig. 8.7. A: the geomorphology of the area around Gilwern showing location of boreholes and section line X-Y. B: cross-section through the cross-valley moraine system at Gilwern along the line of boreholes shown on A

edge of the town correlate with a very complex area of arcuate moraine ridges, intervening channels and large kettle basins in an extensive area of kame-moraine topography on the southern margin of the valley near Llanfoist. These arcuate moraines represent another ice-marginal retreat position (Fig. 8.6; Stage 2). The outer part of this area is overlain by the toes of two small alluvial fans draining from cirque basins located on the rim of the Carboniferous Limestone escarpment below Blorenge Mountain.

The valley of the Afon Gafenni, which is a tributary of the Usk draining south towards Abergavenny (Fig. 8.6), currently carries a diminutive stream. The upper part of the valley is blocked by a very large, compound arcuate moraine at Llanvihangel Crucorney (Lewis, 1970a, Fig. 7.2), the outer face of which grades south into a large ice-front alluvial fan and sandur. To the rear of the moraine the Afon Honddu, following the arc of the moraine ridge, completes an almost perfect U-turn in flow direction from south to north. This suggests that, prior to the creation of this moraine, drainage from the upper part of the Afon Honddu flowed south, via what is now the Afon Gafenni, towards the Usk at Abergavenny.

The Black Mountains

The Honddu valley, upstream of the Llanvihangel Crucorney moraine (Fig. 8.6), flows south-eastwards down the dip slope of the Black Mountains in a trough shaped valley, the lower part of which is known as the Vale of Ewyas. The sides of the valley have been markedly oversteepened, while extensive glacial deposits exist downstream of Capel-y-ffin (BGS, 2002), indicating that glacier ice flowed down the valley and its main Nant y Bwch tributary from the uplands of the Black Mountains. Ice from mid-Wales may also have escaped south from a mid-Wales ice-sheet and breached the escarpment of the Black Mountains at Bwlch yr Efengyl (the Gospel Pass, which is the col between the Honddu and the Wye valleys south of Hay-on-Wye; Fig. 8.1) to flow down the Honddu. Erratics from mid-Wales, which would confirm this possibility, have (as has already been stated in this chapter) not yet been identified in the Honddu valley.

Extensive landslide deposits exist on the west side of the Honddu valley beneath the cliffs of Tarren yr Esgob ('Cliff of the Bishop') in the vicinity of Capel-y-ffin (SO 240310). Continuing slope adjustment occurs on the east side of the Vale of Ewyas under the spectacular Darren immediately north of Cwmyoy (SO 296245). The church at Cwmyoy was built on an unstable landslide slope and has been adversely affected by slope movements: the nave tilts at different angles to those of the rest of the building! The marked oversteepening of the sides of the Vale are evidence of erosion associated with the movement of a powerful glacier down valley.

Glaciers also flowed southwards, down the parallel valleys of the Grwyne Fawr and the Grwyne Fechan. Diamict containing occasional striated clasts, all of which were derived from the Devonian rocks of which the surrounding areas of the Black Mountains are composed, was exposed in 2002 in a road cutting some 50 m upvalley of the gate that leads into the forest plantation near Hermitage (SO 228252) in the valley of the Grwyne Fechan. These deposits indicate that glacial ice formerly flowed down the valley from adjacent higher ground. Cross valley moraines exist lower down the valley, in the vicinity of Llanbedr, where kame topography exists.

An area of kame-moraine exists in the lower portion of the Grwyne Fawr valley, upstream of the Grwyne Fawr/Cwm Coed-y-cerrig trough. Ice previously spilled across that trough to deposit spectacular kame-moraine south of Bettws (SO 300185), in the valley west of Bryn Arw (Fig. 8.6) that is the continuation of the line of the Grwyne Fawr south of the fault guided trough that cuts across the original drainage line. The kame topography, which marks the outermost morphologically identifiable limit in this area of ice originating in the Black Mountains, is evident between Llwyn-gwyn (SO 303187) and the road east of Gott (SO 299185). A sandur/alluvial fan grades towards the floor of the Gafenni valley from the kame region.

In order for a glacier from the Grwyne Fawr to cross the Grwyne Fawr/Cwm Coed-y-cerrig trough, and spill into the Bettws area, the ice must have been in excess of 40 m thick on the floor of the trough.

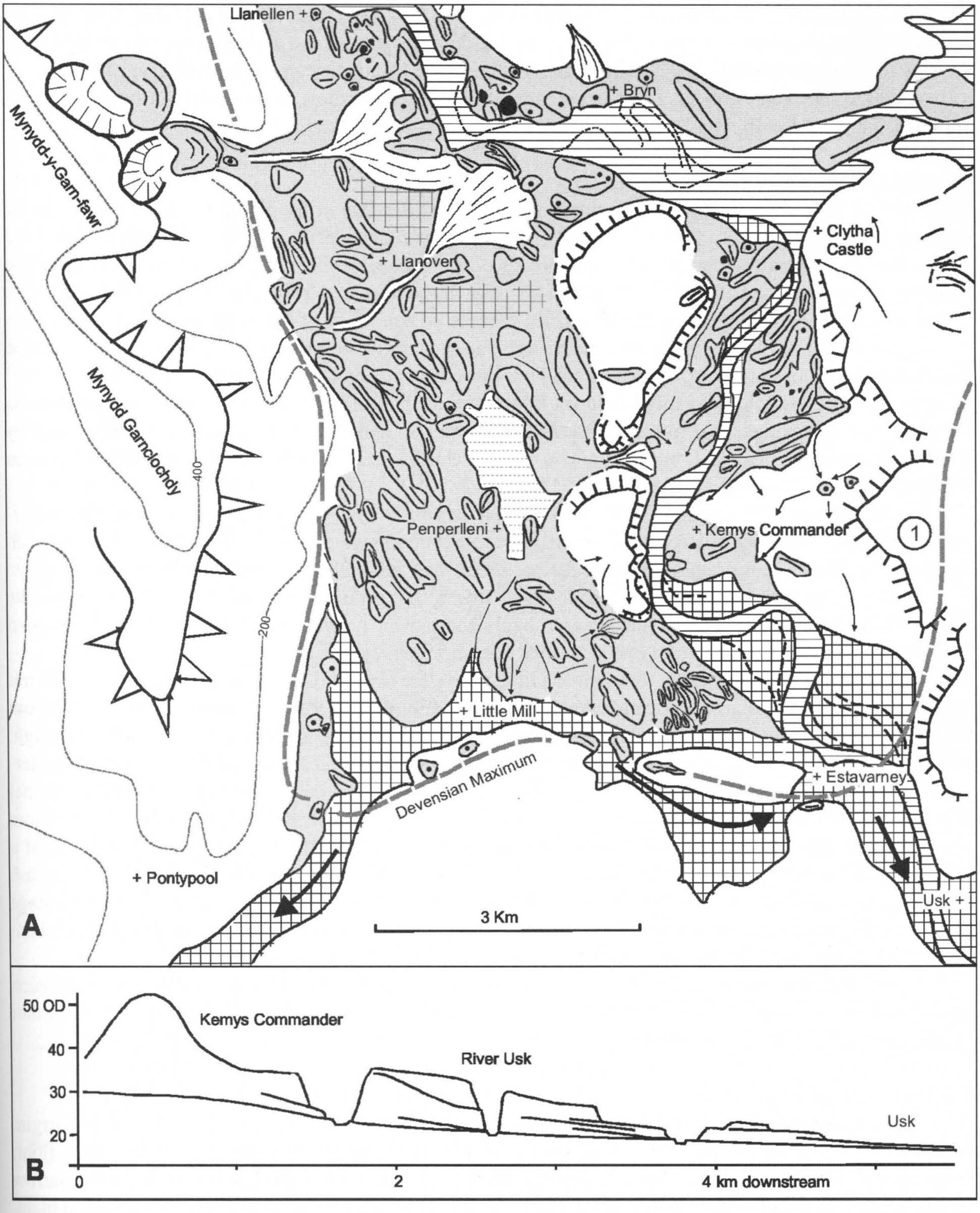

Fig. 8.8. A: geomorphological map of the area between Abergavenny and Usk. For location see Fig. 8.4. For Key see Fig.8.6. The area around Clytha Castle is not mapped in detail and shows moraine crests only. B: section between Kemys Commander and Usk (see A) showing the relationship between sandur surfaces. (Redrawn after Williams, 1968a)

Whether diffluent ice from the Usk valley forced its way into the trough from the Crickhowell area, as Lewis (1970a) suggested, is uncertain. The glacial deposits within the trough may owe their origin entirely to glaciers flowing off the high ground of the Black Mountains and down the Vale of Ewyas.

A cirque moraine exists on the western side of the Honddu valley, as will be described later when similar features in the Brecon Beacons and adjacent areas are discussed.

The middle Usk valley: Abergavenny to Usk

The Usk valley south of Abergavenny widens rapidly along the axis of the Silurian Usk inlier, especially south of Llanellen (Fig. 8.8). Characteristically, when a large valley glacier system passes from a restricted valley to a more open region, the ice-front spreads out to become a piedmont glacier and kame-moraine topography becomes widespread. Subsequent incision by sandur and post-glacial alluviation processes occurs only at limited points along the former ice-margin, so that much of the kame-moraine is preserved.

Downstream of Llanellen the floor of the Usk valley is occupied by a very complex arcuate moraine system composed of numerous, closely-spaced but discontinuous moraine ridges and intervening channel systems in a belt five kilometres wide and six kilometres deep. The continuity of many of the moraines is interrupted by rock ridges that rise along the flanks of the Usk anticline in the centre of the valley and by a steep escarpment to the east, but the arcuate form is clearly seen on both flanks of the valley between Llanover and Clytha Castle (Fig. 8.8 A).

South of Llanellen, where the valley first widens, there is a complex area of ice-disintegration topography, including a number of large kettle basins. The morainic topography is partially buried, further south, by two large alluvial fans draining from cirque basins on the Carboniferous escarpment west of the valley.

East of Penperlleni there is an extensive linear strip of flat ground, underlaid by thick laminated silts and bounded by moraine ridges, that was probably the site of a temporary ice-marginal lake trapped between the retreating ice-front on one side and older moraines on the other.

The outermost moraine ridge terminates a little above the town of Usk and marks the maximum limit reached by Late Devensian ice advance down the valley (Fig. 8.8 A; Stage 1). Meltwater drainage from the ice-margin at its maximum was complex and the outermost moraine is fronted by a flat-floored trough, 500 m wide by 4 km long, running east from Little Mill towards Great Estavarney. The trough is underlain by more than 18 m of coarse gravel and marks the position of a major ice-marginal sandur channel that collected drainage from exit tunnels in the ice and directed it eastwards to join a similar, but more direct, sandur channel draining the eastern side of the ice-margin north of Estavarney. On the west a similar sandur trough passes south from Little Mill towards Pontypool. Neither of these major outwash conduits carry more than diminutive drainage at present and both were active only at the maximum stage of glaciation. As the ice-margin retreated from its maximum the ice-marginal sandur trough was abandoned and meltwater drainage was concentrated on the eastern side of the valley.

South of Kemys Commander the modern flood plain is flanked by a flight of sandur terraces, each graded to a particular retreat ice-marginal moraine, and subsequently incised and abandoned on further retreat (Williams, 1968b; Fig. 8.8 B).

The lower Usk and Nant Olwy

Drift deposits south of the town of Usk are limited, as the area is outside the Late Devensian ice-margin. In the lower Usk (Fig. 8.9 A) drift is restricted to extensive Holocene alluvium across the valley floor, small alluvial fans on the valley margins, some residual pre-Devensian diamict on hill slopes and a narrow sandur terrace on the western margin of the valley at heights of up to 10 m above the modern flood plain. The sandur probably formerly occupied the entire valley floor but has been reduced in extent by subsequent incision. Around Newbridge-on-Usk a fragmentary series of higher terraces occur, up to 30 m above the flood plain.

Very limited subsurface information is available for the region, but to the south, near Newport, Williams (1968b) recorded thick Holocene alluvium overlying sands and gravels of probable pro-glacial origin. Depths to bedrock in the Newport area indicate that the unconsolidated sediments partly occupy buried channel systems, which were probably cut during the Devensian cold stage, when sea levels were lower than at present. The lower Usk valley thus seems to have acted as a major outwash distributary from the Late Devensian ice-margin to the north.

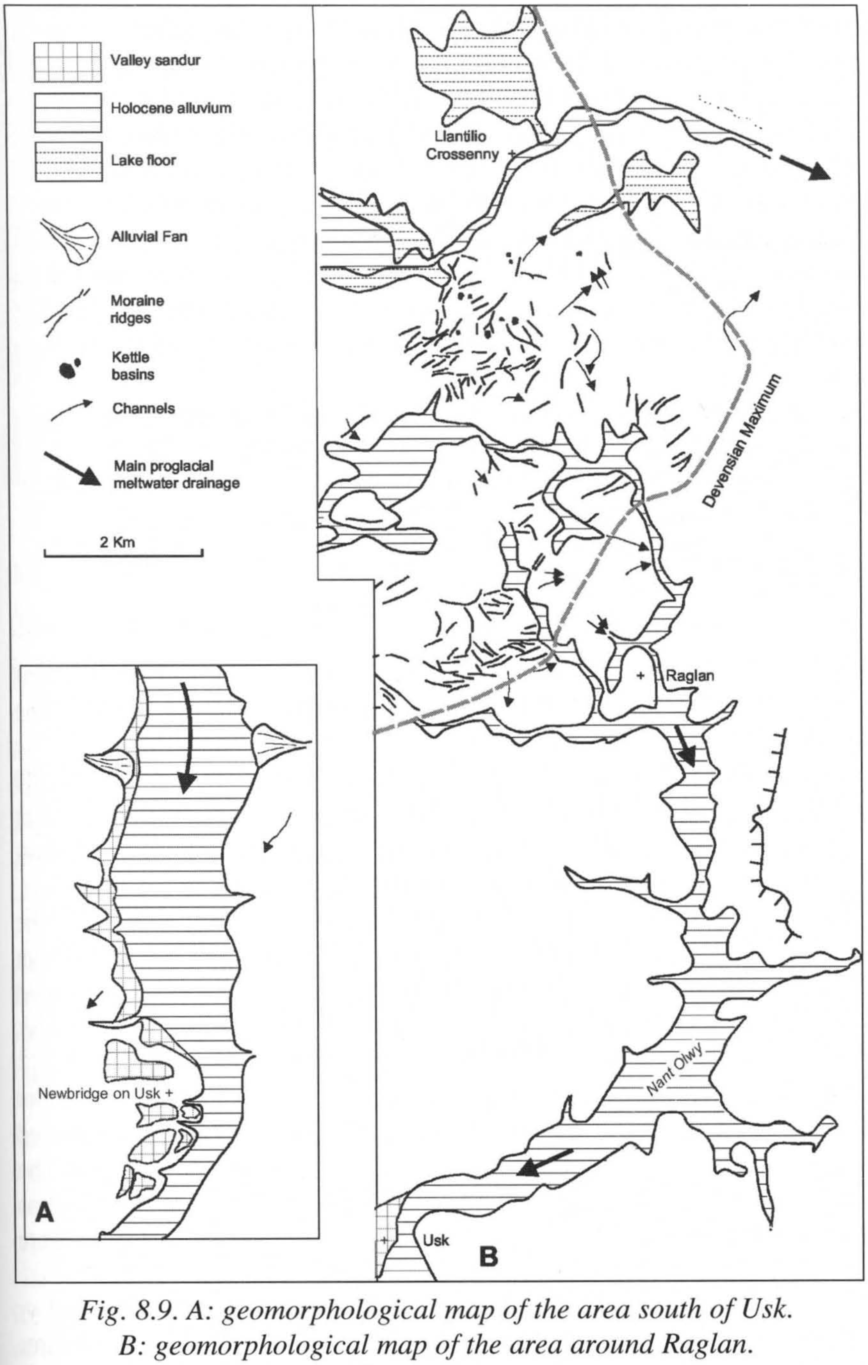

Fig. 8.9. A: geomorphological map of the area south of Usk. B: geomorphological map of the area around Raglan. Note that moraine crests only are shown. For locations see Fig. 8.4. (Redrawn after Williams, 1968a)

The area east of Abergavenny, between Usk and Llantilio Crossenny, lacks drift maps and has not been mapped geomorphologically in detail. A map based on the work of Williams (1968a) is shown in Fig. 8.9 B and identifies ridge crests, kettle basins and channels. These are sufficient to establish that the Devensian maximum limit south of Abergavenny continues north-east towards Raglan and then north towards the valleys draining the southern margin of the Black Mountains. It also establishes that the main marginal drainage in this area fed south into the headwaters of the Nant Olwy, an otherwise diminutive stream, the valley of which is floored with extensive Holocene alluvium, that joins the Usk at the town of Usk. Haslett (2003) indicates that a grey-brown sandy clay deposit, that may be head, underlies the alluvium and is sometimes separated from it by a palaeosol and an overlying fenwood peat. Drainage also passed eastwards along the Afon Troddi (River Trothy), a tributary of the Wye, to the east. Extensive areas of laminated clay near Llantilio Crossenny indicate the existence of large ice-marginal lake systems during early stages of retreat from this maximum ice limit.

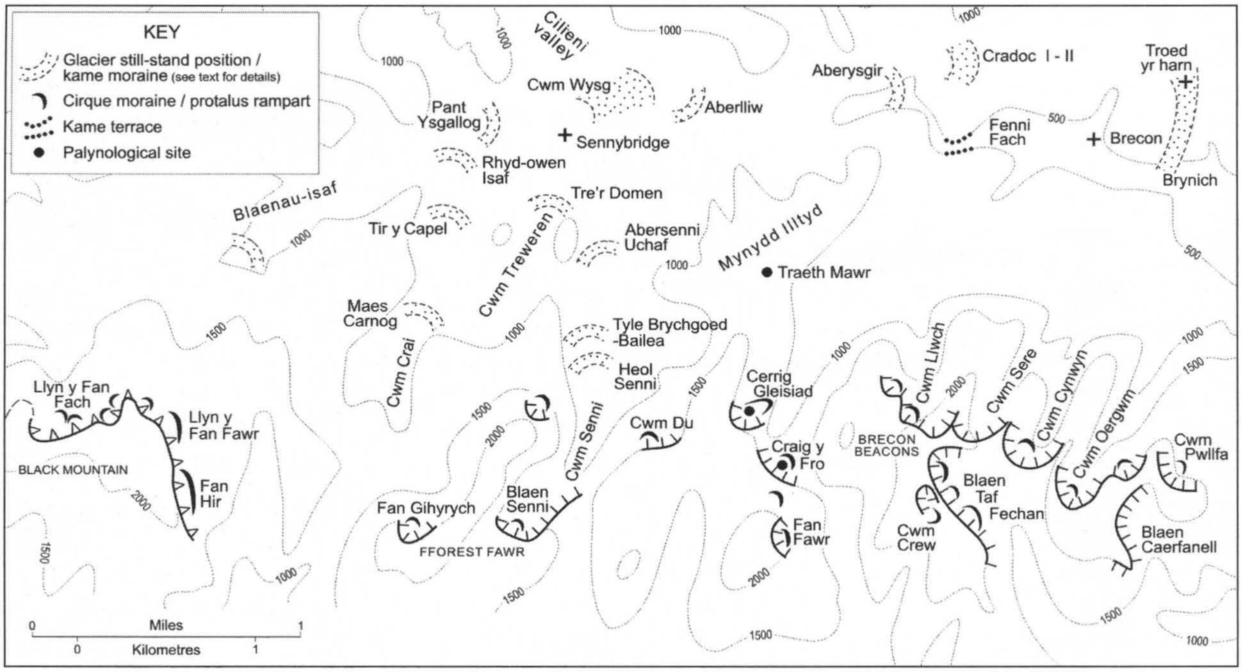

Fig. 8.10 Ice-marginal positions in the upper Usk and its right-bank tributary valleys, cirque moraines/protalus ramparts in the Black Mountain/ Fforest Fawr/Brecon Beacons area and palynological sites at Traeth Mawr, Cerrig Gleisiad and Craig y Fro. Contours in feet. (Redrawn after Ellis-Gruffydd (1972) with additions and emendations)

The upper Usk valley

This is defined as the Usk valley upstream of Talybont. Parts of the valley have been mapped by Williams (1968a), Lewis (1970a, b) and by Ellis Gruffydd (1972, 1977). The drift deposits of the area were in the process of being remapped by the British Geological Survey in 2002/3.

Both Williams (1968a) and Lewis (1970b) identified a morainic system in the Groesffordd region (SO 076280), north of Llanfrynach and some 3 km east of Brecon, where kame-moraine topography is evident on the northern side of the valley. Melt water channels south of Groesffordd, aligned towards the south-east, apparently carried drainage from the main kame-moraine area downstream, to the main Usk valley.

South of the Usk, on the flanks of the tributary valleys of the Cynrig and Menascin, which join the Usk west and east of Llanfrynach respectively, lie other meltwater channels. The channels near Tynllwyn (SO 063240) and opposite Cantref church (SO 058254), start and end abruptly and probably formed subglacially, under ice from the uplands of the Brecon Beacons that extended northwards into the Usk valley. These channels probably formed when the Usk glacier ended down-valley of Llanfrynach.

Excavations during the building of the Brecon by-pass, in the vicinity of the present road roundabout near Brynich (SO 068278; Fig. 8.10), at the junction of the A40 and the A470, exposed matrix supported diamict containing Old Red Sandstone clasts, some of which were more than three metres in length. The diamict overlay bedrock and probably formed through deposition beneath a glacier fed from source areas south and west of Brecon. No erratics from areas of Silurian and Ordovician rocks north of Brecon were identified in the diamict.

A spectacular area of kame topography occurs upslope of Brynich, immediately west of Troed yr harn (SO 066298; Fig. 8.10) where it is best seen in the fields downslope of the B4602 road. Exposures adjacent to that road when it was being improved revealed only Old Red Sandstone clasts, suggesting that

Fig. 8.11 The Tyle-crwn meltwater channel. (Photo: C. A. Lewis)

the ice responsible for the kame topography came essentially from the Brecon Beacons/Fforest Fawr uplands, rather than from northerly sources where other rocks occur.

Unconsolidated deposits are widespread north-east of Troed yr harn, on the plateau below the uplands of Yr Allt (SO 085307). Impressive meltwater channels lead from this plateau into the Dulas valley, which leads into the Llynfi and thence into the Wye (Fig. 8.1). They include the channel in which the house and buildings of Tyle-crwn farm (SO 096327) is situated (Figs. 8.1, 8.11). The Tyle-crwn channel starts suddenly on the slope above the farm buildings and ends above the level of the floor of the Dulas valley. This suggests that it formed subglacially, carrying meltwater from the plateau ice-mass downslope and into ice that occupied the Dulas valley. The meltwater then presumably drained into the Llynfi valley and thence into the Wye.

The Brecon Beacons/Fforest Fawr ice formerly flowed eastwards down the Dulas valley, leaving lateral moraines on the southern side of that valley between Llanfilo (SO 120332) and the vicinity of Tregunter (SO 135339). The glacier terminated in a ridge some 14 m high, at an altitude of 160 m, located south and upslope of the Tregunter farm dams (around SO 130335, where the ridge is partially wooded; Figs. 8.2, 8.12).

In 2002 a gravel pit near the summit of the ridge exposed 4 m of faceted and striated Old Red Sandstone clasts in a sandy and finer matrix. No clasts that might have come from the upper Wye valley were identified. The slopes of the ridge are steep, exceeding 30°, and appear to have formed as ice-contact slopes. Portions of the ridge that are lobate in plan probably formed as a result of minor ice advances in which the ice-front pushed against, and steepened, the ridge. They indicate that the ridge formed in association with ice that was not entirely dead (i.e. decaying and motionless). The ridge itself was formed by ice-marginal deposition of sediments that were mainly carried out of the ice by meltwater. The inner side of the ridge terminates in a grass-covered depression that was formerly occupied by the tongue of ice that was responsible for ridge formation.

Fig. 8.12 The ice-contact ridge south of Tregunter Farm. Glacial ice occupied the ill-drained flat area to the left of the ridge, pushing against its inner (tree covered) face to form a steep ice-contact slope. Sediment from the melting glacier, much of which was transported by meltwater, was deposited at and adjacent to the ice front to form this spectacular depositional ridge. This landform deserves protection as a site of major scientific interest. (Photo: C. A. Lewis)

On the north side of the Dulas valley, near Felin-newydd (SO 117359) and downslope of the A470, lies an area of kame and kettle topography at a similar altitude to the dead-ice ridge at Tregunter. Meltwater channels lead eastwards from this area towards Bronllys (SO 142350) and the Llynfi-Wye rivers (Fig. 8.2). Williams (1968a) and Lewis (1970a) considered that the Dulas valley had been invaded by ice from the Wye valley, but the lithology of the deposits exposed in the Tregunter ridge suggests that they interpreted the morphological evidence incorrectly. The Tregunter dead-ice ridge and the kame and kettle deposits near Felin-newydd probably mark the terminus of an ice advance from the Brecon (Usk valley) area .

Fossiliferous Silurian erratics in diamict, exposed in 2002 in excavations for the erection of a farm building at Porthamel (SO 159350), near Talgarth in the Llynfi valley (Fig. 8.2), (to which the Dulas is a tributary), indicate that ice from mid-Wales invaded the Llynfi downstream of Bronllys, although the correlation between that ice and the ice responsible for the Tregunter/Felin-newydd deposits has not been established. Neither have the latter deposits been correlated with any of the ice-marginal stages in the Usk valley.

Ellis-Gruffydd (1972) mapped seven 'moraines and associated pro-glacial and ice-contact forms' in the Usk valley upstream of Brecon (Fig. 8.10). He believed that they relate to still-stands superimposed upon the recession upvalley of the Usk glacier, or possibly to 'slight readvances superimposed upon a general retreat.'

The nearest of these features to Brecon is the area of kame topography on which the golf course was built at Cradoc (the Cradoc Moraine, I and II). Excavations associated with construction of the golf course showed that the glacial deposits at that site contained Old Red Sandstone clasts, such as might have come from the upper Usk valley and from the southerly tributaries of that valley that drain the high ground of the Fforest Fawr and the Bannau, on the border with Dyfed. No Silurian or other clasts from the Eppynt or other areas north of Brecon were identified. This indicates that the kame topography formed in association with the Usk valley glacier as ice from the upper Usk pushed into the Cradoc-Penoyre col north of Brecon, where it stagnated (Lewis, 1970b). The kame topography is at an altitude of some 220 m and lies about 70 m above the present floor of the Usk valley at Aberysgir, indicating that the Usk glacier was at least 70 m thick at the time that the kame sediments were deposited at Cradoc.

Ellis-Gruffydd (1972) considered that a terrace at Fenni-Fâch (SO 020289) also related to the main Usk valley glacier at the time that the Cradoc kames were forming. Subsequent ice-marginal positions in the upper Usk valley, he believed (Fig. 8.10), were at Aberysgir (SO 000297), Aberlliw (SN 954295), Cwm wysg-ganol, Cwm wysg-uchaf and Pant Ysgallog, at all of which (except for Cwm wysg-ganol) kame and kame terrace topography exists. Lewis (1970b) showed the Cwm wysg-ganol and -uchaf features as one and the same moraine and stated that lateral deposits of this moraine were traceable to both the Senni and Cilieni valleys, on either flank of the Usk. Ellis-Gruffydd (1972) maintained that subsequent to the formation of the Pant Ysgallog moraine glacial ice disappeared from the Usk valley west of the confluence of the Crai valley with that of the Usk. The implication of his statement is that the Usk glacier upvalley of Brecon was fed mainly by ice from the Fforest Fawr, which was limited to the Senni, Treweren and Crai valleys when the glaciers in those valleys were no longer able to reach the Usk.

Ellis-Gruffydd (1972) correlated cross-valley kame topography in the Senni, Treweren and Crai valleys, believing that those features marked the same retreat stages in each valley. In the Senni valley he identified former glacier still-stand positions at Abersenni uchaf (SN 932262); Tyle Brychgoed-Bailea (where kame topography does not exist; SN 925245); and at Heol Senni (SN 925234), where the hamlet is situated on the kame-moraine (Fig. 8.13). Only one still-stand position is identifiable in the adjacent valley of Cwm Treweren: at Tre'r Domen (SN 916267) where kame topography exists on the floor and lower sides of the valley. Road works on the eastern side of the valley formerly exposed sections in this sediment, which is of local provenance and derived from areas upvalley of the moraine. Ellis-Gruffydd (1972) identified three glacier still-stand positions in Cwm Crai: at Rhyd-Owen-isaf (SN 895281); Tir-y-capel and Maes Carnog. At all three sites kame and kame terrace topography exists. Lewis (1970a) also recorded a still-stand position near Blaenau-isaf (SN 845258) in the Hydfer valley, west of Cwm Crai, although Ellis-Gruffydd (1972) wrote that this does 'not exist.'

Both Lewis (1970a,b) and Ellis Gruffydd (1972,1977) agreed that the Usk glacier above Brecon was fed by ice from the Fforest Fawr. Lewis (1970b) also remarked on the morainic topography at Pentre'r-felin (SN 915304) in the lower Cilieni valley, which is a left bank tributary that joins the Usk immediately downstream of Sennybridge, and on deposits at Pentre-bach (SN 906330) 'and odd patches of drift further upstream' in the Cilieni valley. He suggested that they might have been formed by mid-Wales ice

Fig. 8.13 The low ridge on the valley floor in the middle distance is a kame moraine, on which the hamlet of Heol Senni is located. This landform developed at a former ice front of the glacier that occupied the upper portion of Cwm Senni towards the end of the Late Devensian. The view looks down valley, towards the inner side of the kame moraine. The flattish area grazed by sheep in the foreground was covered by the glacier when the ice front lay at Heol Senni, as was the rest of the headward portion of Cwm Senni and the adjacent uplands. (Photo: C. A. Lewis)

from the Vale of Irfon breaching the col at the head of the Cilieni valley and thence passing into the Cilieni and possibly entering the Usk. Other morainic features in the valleys of the Eppynt indicate that they were formerly occupied by ice moving southwards towards the Usk, but it is not certain whether this northern ice contributed to the Usk glacier when it was restricted to the upper Usk valley.

During the decay of ice from the Senni valley, that had extended as a piedmont glacier onto the Mynydd Illtyd plateau on the northern side of the Fforest Fawr, hummocky kame-like topography was formed on that plateau. This suggests that much of the piedmont ice decayed in situ, as dead (motionless) ice. Depressions within the hummocky terrain still contain shallow pools of water and extensive organic deposits, such as that of Traeth Mawr (SN 968255) which has been subjected to palynological analyses (J. J. Moore, 1970; Walker, 1982) and radio-carbon dating (Walker, 1980).

Deglaciation and subsequent cirque glaciation

Deglaciation

Evidence from four sites within the Wye and Usk valleys, and their immediate surroundings, indicates when those regions were deglaciated. North of Hay-on-Wye, at Rhos goch common (SO 195484; Fig.8.1) in the Bach Howey valley, an ill-drained area contains organic deposits that began to form in the Late Glacial (Bartley,1960). These deposits rest on what Bartley (1960) considered to be glacial till. They include pollen indicative of a Late Glacial interstadial vegetation. Although no absolute dates have been obtained from Rhos goch, glacial ice must have melted prior to a Late Glacial interstadial at that site.

Llyn Mire (Cors y Llyn, SO 015553) is located about 2 km south of Newbridge-on-Wye and contains organic deposits that P. D. Moore (1978) believed accumulated in 'a kettle hole in the till.' They lie about 1 km north-east of the kame-like mounds and the esker-like feature already described from near Penmincae, and are probably part of the same ice-marginal topography, marking a still-stand or minor advance in the overall retreat of the Wye glacier. The lower organic sediments at Llyn Mire evidence a typical Late Glacial vegetational sequence that includes an interstadial. Although no absolute dates have been obtained, the site must have been free of glacier ice before that Late Glacial interstadial.

On the southern side of the Usk valley, at Traeth Mawr (SN 968255; Fig. 8.10) on Mynydd Illtyd (as already described), a thickness of over 627 cm of organic deposits exists within a dead-ice depression. J. J. Moore (1970) showed that they evidence a Late Glacial vegetation sequence, passing up into Flandrian (Holocene) organics. Walker (1980, 1982) re-examined the organic deposits. He concluded that the basal organic deposits are 'likely to represent a pioneer stage in vegetational succession on to bare and unstable substrata that had recently been vacated by Late Devensian ice.' Above them, stratigraphically, lies palynological evidence of a Late Glacial interstadial that was followed by stadial conditions.

Walker (1980) obtained a radiocarbon date that calibrates to 13,822 BP (within a one sigma range of 14,006–13,637 BP; Fuls, 2003) for the base of the Late Glacial interstadial deposits, indicating that the site had been deglaciated before that date. (The remaining dates shown in this chapter have been calibrated, unless stated otherwise). Walker considered that the uncalibrated date (11,160 +/- 140 BP) was 'over 1,000 yr younger than age determinations from comparable Late-Glacial Interstadial deposits in Scotland, the Lake District, North Wales and Cornwall', possibly due to the incorporation of younger carbon into the sediment that was dated.

During the 1960s Lewis and J. J. Moore cored a basin at Bryniau Gleision (SO 086155; Fig. 8.6), at an altitude of 475 m on the plateau above the head of Dyffryn Crawnon, which is a right bank tributary of the Usk. Moore identified palynological evidence of at least one Late Glacial interstadial complex in the deposits (Lewis, 1970a), but never published his findings. The site was re-investigated by Robertson (1988), who confirmed that a Late Glacial interstadial is represented at that site.

Robertson (1988) obtained a radiocarbon date that calibrates to 20,023 BP (within a one sigma range of 20,598-19,448 BP; Fuls, 2003) for the base of the interstadial, which suggests that the Bryniau Gleision plateau was ice free by that time. He considered the date to be too old, due to the organic sediments from which it was derived being from a limestone area and consequently subject to the effects of 'hard-water' due to uptake by plants of carbonate-rich water, resulting in errors in radiocarbon dating (Lowe and Walker, 1997). Robertson suggested, on the basis of age determinations from similar pollen horizons elsewhere in Britain, that a date of 13,000–14,000 would be more probable. Whether the radiocarbon date or Robertson's suggested age-range is correct is unknown, but the uplands of Bryniau Gleision were certainly deglaciated before the Late Glacial Interstadial evidenced there, probably by 14,000 BP and possibly as early as 20,000 BP.

The evidence from the four sites so far investigated in the Wye/Usk region indicates that the landscape was deglaciated before 14,000 BP. Deglaciation may even have occurred as early as 20,000 BP on the limestone plateaux south of the middle Usk, if the suspect evidence from Bryniau Gleision is correct. The morphologically 'fresh' looking glacial landforms in the region suggest that the last ice-sheet/valley glacier glaciation of the area was Late Devensian in age. No sediments have yet been found within the area covered by this chapter to indicate the maximum age of that glaciation or of any earlier glaciation of the region.

Subsequent cirque glaciation

Mounds of unconsolidated material near the heads of valleys such as Cwm Crew and Cwm Oergwm in the Brecon Beacons, and essentially linear ridges such as those under Fan Fawr in the Fforest Fawr and Fan Hir near the borders of the Fforest Fawr and Black Mountain, bear witness to cirque glaciation subsequent to ice-sheet decay in those uplands (Fig. 8.10). Evidence of cirque glaciation north of the Old Red Sandstone escarpment, but within the catchments of the Usk and Wye, has not yet been presented.

Reade (1894) was the first to draw attention to the cirque moraines of the area, in a paper on Cwm Llwch in the Brecon Beacons (Fig. 8.14). Subsequently Robertson (1933) suggested that some of the

Fig. 8.14 Llyn Cwm Llwch, Brecon Beacons. Notice the moraine, which encloses the lake, and that the lake lies in the most shaded (and hence the coldest) portion of the valley-head, where there is least incoming solar radiation. The cirque glacier responsible for deposition of the moraine, and for cutting the over- deepened basin occupied by the lake, lay between the moraine and the shadowed area of the backwall. Reade (1894-5) wrote that '... the sun has traced out this moraine, and settled its alignment and position in the larger Cwm or valley'. (Photo: C. A. Lewis)

debris accumulations in the mountains north of Merthyr Tydfil were glacial moraines while others had accumulated at the foot of large snow patches (as protalus ramparts).

The cirque moraines and protalus ramparts of the Brecon Beacons were mapped by Lewis (1966, 1970a,b). Ellis Gruffydd (1972, 1977) remapped the area and extended coverage westwards to the Fforest Fawr and Mynydd Du. Robertson (1988) examined part of the area already covered but also added new information from the Black Mountains. Shakesby (1992, 2002) presented an overall synthesis of the information amassed by previous workers. He also published, with Matthews, two detailed studies of restricted areas in the Brecon Beacons/Fforest Fawr (1993, 1996). In 2001 Carr presented a glaciological approach for the discrimination of small ridge systems in the Brecon Beacons.

Fig. 8.10 depicts the location of depositional ridges and mounds in the Brecon Beacons/Fforest Fawr/ Black Mountain uplands south of the Usk valley. In such valley heads as those of Cwm Cynwyn and Cwm Oergwm there are a number of arcuate ridges on the valley floors that have been interpreted as glacial moraines deposited by cirque glaciers. There is also a small depositional ridge on the south-western side of the valley head at Cwm Oergwm that may well have formed as a protalus rampart.

The Late Glacial, as evidenced by deposits within the Traeth Mawr basin (Walker, 1980, 1982), consisted of an early phase in which the ground was exposed by deglaciation, an interstadial beginning by about 14,000 BP and ending around 12,450 BP in which there were a number of climatic oscillations, and then a stadial in which mainly inorganic sediments were deposited. The stadial was terminated by Post-Glacial, Holocene, conditions, that began at about 11,300 BP. This Younger Dryas Stadial is well-known in Britain and Shakesby and Matthews (1993), in relation to their work within the Brecon Beacons National Park, refer to it as the Loch Lomond Stadial.

Within the mountains, south of Traeth Mawr, Walker (1980) obtained radiocarbon dates from the lowest organic sediments within the basin that is enclosed by the cirque moraine at the foot of Craig y Fro (SN 973207), in Glyn Tarell. He also obtained radiocarbon dates for the equivalent sediments in Cwm Cerrig-gleisiad. In both cases Walker believed that organic sedimentation

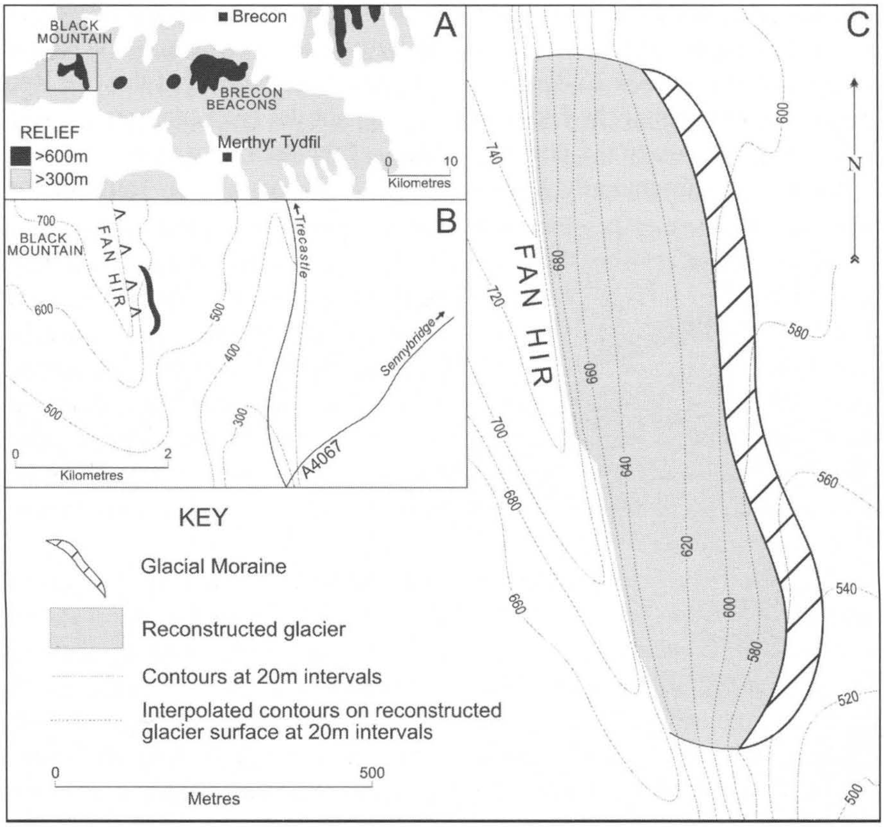

Fig. 8.15 Reconstruction of the former cirque glacier at Fan Hir, Black Mountain. Notice how the glacier was located in the area shaded from the afternoon sun by the uplands of Fan Hir, off which snow was probably blown to accumulate on the glacier surface. (Redrawn after Shakesby and Matthews, 1993)

began in the Holocene, and dates to about 11,400 BP at Craig y Fro and to 12,769 BP at Cwm Cerrig-gleisiad. (As already stated, these are calibrated dates, according to the calculations of Fuls, 2003). Walker therefore concluded that both basins were occupied by glaciers during the Loch Lomond Stadial, as Lewis (1966, 1970a, b) had suggested. The calibrated date from Cwm Cerrig gleisiad is puzzlingly old and apparently pre-dates the Younger Dryas/Loch Lomond Stadial.

Lewis (1966, 1970a, b) suggested that some of the cirques, such as Cwm Cerrig-gleisiad, where there is an outer arc of indistinct ridges, might have harboured cirque glaciers at an earlier stage of the Late Glacial as well as during the Loch Lomond Stadial. Shakesby and Matthews (1996), however, showed that the outer ridges at Cwm Cerrig-gleisiad owe their origin to landsliding associated with collapse of part of the backwall of the cirque. They believed that the steep-sloped ridges close to the southern back-wall of the cirque, and within the limits of the less obvious outer ridges, were of glacial origin and formed during the Loch Lomond Stadial.

At Fan Hir (SN 834198), on the extreme west of the Fforest Fawr, there is a ridge that is circa1.2 km long and that is separated from the escarpment of the Black Mountain that backs it to the west by a 'gully-like hollow' (Shakesby and Matthews, 1993). The ridge is aligned essentially parallel to the escarpment and varies in height from 25 m to less than 3 m at its northern extremity. It has proximal and distal slopes of up to 32°. Towards its southern extremity the ridge curves towards the escarpment (Fig. 8.15). Some of the clasts within the ridge are striated, indicating that they have been transported in a glacier system. Shakesby and Mathews (1993) therefore believe that the ridge formed as a moraine in association with a small glacier that was located beneath the escarpment during the Loch Lomond Stadial. Their reconstruction of this palaeoglacier depicts it as covering an area of 0.28 km^2 and having an equilibrium line altitude (ELA) of 623 m. (The ELA is the dividing line between the accumulation and the ablation area on a glacier).

Robertson (1988) showed that, during the Loch Lomond Stadial, one small glacier existed in the Black Mountains in addition to those that had already been identified on the Brecon Beacons, Fforest Fawr and Black Mountain. This was at Tarren yr Esgob, where there is a moraine below the east-facing cliffs on the west side of the Vale of Ewyas near Capel-y-ffin (SO 240315). The glacier covered an area of 0.07 km^2 and had an ELA of 443 m. The other Loch Lomond Stadial glaciers of the region varied in size from that of Tarren yr Esgob to that of Craig Cerrig-gleisiad, which covered 0.38 km^2, while the glacier of Llyn y Fan Fach was almost as large as that of Cerrig-gleisiad. The average glacier size was 0.17 km^2, compared with an average of 0.5 km^2 in Snowdonia (Gray, 1982) and 0.85 km^2 for the Lake District of England (Sissons,1980). Obviously, therefore, environmental conditions were only just severe enough for the existence of small cirque and valley-side glaciers in the Usk/Wye uplands region during the Loch Lomond Stadial.

Outside the limits of the Loch Lomond Stadial glaciers there was active reworking by periglacial processes of the already deposited glacial and other unconsolidated sediments. This produced stepped slopes, with flights of solifluction terraces on them, as between Fan Gyhirych and the A4067, south of Crai (SN 895210). On steep slopes under the escarpment of the Brecon Beacons, Black Mountain and elsewhere, talus accumulated. The absence of ice-wedge casts and pingo remnants, however, suggests that permafrost was uncommon in the area during the Loch Lomond Stadial, and that the glaciers and snow beds that existed in the region may have owed their existence largely to snow-blow rather than to severely cold climatic conditions.

Conclusion

Mid-Wales was probably glaciated on at least two occasions during the Pleistocene, but evidence for the earlier glaciation is forthcoming only for areas outside the limits of this chapter. The glacial sediments and landforms discussed in this chapter appear to be of Late Devensian age. The scale of glaciation during the

Late Devensian varied from ice-sheet to cirque glacier in size. During deglaciation the mid-Wales ice-sheet shrank to become two major valley glaciers, those of the Wye and the Usk. The oscillations of those glaciers are imperfectly known, as are the limits of mid-Wales and Brecon Beacons/Fforest Fawr sourced ice. The age of deglaciation, especially in the Wye drainage basin, is a matter for debate and much absolute dating is still needed. There is therefore much scope for research in the region covered by the Usk and Wye rivers and their tributaries.

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The glaciations of Wales and adjacent areas

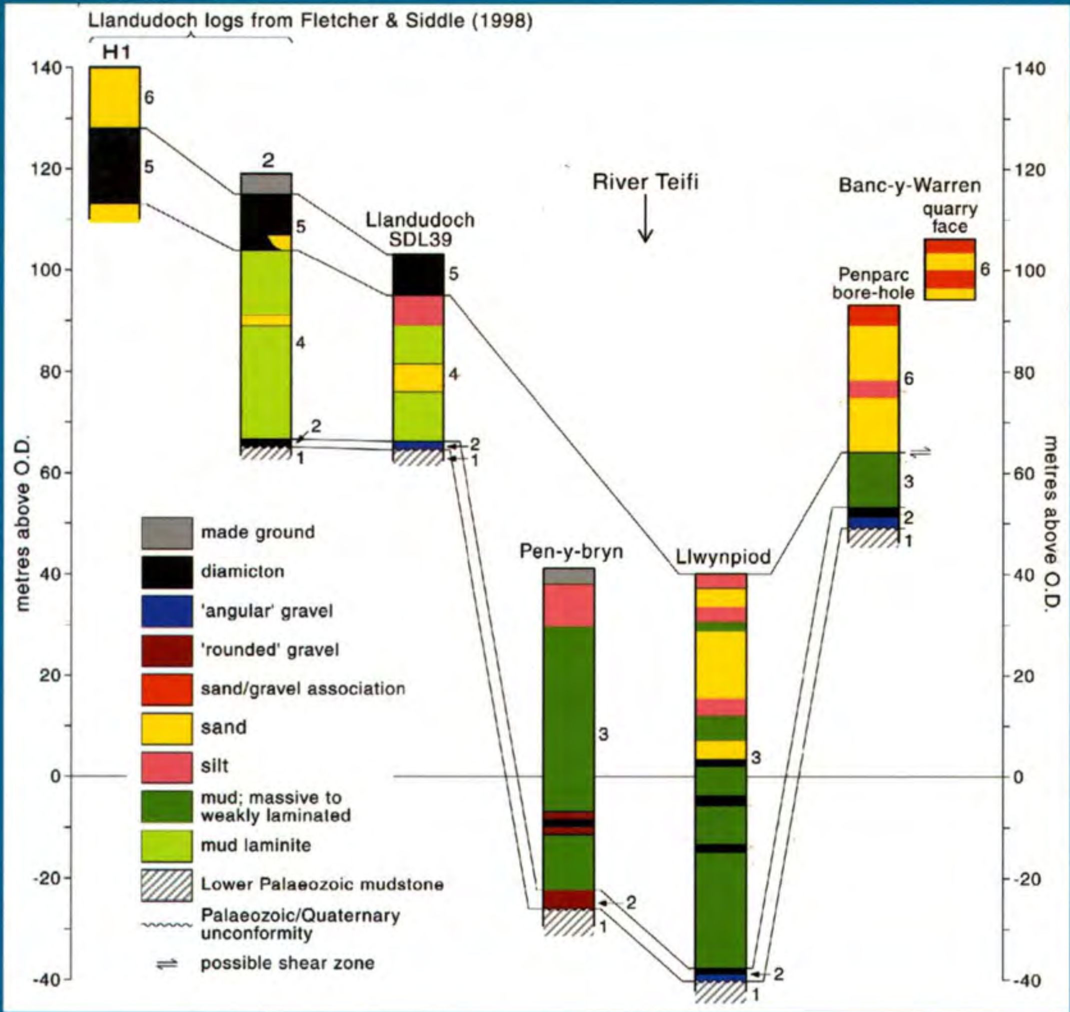

Fig. 7.4 Simplified stratigraphic logs for cored buried valley-fill successions in the lower Afon Teifi region. (From Hambrey *et al.* (2001); figure reproduced with the permission of Wiley Interscience.)

Fifteen leading Geographers and Quaternary scientists have combined in this book to present the latest information on the Quaternary development (and particularly the glaciations) of Wales, the Cheshire-Shropshire lowlands, Severn valley, South West Peninsula, the east coast of Ireland and the Irish Sea and adjoining Celtic Sea basins. This area is divided into ten regions, each of which is the subject of a chapter by one or more specialists on that region.

The initial chapter provides an introduction to the development of glacial studies and shows how events in Wales and adjacent areas relate to global and astronomic systems. The second chapter provides an introduction to glacial deposits, Quaternary stratigraphy and Welsh climate history.

The book is interdisciplinary in nature, since an understanding of glacial events, of glacials and interglacials, stadials and interstadials, requires interdisciplinary collaboration. The main purpose of the text is to provide a compendium of existing knowledge on the glaciations of Wales and surrounding districts so as to provide readers with the latest information on the topic and to form a benchmark for future studies.

The book is clearly written, illustrated by over 100 Figures and Tables and is intended for the general well-educated reading public who are interested in their physical environment as well as for specialists. This book should enable university and senior school students, especially of Geography, Environmental Science, and Geology, to gain a sound appreciation of the evolution of the glaciated landscapes of Wales and surrounding areas.

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