THE UNIVERSITY OF HULL

River Planform, Soil Stratigraphy and the Temporal and Palaeoenvironmental Significance of Terraced Valley Fill Deposits in Upland Scotland, with Specific Reference to Glen Feshie, south-west Cairngorms

being a Thesis submitted for the degree of

Ph. D.

in the University of Hull

by

Melanie S.E. Robertson - Rintoul, (B.A. Hons)

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CONTAINS PULLOUTS

When I have seen....

the firm soil win of the watery main,
Increasing store with loss, and loss with store;
When I have seen such interchange of state,
Or state itself confounded to decay;
Ruin hath taught me thus to ruminate...

Wm. SHAKESPEARE Sonnet No. 54

SUMMARY

Summary of Thesis submitted for Ph. D. degree
by Melanie S.E. Robertson - Rintoul

on

River Planform, Soil Stratigraphy and the Temporal and Palaeoenvironmental Significance of Terraced Valley Fill Deposits in Upland Scotland, with Specific Reference to Glen Feshie, south-west Cairngorms

River terraces form prominent features of the valley floor morphology of many upland valleys in Scotland. Little is known, however, about valley floor landform development in many of these valleys. Previous studies have generally explained the terraces as the outwash products of meltwaters from the most recent cold periods. Detailed investigation of a major valley in the Scottish Highlands challenges, for at least one site, this well established concept and proposes the occurrence of at least three phases of Holocene terrace development.

The correlation of terrace surfaces has traditionally been based upon the construction of height-range diagrams. An alternative approach to terrace correlation and dating is developed in this study using data from Glen Feshie, south-west Cairngorms. Terrace fragments are numerically classified and objectively grouped using quantitative soil-stratigraphic data. Principal Components Analysis and a hierarchical clustering technique numerically define five soil-stratigraphic units and place these on a relative time scale. Various methods of absolute dating control permit association of these units with five phases of terrace development. These are placed at 13,000, 10,000, 3,600, 1,000, 80, radiocarbon years BP.

Comparison of palaeochannel networks preserved on the terrace surfaces suggests that these phases of terrace development have been associated with changes in channel pattern morphology. A unified approach to analysis of channel pattern morphology is developed and from this a new technique for palaeohydrological interpretation of gravel-bed streams. A segment density index is developed which allows total sinuosity to be predicted from just a part of the braided channel network. Application of these techniques to the Glen Feshie terraces demonstrates a trend for an overall decrease in discharge from the oldest terrace surfaces to the present day.

Assessment of these landform changes within the context of known environmental fluctuations in the Cairngorms suggests that the early-mid Holocene was a period of relative landscape stability while the late Holocene was characterised by increasing instability. These changes may have been associated with the changes in river behaviour. However, spatial variation in the depth of the fill/bedrock interface may produce a discontinuous river response to changing environmental conditions.

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CHAPTER 1

INTRODUCTION

In many drainage basins in upland Scotland there is a late Pleistocene legacy of stored glacial and glaciofluvial sedimentary landforms, these occupying the valley floors and lower valleyside slopes as valley fills. The landforms which comprise such fills are generally sensitive to changes in climate, vegetation and sediment supply (Church and Ryder, 1972). In Scotland they have experienced erosion and reworking by hillslope and river processes since their deposition at the end of the Late Devensian cold period about 13,000 years ago (Price, 1980). Episodic erosion, reworking and evacuation of sediment from the fill in many of the valleys in upland Scotland has resulted in spatially complex patterns of incision and aggradation and consequent changes in valley floor sediment surface elevation. These changes have been recorded in the form of gravelly river terrace sequences. As a result, river terraces, often found in association with tributary alluvial fans, form prominent features of the valley floor morphology of many upland valleys in Scotland.

A traditional model of Scottish environmental change and landform development has emphasised the contrast between the geomorphically active cold periods, the Dimlington (Late Devensian) and Loch Lomond (Zone III) Stadials, and the relatively inactive postglacial (Price, 1980). Discussing the terrace sequences of highland Scotland remote from the effects of marine base-level changes Sissons (1967, 1976, 1982) has perpetuated this argument and extended it to the fluvial environment. Sissons was initially discussing the development of terraces in the context of the formation and drainage of ice-dammed lakes, but subsequently extrapolated his model of terrace development to include almost all upland Scotland river terrace sequences (Sissons, 1982). He suggested that most river terraces in upland areas are outwash terraces deposited by powerful proglacial rivers associated with waning ice margins. He further proposed that the floodplains bordering many of the rivers were also formed proglacially and have been infrequently inundated by floods of diminished Holocene streams. In the vicinity of the Gaick Plateau, central Grampian Highlands, Sissons describes outwash terraces in many of the valleys radiating from the plateau. The low-level terraces in Glen Tromie, Glen Truim and the Allt Lorgaidh, a tributary to the River Feshie, are all described as outwash terraces formed about 10,000BP in response to deglaciation of the Loch Lomond ice cap. The merging of the terraces with hummocky moraine is seen as evidence for the age of the landforms, despite the presence of several terrace surfaces above present river level.

Outwash terraces undoubtedly form prominent elements in the terrace sequences in many valleys in upland Scotland, both in the context of the ice-dammed lake and in valleys which have not experienced glacial lake drainage. For example, Sissons (1979,1982) describes sequences of terraces in Glen Spean, Glen Roy and Strath Bran. These terraces have formed in response to falling levels of ice-dammed lakes as glacier wastage proceeded. Many of these terraces terminate in ice-contact slopes or merge with drift mounds thus providing convincing evidence of formation near a glacier margin. Kirby (1969) identified twelve stages of downcutting in the episodic evolution of the River Esk, Midlothian; eleven of these stages were related to the northward retreat of glacier ice. Each glacier retreat stage is marked by a terrace level terminated upstream by an ice-contact front.

Sissons' hypothesis of a proglacial origin for many of the terrace sequences in highland Scotland has received support from several authors. Young (1977, 1978) suggests that many of the terrace suites in valleys tributary to the Spey have been produced by meltwaters from deglaciation of the Late Devensian ice. These include the terrace sequences of the Truim, the Tromie, the Calder and the Dulnain. Ballantyne (1984) notes that the evidence presented by Sissons confirms the importance of fluvial activity during the Loch Lomond Stadial. He further notes that this evidence lends support to a Loch Lomond age for many relict alluvial fans and river terraces in upland valleys outside the limits of Loch Lomond Stadial glaciers.

Recent work has led, however, to the development of the concept of an unstable late glacial/early Holocene, and a stable early-mid Holocene followed by an unstable late Holocene for valley floor landform development in some upland valleys. Evidence from the Howgill Fells, Cumbria (Harvey et al., 1981) and the North York Moors (Richards, 1981; Richards et al., in press) may point to at least two phases of Holocene valley alluviation coincident with vegetation changes. A date of 6270 radiocarbon years BP from in-situ alder stumps buried by 1.5m of fluvial sands and gravels in Dovedale Griff, North York moors, suggests an early mid-Holocene date for one phase of valley alluviation in Dovedale Griff in the North York Moors. A date of 900 radiocarbon years BP on wood remains from an alluvial fan deposit in another valley in the North York Moors, Jugger Howe Beck, suggests a second phase of alluviation in the late Holocene. This age is coincident in time with that suggested by

Harvey et al. (1981) for a phase of late Holocene terracing in the Howgill Fells. This late Holocene phase of terracing may be attributed to vegetation and landuse changes following settlement of the area in Viking times. Similarly, Crampton (1969) has attributed late Holocene terracing in the upland valleys of southeast Wales to increased sediment supply consequent upon forest clearance and increased runoff due to climatic deterioration.

Recent palynological studies undertaken in the vicinity of the Cairngorms (O'Sullivan, 1974a, 1974b, 1976; Walker, 1975b; Rapson, 1984, 1985) support findings from upland England of late Holocene vegetation changes which may reflect both forest clearance by man and climatic deterioration. To date, research concerning the effects of such environmental changes on valley floor landform development has been lacking in upland Scotland. Consequently, the possibility of low-level river terraces related to hydrological and sediment supply changes consequent upon such Holocene environmental changes has received very little attention in the literature. This lack of research concerning the development of Holocene landforms in upland Scotland may be because the evaluation of Holocene landform development in many parts of upland Scotland has been hindered by the absence of dated reference levels (Walker and Lowe, 1980).

The differentiation between the effects of climatic changes or man-induced vegetation changes is difficult to determine unequivocally. However, the hydrological and sediment supply changes which must have accompanied the late Holocene vegetation changes in the Grampian Highlands would be expected to have had some effect on slope and fluvial processes, and, as in upland

England, evidence for these would be expected to be recorded in changes in the level of the valley floor sediment surface.

Bluck (1976) and Ferguson and Werritty (1983) have described modern floodplain development for active gravelly rivers in upland Scotland. Evidence of the activity of modern floodplain development has been carried into the historical context by McEwan (1985) who describes historical river channel changes in both Speyside and Deeside over the past 250 years. McEwan cites evidence for both changes in braiding intensity as a response to increased storminess, and the presence of abandoned flood surfaces. River incision has left some of these surfaces as the most recently abandoned valley floor surfaces in the upland valleys. Such evidence of modern floodplain development and the creation of recent low-level terraces questions the concept of the widespread occurrence of proglacially-formed floodplains in upland Scottish valleys, and suggests the possibility of valley floor sediment surface elevation changes in response to the changing environmental conditions of the Holocene.

Comparison of the stratigraphic relationships between fluvial surfaces within one river system and ultimatly between river systems, directly raises the issue of the correlation of alluvial deposits. Methods for correlating terrace fragments and establishing their relative position in time have traditionally been morphological. The primary criterion for correlating terrace fragments has been longitudinal continuity of the terrace remnant, and an associated tendency for remnants of one terrace level to occur at a similar relative height above present river level (Leopold, Wolman and Miller, 1964). Correlation of terrace fragments into surfaces of one age on this basis is usually made by the plotting of a height-range diagram.

The establishment of a chronology for correlated fluvial surfaces has generally been based on radiometric and palaeo-ecological data where available, and on the assumption that higher terraces are older. As a consequence of the lack of dated reference levels, the establishment of a chronological framework within which to interpret the terrace sequences of upland Scotland has been particularly difficult. It has therefore been based on morphological data derived from levelling of terrace fragments and construction of relative time scales from height-range diagrams (Kirby, 1969; Young, 1976, 1977, 1978; Sissons, 1979, 1982, Sissons and Cornish, 1982). The chronology of evolution has been provided by the physio-graphic association of some terrace fragments with moraines and ice-contact deposits from the Dimlington and Loch Lomond Stadials.

The conclusions of both Sissons (1979, 1982) and Young (1977, 1978) concerning the development of the terrace sequences in Glen Spean, Glen Roy, Strath Bran and Strathspey were based on analyses of height-range diagrams constructed from field-levelled elevation data points. A similar approach has been adopted in a study of the terraces in Glen Feshie, which are unrelated to glacial drainage. Young (1976) conducted a morphometric analysis of the terrace sequence in Glen Feshie. On the basis of his height-range diagram he identified four groups of fluvial terraces. The highest terraces were suggested to have been formed proglacially by meltwaters from downwasting of the Late Devensian ice sheet about 13,000BP. Two groups of low-level terraces have been identifed on the basis of the height-range diagram. These lower terraces are suggested to have been formed by meltwaters from the Zone III or Loch Lomond deglaciation, about 10,000BP. The age and origin of the low-level terraces has been proposed on the basis of the suggested merging of the low-level terraces with the toe of a Loch Lomond age alluvial fan.

However, Thornes and Brunsden (1977) remark that the crudest and often most unreliable techniques for correlation and relative dating of geomorphic surfaces are those involving height data alone. Morphological correlations of alluvial surfaces may produce correlations and chronologies of evolution that are misleading or even incorrect (Chapter 4). As correlation and relative dating of terrace fragments provide the basis for interpretation of the terrace sequence, the criteria used for correlation and relative dating are fundamental to:-

(1) the reconstruction of elevation changes in the valley floor sediment surface;
(2) the palaeohydrological reconstruction of the former channel systems preserved on the terrace surfaces;
(3) the subsequent palaeoenvironmental interpretation of the terrace sequences;.
(4) correlation of terraces between tributaries and main valleys; and inter-basin correlation of terraces of the same age.

If the initial criteria used to establish the stratigraphic relationship between fluvial surfaces in the valley floor sequence produce correlations that are incorrect then the subsequent chronology of evolution and interpretation of the development and environmental context of a terrace sequence must also be questioned. Criteria additional to morphological data may be needed to provide a more satisfactory basis for terrace fragment correlation and the subsequent interpretation of the terrace sequence.

Particularly because it is now evident that complex relationships exist between phases of cutting and filling in different parts of a drainage basin, correlation and dating of river terraces is more usefully based on the stratigraphic relation of the terraces to dated deposits or age-calibrated parameters measuring the degree of post-depositional weathering of the terrace deposits (Born and Ritter, 1970). When weathering profiles and residues are used for correlation and relative dating of geomorphic surfaces the "somewhat dubious methods based on morphology alone give way to the more certain methods of stratigraphy" (Thornes and Brunsden, 1977 p.32).

Stratigraphy involves the study of the form, geographic distribution, chronological ordering and correlation of strata. Soil stratigraphy is that branch of stratigraphy that deals with the chronological ordering of soil profiles (Finkl, 1980). Soil profiles develop as a result of the operation of the soil-forming processes through time. In consequence an investigation of the degree of soil profile development on the surfaces of alluvial landforms created by episodic landform development facilitates correlation of landform elements belonging to the same age-range. In order to correlate landforms of the same age-range using degree of soil profile development, measurement is required of physical and chemical properties of the soil profiles that are diagnostic of the soil-forming processes and their temporal development. These measurements then provide a strong quantitative basis for correlation and relative dating of alluvial deposits. If calibration of the sequence of soil profiles with some absolute dating control is possible then the sequence of soil profiles may be placed on an absolute time scale. Data may then be obtained on the length of time since stabilisation of alluvial landform surfaces and the lengths of time between periods of stability. Such a calibrated sequence of soil profiles may then be used to predict the ages of undated terrace sequences in other basins with similar environments.

The lack of detailed study of Holocene valley floor landform development in upland Scotland, the difficulties of correlating terrace fragments using height-range methods and, as Ferguson (1981) notes, the dearth of knowledge concerning the palaeo-hydrological implications of the terraced fluvial sediments, provided the impetus for the present study. This study has the following major aims:-

(1) to re-examine the concept that most of the Scottish river terraces in upland valleys are outwash terraces related to deglaciation at the end of the most recent cold periods;

(2) to re-examine the traditional morphological methods of terrace fragment correlation and relative dating;

(3) to test a multivariate statistical method of terrace correlation based on the principles of soil stratigraphy;

(4) to develop and test a new appoach to the palaeo-hydrological analysis of gravelly fluvial deposits.

Glen Feshie in the south-west Cairngorms was selected for detailed investigation for the following reasons:-

(1) The valley has a well developed sequence of terraces which range in height from 25m to about 1.5m above present river level.

(2) The terrace deposits consist of coarse to very coarse massive gravel units. These may be regarded as forming a

more or less constant parent material for surface soil development. This is an important prerequisite for any quantitative soil-stratigraphic analysis.

(3) The valley is well situated with reference to sites where detailed palynological investigations have been carried out for the Cairngorms area. The history of environmental change since deglaciation as it is recorded in bog and lake sediments is therefore well documented.

(4) All the extensive terrace fragments possess well developed palaeochannel networks on their surfaces which can be used for palaeohydrological analysis.

(5) Modern floodplain development is currently being investigated in Glen Feshie (Ferguson and Werritty, 1983) and data are therefore available for contemporary river processes. This permits comparison of past river processes, as reconstructed from the fluvial sediments, with the processes of the contemporary river channel.

An alternative methodology to the traditional height-based method of river terrace correlation is therefore developed in this study, utilising data from the Glen Feshie terrace sequence (Chapter 3). The terrace fragments in Glen Feshie are numerically classified and objectively grouped using multivariate soil-stratigraphic data collected from the soil profiles developed on the terrace surfaces. The surface soils developed on the terrace fragments are shown to form a genetically related podzolic sequence of soils in different stages of soil profile development. Some absolute age control enables the chronosequence of soils to be tentatively age-calibrated and a reassessment made of the ages of the river terraces and the development of the terrace sequence.

An alternative method of correlating terraces using several morphometric variables is also examined, in Chapter 4. The results of this analysis are compared with that from the more traditional height-range method as carried out by Young (1976) in Glen Feshie. Both of these methods are then compared with the soil-stratigraphic approach to terrace correlation and relative dating.

Very little research has been undertaken that considers the palaeohydrological implications of terraced fluvial sediments in upland Scotland, although Maizels (1983a, 1983b) has attempted to extend the model of proglacial terrace formation by carrying out palaeohydrological studies of the terrace sediments and palaeochannels. In her model, terracing of valley fill deposits is suggested to occur during a period of deglaciation in conjunction with a reduction in braiding intensity and a fluctuating balance of water and sediment supply around a threshold state. The model has been applied to the terrace levels of the River North Esk, south-east Grampians. As with other terrace sequences in the Grampian Highlands there is little absolute dating control for the terraces in the sequence, and it is assumed that the majority of the terraces are outwash surfaces.

The development, evaluation and application of palaeohydrological techniques to the analysis of change in fluvial landscapes over timescales ranging from over 10,000 years to a few hundred years addresses issues that are central to geomorphology as a discipline concerned to explain the processes of landform development (Richards et al. in press). The immediate problem in palaeohydrological investigation is to develop and test workable relationships between parameters defining the flow and sediment transport regimes, and variables of channel form and sedimentology, of present day streams. These relationships may then be applied to past river channel traces and sediments in an attempt to evaluate the hydrological context of fluvial landform development over the longer time scale.

Current palaeohydrological investigations are usually approached using one of two methodologies. The methodology used depends on the channel pattern morphology of the fossil stream channels in question. These are :-

(1) the application to fossil channel forms of empirical relationships developed from present day streams between discharge and parameters of channel form, such as meander wavelength or channel cross-sectional area; this channel-scale approach has been applied to meandering palaeochannels;

(2) a palaeohydraulic approach which involves the application of hydraulic relations between clast size and various parameters of the flow, to individual clasts from former channel system deposits. This approach provides estimates of instanteous velocity and flow depths and has been applied to coarse-grained flood deposits and to gravel-bed braided stream deposits.

The first method reconstructs discharges of a relative narrow frequency range. However, the relationships developed for meandering streams require careful application both in the selection and measurement of channel form, and in the application of the appropriate equation for the river type being examined (Chapter 5). The second approach has been evaluated in the context of bedrock canyon flash-flood deposits (Costa, 1983) and has been used widely to reconstruct former flows at the stage of incipient gravel motion for former gravelly braided channels. However, inherent within this palaeohydraulic approach are a number of fundamental assumptions which, in the light of recent research, cannot be supported, and, in particular, the method does not provide reliable information on the frequency of reconstructed flows. This is discussed in Chapter 5 where methods of palaeohydrological and palaeohydraulic reconstruction in common usage are tested using data from the present River Feshie.

Recent discussions of the facies approach to the analysis of fluvial sediments, as a means of reconstructing past fluvial environments, suggest that prediction of channel types from small-scale vertical lithofacies analysis is unrealistic. Rather, it is argued that attention should be directed at a larger scale, comparable to the overall channel dimensions. This more geomorphological approach involves consideration of macrobar forms (Bridge, 1985; Miall, 1985). Miall suggests that fluvial styles, or the channel pattern morphology of past rivers, should be classified by a combination of their architectural elements, together with channel sinuosity and braid parameter. Bridge reiterates the necessity to reconstruct past river form using macrobar forms and further suggests that past discharges must be quantitatively reconstructed from channel/bar and channel-fill deposits. Closely paralleling the views expressed in the sedimentological literature, Rotnicki and Borowka (1985) have suggested that investigations of gross channel pattern may prove more valuable in a palaeohydrological context than detailed reconstructions from grain size parameters of the deposit.

At present there does not exist a quantitative relationship between channel pattern morphology at the channel scale and the controlling variables of channel pattern development which can be applied to both meandering and braided channels and which allows differentiation of different degrees of braiding. Rather, fluvial geomorphologists have tended to discuss the differences between channel pattern types through discriminant analysis of nominally classified pattern types. However, thresholds between channel pattern types are emphasised by discriminant functions separating preclassified data clusters.

Bridge (1985) notes that the sedimentary processes of different channel pattern types form a continuum, and that the similarities between their deposits are more significant than their differences. There should then be a complete gradation of channel geometry, flow, and sedimentary processes between channel patterns with this depending on the imposed discharge and sediment load. It is argued in the present study that the discriminant function approach to channel pattern types has inhibited the development of a general model which can relate channel pattern morphology as a continuum to the dominant controls of channel pattern development, at least in free alluvial channels.

In Chapter 6, therefore, a unified approach to channel pattern morphology is adopted. The physical basis for such an approach is examined in the light of previous analytical and theoretical studies, and a new empirical relationship is developed between a dimensionless parameter of channel pattern (total sinuosity) and three of the controlling variables of channel pattern development.

Having established the existence of a channel pattern continuum for free alluvial gravel-bed rivers a new approach to the palaeohydrology of gravel-bed streams is developed, at the larger, channel scale as opposed to the scale of the individual grain. In Chapter 7 a new regression relationship is derived which allows prediction of the stream power and ultimately a flood approximately equal to the 2-2.3 year flood of past channel systems given the total sinuosity of the former channel system and a measure of particle size.

One of the most difficult problems to overcome when attempting a palaeohydrological analysis of past braided streams is the uncertainity of the width of the active channel zone. Because of the uncertainty as to the extent of post-depositional reworking of higher terraces, total sinuosity cannot be measured directly from extant channel traces. In Chapter 7 an examination of several parameters of braided stream networks shows that the total sinuosity of a braided stream is closely and linearly correlated to a dimensionless measure of braided stream networks. This measure, the segment density index, may be used to predict the total sinuosity of the whole former active channel zone from fragmentary braided channel traces on terrace surfaces, on the assumption that the fragments are random samples of the former more extensive system. Confidence limits may be attached to the estimate of total sinuosity thus obtained, by assuming that the measured palaeochannels are a random sample from a normal population.

These two new relationships may be used to predict the stream power and flood discharge approximately equal to the 2-2.3 year flood from only three parameters, total sinuosity, a measure of particle size, and a measure of the terrace slope. This method is tested using data from the upper braided reach of the River

Feshie in Chapter 7 and is then applied to the palaeochannels of former generations of the River Feshie in Chapter 8. The palaeoenvironmental implications of the former channel systems are discussed.

This thesis seeks to provide a rigorous methodology for correlation and relative dating of river terrace fragments and the palaeohydrological analysis of terraced gravel-bed braided stream deposits. It also provides an inte grated analysis of the environmental and chronological context of valley floor landform development in one major valley in the Scottish Highlands during the Holocene. The methodology developed in the thesis should allow further work to continue in which examination of terrace sequences between tributaries and mainstreams, and between valleys can take place.

CHAPTER 2

GLEN FESHIE : THE STUDY AREA, LANDFORMS AND SEDIMENTS

2.1 Location

Glen Feshie is a deep south-north trending valley that forms the western boundary of the Cairngorm Mountains (Figure 2.1). The upper River Feshie, downstream from the River Eidart, flows westward through a deep glacial breach (Linton, 1949) and is confined by bedrock for much of its course. At about 400m above sea-level the River Feshie turns north into the main Feshie valley to flow across valley fill deposits until, just below the lower braided reach (Figure 2.1), and at about 260m above sea-level, it leaves the deposits of the valley fill and enters a deep, 1km long meltwater gorge cut through bedrock. This gorge has represented the base-level for the River Feshie since its formation at the end of the Late Devensian cold period, the Dimlington Stadial (Young, 1975).

The River Feshie is a major, fifth order tributary of the River Spey and drains about 240km^2 of the western Cairngorms. Most of the drainage basin lies at 600-1000m above sea-level. The valley floor of the main south-north trending valley falls from about 400m at the upper braided reach to 260m at the lower braided reach and has a gradient of 0.012. Five tributaries enter the River Feshie along the length of the study reach. These are the second order tributaries of the Allt Lorgaidh, the Allt Coire, the Allt Garbhlach, and the Allt Fhearnagan, and the fourth order Allt Chomhraig (Figure 2.2). Stream orders are derived from the 1:25 000 Ordnance Survey maps.

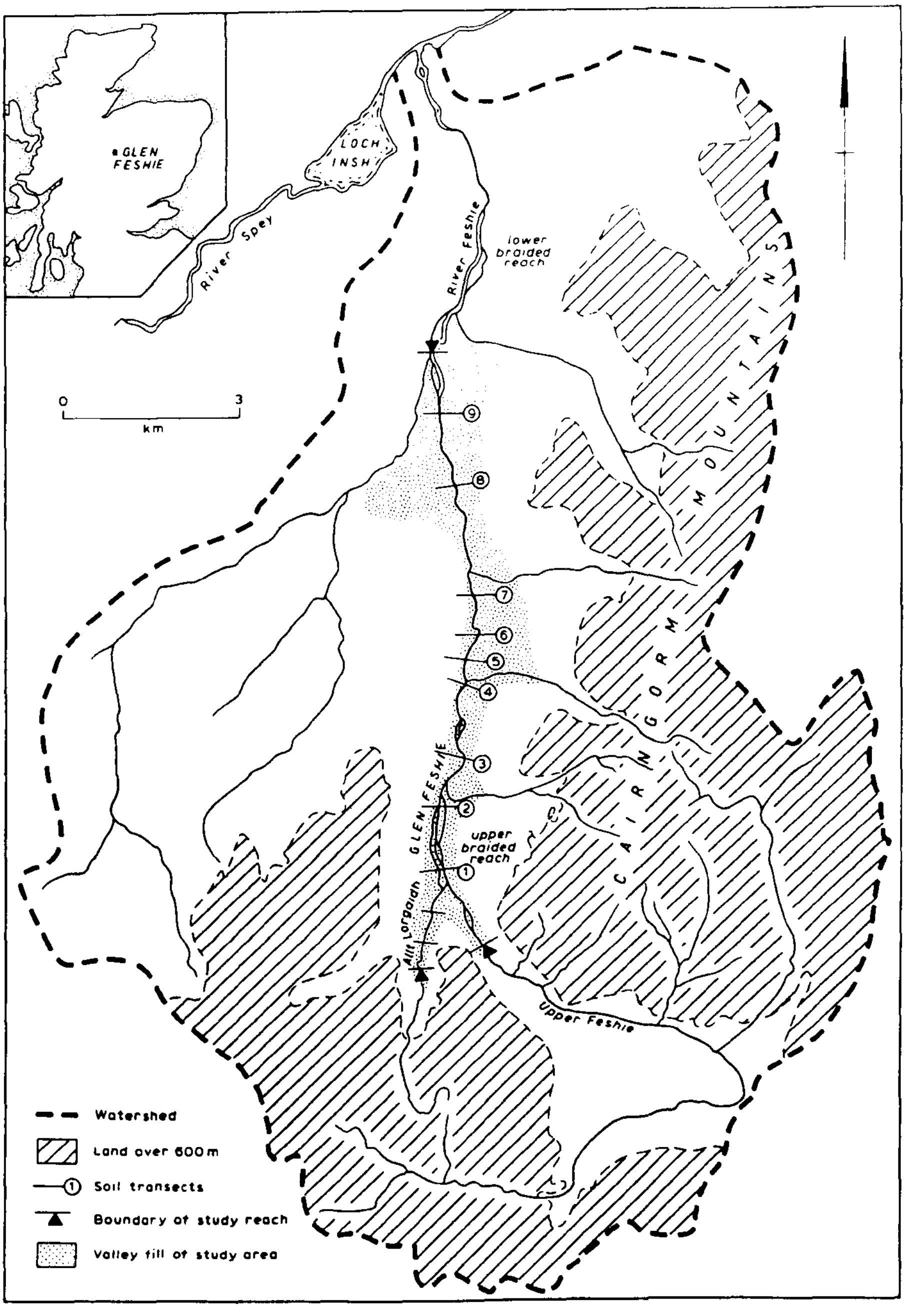
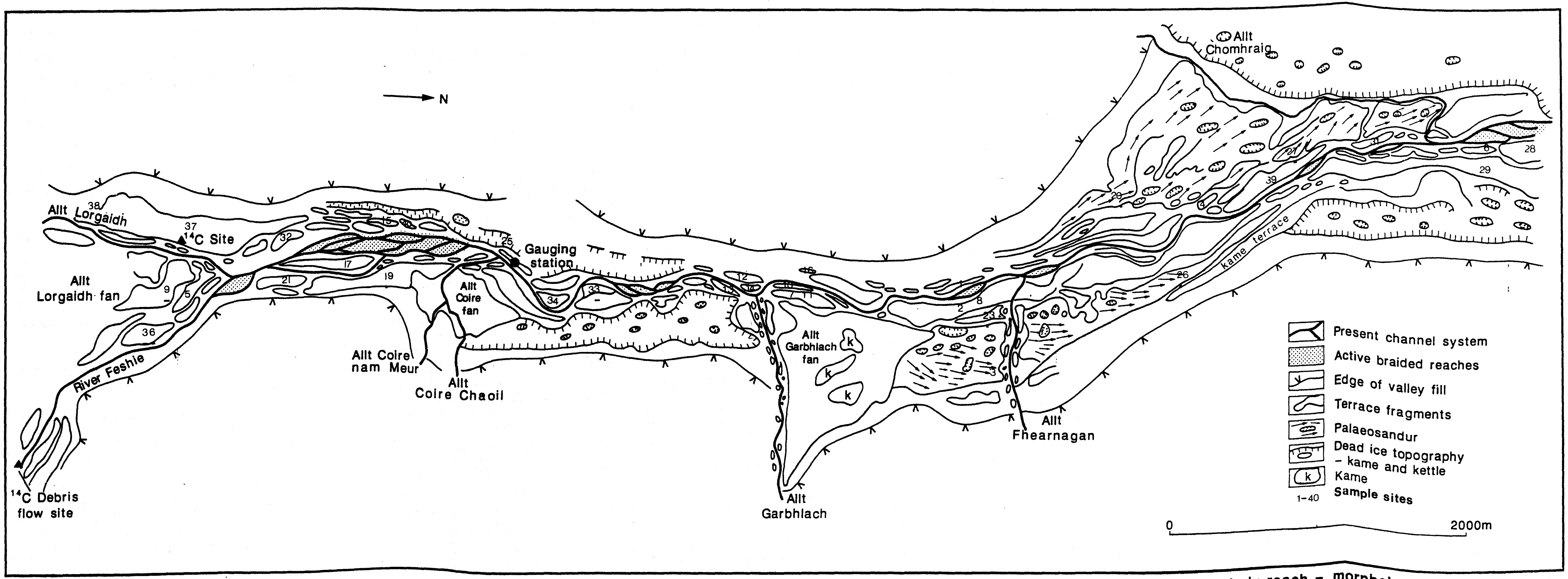


Figure 2.1 The Location of Glen Feshie and the Position of the Study Reach



The Glen Feshie study reach - morphological features of the valley fill

Figure 2.2

The study area is located in the main south-north Feshie valley and comprises a twelve kilometre reach of the glen extending from the upstream end of the upper braided reach in the south to the lower braided reach in the north (Figure 2.1). This reach was selected for study because it comprises that reach of the River Feshie which has developed downstream of the rock controlled upper Feshie and upstream of the rock gorge. The Glen Feshie terrace sequence experiences its maximum development in this 12 kilometre reach, exhibiting up to five levels above the present river in some locations. This reach is therefore likely to express more fully the postglacial development of the valley floor landforms than the reaches upstream and downstream which are interrupted by rock outcrops. The reach is also sufficiently long to enable assessment of downstream changes in the configuration of the terrace sequence, and its deposits and palaeohydrology. This reach was an optimal choice for a soil-stratigraphic study of the surface soils developed on the terrace surfaces. This is because parent material lithological and sedimentological composition and particle size, and site drainage conditions, are relatively constant throughout the reach. Nine sections across the valley fill deposits were located at more or less evenly spaced distances down the study reach (Figure 2.1). The exception to this was the reach between cross-sections 7 and 8 where permission to sample the terrace gravels and surface soils was not given. These nine cross-sections formed the locations of the sampling sites for the soil stratigraphy and the terrace sediments.

2.2 Climate, vegetation and geology

Glen Feshie has a mean annual rainfall of about 1050mm and is in the extremely humid sub-division of Scotland with an average annual potential water deficit of 0mm (Birse and Dry, 1970). Glen Feshie has accumulated temperatures of 875-1100 day-degrees C. Accumulated temperature is the integrated excess or deficiency of temperature with reference to a fixed base temperature. The base temperature which has been used for Scotland is 5.6 degrees C, approximately the level at which plant growth commences (Birse and Dry, 1970). Accumulated temperatures of 875-1100 day-degrees C places Glen Feshie in a cool temperature zone. The whole of the study area lies within the most continental part of Scotland. Glen Feshie lies in the lower and upper oroboreal sub-zones. The prefix oro denotes change to cooler thermal conditions due to altitude and the term oroboreal is used for the boreal zone in mountains (Birse and Heslop, 1970). In some years an annual freezing cycle may occur with ground freezing to depths of 50cm (Ragg and Bibby, 1966). Glen Feshie lies in a zone experiencing very severe winters and high exposure with 230-470 day-degrees C of accumulated frost (Birse and Robertson, 1970). Cyclonic and summer convectional rainstorms are common. Duration of winter snow-lie generally exceeds 100 d y^{-1} at 600m. Very strong winds are common (Pears, 1968).

The most extensive vegetation community in Glen Feshie is dry *Calluna* moor or of *Calluna vulgaris*, *Vaccinium myrtillus* and hypnaceous mosses, and lichen-rich *Calluna* moor; there are occasional areas of *Agrostis-Festuca* acid grassland; in the upper braided reach remnants of the primitive Caledonian woodland still persist on the lower valleyside slopes.

Most of the study area is underlain by acid Moinian schists, from which the bulk of the valley fill deposits are derived. Granite outcrops in the north and north-east corners of the study area.

2.3 The River Feshie

The channel pattern morphology of the present River Feshie is variable in a downstream direction. This variability is in association with the alternate occurrence of sedimentation zones where the river is laterally unstable and is subject to channel avulsion. Such reaches are connected by stable transport reaches where the river is either confined by resistant valley fill deposits or locked into bedrock (Plate 2.1). In reaches where the river is laterally unstable the channel pattern morphology is braided (Plate 2.2). Two main braided reaches occur in the study area, the upper braided reach and the lower braided reach (Figure 2.1 and 2.2). The number of channels at a cross-section varies in both reaches but averages between 4 and 5 channels. A third area of braiding occurs downstream from the gauging station. This braided reach is restricted in lateral extent by a narrowing of the valley fill deposits as a result of more resistant till underlying the easily reworked glaciofluvial deposits (Plate 2.3). Between the braided reaches the river is a singlethread channel with occasional mid-channel bars (Plate 2.4). In these more stable reaches, the channel exhibits a pattern typical of confined meandering in other upland valleys in Britain (Lewin and Brindle, 1977; Ferguson, 1981; Milne, 1983). In the confined reaches of the River Feshie the loops of the river have become stabilised in either bedrock or the confining walls of more resistant valley fill deposits. For example, in Plate 2.5 the meander has become pinned into the bluff of the valley fill.

At the downstream end of the upper braided reach (Figure 2.2)

Plate 2.1

Local rock-controlled reach, River Feshie

Plate 2.2

Upper braided reach, River Feshie

Plate 2.3

Confined braiding in the River Feshie

Plate 2.4

Single-thread channel configuration
with mid-channel bar

Confined meandering, River Feshie

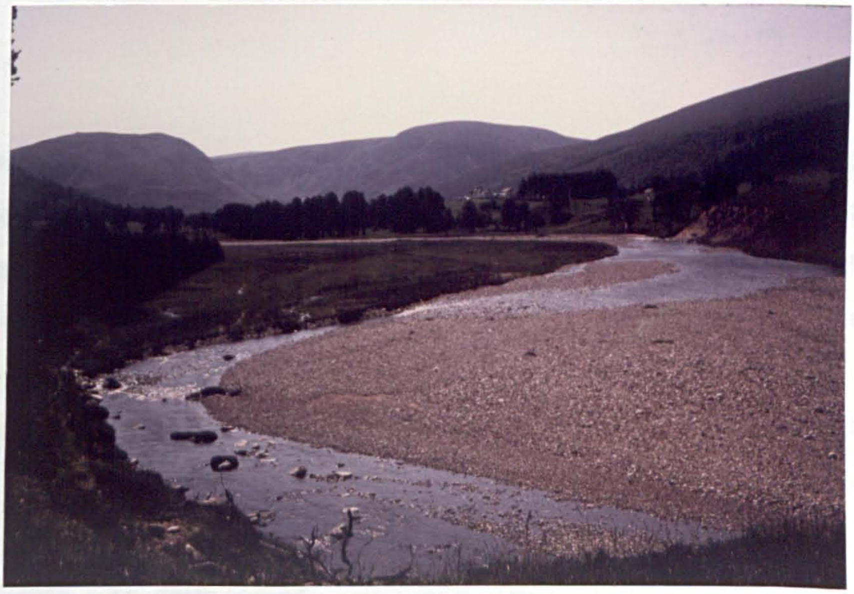


Plate 2.5

Plate 2.6

Channel avulsion in the upper braided reach

St. Andrews University installed an autographic water-level recorder in 1978. All quoted discharge data relating to the upper braided reach has been taken from data gathered from the recorder by Dr. A. Werritty, University of St. Andrews, and Dr.R. Ferguson, University of Stirling. The River Feshie has a mean flow of about 3-4m^3s^{-1} at the upper braided reach and a mean flow of about 8m^3s^{-1} at the lower braided reach (Ferguson and Werritty, 1983). The flood equivalent to the mean annual flood is estimated to be about 80-90m^3s^{-1} at the upper braided reach and has been estimated to be about 127m^3s^{-1} at the lower braided reach (Chapter 8). The 1.5 year flood at the upper braided reach is about 40-60m^3s^{-1}.
Large floods in the upper braided reach exceed 100m^3s^{-1} , and reach up to 200m^3s^{-1} at the lower braided reach. Most floods occur after prolonged frontal rainfall in autumn and winter, but some follow convective storms in the summer. Diurnal snowmelt peaks are common in spring. Discharges of 20-30m^3s^{-1} are common with 50 floods of this magnitude or higher occurring during the first three years of gauging.

With this hydrologic regime and the relatively steep valley slope of 0.012, the River Feshie is a powerful stream. Specific stream power at the upper braided reach may be calculated from

$$\Omega = \rho gQs/w \qquad 2.1$$

where Ω = stream power in watts per metre (W m^{-2})

ρ = density of water (1000kg m^{-3})

g = acceleration due to gravity

Q = discharge

s = slope

w = channel width

and is approximately 105W m^{-2} . Thus the upper braided reach of the River Feshie is a high powered, actively braided, low sinuosity channel according to the classification of Ferguson (1981). With this stream power the River Feshie is actively reworking its valley fill sediments in the laterally unstable, braided reaches (Ferguson, 1981). In the upper braided reach annual bank erosion may exceed 17% of channel area (Ferguson, 1981). Bedload transport is considerable and extensive during floods. The river transports cobble-sized bedload, not only within its channels but also over vegetated floodplain areas (Plate 2.6). The spade in Plate 2.6 marks the channel-floodplain interface. Channel avulsion in the braided reaches is a common process, recent major channel alignment changes having occurred in 1976-7 and again in late 1985. Plate 2.6 shows the channel and bar complex abandoned by the avulsion which occurred in late 1985. Avulsion appears to have been triggered in this instance by the upstream choking of the channels with sediment. Here, sediment had aggraded in the formerly active area by at least 50-75cm, up to and above the level of the floodplain. Channel avulsion in the River Feshie may also occur by chute incision (Ferguson and Werritty, 1983). Bar development and channel changes in the upper braided reach have been investigated by Ferguson and Werritty (1983). They suggest that bar development in the upper braided reach takes place through the progradation of alternate diagonal bars in combination with the processes of bank erosion and channel avulsion. The latter processes represent erosional interruptions to bar progradation and can result in the conversion of mid-channel bars to lateral bars, or lateral bars to medial bars by chute incision. The bars in the upper braided reach are actively migrating. For example, bars near the downstream end of the upper braided reach advanced intermittently by a total of about 50m between 1976-1981. Maximum advances occur during major floods (Ferguson and Werritty, 1983). Stacking of these diagonal and lateral bars produces extensive compound bar complexes that eventually accrete onto the floodplain elements. Such a mechanism has also been advanced to explain the building of the active channel and floodplain zones more generally in gravelly, low sinuosity streams in Scotland (Bluck, 1976, 1979).

2.4 The Glen Feshie Valley Fill

Glen Feshie is typical of the valleys in upland Scotland in that it contains large accumulations of glaciofluvial and glacial deposits disposed as a valley fill. This occupies the valley floor and the lower valleyside slopes up to an altitude of about 600m (Sissons, 1974; Young, 1975, 1976). The fill was created at the end of the Dimlington Stadial as the ice sheet in Glen Feshie downwasted in-situ (Young, 1975). Ice wastage in Glen Feshie and the drainage of meltwaters out of the glen towards the River Spey has been discussed by Young (1975).

There is no evidence of a Loch Lomond Readvance ice limit in Glen Feshie. The limit for the Readvance in the central Grampians, the Gaick Plateau ice cap, has been placed in the upper valley of the Allt Lorgaidh (Sissons, 1974a) where it is marked by the abrupt termination of hummocky moraines.

The present channel and the prior river channels that built the terrace deposits have had relatively limited contact with the basin hillslopes because of the presence of the valley fill. This limited coupling of the slope and river subsystems is a common phenomenon in valleys in upland Scotland. Present river channels are typically separated from the valleyside slopes either by valley fill deposits or extensive, vegetated floodplains (Milne, 1980; Werritty, 1982). The immediate environment of past channels in Glen Feshie has been the valley fill deposits, and for the present channel, is the terraces and floodplain. Since its deposition the Glen Feshie valley fill has been subjected to reworking, excavation and episodic incision by the River Feshie and its tributaries, so that the present channel is well incised into the valley fill. The thickness of the fill in Glen Feshie may be considerable, the interface between the deposits and the bedrock being buried in the upper and lower braided reaches. In the middle reaches of the river this interface has been exposed by the river after nearly 40m of incision.

Three groups of landforms comprise the major geomorphological features of the Glen Feshie valley fill in the study area (Figure 2.2). These are:-

(1) kame and kettle landforms and associated with these, an areally extensive, dissected palaeosandur now forming a high terrace surface;
(2) tributary valley alluvial fans;
(3) and an extensive suite of gravel terraces. It is this sequence of terraces which form the subject of this investigation.

The extent of these landform assemblages within the study area coupled with the large amount of planimetric and height detail required for various aspects of this study necessitated a stereo-photogrammetric survey of the Glen Feshie valley floor landforms.

2.5 Survey and photogrammetric mapping of the Glen Feshie landform sequence

The photogrammetric survey of the Glen Feshie valley floor landform assemblages was facilitated by the availability of a first order plotting instrument, a Kern PG2-L, and the existence of high quality, large-scale aerial photography. The purposes of the photogrammetric plotting were as follows:-

(1) The production of a map of the major geomorphological features of the study area, Figure 2.2.

(2) The production of accurate planform plots of the landform assemblages in the study area to provide base maps for the sediment survey and work on soil stratigraphy (Chapter 3).

(3) The production of accurate plots of the palaeochannel networks on the surfaces of the terrace fragments. This was required for the palaeohydrolgical calculations carried out in Chapter 8.

(4) The accurate plotting of the areas of the terrace fragments. The area of the terrace fragment was also required for the palaeohydrological calculations in Chapter 8.

(5) The generation of altitude matrices required for the computation of the valley gradients of each terrace fragment. These data were required for morphometric analysis of the terrace fragments (Chapter 4) and for the palaeohydrology computations in Chapter 8.

Stereo-photogrammetry is the science of obtaining geometric measurements by means of stereopairs of aerial photographs. The technique uses the bundles of rays that have made the photographs for the plotting rather than the photographs themselves.

Stereo-photogrammetry involves the exact reconstruction of these bundles of rays using a pair of aerial photographs and a perspective centre incorporated into the stereoplotter (Zorn, 1977). The photogrammetric plotting instrument in this study, the Kern PG2-L, is a restitution instrument which enables the exact mechanical reconstruction of the perspective bundle of rays via a universal joint and a space rod. It may be regarded as an analogue computer that solves continuously the mathematical equations linking the photo-coordinates and the ground coordinates of points in a stereomodel formed from two aerial photographs (Yzerman, 1963; Ritchie et al., 1977).

The Kern PG2 has a 4X enlargement capability from photo-scale to map-scale. The aerial photographs used were flown by Meridian Airmaps in 1967 and are the 50/67 survey flight. The contact scale of the photographs is 1:7 500. The photography was obtained using a wide-angle camera, with a focal length of 152.05mm and with a 23x23cm^2 format. The flying height was approximately 1400m. Paper prints were used for the ground survey work and diapositives were used for all the photogrammetric plotting. A plotting scale of 1:5 000 was used for the final map production. This provided a suitable scale both for plotting of planimetric detail and for the generation of altitude matrices comprising rectangular grids of spot heights on the terrace fragment surfaces.

The plotting machine has a spot heighting precision of better than 0.02% of the flying height, so that spot-heighting accuracy was about +/-25cm to 30cm on the ground. The planimetric plotting accuracy for establishing definite points of detail is +/-0.2mm, which at a scale of 1:5 000 is 1 metre on the ground.

The exact reconstruction of the bundles of rays that have made the photographs is carried out on the PG2 using set inner orientation and relative orientation procedures. By these procedures bundles of rays reconstructed from overlapping photographs are given the relative position they occupied with respect to each other during exposure. Corresponding rays, that is rays originating from the same terrain point, will intersect. All the locus intersections together form an exact model of the terrain. Relative orientation for the Glen Feshie stereomodels required the use of special orientation techniques because of the large relief amplitude in the glen. The procedures followed for this were those outlined by the International Institute for Aerial Survey and Earth Sciences (Zorn, 1977).

The terrain model is subsequently scaled in three dimensions using set procedures of absolute orientation. In order to carry out the absolute orientation procedures ground control points are needed for each stereomodel to be used in the plotting programme. The minimum requirements for absolute orientation are two plan control points and three height control points per stereomodel. However, a fourth height control point is also required to allow for checking of errors, so four points, one in each corner of each stereomodel were required. Twelve stereomodels were required for the study area.

The ground control points for the twelve stereomodels had to be established by a field survey programme. This survey programme was carried out using the standard methods outlined in Pugh (1975). A levelling programme was undertaken to establish temporary bench marks for the triangulation networks. Closed traverses were run from Ordnance Survey benchmarks at the downstream end of Glen Feshie up to the Allt Lorgaidh and temporary benchmarks cut every kilometre upstream. The plan coordinates, tied into the National Grid, were calculated for each benchmark. The final closing error of the traverse network was 0.04m. The ground control points for each stereomodel were natural features identified on the aerial photographs and checked in the field for intervisibility between stations. A Hewlett-Packard 3800B electronic distance meter was used to establish the distance of the baseline for the triangulation network. All subsequent distances for the ground control were established by radiation. A Kern DKM2-A 1 second theodolite was used to establish the horizontal and vertical angles required in the heighting and distance measurements of the triangles. All rounds of angles were measured on Face Left and Face Right. All heights were established using trigonometric heighting. Closing errors were calculated for all spatial coordinates using the coordinates of the benchmarks as closure points. The closing error for each triangle did not exceed 25cm which is well within the plotting accuracy of 1m. Closing errors for the heights were all calculated relative to the temporary benchmarks and were all less than 10cm. The whole survey and subsequent photogrammetric plotting was orientated to Grid North.

2.6 The Landform Sequences in Glen Feshie

Downwastage of the ice in Glen Feshie produced the glacio-fluvial landforms which make up the original valley fill (Young, 1975). These comprise glacial drainage channels, kame terraces, kame and kettle landscapes, and pitted outwash deposits (Figure 2.2). The discontinuous kame terraces lie about 40m above present river level (Plate 2.7). They occur on the east side of Glen Feshie. The kame terraces lie about 10m above the level of the kame and kettle topography and are therefore likely to have

Kame terrace, Glen Feshie

Plate 2.7

Plate 2.8

The Glen Feshie pitted outwash and low level terraces at the confluence of the Allt Fhearnagan formed between the ice margin and an ice free slope (Young, 1976). Immediately downstream of the Allt Garbhlach fan and extending to the meltwater gorge are the extremely extensive deposits of a dissected valley palaeosandur. The palaeosandur is pitted with kettle holes between which fossil braided channel networks can be clearly traced, thus making it a typical pitted outwash type of valley sandur (Plate 2.8). Pitted outwash deposits are characterised by the presence of many kettle holes which interrupt the surface of the sandur. Valley sandar develop between valley walls and their development is closely related to the provision of meltwater and debris in a relatively restricted location (Krigstrom, 1962). Pitted outwash or kettled sandar are both proglacial and ice-contact features (Price, 1973). They are pro-glacial in that they form in front of the melting ice, but are ice-contact in that the ustream end of the sandur is in contact with the waning ice margin. The eventual surface of the pitted outwash is the product of the modification of the initial sandur surface by the melting out of glacier ice leading to the development of kettle holes. The palaeosandur, developed in a proglacial rather than ice marginal situation, represents the highest fluvial surface in the Glen Feshie terrace sequence. At the level of the Allt Fhearnagan it is 24m above present river level.

The pitted outwash terminates upstream in kame and kettle deposits, (dead ice topography). The kettles in the kame and kettle topography are not linked by fossil channels, as are those of the pitted outwash deposits. Some of the deposits are clearly stratified (Plate 2.9) with alternating layers of sand and gravel whilst some of the kame and kettle deposits are underlain by till deposits. Kame and kettle deposits are ice-contact deposits which mark the terminal zones of valley

Kame and Kettle deposits, Glen Feshie

Plate 2.9

glaciers from which the outwash rivers issued (Sissons, 1967). Kame and kettle forms occupy much of the valley floor upstream from the Allt Garbhlach and extend into the upper braided reach (Figure 2.2).

Separating the kame and kettle upstream of the Allt Garbhlach and the palaeosandur is the Allt Garbhlach alluvial fan (Plate 2.10). Partially buried by the fan deposits, but projecting above the surface of the fan, are several kames. The fan deposits have buried the ice contact slope between the sandur and the kame and kettle deposits upstream. Young (1976) suggests that this fan was probably deposited during the later stages of the deglaciation of the late Devensian ice sheet at the end of the Dimlington Stadial (Young, 1976).

Lying some metres below the level of the outwash surface are a group of low-level terraces, areally extensive and never exceeding about 5m above present river level. These low-level terraces are present along the whole length of the River Feshie within the study area. Figure 2.3 shows nine cross-profiles of the terrace sequences as they are present along the length of the study reach. Each cross-profile is sited at the soil transect sites marked in Figure 2.1. The cross profiles of the terraces show that the terrace fragments along the length of the River Feshie are generally unpaired, and vary both in number of terrace levels present and in height above present river level. The terrace fragments are discontinuous from reach to reach (Figure 2.2).

Plate 2.11 and cross-section 1 shows the terraces at the upstream end of the upper braided reach. Here three terrace levels can be seen. The terraces vary from about 1.2m above

Plate 2.10

The Allt Garbhlach fan

Plate 2.11

Low level terraces of the upper braided reach

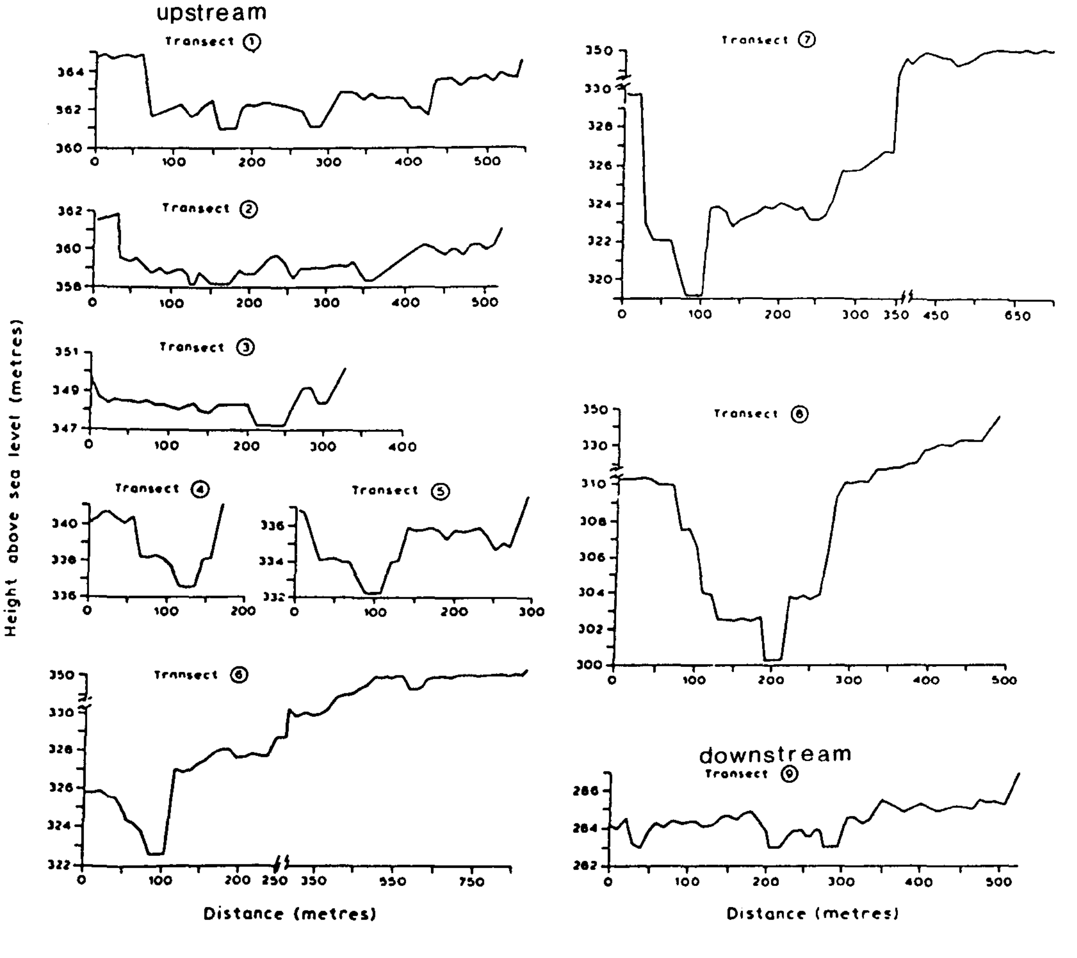


Figure 2.3 Cross Profiles of the Terrace Sequence - viewed downstream

present river level to nearly 4m above present river level. In this reach, where the valley floor is almost 700m wide, the terrace fragments are laterally extensive. At the downstream end of the upper braided reach the Allt Coire fan merges with the higher of the low-level terraces on the east bank of the River Feshie. At the Allt Coire fan the valley floor narrows and river becomes confined by kame and kettle deposits which are underlain by relatively resistant till deposits. Plate 2.12 and cross-sections 4 and 5 show the terrace sequence at the level of the Allt Garbhlach fan. Here the valley floor is effectively narrowed by the fan and is about 350m wide. The terraces in this reach are more limited in lateral extent. In contrast to the upstream braided reach only two terrace levels are present. On the east bank and abutting the base of the fan is an extensive higher surface and a restricted lower level. Conversely, a restricted higher terrace and an extensive lower terrace are present on the west bank. Braided palaeochannels are evident on the surfaces of both extensive terrace fragments. The terrace fragments range in height above present river level between 1.2m-4.2m. Plates 2.4 and 2.8 and cross-sections 6 to 8 show the terrace sequence at the reach extending from the Allt Garbhlach fan downstream of the Allt Fhearnagan; again a variation in terrace levels is evident. It is at this reach of the River Feshie that the palaeosandur appears and forms the highest fluvial terrace in the fill. Three low-level terraces may be seen. On the east bank, below the outwash surface is the highest of the low-level terraces. This is about 5m above present river level, is areally restricted and merges with the eroded remnants of a small alluvial fan. About 1.5m below the surface of this fragment is a more extensive surface possessing well marked palaeochannels. Approximately 1m below the level of this fragment, on the west bank is a lower surface again also

Low level terraces below the Allt Garbhlach fan

Plate 2.12

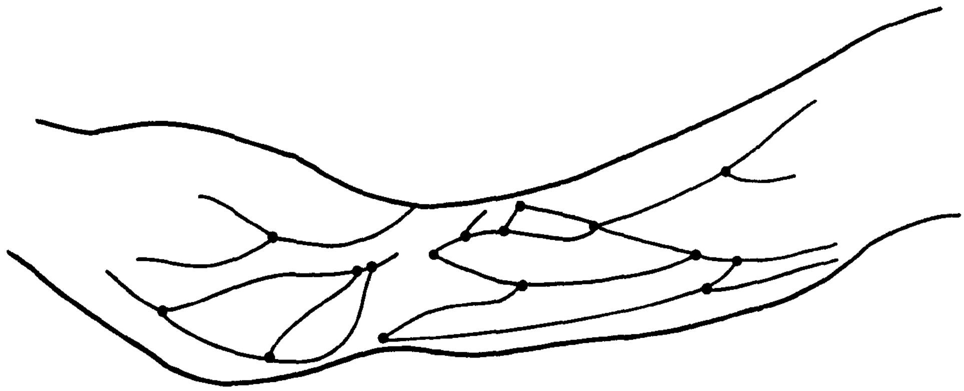
Plate 2.13

Modern bar deposits, upper braided reach possessing well marked palaeochannels. Transect 9 is a profile through the terrace levels at the lower braided reach. Here three low-level terracs are in evidence.

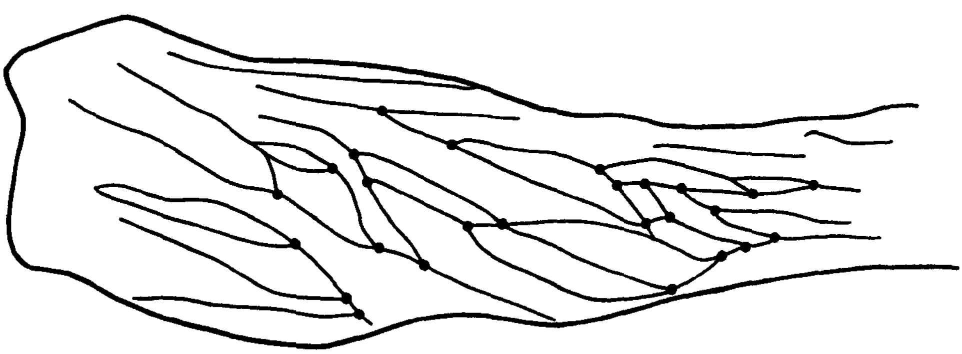
All of the more extensive low-level terrace fragments and the dissected palaeosandur surfaces exhibit complex braided palaeochannel networks on the terrace surfaces (Plates 2.4, 2.5, 2.10, 2.11 and 2.12). In Glen Feshie the general lack of fines in the catchment has prevented the infilling of palaeochannels, so that they are preserved as distinct topographic lows, not only on the present floodplain but also on the terrace surfaces. The palaeochannel networks have been plotted photogrammetrically for each terrace fragment. Figures 2.4 - 2.5 are plots of some of the palaeochannel networks on the terrace surfaces.

2.7 The Terrace Deposits : Stratigraphy

Exposures of the terrace sediments are limited in extent but those that can be observed suggest that sedimentary structures are poorly developed in the Glen Feshie terrace sediments. Plate 2.13 is an exposure in the modern bar deposits of the upper braided reach and Plates 2.14-2.19 are exposures ofthe deposits comprising the low level terraces; these exposures exhibit the main sedimentary characteristics of the modern stream and terrace sediments. Plate 2.14 is an exposure in the upper low-level terrace of transect 9; Plate 2.15 is an exposure of the first level terrace of the upper braided reach at cross-section 2; Plate 2.16 is an exposure of the second level of the terraces in the upper braided reach at cross-section 2; Plate 2.17 is an exposure of the upper surface at cross-section 4; Plate 2.18 is an exposure of the upper surface below the Allt



Palaeochannels - fragment 2

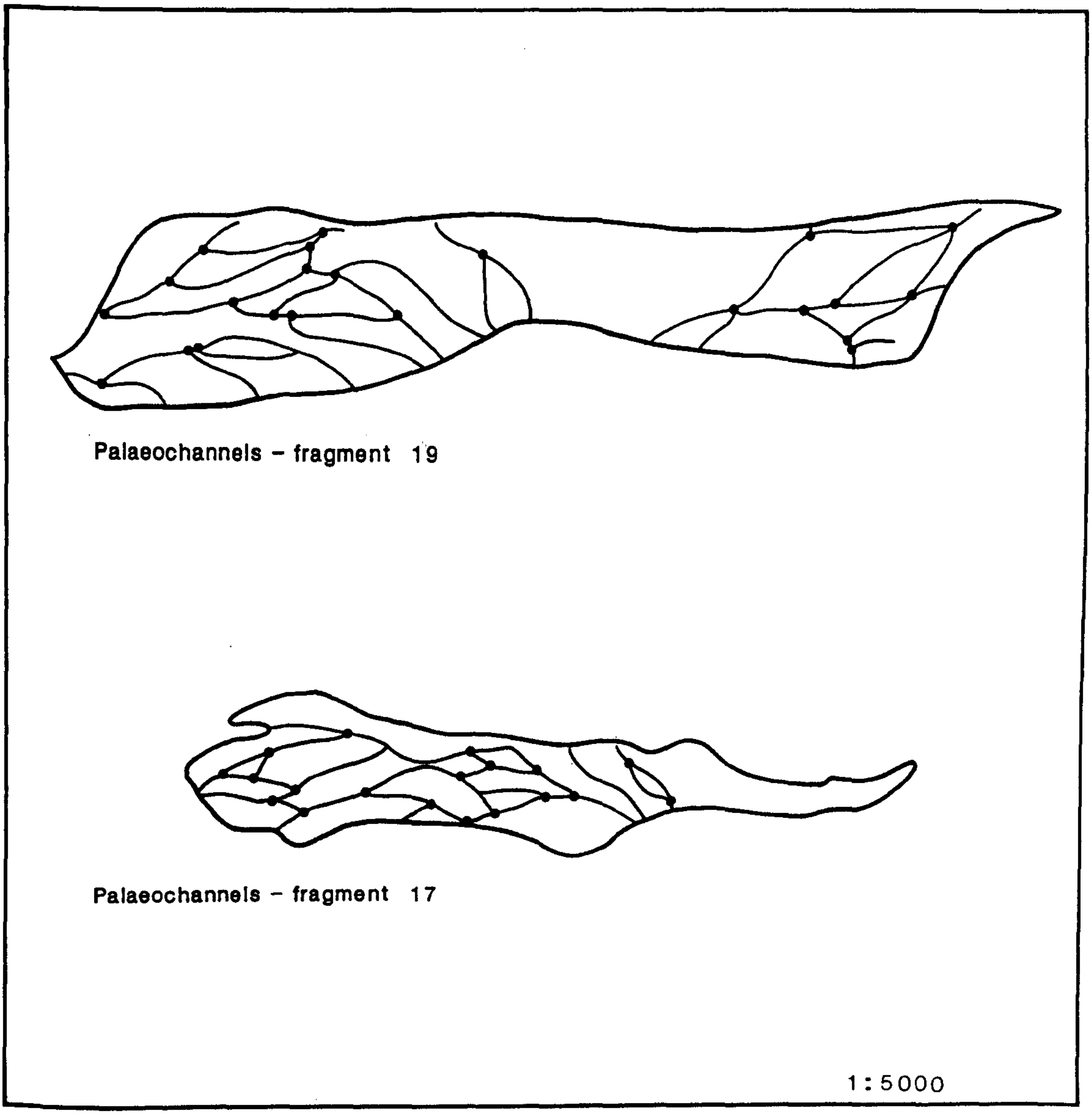


Palaeochannels - fragment 3

1:5000

River Feshie braided palaeochannels

Figure 2.4



River Feshie braided palaeochannels Figure 2.5

Garblach fan, cross-section 5; the particle size of this exposure was too coarse to allow freshening of the site; Plate 2.19 is an exposure of the second terrace level at cross-section 6.

The terrace gravels are imbricated, clast-supported gravels. They possess a secondary population of fine gravel and sand which forms an infill in the interstices between the clasts. In some of the exposures a crude horizontal stratification may be evident, but all such strata are laterally discontinuous. This stratification takes the form of units of imbricated cobbles separated by units of smaller clasts possessing a gravel/sand infill as in Plates 2.14 and 2.15. The smaller clast units may be 20-30cm thick. Occasional lenses of finer, matrix-supported gravel are found in some of the terrace sediments again as seen in Plate 2.14 and in the modern bar sediments (Plate 2.13). More commonly, however, the gravel units comprising the bulk of the terrace deposits for all the low-level terraces are up to 1 metre thick and comprise poorly sorted imbricated and clast-supported gravels which are massive in structure. These units are seen in Plates 2.16, 2.17, 2.18 and 2.19. In general the terrace gravels do not exhibit a fining-upward sequence, although this may occur locally in some exposures. As can be seen in all the exposures large clasts are just as likely to occur at the top of the deposit. These massive and crudely-bedded gravels are typical of the Gm (massive gravel) lithofacies of Miall (1978). There is no observable difference between the stratigraphy of the deposits between the low level terraces.

The braid bar and channel assemblages of the low-level and palaeosandur surfaces coupled with the Gm lithofacies of the

Plate 2.14

Terrace deposits, cross-section 9

Plate 2.15

Terrace deposits, cross-section 2

Terrace deposits, cross-section 2

Plate 2.16

Plate 2.17

Terrace deposits, cross-section 4

Plate 2.18

Terrace deposits, cross-section 5

Plate 2.19

Terrace deposits, cross-section 6

terrace sediments are also typical of the major low sinuosity gravelly river as defined by the architectural-element analysis of Miall (1985). Miall argues that fluvial styles, or channel pattern morphologies, may be classified by a combination of their architectural elements, the channel sinuosity and a braid parameter. The braid parameter has been defined by Rust (1978) and is the number of braid bars per mean wavelength. The parameter is not dimensionless but may be used to indicate degree of braiding within one braided stream. Reference to the plots of the palaeochannels on the terrace surfaces in Glen Feshie indicates that the braid parameter for the terrace sediments would be high. The Glen Feshie deposits with their low sinuosity (channel distance/valley axis distance), medium-high braiding parameter and Gm lithofacies are thus typical of Miall's category 3 fluvial style, the major low sinuosity gravelly rivers.

2.8 The Terrace Deposits : Particle Size

Particle size data were required in this study from both the modern river and from the terrace sediments for the following purposes :-

(1) Since the St. Andrews University gauging station is at the downstream end of the upper braided reach, discharge data are available only for the upper braided reach in the River Feshie. In order to compare the palaeodischarge estimates made for the terraces in Chapter 8, with the flood discharges of the modern Feshie, flood discharges for the remaining four study reaches were calculated from 4 cross-sections of the River Feshie using a modified slope-area method of discharge estimation (Richards *et*

al., in press). This is a method developed to estimate flood discharges for morphologically complex sections. The method requires an estimate of the Darcy-Weisbach friction factor (f) from

$$1/(f)^{\frac{1}{2}} = 1.16 + 2.0 \log (d/D_{84}) \qquad 2.2$$

where

d = depth in metres

D_{84} = 84th percentile particle intermediate axis

and therefore necessitates an estimate of D_{84} of the modern stream gravels for each cross-section used for discharge reconstruction.

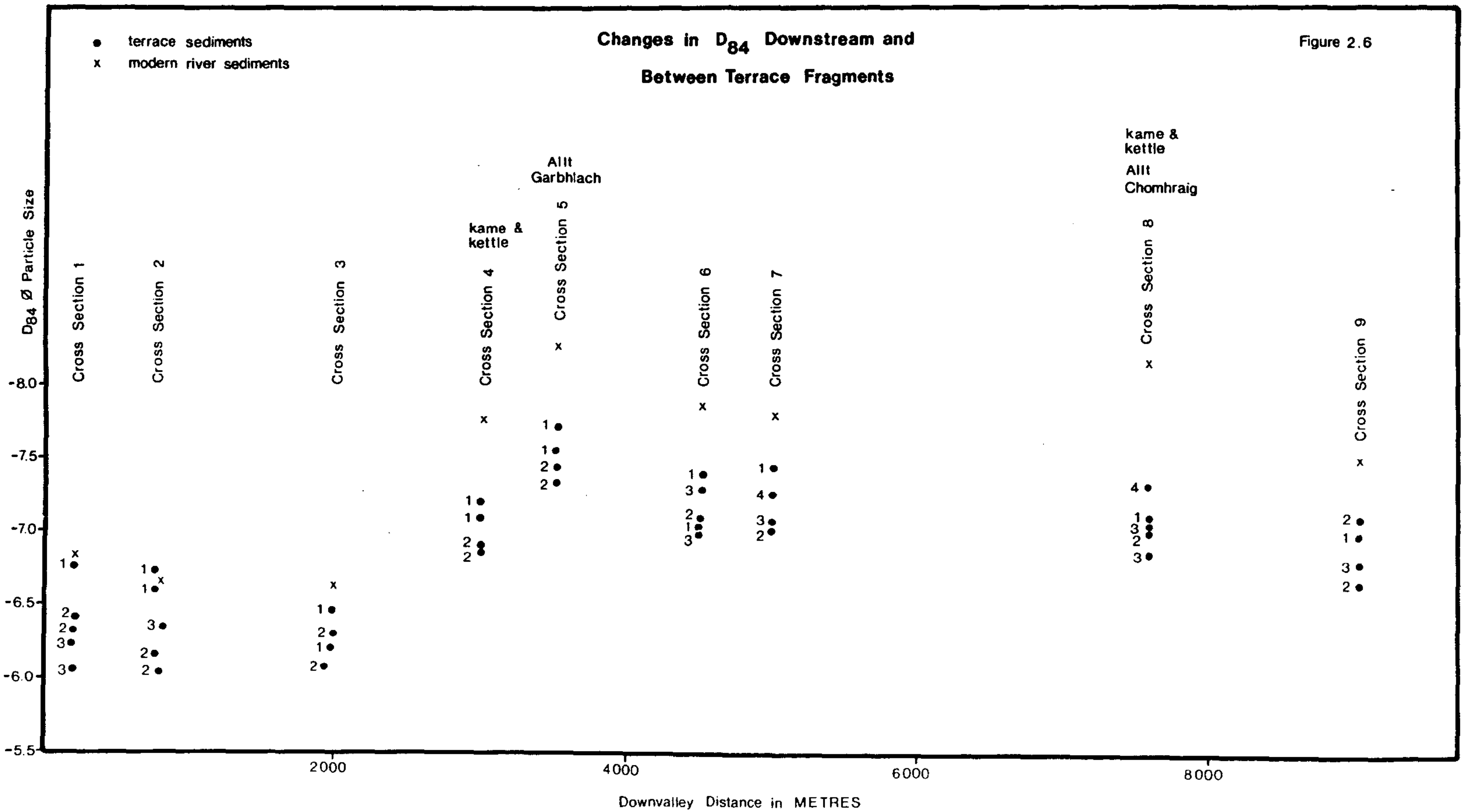
(2) Particle size data for the terrace sediments were required for the palaeohydrological analysis carried out in Chapter 8. This analysis utilises the relationship developed in the present study for a quantitative index of channel pattern, stream power and grain size (Chapter 6). This relationship has been developed using data from 40 gravel-bed streams for which an estimate of D_{84} is available. An estimate was therefore also required of D_{84} for the terrace gravels, as an input to the palaeohydrological analysis. It was also necessary to adopt a method of sampling both the terrace sediments and the modern stream sediments that was as comparable as possible.

The aim of the sediment sampling for both the modern river sediments and the terrace sediments was to obtain a measure of the resistance to entrainment and transport of the clasts. The channel beds and bar surfaces comprise a cobble and gravel-sized clast-supported population with a smaller secondary population of finer material infilling the void spaces of the framework on the bars. Surface imbrication was universal on both channel beds and bar tops. For the modern river sediments the surface layer of clasts is the main point of interest and therefore surface-orientated sampling methods are appropriate (Kellerhals and Bray, 1971). Accordingly, as the Wolman method of surface sampling individual particle diameters on a grid basis gives a measure of the roughness which directly affects the flow within the channel (Wolman, 1954), this method was adopted. Adoption of this method has the further advantage that the published grain size data used in the analysis of Chapter 6 was also collected using the Wolman method. The Wolman method was developed for coarse-bed channels and is based on an analysis of the relative area covered by particles of given sizes. Sampling consists of measuring the intermediate axis of 100 clasts picked from the channel bed on a grid system: the method is essentially a systematic random sampling strategy. Wolman (1954) and Kellerhals and Bray (1971) have demonstrated the reproducibility of the method in terms of repetitive sampling at the same site.

The bulk of the terrace sediments comprise massive, clast-supported cobbles and gravels. As such the method for sampling D_{84} of the clasts in the terrace sediments which is most comparable to the Wolman count method for surface clasts is likely to be a method based on grid sampling. The following approach was therefore used to sample the clasts in the exposures of the terrace gravels. Each exposure was cleaned and steepened to an angle close to the angle of repose. A grid was constructed which was a square metre and which was divided into 10cm squares, so that there were 100 grid intersections. The grid was placed at random on the exposure and the clast which fell under each grid intersection was selected for sampling. Approximately 100 clasts were measured from each sample. The A, B and C axes of each of the clasts were measured using calipers. Sampling on a randomly located grid ensured that the sample was a statistical representation of the distribution of sizes in the gravel unit. This method, carried out on a steep exposure is a reasonably comparable method to the Wolman count method on a horizontal surface, but with a closer spacing between the grid intersections. In all 40 exposures were available with which to measure the clast size of the terrace sediments. These exposures included samples from most of terrace levels for the 9 cross-sections in the study reach. Some duplicate sampling was possible for a few of the fragments, but insufficient exposures were available to attempt an assessment of the spatial variability of clast size between sites for one terrace fragment. The analysis of Mosley and Tindale (1985) shows that a considerable number of samples and very large sample sizes would be required for such an assessment. They suggest that the spatial variation of sediment calibre in gravel-bed rivers requires bulk samples of around 1 tonne at 228 locations to estimate the mean +/-10% with 95% confidence, for representative estimates to be made of average reach size.

A Fortran programme was written to derive cumulative percentages of clast sizes from the gravel count data. These were derived for both the modern river and terrace sediments and D_{84} was computed. The standard deviation, a measure of sorting, was calculated for each sample of both terrace and modern river sediments using the method of moments (Folk, 1974).

Figures 2.6 and 2.7 show the downstream and between level variations in D_{84} and sorting for the gravels of the modern stream and the terraces. The data points are plotted by cross-sections and the level of the terrace fragment above present



terrace sediments
modern river sediments
Changes in D_{84} Downstream and
Between Terrace Fragments
Figure 2.6
D_{84} Ø Particle Size
Cross Section 1
Cross Section 2
Cross Section 3
kame & kettle
Cross Section 4
Allt Garbhlach
Cross Section 5
Cross Section 6
Cross Section 7
kame & kettle
Allt Chomhraig
Cross Section 8
Cross Section 9
-8.0
-7.5
-7.0
-6.5
-6.0
-5.5
2000
4000
6000
8000
Downvalley Distance in METRES

river level is indicated. Thus, level 1 at a cross-section is the highest terrace above present river level and level 4 would be the lowest at that cross-section. The discontinuous nature of the terrace fragments between reaches means that the highest level of one cross-section is not necessarily the same terrace level as the section immediately downstream.

Figure 2.6 shows the variation in D_{84} downstream and between terrace levels for the terrace and modern river sediments. The plot shows that there is not a simple pattern of reduction in the size of the bed material of either the terrace sediments or the modern stream sediments with distance downstream. A decrease in size of bed material (D) with distance downstream (L) is a common feature of many alluvial streams and can be expressed by

$$D = D_o e^{-(kL)} \qquad 2.3$$

where D_0 = grain size at L=0
k = coeffecent of abrasion.

This expression describes the exponential decrease of particle size downsteam as a result of abrasion, and is attributed to Sternberg (Church and Kellerhals, 1978). Equation 2.3 may be reasonably appropiate for describing the change in particle size downstream for rivers flowing on gravels (for example, Church and Kellerhals, 1978; Knighton, 1982). Incorporating the influence of both abrasion and sorting (Knighton, 1980) the expression

$$D = D_o e^{-(k_1 + k_1)L} = D_o e^{-(k_3 L)} \qquad 2.4$$

where D_0 = grain size at L = 0

k_1 = coeffecent of abrasion

k^2 = coefficient of sorting

k_3 = combined abrasion and sorting coefficient

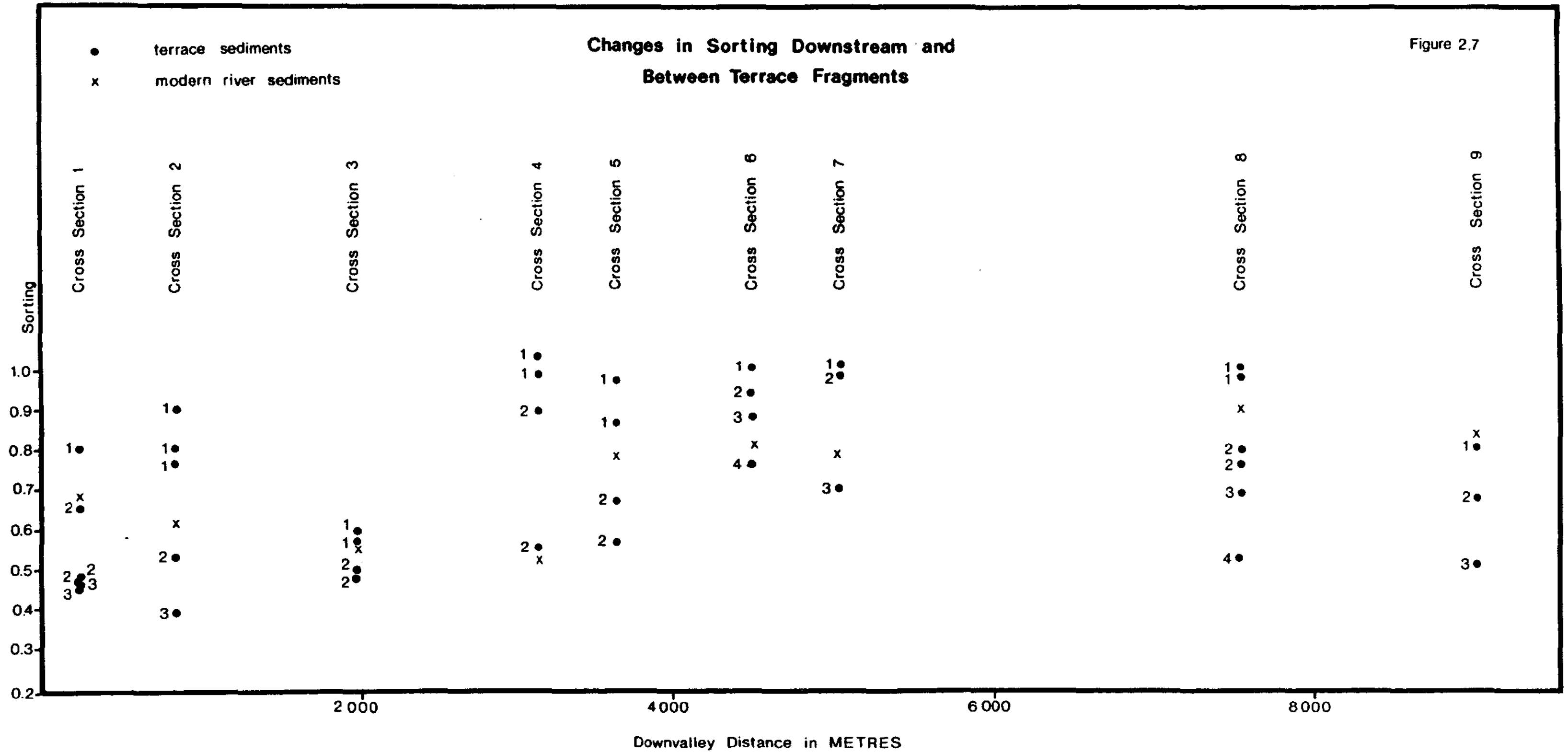
The empirical equation describing an exponential decrease of particle size downsteam combines the results of abrasion and progressive sorting, which cannot readily be disaggregated (Tanner, 1971; Rana et.al., 1973). Downstream fining of particle size in the gravelly Squamish River, British Columbia, has been observed by Brierley and Hickin (1985), who suggested that for this river the curves demonstrating the grain size decline are better described by power functions. These reflect a very rapid decline in particle size immediately downstream of the major sediment source. However, a number of studies have shown that for gravel-bed streams the overall decline in particle size with distance downstream may be complicated by the introduction of fresh material of anamolous size from coarse tributary and valley fill inputs (Church and Kellerhals, 1978; Knighton, 1980; Inglis and Werritty, 1985). The introduction of fresh material from bank and tributary sources produces increases in grain size below junctions. The magnitude of the increase is related to the relative sizes and gradients of the main stream and tributary at each confluence (Knighton, 1984).

Three clusters of data points can be discerned in the plot of D_{84} against distance downstream for the Glen Feshie terrace and modern stream sediments. These are the data points which comprise the 3 cross-sections of the upper braided reach samples (cross-sections 1-3), the four cross-sections which comprise the middle reaches of the river (cross-sections 4-7) and the two cross-sections which comprise the lower reaches of the study area (cross-sections 8-9). For the terrace sediments of the three cross-sections comprising the upper braided reach D_{84} varies between -5.9phi and -6.65 phi (60-101mm). For the cross-sections comprising the middle reaches D_{84} varies between -6.8 phi and -7.53 phi (112-185mm) whilst D_{84} for the lower reaches varies from -6.8 phi to -7.13phi (112-140mm). The coarsest clasts occur in the terrace sediments of the middle reaches of the river, with the lower braided reach terrace fragments possessing coarser clasts than the terrace sediments of the upper braided reach. D_{84} for each group of cross-sections does show some tendency to fine downstream, although complications are introduced into the plot because of the variability in particle size between terrace levels at a cross-section. Generally, the coarsest clast sizes belong to the higher terrace levels, with clast sizes decreasing with height above present river level. Exceptions to this are cross-sections 8 and 9.

The trend for the cross-sectional data of the terrace deposits to fall into three groups is mirrored in the trend for D_{84} to vary with distance downstream for the modern stream sediments. At least two, and possibly three fining cycles may be discerned, although the number of data points is rather limited. The smallest clast sizes occur in the upper braided reach, as was the case with the terrace sediments. A fresh fining cycle may be apparent from cross-section 8, just upstream of the lower braided reach. The largest clast sizes occur in the middle reach (cross-sections 4 to 8) of the river, as with the terrace sediments. In this reach the river bed is paved and the channel relatively immobile, acting as a transport reach. The clast size of the river sediments in this middle reach is significantly larger than that of the terrace sediments, suggesting that the paved sediments could be lag deposits of clasts eroded from the terrace sediments and too large to be moved under present day hydrological conditions. This is in contrast to cross-sections 1-3 and 9 where there is no significant coarsening of the modern river sediments from the terrace sediments. At these cross-sections the clasts were sampled from mobile beds with the braided channels experiencing active bar migration.

The trend for both the modern stream sediments and the terrace sediments to fall into three groups is explicable with reference to the locus of coarse sediment inputs into the stream. At the level of cross-section 4 the River Feshie has incised some 20m through extremely coarse glaciofluvial deposits of the kame and kettle topography (Figure 2.2). Immediately upstream from cross-section 5 the Allt Garbhlach tributary enters the River Feshie contributing boulder-sized material into the mainstream system. At cross-section 8 the Allt Chomhraig tributary enters the River Feshie. This tributary has incised through coarse glaciofluvial deposits and has contributed coarse sediment to the terrace deposits downstream from the tributary entrance. This coarse input has been coupled with that from the very coarse sediments of the kame and kettle deposits which the main river has undercut in the past. Plate 2.20 shows the nature of the coarse inputs at one site where the present River Feshie is undercutting a bluff comprising till capped by glaciofluvial deposits.

The deposits of the terraces and present channel are derived from slope, tributary and channel boundary sources. The particle size distribution at any location depends of the characteristics of the initial input and on the nature and rate of subsequent modifications to that input. Reference to Figure 2.7 shows that as the River Feshie has reworked the original valley



terrace sediments
modern river sediments
Changes in Sorting Downstream and
Between Terrace Fragments
Figure 2.7
Cross Section 1
Cross Section 2
Cross Section 3
Cross Section 4
Cross Section 5
Cross Section 6
Cross Section 7
Cross Section 8
Cross Section 9
Sorting
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
2000
4000
6000
8000
Downvalley Distance in METRES

Plate 2.20

Coarse sediment input, River Feshie

fill sediments, the deposits which have been incorporated into the terrace deposits have become progressively better sorted. The terrace deposits show a consistent decrease in the value of the sorting coefficient from level 1 to the lower levels for each section.

The similarity in the particle size of the terrace sediments as well as the progressive improvement in sorting from the outwash surfaces to the lower terraces suggests that these coarse deposits have been eroded, reworked and incorporated into the terrace deposits of the River Feshie since deposition of the valley fill about 13,000BP. The lateral activity of the incising stream is attested by the sinuous nature of the bluffs of the valley fill and terrace scarps (Figure 2.2). Thus, as is typical of valleys that have been affected by proglacial and glacial depositional processes, much of the coarse component of the available sediment is inherited from the original valley fill deposits (Church and Ryder, 1972; Lewin, 1983). The terrace deposits, disposed as bar and channel complexes, therefore represent sediment storage zones in stages of the flushing of the sediments from the fill of the mainstream and tributary valleys.

CHAPTER 3

A QUANTITATIVE SOIL-STRATIGRAPHIC APPROACH TO THE CORRELATION AND DATING OF RIVER TERRACE FRAGMENTS

3.1 Introduction

Over the past twenty-five years the growing interest in the study of Late Quaternary environments has been accompanied by an increase in the number of studies of soil chronosequences. These have been conducted by pedologists, to better understand soil genesis and the development of soil profiles with the operation of the soil-forming processes, and by geomorphologists, to determine the ages of sedimentary landforms using degree of surface soil profile development.

The soil is surface-weathered material affected by processes of organic matter accumulation and decomposition, vertical and lateral translocation in solution and suspension, and the development of horizons subparallel to the earth's surface (Richards et al., 1985). Soil profiles are continually forming and changing. They are variable spatially in response to the soil-forming factors of climate, organisms, relief and parent material, and temporally in response to the length of time available for the development of the horizons comprising the profile. The examination of the temporal variation of soil properties and the rate of operation of the soil-forming processes is carried out through soil chronosequence studies.

The correlation and relative dating of landforms using degree of soil profile development forms the rationale for soil stratigraphy. Soil-stratigraphic units form, in situ, on deposits as pedological processes work downwards on fresh and unweathered (Cn) horizons (Mahaney, 1978). To be stratigraphically significant a soil-stratigraphic unit should have physical and chemical characteristics and stratigraphic relations that permit its consistent recognition and mapping. It possesses distinctive pedological characteristics and has specific relations in time to the underlying deposits and other soil-stratigraphic units (Finkl, 1980). It is therefore important that the physical and chemical characteristics of the soils should be described as precisely as possible in order to provide a strong quantitative basis for correlation and dating purposes. In order to be used for predicting the ages of undated landforms the soil-stratigraphic unit must be independently dated in at least some of its locations. Thus the aim of soil stratigraphy is to recognise significant soil-stratigraphic units, to rank them chronologically and to use them for prediction of the ages of undated landforms.

In this chapter an alternative to the traditional height-based method of river terrace correlation and relative dating is presented. Terrace fragments are numerically classified and objectively grouped using quantitative soil-stratigraphic data derived from surface soils developed on the Glen Feshie terrace fragments. Principal Components Analysis is used to quantify pedological data collected from the soil profiles developed on the terrace surfaces. The scores on the first principal component for each soil profile are then grouped into soil-stratigraphic units using a hierarchical clustering technique. The soils developed on the terrace fragments are shown to form a podzolic sequence of soils in different stages of profile development, so forming a chronosequence of soils. Some absolute dating control enables the clusters of soils to be tentatively age-calibrated and a re-assessment made of the ages of the river terraces in Glen Feshie.

3.2 Soil Chronosequences : Previous Studies

Soil chronosequence studies may be divided broadly into three main groups:-

(1) Those studies which embrace an age-range in excess of the postglacial period, 15,000 years to more than 3 million years, (for example Ruhe, 1956; Birkeland, 1978; Bockheim, 1979a, 1979b)

(2) Those studies that encompass a time span of the post-glacial, approximately the last 10,000 years, (for example Walker, 1962; Franzmeier and Whiteside, 1963; Franzmeier et al., 1963; Mahaney, 1974, 1978; Birkeland, 1984; Ellis and Richards, 1985).

(3) Those studies considering a time span of 1,000 years or less (for example Crocker and Major, 1955; Crocker and Dickson, 1957; Ugolini, 1966, 1968; Jacobson and Birks, 1980; Mellor, 1985).

Alexander (1974) examined the soils on eight terrace levels along the Truckee River, Nevada. The oldest terraces are older than 0.4 million years and the youngest may be younger than the last major stadial (Wisconsin). Iron was extracted from the B horizons of the surface soils developed on the terrace surfaces. The results showed an increase in the amount of iron in the terrace soils up to the third terrace level, with a gradual decline in iron from the third to the sixth level. The change in the trend of iron accumulation occurs in association with a change in the nature of the parent material of the terrace soils. Alexander suggests that the slightly lower amounts of iron in the higher terrace levels may be due to variation in the parent material of the higher terraces, as well as to possible eluviation of iron from the B horizons. However, it is possible that if Alexander had examined extractable iron using the ratio of B:C extractable iron, the problem of variation in iron content inherent in the parent material would have been overcome and trend of increasing iron with age of the terraces may have been continued. A trend of increasing amounts of iron, as well increased soil reddening, in terrace soils has been demonstrated by Torrent, Schwertmann and Schulze (1980), for terraces in Spain, although no dates are available for the terrace sequences.

Eleven soils formed in till parent materials in northeastern Pennsylvania were studied by Levine and Ciolkosz (1983) to determine changes in soil profile development in tills ranging in age from about 15,000BP to greater than 75,000BP. This study showed that extractable iron, aluminium and clay content of the B horizons of the soils increased with increasing age of the soil. This study also suggested that movement of the zone of maximum iron oxide accumulation deeper in the soil is a distinctive age trend of soil formation in well-drained soils.

In an early study, Walker (1962) examined the development of soils on four river terraces in New South Wales. These terraces ranged in age from 29,000BP for the oldest terrace, terrace 4, to a modern age for terrace 1. Very little soil development had taken place on terrace 1; the terrace 2 soils, about 390BP in age, exhibited shallow profiles with only A and C horizons. The terrace 3 soils, at 3,740BP, show an increase in soil depth and A/ B/ C horizons. The oldest soils show a A1/ A2/ B1/ B2/ C horizon development. There is an increase in clay content with depth and development of clay skins as the soils increase in age from the terrace 1 to terrace 4 soils.

Considering the development of podzolic sequences of soils up to 10^4 years Franzmeier and Whiteside (1963), Franzmeier et al. (1963) Mahaney (1974, 1978), Mahaney et al. (1981) and Ellis and Richards (1985) have examined the development of podzols over the time span of the Holocene. These studies are considered in more detail below.

Developing Holocene chronofunctions in the Southern Alps of New Zealand, Birkeland (1984) found progressive changes with time in soil morphology. He also found increasing amounts of organic carbon in the A horizons and some of the B horizons of older soils, accumulation of free iron and aluminimum in the B horizons of the soils, and an increase in the amount of clay in the profiles.

Very few chronosequence studies have been carried out on the soils on British terrace sequences. Recently however, Harvey et al. (1984) have examined the surface soils developed on three ages of alluvial surfaces in the Howgill Fells, Cumbria. Progressive change in soil profile morphology is observed from the late Pleistocene solifluction terrace to the low terrace surface, which is between about 2,500-1,000BP in age. The high terrace between the solifluction terrace and the lowest surface probably dates from the late Pleistocene rather than the Holocene. The soils developed on the surfaces are podzols, but with the lowest surface showing minimal soil development. The soil on the lowest surface possesses an organic-rich surface horizon, with an incipient bleached or Ea hozizon, and an oxidised C or B/C horizon. The podzol on the high terrace is a well developed soil with an O/ Ea/ Bh/ Bfe/ C horizon profile. The soil profile developed on the oldest surface is shallower , but has a thicker Ea horizon followed by a Bs horizon.

A number of chronosequence studies with age ranges of less a than about a 1,000 years have reported significant trends in physical and chemical soil properties. These include investigations carried out on Neoglacial and related deposits in Alaska and neighbouring areas of the Canadian Rockies (for example, Crocker and Major, 1955; Crocker and Dickson, 1957; Ugolini, 1966, 1968; Jacobson and Birks, 1980) and in Norway, (Mellor, 1985). Similar investigations have been conducted on prograded beach and dune deposits in Canada (for example, Sondheim et al., 1981), whilst Harvey et al. (1984) have examined degree of soil profile development of soils on modern valley floor deposits spaning an age range of 140 years in the Howgill Fells, Cumbria. The majority of these studies have reported significant and rapidly adjusting age-related trends in organic matter or organic carbon content, pH, cation exchange capacity, exchangeable bases, carbonate content and base saturation. However, there are few reports of age-related trends in free iron and aluminium, rates of adjustment of these properties appearing to be much slower, as suggested by Birkeland (1974).

3.3 Soil Stratigraphy and its Application to the Correlation and Dating of Geomorphic Surfaces : Previous Studies

Late Quaternary deposits have been correlated and assigned relative and absolute ages using soil-stratigraphic units derived from a combination of radiocarbon dating and degree of soil development (for example, Mahaney 1974, 1978; Mahaney and Fahey, 1976; Mahaney et al.,1981).

Thus in the Wind River Mountains, Western Wyoming, Mahaney (1978) examining podzolic soils exhibiting varying degrees of profile development, identified five soil-stratigraphic units based on differences in field morphology, % organic matter in the organic horizons of the soils, textural characteristics, clay-mineral assemblages and selected soil chemical parameters, including free iron and pH. The ages of the deposits on which the soil-stratigraphic units are developed range from about 100BP to about 100,000BP. Correlative soil-stratigraphic units have been found in adjacent areas including the Teton Range, the Big Horn Mountains and the Front Range in Colorado.

In the Ventura Basin, southern California, Rockwell et al., (1985) have constructed a chronosequece for soils developed on river terraces that have been displaced by Neotectonic folding and faulting. The soils develop thick A horizons and clay-enriched argillic or cambic B horizons. A number of morphological soil properties, including total soil depth, B horizon thickness, texture and the presence of clay films, together with several chemical properties, including calcium carbonate content and exchangeable bases as well as particle size characteristics were used to assess degree of soil profile development. As with the studies cited above, the soils were age-calibrated using various measures of absolute age determination. The soils range in age from 200,000BP to the present. The derived chronosequence of soils was then used to date and correlate geomorphic surfaces of fluvial origin which have been displaced by tectonic activity.

In the studies cited above grouping of deposits using soil-stratigraphic units has generally been qualitative. Recently however, quantitative techniques have been used to group landforms into units of different relative age using relative weathering variables. Relative weathering variables measure the degree of post-depositional alteration of surficial deposits. This is assumed to vary in proportion to the length of time during which the weathering processes have been operational. Thus Mills and Wagner (1985) have used a number of weathering indices to relatively date and correlate discontinuous, unpaired river terrace fragments along the New River of southwest Virginia.

Such a multi-parameter approach to relative age dating is useful since it reduces the possibility of error in subdivision of deposits due to local site variation in any single parameter (Burke and Birkeland, 1979). Dowdeswell (1982) and Dowdeswell and Morris, (1983) have shown that groups of glacial landforms of similar relative age may be differentiated using a number of relative age indictors including weathering rinds and lichen cover and size. They used Principal Components Analysis to transform these weathering variables into a fewer number of significant components. The scores on the significant components were clustered and Dowdeswell (1982) produced four major groups of deposits. These were tentatively age-correlated with the four groups of Late Quaternary deposits considered by Mahaney (1978) in Colorado, giving 100-300BP; 950- 1,850BP; 3,000-5, 000BP and about 10,000BP as the age ranges for the groups of landforms.

3.4 The Glen Feshie Soils : Environmental Conditions

The present climate of Glen Feshie was discussed in Chapter 2. This discussion was based on the bioclimate map of Scotland published by the Macaulay Institute for Soil Research (Birse, 1976). The map combines a variety of data to give a map of the bioclimatic sub-regions of Scotland in terms of :-

(1) Thermal zonation - by accumulated temperature and exposure

(2) Moisture balance - directly from potential water deficit and potential summer water surplus

(3) Oceanicity - based on accumulated frost.

This scheme is a useful one to use when considering soil development as it is based on the temperatures for plant growth.

The most extensive vegetation community on the terraces where the soils were sampled is dry Calluna moor of Calluna vulgaris, Vaccinium myrtillus and hypnaceous mosses, and lichen-rich Calluna moor; occasional areas of Agrostis-Festuca acid grassland occur; in the upper braided reach remnants of the primitive Caledonian woodland still persist in some areas with a ground vegetation of Calluna vulgaris and lichen.

Most of the study area is underlain by acid Moinian schists, from which the bulk of the valley fill deposits are derived. Granite outcrops in the north and north-east corners of the study area. The rock type of the terrace gravels predominately comprises mica schist clasts with about 10-15% of granite clasts.

Quantitative classification of soil profile development and a stratigraphic correlation based on degree of soil profile development should take into account the texture of the parent material. The sediments of the terraces comprise the parent material of the soils. As a prerequisite to an examination of soil development on the terrace surfaces it is essential to assess the intrinsic textural variability of the parent materials. Variations in these properties may account for important differences in the subsequent operation of weathering and translocatory processes. Such differences will dictate the degree to which comparisons of these processes between profiles will be meaningful.

Gravel units comprise all the Glen Feshie terrace deposits (Chapter 2). The gravels are clast-supported but have a fine gravel and sand infill. The gravel units continue to the surface at all but five of the soil sample sites. As a more or less constant gravel content to the surface establishes uniformity of parent material in stream gravels (Birkeland, 1974) the terrace soil parent materials may be regarded as forming constant parent materials for the development of the surface soils. These gravel units form coarse to very coarse textured, freely drained parent materials for the surface soils. Five of the sample sites, (sites 1,2,8,21 and 28 in Figure 3.2) were sampled on freely drained sandy parent materials.

In clastic parent materials which comprise coarse to very coarse gravel, variation in the matrix size material forming the infill around the clasts may affect the rate of leaching. It is therefore important to establish the homogeneity of this size fraction also (Birkeland, 1974). Table 3.1 gives the percentages of sand, silt and clay for the 35 gravelly parent materials of the 40 sites sampled from the five groups of terraces shown in Figure 3.2.

Table 3.1

Group 1	Group 2	Group 3
% sand = 96.2%	% sand = 97.8%	% sand = 98.4%
% silt = 3.0%	% silt = 1.5%	% silt = 0.9%
% clay = 0.9%	% clay = 0.4%	% clay = 0.2%

Group 4	Group 5
% sand = 98.7%	% sand = 99.0%
% silt = 0.8%	% silt = 0.6%
% clay = 0.2%	% clay = 0.1%

In order to test the null hypothesis that there is no significant difference between the matrix content of the five groups an analysis of variance was carried out using the percentage of sand for each sample for the five groups. The analysis of variance table is given in Table 3.2.

TABLE 3.2

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	4117	4	1029	0.5297
Within gps	0.3303E+05	30	1943	
Total	0.3714E+05	34	1769	

The 5% point for $F(4,30)$ is 2.69, so that the null hypothesis of no significant differences between the groups is accepted.

As the terrace sediments which constitute the soil parent materials exhibit a high degree of homogeneity, any differences in soil profile development are unlikely to be due to variation in textural properties between terraces.

3.5 Soil sampling sites

The surface soils used to examine degree of soil profile development and to derive the soil-stratigraphic units were sampled at 40 sites across the 9 valley profiles within the study reach of the main Feshie valley (Figure 2.1). Some of the soils on the extensive terrace fragments were sampled at more than one site to ensure continuity of soil profile development within the terrace fragment. Soils were also sampled at 2 cross-sections in a tributary to the River Feshie, the Allt Lorgaidh (Figure 2.1). In addition, over 50 exploratory soil pits were dug and soil profile descriptions made on the terrace fragments to ensure continuity of soil profile development for each terrace fragment. Chemical analyses were not performed on these additional pits. The soils were all sampled on former bar complexes so that continuity of topographic location, and therefore site drainage conditions, was maintained. No sample sites had impeded site drainage conditions. None of the sub-soils exhibited mottling which might be indicative of fluctuating water table effects.

Generally, some soils may receive influxes of parent material at the same time as soil formation is continuing. Such soils have been termed cumulative soil profiles (Nikiforoff, 1949). They may be recognised through discontinuities in the depth functions of soil properties or through changes in grain size characteristics with depth (Gerrard, 1981). River terrace soils may be liable to soil formation and concomitant deposition of overbank flood deposits which may produce sediment stratification, and cumulative soil profiles. Such soils have been described from the Severn lower terraces (Hayward and Fenwick, 1983).

Very little evidence of cumulative soil profile development was observed on the Glen Feshie soils. No textural or chemical discontinuities occurred with depth in any of the soil profiles analysed. Absence of cumulative soil profiles on the terrace surfaces in Glen Feshie may be indicative of rapid abandonment of the former floodplain surfaces through channel avulsion. This may mean that the former active zones of the stream had been removed from the effects of flooding and from potential sources of fine material that contribute to the accumulation of cumulative soil profiles. Such an hypothesis is supported by the findings of Bluck (1976). Discussing sedimentation in some Scottish gravel-bed rivers, Bluck has described the removal of bar and channel complexes further and further from the main channel zone as a result of channel avulsion as gravel-bed streams migrate across the valley floor. In these Scottish rivers the sediment is often very coarse and in situations where there are a number of channels on the floodplain the chances of much sediment of sand and silt size being deposited on the abandoned bar surface is greatly reduced. The abandoned bars may then never accumulate an appreciable thickness of fine sediments.

The soil profiles sampled were described in the field according to a procedure based on that of Hodgson (1974). Soil colour was estimated from Munsell Colour charts using moist soil samples. The soils have been classified in consultation with members of the Macaulay Institute for Soil Research, using the classification scheme of the Soil Survey of Scotland. However, with some of the younger soils it is difficult to apply a classification scheme satisfactorily as they are still in the process of horizon differentation and development. Some of the very young soils are thus classified as mineral alluvial soils. In discussing the soil profiles the terms A, E, B and C are employed with respect to an initial working horizon nomenculture. They are used in a general sense and are based on visual criteria.

Bulk samples for laboratory analysis were collected on an horizon basis. Detailed physical and chemical laboratory analyses were conducted on each horizon sampled. Samples were analysed for soil reaction (pH), organic carbon, and iron and aluminium content using the procedures outlined by Bascomb (1974). Iron and aluminium content was measured by atomic absorption spectrometry. In order to ensure reproducibility of results in the chemical extractions each sample was analysed twice. In all over 1,000 extractions were performed. Aluminium is difficult to detect using atomic absorption spectrometry. It was found that duplicate and sometimes triplicate samples gave quite varying results, so for this reason the aluminium data were not included in the final statistical analysis. Particle size analysis of the fine, <2mm fraction was carried out for the parent materials of each soil profile site. A comparative scanning electron microscope study was made to further investigate some of the interpretations of the horizon development of the soils. Details of the general laboratory procedures for all the analyses are given in Appendix 1.

3.6 The Soil-Stratigraphic Parameters

The properties used for differentiation of the soil-stratigraphic units are determined by the type of soil that has developed through the operation of the soil-forming processes for the particular location under consideration.

The combination of the coarse textured, base-poor gravels of the Glen Feshie terrace deposits, the present climatic and vegetation characteristics described above, and the postglacial climatic and vegetation characteristics, (Birks and Mathewes, 1978; Walker, 1984) suggest the likelihood of podzolisation as the dominant pedogenic process (Romans and Robertson, 1983; Gauld, 1982; Reiger, 1983). This is confirmed by the field mapping of Birse and Heslop (1970) for the Soil Survey of Scotland. Iron humus podzols have been identified on the glaciofluvial deposits in Glen Feshie and immature podzols have been described on several of the low-level terrace fragments.

The processes of podzolisation include:-

(1) the processes of organic matter accumulation and decomposition;
(2) the production of chemically active organic acids in the organic-rich surface horizons;
(3) the attack by these acids on silicate and other minerals in the upper part of the soil;
(4) a combination of the acids and the mineral decomposition products;
(5) a downward migration of the resulting organo-metallic complexes;
(6) the precipitation of the complexes to form a sesquioxide enriched B horizon;
(7) the migration of iron and aluminium as mixed inorganic hydroxide sols;
(8) the precipitation of these sols in the B horizon as

coatings around mineral grains;
(Anderson et al., 1982; Farmer and Fraser, 1982; Reiger, 1983).

It has been thought that the translocation of iron and aluminium from the upper horizons of the soil as organo-metallic complexes was the primary mechanism for the formation of the illuvial B horizon in podzols. Additionally, the accumulation of iron as a result of progressively greater in-situ weathering in the B horizon relative to the upper Ea horizon was thought to contribute to intensification of the podzolic B horizon.

Recently however, it has been shown that for Scottish podzols developed on coarse parent materials, translocated sesquioxides are present predominately in inorganic forms in the B horizons of the podzol profiles examined (Anderson et al., 1982; Farmer and Fraser, 1982; Farmer et al., 1985). It is suggested that the first stage in podzolisation involves the formation of a Bs horizon characterised by inorganic deposits of allophane, imogolite, and proto-imogolite type material. In such material there is a complete breakdown of the feldspar structure of mineral grains to form allophane-amorphous material, and inorganic amorphous hydroxides of iron. The imogolite materials formed in the Bs horizons can be deposited only from solutions containing a positively charged hydroxy-aluminium silicate complex. These positively charged sols can form from aluminium and iron brought into solution by non-complexing organic and inorganic acids. These same mechanisms, which dissolved the fresh and weathered minerals in the A_2 or Ea horizon, can continue to operate on the deposited oxides and hydroxides of iron and aluminium in the B horizon, thus deepening the Ea horizon and building up the sesquioxide concentration and depth of the Bs horizon. With further development of the organic surface layers colloidal and soluble organic acids can descend through the Ea horizon until they encounter the sesquioxide-coated surfaces of the Bs horizon. The precipitated acidic organic matter at the top of the Bs horizon can then act as trap for remobilized iron and aluminium thus forming the Bh horizon. These various processes contributing to the development of the podzolic profile may alternate seasonally rather than be widely separated in time.

Considering the processes involved in podzolisation it would be expected that as podzol profiles developed through time a number of parameters of podzolisation should show distinctive age trends. Although there have been no chronosequence studies of podzols in Scottish soils, a number of workers elsewhere have examined the development of podzols with increasing length of time available for the operation of the soil-forming processes. These workers have shown that intensity of podzolisation of soils is a function of the length of time available for podzolisation.

Franzmeier and Whiteside (1963) presented soil chemical data in the form of depth functions for a podzolisation sequence of soils in Michigan. Age-related trends in the depth functions are discussed. Franzmeier and Whiteside show a depletion of organic carbon from the bleached (Ea) horizon and subsequent accumulation in the B horizon, a feature which required between 3,000 and 8,000 years to develop. Maximum amounts of free iron and aluminium were observed to occur in the B horizon with intensity of the maxima increasing with soil age. Horizon differentiation became more pronounced with the age of the soil.

Examination of podzolic soil chronosequences in the Colorado

Front Range (Mahaney, 1974), Wind River Mountains (Mahaney, 1978) and the Cascade Range, Mahaney et al. (1981) has demonstrated significant increases in organic matter content in surface organic horizons with increasing horizon age. In the Colorado Front Range organic matter content showed rapid increases during the first 2,000 years of soil development. Similarly, pH reached a steady state on surfaces varying in age from 1,000 years in the Colorado Front Range to 3,000 years in the Cascade Range. Total iron content of the B horizons of the soils was found to increase steadily with the age of the soils, with the maxima of iron accumulation occurring in the B horizons. The North American studies have also shown time related increases in total solum depth, depth of the B horizon and development of horizons in the podzolic chronosequences. Thus, in the Wind River Mountains soils of the age range 300-100BP have only weakly developed soil profiles with C1/ Cn profiles about 20cm in depth. Soils in the age range 2,000-1,000BP age exhibit an A1/ Cox/ Cn horizon sequence and range in depth from 40-60cm. Showing more complex soil profile development, soils with an age range of 5,000- 3,000BP have soil depths which reach up to 80cm and an A1/ B2ir/ B3/ Cox/ Cn horizon sequence. Soils of about 10,000 years BP exhibit soil depths of about 100cm and well developed O1/ A11/ A12/ B2irh/ IICox/ IICn profiles.

More recently, Ellis and Richards (1985) examined a chronosequence of podzols developed on landslip deposits in Ulvadalen, west-central Norway. These deposits ranged in age from about 20BP to 9,000BP. Organic carbon content in the organic-rich surface horizon increased rapidly during the first 500 years of soil development as rapid vegetation colonisation occurred on the surfaces. Subsequently, the rate of increase slowed, suggesting an approach towards steady state. Organic carbon in the surface horizon may therefore be a rapidly adjusting soil property, attaining a steady state within 10^3 years (Yaalon, 1971). Since sesquioxides are mobile during podzol development, these should exhibit temporal trends. The Ea:B horizon ratio of both pyrophosphate and dithionite-extractable iron and aluminium showed a decrease with increasing age. These trends were interpreted as being due to the progresssive eluviation of sesquioxides from the Ea horizon and subsequent deposition in the B horizon of both organically-bound and inorganic sesquioxides. Sesquioxides appear to represent slowly adjusting soil properties which show no evidence of reaching a steady state within 10^4 years (Birkeland, 1974). Total solum thickness and depth of both the Ea and the illuvial B horizons show increases with the age of the deposit.

Franzmeier and Whiteside (1963b), Mahaney (1978), and Ellis and Richards (1985) have demonstrated several parameters of the processes of podzolisation which are age dependent and which show continuous change over a 10^4 years time scale. The morphological parameters of podzolisation that are recognisable in the field and which are age-dependent include total solum and B horizon thickness, horizon development and horizon colour. Chemical properties of the podzolic chronosequences which are age-dependent include depletion of sesquioxides from the lower A horizon and concentration of sesquioxides in the B horizon. Further, organic matter accumulation and decomposition, and concomitantly, the percentage of organic carbon, in the organic-rich surface horizon progressively increases with soil age.

The soil properties used for the derivation of the soil-stratigraphic units on the Glen Feshie terraces were selected after a consideration of the appropriate parameters of podzolisation. Those selected included:-

(1) Total solum depth;
(2) B horizon thickness;
(3) Percentage of organic carbon in the surface organic horizon;
(4) Pyrophosphate and dithionite-extractable iron content in each horizon;
(5) pH;
(6) soil colour;
(7) horizon differentiation.

Pyrophosphate and dithionite-extractable iron contents were evaluated for all of the samples collected. The two extractions were performed on separate sub-samples and analysed by atomic absorption spectrometry. The two extractions were performed in order to separate the iron compounds associated with inorganic processes (dithionite-extractable) from those bound up in organo-metallic complexes (pyrophosphate-extractable). As indices of translocation, the ratios of the quantity of iron in the lower A:B and B:C horizons were derived for each profile. The ratios assess the relative concentration of iron in the illuvial B horizon of the soils, and allow relative variation between profiles to be compared by eliminating the effect of small-scale variability in iron content of the parent materials.

3.7 Numerical Analysis

(a) Principal Components Analysis

Principal Components Analysis was performed on the data matrix of soil variables. Ten variables were used in the analysis for each of the 40 sampled soil profiles developed on the terrace surfaces. These 10 variables included both morphological and chemical properties of the soils. The variables were total solum depth, B horizon thickness, % organic carbon in the surface organic horizon, the lower A:B ratios of pyrophosphate-extractable iron, dithionite-extractable iron, and total iron, the B:C ratios of the three iron parameters and the ratio of A:B horizon soil pH. This gave a 40x10 matrix.

Principal Components Analysis has been used as a major analytical tool in a variety of geomorphological investigations (for example, Mather and Doornkamp, 1970; Gardiner, 1979; Dowdeswell, 1982; Dowdeswell amd Morris, 1983; Robertson-Rintoul, in press) as well as in a number of pedological investigations (for example, Webster, 1977; Jacobson and Birks, 1980; Sondheim et al., 1981; Richardson and Bigler, 1984). In pedological investigations this may be particularly useful where a number of parameters are required to represent the soil-forming processes. Thus Webster used 15 soil properties from 85 soil sites to quantify the continuous variation in soil profile characteristics with variation in the water regime and texture of the soil. Several variables representing water regime characteristics and texture were measured and the raw data used as input for a Principal Components Analysis. Two significant components were derived which allowed continuous description of the water regime and texture of the soils. Sondheim et al.(1981) studying a podzolic chronosequence on a Canadian prograded beach showed a dominant trend of increasing intensity of podzolisation with age for podzols developed on five beach ridges of different age. This trend was represented by one principal component, the scores for each soil site quantifying degree of podzolisation of the soils. The derived components from a Principal Components Analysis may thus relate to specific pedogenic processes.

Principal Components Analysis is a mathematical technique which transforms the original data matrix into a new data matrix of principal component scores (Daultrey, 1976). This is valuable for a number of reasons. First, a transformation of the original point distibution to principal components will eliminate the redundancies incurred when several variables display a single pattern of concomitant variation. Each pattern of intercorrelated variables is replaced by a single principal component which represents the pattern. The original data matrix can therefore be described approximately in terms of a smaller number of principal components. Second, the derived principal components or dimensions in the Principal Components Analysis are mutually orthogonal and therefore uncorrelated. The derivation of the principal components therefore eliminates any collinearity problems that may arise in subsequent statistical analyses. Third, in the Principal Components Analysis the raw data matrix is standardised. This ensures that all the variables have equal variance characteristics and equal potential for contributing to the extracted components. This procedure is essential if it is desirable to compare the distribution of one variable with another when the variables are expressed in different units of measurement.

Principal Components Analysis maximises the variance of the first component from the basic correlation matrix, with sucessive components extracting the maximum amount of residual variance. This mathematical technique is therefore an elegant method by which associations within a data matrix, comprising several intercorrelated variables, may be parsimoniously expressed.

Principal Components Analysis is also of interest because it may be used to reveal structure within a multivariate data set. The derived components from the analysis are a function of the original variables. From this point of view the salient characteristic of Principal Components Analysis is the maximisation of the variance on the first component. Any systematic variation in the group of initial variables may then be represented numerically on a single continuous scale by calculating the principal component scores for the first component. An analysis of the pattern of the structure of the scores on subsequent significant components may yield secondary trends superimposed on the dominant trend represented by the first component. For example, Sondheim _et al_. (1981) found that their second component indicated systematic variation in the distribution of cations in the soils with increasing distance from the sea. This trend was separated from the dominant component which represented increasing podzolisation with age of the soil.

(b) Results

The derived components or eigenvectors, and the eigenvalues for the Glen Feshie soils data set are presented in Table 3.3.

TABLE 3.3

Eigenvalues and cumulative proportion of total variance for 3 principal components

1	6.0014	0.60014
2	1.4013	0.74027
3	1.0487	0.84515

Principal Component Loadings Matrix

		Component 1	Component 2	Component 3
1	total depth	0.7879	0.0062	0.5662
2	depth of B	0.7920	0.0336	0.5691
3	% O Carbon	0.7098	-0.1212	-0.4651
4	pH	-0.0589	0.9457	-0.0371
5	Fed A : B	-0.8948	0.2572	0.1979
6	Fed B : C	0.8765	-0.2261	0.0849
7	Fep A : B	-0.6785	-0.4023	0.2654
8	Fep B : C	0.7440	0.4546	0.0275
9	Fetot A : B	-0.9069	0.0723	0.2472
10	Fetot B : C	0.9170	-0.0078	-0.0899

Although there is no clear rule for determining significant principal components from the analysis, Kaiser's rule is generally adopted (see for example, Gardiner, 1979; Dowdeswell, 1982; Sondheim *et al*., 1981; Dowdeswell and Morris, 1983; Miles and Norcliffe, 1983). By Kaiser's rule significant components in a Principal Components Analysis are those which account for an eigenvalue not less than unity.

Three of the derived components had eigenvalues of greater than 1.0 and collectively they account for 84.5% of the original variance. Component 1 alone accounts for 60% of the total variance. Components are mathematically independent so elucidation of their physical meaning, established through an appraisal of the eigenvector loadings, should be made on an individual basis (Gould, 1967). The loadings on the first component (Table 3.3) show that the component has strong positive correlations with the three B:C ratios of iron, organic carbon, total solum depth and B horizon thickness. High negative loadings occur on the three lower A:B iron ratios. pH does not load significantly onto the first component. The negative loadings on the lower A:B ratios and the positive loadings onto the B:C ratios represent depletion of iron from the upper soil horizons and an accumulation of iron in the illuvial B horizons where there is a greater amount of iron relative to the C horizons. These trends in the iron variables are indicative of one of the major processes of podzolisation, the downward migration of iron and its subsequent precipitation to form a sesquioxide-enriched B horizon. The positive loading on the organic carbon variable is suggestive of increasing surface horizon organic carbon content in association with the building of the organic-rich surface horizon that is characteristic of podzols. Strong positive correlations of solum depth and B horizon thickness represent increasing soil depth with the development of increasing numbers of horizon, and the thickening of the individual horizons.

The magnitude and direction of the loadings on the first component indicate that this dimension represents a compound index of the variables representing the processes of podzolisation. This first component is therefore an index of the intensity of podzolisation. As a number of studies have shown that intensity of podzolisation is a function of the duration of pedogenesis, then this axis may also be regarded as having a temporal dimension.

The pH variable shows a very high loading onto the second component. Soil reaction in all the soils is generally low due to the base-poor parent materials. However, pH does vary between sites and a plot of the component 2 scores against the ratio of A:B horizon pH helps to elucidate the meaning of the second component (Figure 3.1). There is a clear division between the Calluna heath vegetation and the acid grassland sites. For sites on the heath vegetation scores on component 2 are negative and the value for the pH ratio is also low. For the sites on the acid grassland vegetation, and those with a sandy parent material, the component 2 score is positive and the ratio for pH is higher. This component is separating out a secondary trend of variation in pH with local changes in vegetation.

Positive eigenvector loadings on the soil morphological variables and a negative loading on organic carbon for component 3 indicates variation in vegetation and calibre of parent material. High positive scores occur on sites with sandy parent materials, and/or acid grassland, whilst high negative scores are exhibited by sites with gravelly sola and Calluna heath. Components 2 and 3 therefore isolate local site variation in vegetation and parent material factors from the dominant trend in the original point distribution, that of increasing intensity of podzolisation.

Projecting the component scores for each soil profile onto the first principal axis enables the structure of the first

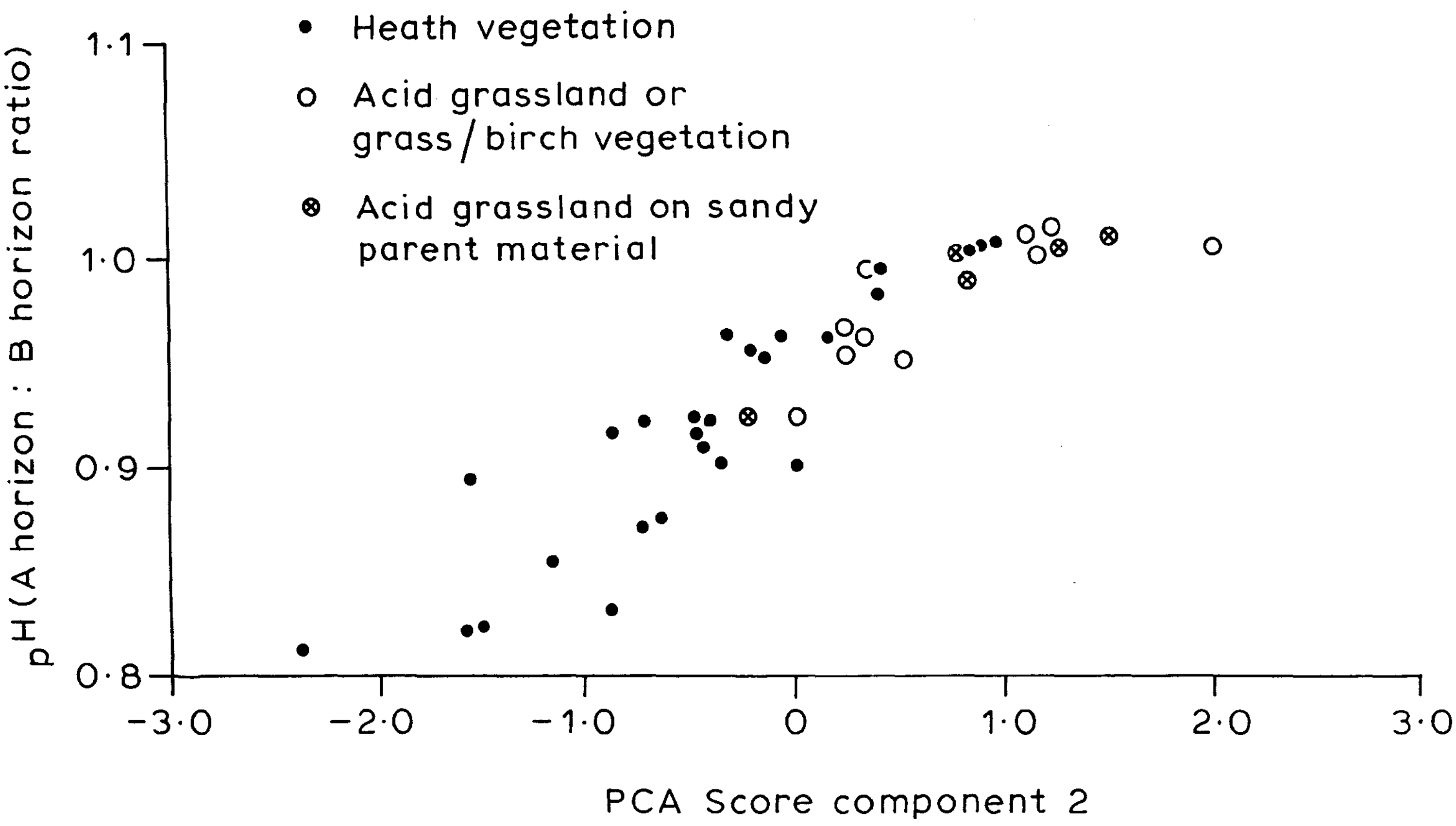


Figure 3.1 Plot of the Component 2 Scores for 40 soil profiles against A : B horizon pH

component to be examined more clearly (Figure 3.2). Here the standardised scores are ranked on a single numerical scale and show clearly the variation in value of the podzolisation component for the 40 profile sites.

It is immediately apparent that the point distribution is dispersed in discrete clusters of scores along the principal axis. As the principal axis represents intensity of podzolisation this arrangement of points suggests that the development of the surface soils on the Glen Feshie terraces has been stepped in time, and that this phased development of the surface soils has been in response to the episodic development of the valley floor landform elements. Five clusters of points may be identified.

(c) Cluster Analysis

Cluster analysis is the modern statistical method of partitioning an observed sample population into disjoint classes in order to produce an operational classificaion of the data. This classification may be used to reveal associations and structure within the data set, and possibly provide the basis for a general model with which to describe sample populations (Wishart, 1978).

Cluster analysis was used here to statistically confirm the identification of the five clusters of points projected onto the first principal axis which represents intensity of podzolisation. A minimum variance type of clustering, which minimises within-group variance and maximises between-group differences, was therefore necessary. Ward's (1963) hierarchical grouping algorithm, minimising an error sum of squares objective function, is probably the best option for finding tight minimum

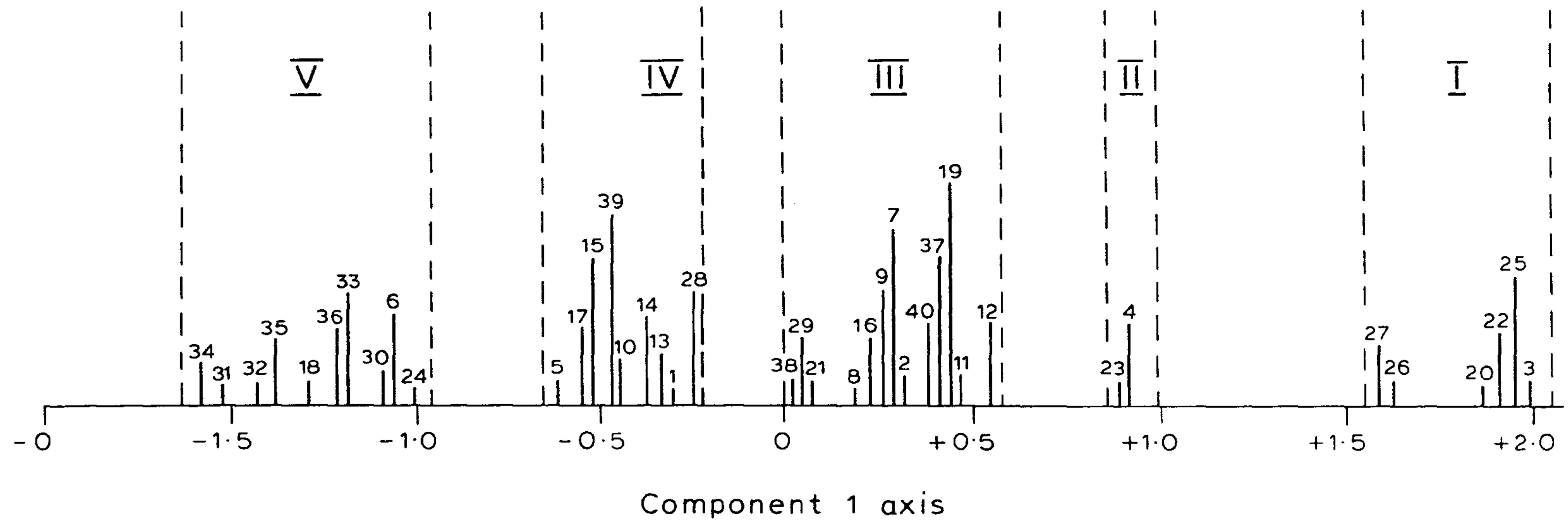


Figure 3.2 Projection of Component 1 Scores for Soil-Stratigraphic Data onto the First Principal Axis

variance clusters (Wishart, 1978) and was therefore used here.

(d) Results

The standardised scores on the first principal component were used as the data input for the clustering procedure. Ward (1963) stated that at any stage in a hierarchical grouping analysis, the "loss of information" which results from the grouping of n points into clusters can be measured by the total sum of the squared deviations of every point from the mean of the cluster to which it has been allocated. At each step in the clustering procedure the union of every possible pair of clusters is considered, and the two clusters whose fusion results in the minimum increase in the error sum of squares are combined. At the beginning of the procedure each individual is regarded as single point cluster. The first fusion combines individuals with the highest similarity, followed by less similar individuals until all the individuals in the data set have been clustered.

Figure 3.3 shows the results of the clustering procedure for the 40 soil sites presented as a dendrogram. The screen test may be used to define a cut-off point for the acceptance of the number of groups in a cluster analysis (Ward, 1963; Wishart, 1978; Miles and Norcliffe, 1984). Plotting within-group variance as a proportion of total variance against number of groups indicates a marked discontinuity where the scores are clustered into five groups (Figure 3.4). Before the discontinuity the error sum of squares increased slowly. Beyond the discontinuity, from five to four clusters, there is a large increase in the error sum of squares. This would suggest, according to Ward's guide to the use of the method, that the level which best minimises the

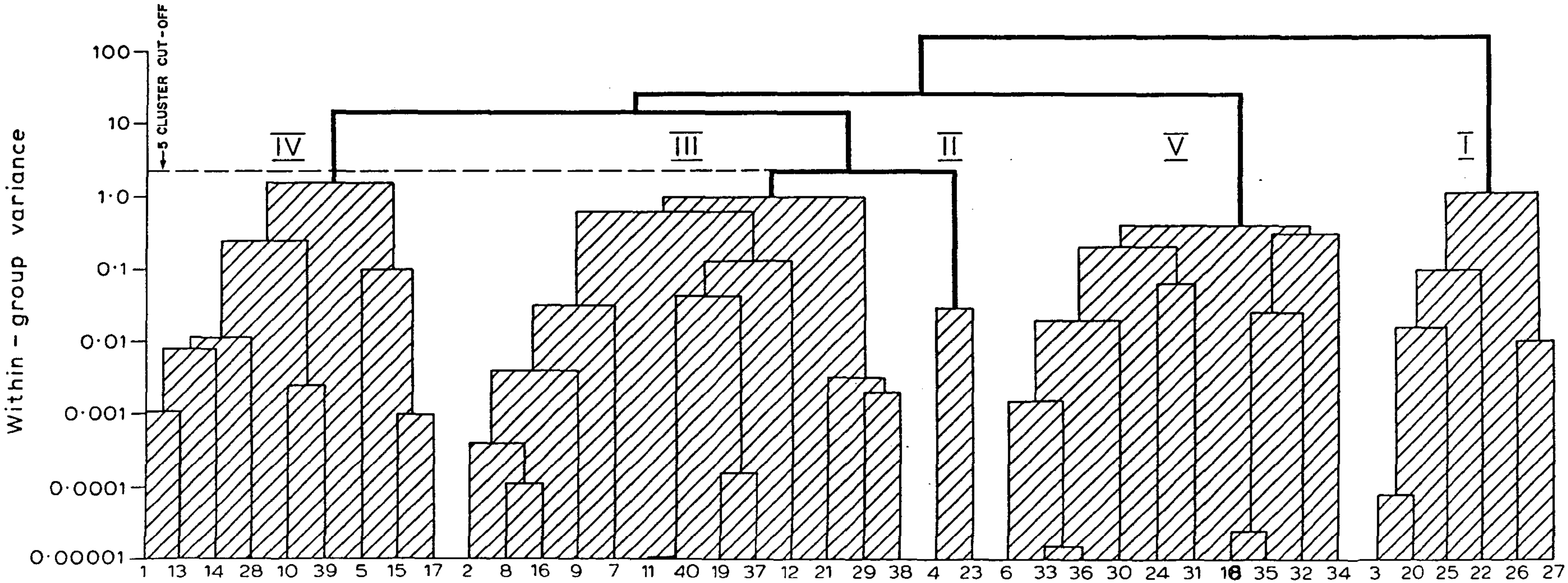


Figure 3.3 Cluster Analysis from Ward's Heirarchical Grouping Algorithm for the Soil Stratigraphic Data

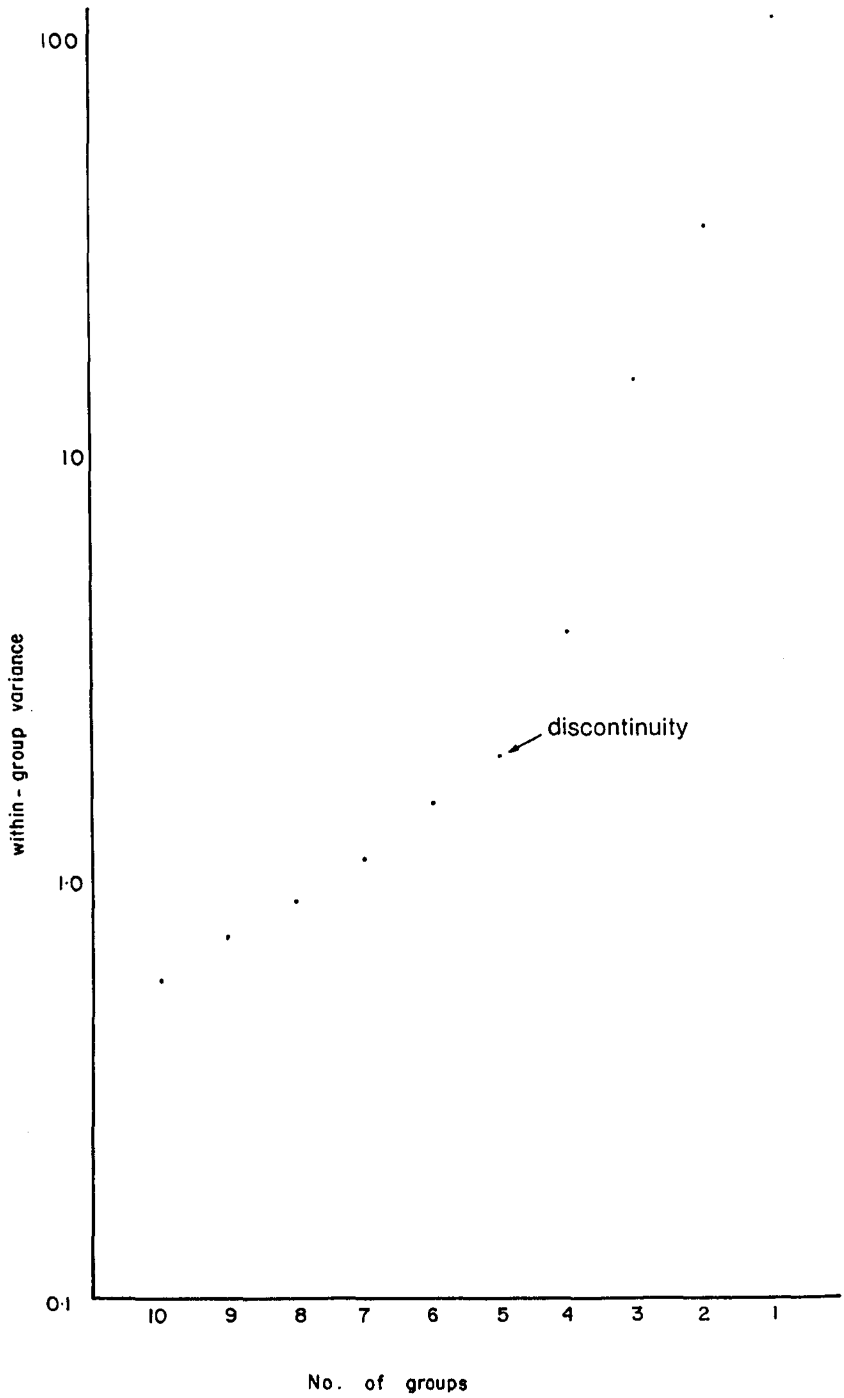


Figure 3.4 Acceleration of Information Loss for Cluster - Analysis of Soil Stratigraphic Data

within-group variance and maximises the between-group differences is the five cluster grouping.

The principal axis diagram (Figure 3.2) may be regarded as graphically presenting the results of the cluster analysis. The plot shows that the cluster analysis has produced the same groups as those that were initially identified from the principal axis plot. The clusters are not only separated on the plot, but they are also reasonably compact, suggesting that the five grouping scheme constitutes an operational classification of the original data matrix.

The statistical significance of the grouping scheme may be tested using discriminant analysis (Bull, 1978; Dowdeswell, 1982). However, if only one principal component has been used in the classification the statistical significance of the grouping scheme may be tested using classical analysis of variance (Dowdeswell and Morris, 1983).

Analysis of variance requires an a priori classification of the sites into groups. These have been established from the cluster analysis presented above. The analysis of variance table is given in Table 3.4 below.

TABLE 3.4

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	31.05	4	7.763	76.63
Within gps	3.34	35	0.1013	
Total	34.39	39	0.9296	

The 1% point for $F(4,35)$ is 3.925 so that the null hypothesis of no significant difference between the groups can be rejected. The five grouping scheme is significant at greater than the 0.01 level.

In this grouping scheme (Figure 3.2) it is notable that profiles sampled on the same terrace fragment were grouped together. These were profiles 7 and 8 (with sandy parent materials) and 11 (with a gravel parent material); profiles 21 (with a sandy parent material) and profile 19 (with a gravel parent material); profiles 26 and 27 and finally profiles 14 and 15.

3.8 Discussion : The soil-stratigraphic units

The soil sites within one cluster have a wide geographical distribution in Glen Feshie and possess clearly defined physical and chemical features which permit their recognition as marker horizons. Finkl (1980) notes that these are the requirements for defining a soil stratigraphic unit (Finkl, 1980). Each cluster of soils may thus be regarded as representing one soil-stratigraphic unit.

The five grouping scheme of the 40 soil sites uniquely locates each cluster of profiles within the frame of reference defined by the first principal component. Because the profiles have been clustered using the scores from the component which represents a compound index of parameters representing the degree of podzolisation of the soil profiles, each group must therefore be characterised by a particular degree of podzolisation. As intensity of podzolisation can be shown to be a function of the length of time available for the operation of the soil-forming processes then the arrangement of the clusters along the first principal axis in Figure 3.2 must be chronological with the five groups arranged on a relative time scale.

(a) The profile morphology of the soil groups

Figure 3.5 is a diagrammatic representation of the soil profile changes between each of the five statistically-derived soil groups. Soil profile morphology can be seen to vary markedly from the group I to the group V soils. Plates 3.1-3.7 show typical soil profiles from each of the five groups. Table 3.5 gives representative soil profile descriptions for soils from each of the groups.

TABLE 3.5

Profile name:	Glen Feshie, 1
Profile group:	Group V
Slope:	0 degrees
Vegetation:	Vaccinio-Ericetum cinereae: lichen-rich heather moor
Soil drainage:	Free
Series:	Alluvial soil, Dryburn
Parent material:	River gravels
Rock type:	Mica schist and granite
Major soil subgroup:	mineral alluvial soil

Ah : 0-3cm; very dark brown 10YR 2/2 medium sand; no mottles; weak fine granular structure; moist; friable; many medium woody and many fine fibrous roots; common medium sub-rounded stones; clear wavy boundary.

BC : 3-7cm; dark brown 10YR 3/3 medium sand; no mottles; single grain structure; moist; loose; few fine

Generalised Soil Profile Morphologies of the Five Soil - Stratigraphic Units with Colour Development Equivalents

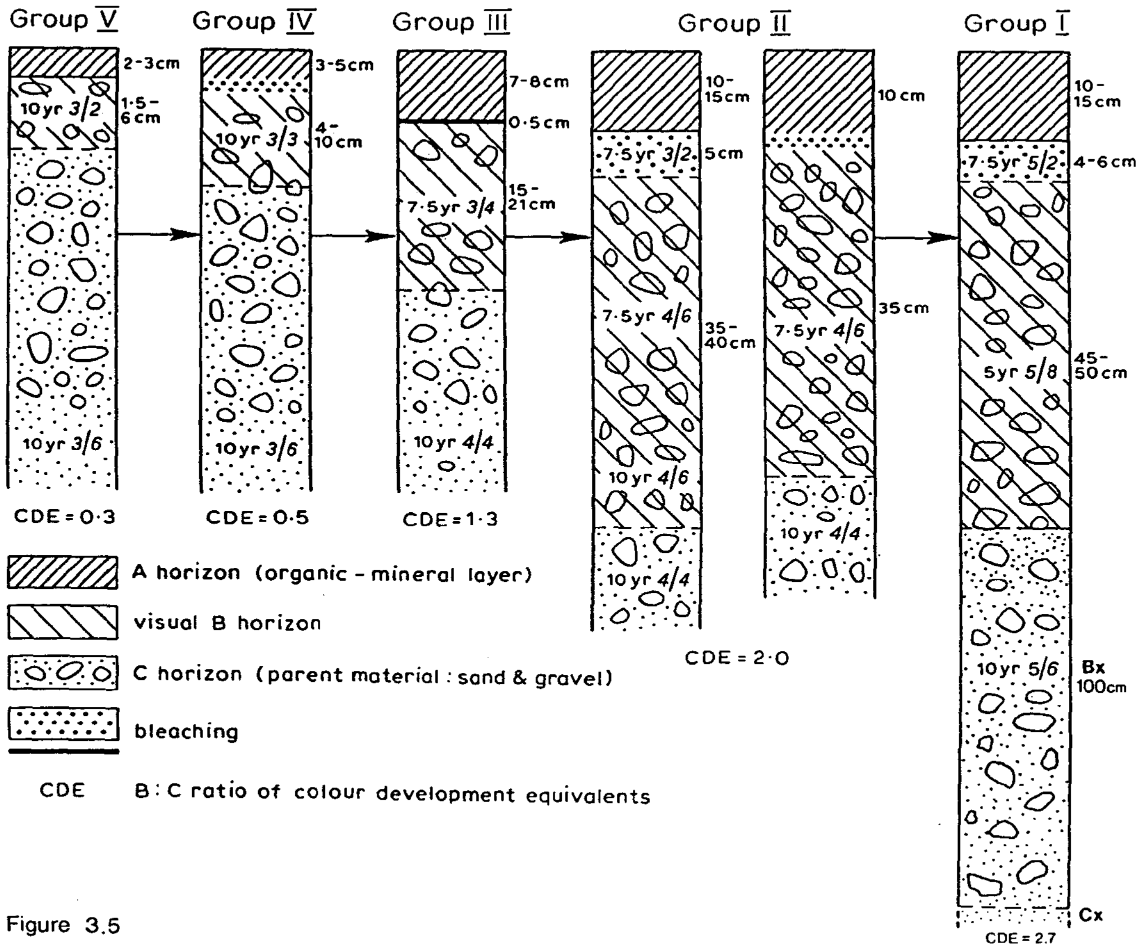


Figure 3.5

Soil profile from the group V soil stratigraphic unit

Plate 3.1

Plate 3.2

Soil profile from the group V soil stratigraphic unit

Soil profile from

the group IV soil

stratigraphic unit

Plate 3.3

Soil profile from

the group III soil

stratigraphic unit

Plate 3.4

Soil profile from

the group III soil

stratigraphic unit

Plate 3.5

Soil profile from

the group II soil

stratigraphic unit

Plate 3.6

Soil profile from the group I soil stratigraphic unit

Plate 3.7

Platy structure of the Bx horizon from the group I podzol

Plate 3.8

fibrous roots; abundant medium sub-rounded stones; gradual wavy boundary.

C : 7-60cm; dark yellowish brown 10YR 4/4 coarse sand; no mottles; single grain structure; moist; loose; few fine fibrous roots; abundant medium sub-rounded stones.

Profile name:	Glen Feshie, 2
Profile group:	Group IV
Slope:	0 degrees
Vegetation:	Vaccinio-Ericetum cinereae: lichen-rich heather moor
Soil drainage:	Free
Series:	Alluvial soil, Dryburn
Parent material:	River gravels
Rock type:	Mica schist and granite
Major soil subgroup:	brown podzolic soil

HA : 0-5cm; very dark brown 10YR 2/2 fine sand; no mottles; weak fine granular structure; moist; friable; many medium woody and abundant fibrous roots; few large sub-rounded stones; clear smooth boundary.

Bs : 5-10cm; dark yellowish brown 10 YR 3/4 loamy fine sand; no mottles; single grain structure; moist; loose; abundant fine fibrous roots; abundant large sub-rounded stones; gradual wavy boundary.

C : 10-60cm; dark yellowish brown 10YR 4/6 medium sand; no mottles; single grain struture; moist; loose; few fine fibrous roots; abundant large sub-rounded stones.

Profile name:	Glen Feshie, 3
Profile group:	Group III

Slope:	0 degrees
Vegetation:	Vaccinio-Ericetum cinereae: lichen-rich heather moor
Soil drainage:	Free
Series:	Alluvial soil, Dryburn
Parent material:	River gravels
Rock type:	Mica schist and granite
Major soil subgroup:	brown podzolic soil

H : 0-7cm; black 5YR 2/1 no mineral content; semi-fibrous; moist; weak fine granular structure; many medium woody and fine fibrous roots; common medium sub-angular stones; sharp wavy boundary.

AEh : 7-8cm; very dark brown 10 YR 2/2 fine sand; no mottles; single grain structure; moist; friable; many fine fibrous roots; abundant medium sub-angular stones; clear wavy boundary.

Bhs : 8-21cm; dark brown 7.5 YR 3/4 loamy fine sand; no mottles; single grain structure; moist; very friable; many fine fibrous roots; abundant medium sub-angular stones; gradual wavy boundary.

C : 21-60cm; dark yellowish brown 10 YR 4/4 medium-coarse sand; no mottles; single grain structure; moist; loose; few fine fibrous roots; abundant large sub-angular stones.

Profile name:	Glen Feshie, 4
Profile group:	Group II
Slope:	0 degrees
Vegetation:	Vaccinio-Ericetum cinereae: lichen-rich heather moor
Soil drainage:	Free

Series: Corby

Parent material: Glaciofluvial gravels

Rock type: Mica schist and granite

Major soil subgroup: Iron humus podzol

H : 0-5cm; black 5 YR 2/1 no mineral content; semi-fibrous; moist; massive structure; abundant fine fibrous and abundant medium woody roots; few medium sub-angular stones; sharp smooth boundary.

AeH : 5-15cm; very dark brown 10YR 2/2 loamy fine sand; no mottles; very weak medium granular structure; moist; friable; many fine fibrous and few medium woody roots; few medium sub-angular stones; clear smooth wavy boundary.

Bhs : 15-25cm; dark reddish brown 5 YR 3/3 medium loamy sand; no mottles; single grain structure; moist; friable; many fine fibrous roots; abundant medium sub-angular stones; gradual smooth boundary.

Bs : 25-45cm; strong brown 7.5 YR 4/6 medium sand; no mottles; single grain structure; moist; very friable; weak induration; weak cementation; many fine fibrous roots; abundant medium sub-angular stones; gradual smooth boundary.

C : 45-100cm; dark yellowish brown 10 YR 4/4 medium sand; no mottles; single grain structure; moist; loose; no roots; abundant medium sub-angular stones.

Profile name: Glen Feshie, 5

Profile group: Group I

Slope: 0 degrees

Vegetation: Vaccinio-Ericetum cinereae: lichen-rich heather moor

Soil drainage: Free

Series: Corby

Parent material: Glaciofluvial gravels

Rock type: Mica schist and granite

Major soil subgroup: iron humus podzol

H : 0-5cm; very dark brown 5 YR 2/2 no mineral content; semi-fibrous; moist; massive structure; abundant fine fibrous and abundant medium woody roots; few medium sub-angular stones; sharp smooth boundary.

AEh : 5-15cm; very dark brown 10YR 2/2 loamy fine sand; no mottles; very weak medium granular structure; moist; friable; many fine fibrous and few medium woody roots; abundant large sub-angular stones; sharp wavy boundary.

Bhs : 15-35cm; dark reddish brown 5 YR 3/2 loamy fine sand; no mottles; single grain structure; moist; very friable; common fibrous roots; abundant large sub-angular stones; gradual wavy boundary.

Bs : 35-50cm; yellowish red 5 YR 5/8 loamy fine sand; no mottles; single grain structure; moist; very friable; weak cementation; few fine fibrous roots; abundant large sub-angular stones; diffuse wavy boundary.

Bx : 50-100cm; yellowish brown 10 YR 5/6 medium sand; no mottles; massive to medium platy structure; moist; friable; medium induration; fine fibrous roots; abundant large sub-angular stones; diffuse wavy boundary.

C : 100-150cm; dark yellowish brown 10 YR 4/6 medium sand; no mottles; massive to very weak medium platy structure; moist; loose; no roots; abundant large sub-angular stones.

The group V soils are weakly developed mineral soils with an <u>Ah/</u>

BC/ C horizonation; the group IV soils show greater horizon differentiation and are brown podzolic soils which comprise HA/ Bs/ C horizons; the group III soils are better developed brown podzolic soils with an H/ A/ Bs/ C horizon sequence. The group II soils are well developed iron humus podzols with an H/ AEh/ Bhs/ Bs/ C horizonation. It should be noted that one of the group II soils (Figure 3.5) appears to conform to the modified podzols developed under acid grassland and birch as discussed by Gauld (1982). These soils lack a bleached AEh horizon, (although they do show the presence of many bleached grains towards the base of the Ah horizon) but have podzolised B horizons that are visually and chemically similar to podzols with bleached horizons. The group I soils exhibit the greatest intensity of soil development. They are also iron humus podzols and typically possess a complex sequence of horizons, H/ AEh/ Bhs/ Bs/ Bx/ C.

The Aeh horizons in the Glen Feshie soils are spatially variable within one soil group. Bleaching frequently occurs in lenses or horizons which exhibit spatial discontinuity. They may also be brown in colour, with moist Munsell Colours of 10YR 2/2 or 7.5 YR 3/2 as a result of the presence of organic material. More typically bleached colours such as 7.5YR 5/2, are found on some of the group I soils.

The changes in horizon sequence described above for the Glen Feshie soils are comparable with those described by Mahaney (1974, 1978) for the Holocene podzolic sequences comprising the soil-stratigraphic units in the Colorado Front Range and the Wind River Mountains respectively. The changes in horizon sequence in the Glen Feshie sites are indicative of a trend of increasing intensity of podzolisation and soil profile development from the incipient podzols of the group V soils to the mature iron humus podzols of the group I soils.

Total solum thickness and B horizon thickness both increase from the group V to the group I soils. Typical values of total solum and B horizon thickness are shown in Figure 3.5. Analysis of variance was used to test the null hypothesis that there was no significant difference in total solum thickness and B horizon thickness between the five soil groups. The analysis of variance table for total solum thickness is given in Table 3.6, and that for B horizon thickness is given in Table 3.7.

TABLE 3.6

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	0.7752E-04	4	0.1938E-04	19.34
Within gps	0.2806E-04	35	0.1002E-05	
Total	0.1056E-03	39	0.3299E-04	

The 1% point for F(4,35) is 3.925 so that the null hypothesis of no significant difference between the groups can be rejected.

TABLE 3.7

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	0.5832E-04	4	0.1458E-04	17.78
Within gps	0.2464E-04	35	0.8214E-04	
Total	0.8297E-04	39	0.2248E-04	

The 1% point for F(4,35) is 3.925 so that the null hypothesis of no significant difference between the groups can be rejected with confidence. The five groups show significantly different soil depths at greater than the 0.01 level. For the Glen Feshie soils both total solum thickness and B horizon thickness are soil properties which are useful for differentiating between soil profiles exhibiting different degrees of soil profile development. This trend of increasing solum depth with intensity of podzolisation is comparable with the findings of Mahaney (1974, 1978) in North America and the Holocene podzolic chronosequence described by Ellis and Richards (1985) from Norway.

Organic matter accumulation appears to be operative at all the profile sites examined, as indicated by the presence of surface organic-rich horizons. As the profile descriptions above indicate, however, there is a marked difference in this horizon between the group V, group IV and groups III-I soils. The group V soils exhibit an Ah horizon, or organic-mineral horizon where the organic material has become incorporated into the upper part of the mineral soil. The group IV soils are characterised by a greater accumulation of organic material at the surface and possess HA horizons. The surface horizon for the groups III through to group I soils comprises an almost exclusively organic layer. In these latter groups the organic rich-layer overlies a mineral horizon (AEh) which visually exhibits progressively greater degrees of bleaching from the group III to the group I soils (Figure 3.5). Degree of bleaching within the group III soils is variable however, with some soils displaying only occasional bleached lenses, rather than a continuous horizon of bleached mineral material. At occasional sites the group IV soils also exhibit some bleaching whilst at others, as with the group V soils, the lower parts of the A horizons exhibit the presence of many bleached grains, possibly indicating a trend towards the development of incipient bleached layers.

Soil colour is another morphological property which may be indicative of intensity of soil development, especially with respect to the development of the visual B horizon (Birkeland, 1974; Bockheim, 1979a). Colour development equivalents (CDE's) as developed by Bockheim (1979a) may be used to indicate degree of soil profile development. Colour development equivalents are calculated as the product of a numerical notation of hue and chroma as given in Munsell Colour Charts. CDE's were designed as a measure of oxidation intensity and were shown by Bockheim to be related to the length of time available for the operation of the soil-forming processes. Larger values of CDE's denote red and brown colour components, which appear to indicate a high degree of soil development. As the value of the CDE's decreases, however, yellow and grey colour components dominate and appear to be more representative of soil parent material colours and therefore the early stages of soil development.

Colour development equivalents were calculated for the 40 soil profiles examined in Glen Feshie. The ratio of B to C horizon CDE's was derived in order to allow for variation in degree of parent material oxidation. Average values for the CDE ratios are given in Figure 3.5. These vary from a value of 0.3 for the group V soils to 2.7 for the group I soils. Analysis of variance (Table 3.8) was again used to test the null hypothesis that there was no difference in soil colour between the groups.

TABLE 3.8

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	19.08	4	4.771	377.7
Within gps	0.2527	35	0.1263E-01	
Total	19.34	39	0.8057	

The 1% point for F(4,35) is 3.925 so that the null hypothesis of no significant difference between the groups can be rejected. The five groups show significantly different ratios of B:C horizon CDE's and perhaps therefore in degree of soil development. The group V and group IV soils have relatively small values for the CDE ratios which may indicate that they are still in the early stages of soil development. The group III through to group I soils exhibit a steady rise in the value of the CDE ratio which may be suggestive of increasing intensity of soil development and hence soil age.

The soil profile descriptions together with the three sets of analysis of variance carried out on the morphological properties of total solum thickness, B horizon thickness and soil colour for the Glen Feshie soils suggests that soil profile morphological properties provide very good discriminators between soils of different groups and therefore stages of soil profile development.

The soils from the group I profiles exhibit some interesting soil macrostructures which are not found in the other four soil groups. These are :-

(1) wavy horizon boundaries

(2) the presence of silt cappings on the clasts

(3) the presence of a well-indurated layer in the lower B horizon, the Bx horizon, which possesses a massive to platy structure and a marked colour change from the horizon above it.

The profile boundaries of the group I soils below the AEh horizons are generally wavy and even convoluted in some locations. Such boundaries may be a consequence of cryoturbation occurring in a periglacial environment (Chartres, 1980).

In the profiles of the group I soils cappings of fine sand and silt-sized material were observed on the upper surfaces of the clasts. These features occurred in the lower B horizons of the soils. Such cappings have been described as relict features in the lower B horizons (B_3 horizons) of mountain soils in north-east Scotland (Romans et al., 1966; Romans and Robertson, 1983). They have been recognised in mature podzols developed on soli-flucted till and in peaty podzols developed on glaciofluvial deposits in the eastern Grampians (Romans and Robertson, 1983). These features are suggested to be relict and represent the earliest stages of soil profile development of soils forming after deglaciation of the late Devensian ice-sheet. It is suggested that silt cappings, and silt droplet fabric seen in thin section, form in soil on freely draining sites where a frozen crust develops in winter. A dry layer forms in winter between the frozen crust and underlying permafrost by migration of water towards the cold surfaces. In spring most of the water content of the frozen surface layer is rapidly removed by lateral drainage and eventually the final basal layer melts to produce muddy droplets. These muddy droplets drain rapidly into the lower soil horizons leaving silt droplets on the upper surfaces of clasts. These features have been recognised in the subsurface horizons of soils in a number of present day cold environments, for example, south-east Iceland (Romans et al., 1980) Norway (Ellis, 1983; Mellor, 1985) and Canada (Mermut and St. Arnaud, 1981), where they have been interpreted as the result of translocation following the melting of ice in the soil. It has been suggested that these cappings form during the early post-depositional stages of recently deglaciated landscapes (Boulton and Dent, 1974; Romans et al., 1980). The presence of these relict features in the B horizons of the group I soils is therefore suggestive of soil formation in a period immediately following deglaciation.

Further evidence of the formation of soil macrostructures during conditions of permafrost comes from the massive-platy structure exhibited by the lower B horizons (the Bx horizon) of the group I soils. This massive to platy structure can be seen clearly in Plates 3.7 and 3.8 in the lower profiles of the podzols. The Bx horizon occurs below the colour change from the Bs horizon. Such horizons, called fragipans, occur in Britain mainly in deposits of late Pleistocene age (Hodgson, 1974), and have been recognised in mountain soils in north-east Scotland where they have been described as dense silty layers (Romans et al., 1966; Romans et al., 1980; Romans and Robertson, 1983). Fragipans have also been described from iron podzols developed on glaciofluvial deposits in the Nethy Bridge area, Strathspey (Anderson et al., 1982). These horizons are compact, dense and frequently brittle indurated horizons which are often found in association with silt cappings. They exhibit both a massive and a platy structure, with the platy structure best developed near the top of the layer. There is usually a marked difference in colour between this horizon and the horizon above it.

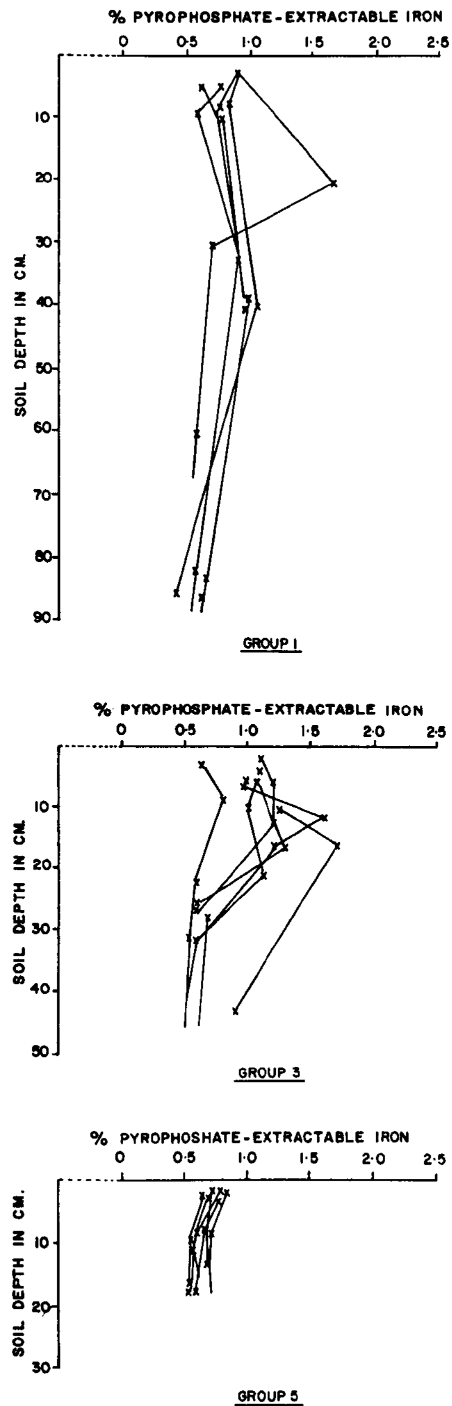
These horizons have been described from environments where the soil is subjected to permafrost conditions (Fitzpatrick, 1956, 1980). Fitzpatrick suggests that these fragipan horizons were initially permafrost characterised by horizontal veins of ice and lenses of frozen soil which were compacted by the growth of the ice. With the amelioration of climate at the end of the Dimlington Stadial the ice melted leaving behind compacted mineral material.

These features have not been observed in the B horizons of the other soil groups. Taken together the structures observed in the B horizons of the group I soils are suggestive of formation in a cold climate environment. Romans et al. (1966) have attributed such features to either a period of formation following deglaciation of the late Devensian ice sheet, or to the phase of periglacial climate which accompanied the Loch Lomond Readvance.

(b) The soil chemistry of the soil groups

Pyrophosphate-extractable iron data provide a measure of the organically-bound iron existing in the form of organo-metallic complexes, whilst dithionite-extractable iron data provide a measure of the inorganic iron released by weathering (Bascomb, 1968; McKeague et al., 1971; Childs et al., 1983; Farmer et al., 1983). The pyrophosphate extractant is specific for organic complexes of iron, whilst the dithionite extracts iron from the inorganic amorphous iron (amorphous hydrated iron oxide) and microcrystalline iron in soil.

Depth function plots of pyrophosphate-extractable iron for the 40 soil profiles are presented in Figure 3.6. For the group III, II and I soils there is a general subsurface peak of



Depth function plots of percentage pyrophosphate extractable iron for the 40 soil sites

Figure 3.6

organically-bound iron, which also occurs in some of the group IV soils. Some of the group IV soils and all of the group V soils exhibit maxima of pyrophosphate-extractable iron in the surface organic-mineral horizon, with a marked decrease in pyrophosphate-extractable iron below this horizon.

This trend of surface maxima of pyrophosphate-extractable iron has been found by Mellor (1985) for youthful soils up to 230 years in age which are developed on Neoglacial morianes in Norway, and by Ugolini (1968) for very young soils in the Glacier Bay area, Alaska. The latter soils range in age up to 250 years. Maxima of iron in the A horizons has also been reported by Harvey et al. (1984) for a very young sequence of soil profiles in the Howgill Fells. The oldest profiles here are 140 years. It is suggested that in all these very youthful soils the organically-related processes tend to be concentrated near the surface. The formation of organo-metallic complexes may be associated with organic matter decomposition and the release of chelates, especially in the surface organic-rich horizons. Pyrophosphate-extractable iron contents may therefore be expected to be highest in these surface horizons. Incipient B horizon development is observable in these immature soils in Glen Feshie. This is evident both in the form of colour changes relative to the C horizon and slighly higher values of pyrophosphate-extractable iron in the B horizon relative to the C horizons.

Environmental contrasts will influence the time required for the development of a visually and chemically identifiable podzolic profile (Ellis, 1980). These factors will include vegetation and parent material conditions, climatic factors and the amount of iron already present in the parent material. In northern

Norway a visually and chemically distinguishable podzolic profile may develop in 250-1000 years, whereas in west-central Norway such a profile can develop in about 55 years. Thus in both the Norwegian soils and the Glacier Bay soils incipient B horizon development has been noted in the very young soils below the surface horizon showing the maximum of pyrophosphate-extractable iron. However this incipient B horizon development has not been detected in about 140 years of soil development in the Howgill Fells.

The similarity in the trends of organically-bound iron for the very young soils described by Ugolini (1968), Mellor (1985) and Harvey et al. (1984) with the observable trends in the group V soils for Glen Feshie supports the Glen Feshie soil profile morphology data and would imply very young soils at the stage of incipient podzolisation, perhaps no older than about 250 years.

Analysis of variance was used to test the null hypothesis that there is no significant difference between the five groups of soils for the lower A:B and B:C pyrophosphate-extractable iron. Table 3.9 gives the analysis of variance table for the A:B ratio and Table 3.10 that for the B:C ratio.

TABLE 3.9

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	24.84	4	6.211	17.10
Within gps	10.17	35	0.3631	
Total	35.01	39	1.094	

TABLE 3.10

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	148.7	4	37.17	4.44
Within gps	243.0	35	8.379	
Total	391.7	39	11.87	

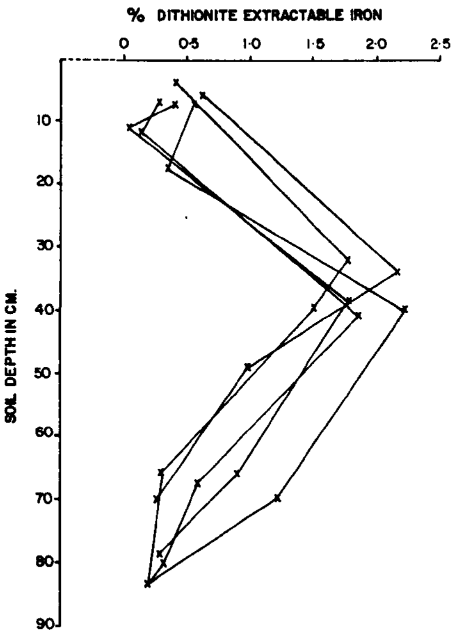
The 1% point for $F(4,35)$ is 3.925 so that the null hypothesis of no significant difference between the groups for both sets of analysis can be rejected. The five groups show significantly different ratios of pyrophosphate-extractable iron at greater than the 0.01 level.

For some of the group III, II and group I, and some of the groupIV soils the ratios of organically-bound iron are indicative of translocation and subsequent illuviation of pyrophosphate-extractable iron. Ratios of less than one for the lower A:B pyrophosphate-extractable iron are indicative of accumulation of organically-bound iron in the B horizons of the soils. Ratios greater than one indicate maxima of pyrophosphate-extractable iron in the surface horizons of the soils. Ratios of lower A:B Fep in the group IV soils range from 1.48 to 0.55 indicating that some of the soils have a surface maximum of organically-bound iron. Ratios for the group III soils range from 0.33 to 1.0, with only two profiles having a ratio of 1.0. Thus for the majority of the soils in this group translocation of organically-bound iron is probably taking place. The ratios for the group II soils average 0.70 and those for the group I soils, 0.76, indicating the translocation of organically-bound iron for all the profiles analysed.

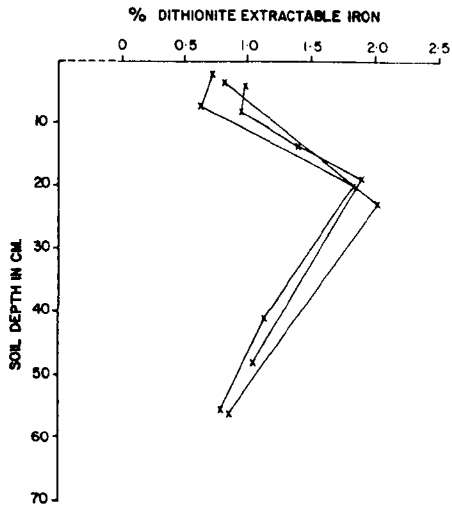
Absolute values for pyrophosphate-extractable iron maxima increase from the group V soils to the group III soils after which the peak values for the organically-bound iron do not increase, rather they decline. This may suggest that in some cases pyrophosphate-extractable iron has reached a steady state by the group III soils. Alternatively, this may also be because the pyrophosphate-extractable iron, in a gel state, crystallises with time and becomes extractable using the dithionite extractant. An hypothesis may be advanced to explain the decline in absoulte values for the pyrophosphate- extractable iron.

A possible explanation for the higher pyrophosphate-extractable iron maxima in the group III soils and the lowest pyrophosphate-extractable iron ratios may be that these soils began to form at a time in the Holocene when there was a change in palaeoenvironmental conditions to a cooler, wetter climate with the development of heath type vegetation. Such environmental conditions would be conducive to the relatively rapid formation of an H horizon whose decomposed humus would be a likely source of organic colloids which could combine with decomposed mineral matter to form organo-metallic complexes.

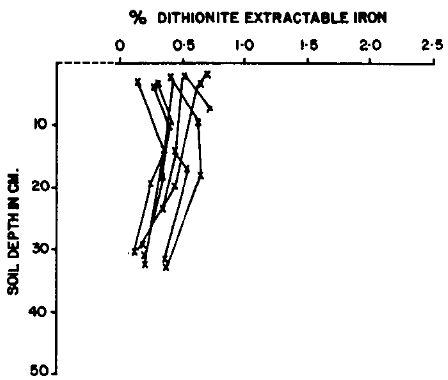
Dithionite-extractable iron is considered to provide a measure of the inorganic iron in the soil profile, in the form of both amorphous hyroxides of iron and microcrystalline iron. The depth functions for inorganic iron for the 40 soil sites are given in Figure 3.7. Total iron represents the sum of both forms of iron, the inorganic iron and the organically-bound iron. Comparison of the depth function plots for both pyrophosphate and dithionite-extractable iron (Figures 3.6 and 3.7) shows:-



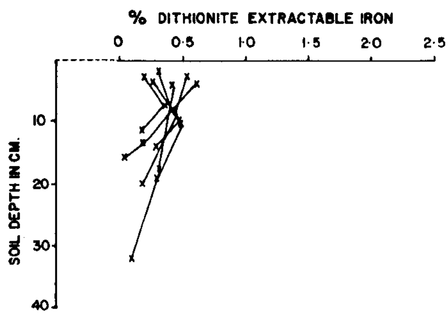
GROUP 1



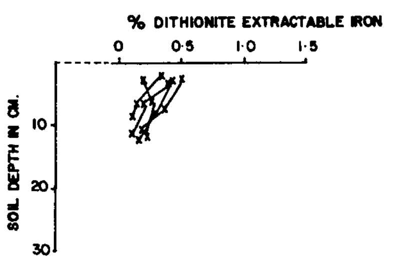
GROUP 2



GROUP 3



GROUP 4



GROUP 5

Depth function plots of percentage dithionite-extractable iron for the 40 soil sites

Figure 3.7

there is a progressive increase in absolute amounts of total iron in the profiles from the group V through to the group I soils;

For most of the soils, apart from the group V soils, the depth function plots of dithionite-extractable iron show an increase in inorganic iron with depth below the surface organic-rich horizon. With the group V soils and a few of the group IV soils iron maxima occur in the surface organic-mineral horizon. This pattern mirrors the trends evident in the pyrophosphate-extractable iron depth functions.

Analysis of variance was carried out for the lower A:B ratios and B:C ratios of dithionite-extractable iron to test the null hypothesis that there is no significant difference between the groups for the iron ratios. The results are presented in Tables 3.11 and 3.12.

Table 3.11

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	6.628	4	1.707	9.282
Within gps	5.762	35	0.1849	
Total	12.91	39	0.369	

Table 3.12

Source	Sum of sqs	DF	Mean SQ	F Ratio
Between gps	111.5	4	27.88	48.81
Within gps	16.65	35	0.5711	
Total	128.1	39	3.88	

The 1% point for F(4,35) is 3.925 so that the null hypothesis of no significant difference between the groups for both analyses can be rejected. The five groups show significantly different ratios of dithionite-extractable iron at greater than the 0.01 level. The iron variables used in this analysis are thus very good discriminators of soil profile development for the Glen Feshie terrace soils.

The depth function plots for the dithionite-extractable iron for the 40 soil sites show a clear progression in the absolute values for the peak concentration of the dithionite-extractable iron from the lowest values for the group V soils to the highest values for the group I soils. This represents an increase in maximum iron values from the incipient podzols of group V to the complex iron humus podzols of the group I soils. This trend is accompanied by a movement of the zone of maximum iron accumulation into the B horizon of the soils. This trend is also accompanied by a movement of the zone of maximum iron accumulation deeper into the B horizons of the soils, both trends also found for the podzols examined by Franzmeier and Whiteside (1963) and Levine and Ciolkosz (1983) in North America. This trend of accumulation of the iron lower into the profile may be demonstrated by examining the values for the ratios of lower A:B horizon dithionite-extractable iron. The mean, standard deviation and coefficient of variation for the A:B ratios of dithionite-extractable iron for the five groups of soils are given in Table 3.13.

Table 3.13

Group	Mean	St.Dev.	C.of Var.
I	0.18	0.06	34%
II	0.38	0.22	57%
III	0.77	0.25	32%
IV	0.98	0.17	17%
V	1.80	0.40	22%

This index measures the relative concentration of iron in the illuvial (B) horizon of each podzolic profile. An index greater than one indicates a maximum accumulation of iron near the surface of the soil whilst an index less than one indicates a depletion of iron from the upper soil layers and an increase in concentration of iron in the B horizon. The data presented in Table 3.13 suggest that translocation of iron in amorphous and microcrystalline form and its subsequent precipitation in the lower soil profile is increasing from the group V soils to the group I soils. This movement of iron within the soil profile is one of the major processes involved in podzolisation. Scanning electron microscope analysis of mineral grains sampled from the B horizons and lower A horizons of soil profiles from the Glen Feshie soils further demonstrates the operation of this process in the genesis of the five soil groups.

(c) Scanning electron microscopy

A combination of scanning electron microscopy (SEM) and energy-dispersive X-ray analysis (EDAX) was used in this study to examine soil genesis and progressive translocation of sesquioxides during pedogenesis for the five soil groups. Several previous studies have established the general technique of examining various soil components with the SEM in conjunction with a variety of back-scattered energy analysers using soil thin sections (for example, Bisdom et al., 1975; Bisdom et al., 1976). However, the studies that have examined the micromorphological features of soils with the combination of SEM and energy-dispersive X-ray analysis have not applied these techniques to analyse the features that are diagnostic of the dynamic nature of podzolisation, particularly the formation of the iron-enriched B horizon characteristic of podzolic soils. Energy-dispersive X-ray analysis is a particularly useful technique for this purpose. Using this technique chemical elements can be determined at sample magnifications of up to 10,000 times under the SEM (Bisdom et al., 1975, Bisdom et al., 1976). Selective spot energy-dispersive X-ray analyses of grain surfaces thus allows a rapid assessment of the elemental composition of mineral fragments adhering to single grains. Identification of chemical elements in this way together with the high resolution of the topographic image of the grains attainable with the SEM makes it possible to estimate simultaneously the composition, and extent, of any precipitation on a grain surface.

Previous SEM studies of pedogenesis at the scale of the external morphology of an individual grain are relatively few in number. Jauhiainen (1977) made environmental interpretations of quartz grain surface textures in Scandinavian podzols developed in contrasting parent materials. Profiles studied were developed in dunes from Finland, Norweigan terminal moraine deposits, and Danish outwash deposits. Many of the observations of differences between the grains in the samples were attributed to the differing nature of the soil parent materials, rather than to the processes involved in pedogenesis. A study by Douglas and Platt (1977) seems to be the only study which has related surface textural features of grains to the length of time available for pedogenesis. The soils used for their study were podzols developed on morainic deposits and glaciofluvial outwash deposits. The deposits ranged in age from Kansan (Anglian) to Wisconsin (Devensian). Samples were taken from the lower A horizons and the lower B horizons of the soils. Both solution and precipitation surfaces were observed. However, in this study differences in parent material were observed to influence changes in external morphology of individual grains. Furthermore, no attempt was made to assess the effects of translocatory processes on the external morphology of individual grains in the B horizons of the soils. The grains were cleaned prior to analysis in order to remove supra-grain precipitates.

Considering the application of aspects of scanning electron microscopy textural analysis to soil description and analysis, Whalley (1985) suggests that discrimination between deposits of varying ages using differences in surface texture should be possible. If careful evaluation is made of the soil chemistry, then it should be possible to demonstrate that the surface texture is the result of the processes operative in the present depositional environment.

In the present study the samples have all been taken from soil profiles which are developed within a homogeneous parent material. The soil profiles are visually and chemically podzolic although the intensity of podzolisation varies from the incipient podzols of the group V soils to the complex soil profiles of the group I soils. Any observed differences in the external morphology of the individual grains should not be caused by differences in parent material or soil-forming pocesses. In an attempt to ensure that the features examined were not inherited from a previous cycle samples were also analysed from several visual C horizons. The Glen Feshie profiles therefore appear to meet the qualifications given in Whalley (1985).

Laboratory procedure

Samples were taken from the lower A horizon and the upper B horizons of the five soils described in Table 3.5. Each profile is a representative member of the five statistically grouped soil profiles. In addition C horizon samples were analysed for the group I, group III and group V soils. A few grammes of each sample were rinsed in distilled water and the organic material floated off. Apart from removing the loose organic material the grains were not pre-treated for the removal of supra-grain precipitates. The mineral sand residue was then rinsed into petrie dishes and dried at room temperature. A monolayer of grains was taken by lightly pressing an Al stub covered with double-sided adhesive tape into the dried sample. The stubs were then coated with gold. This is the standard treatment for preventing electrical charge accumulating on the stub during scanning electron microscopy. Surface texture analysis was conducted using a Cambridge Stereoscan 600 SEM. With a sample stub mounted in the instrument, grains for analysis were randomly selected by use of the micrometer scales on the SEM stage with controls for movement in both the X and Y directions. Each selected grain was examined with the energy-dispersive X-ray analyser to ensure that it was of quartz composition. Examining only quartz grains was considered necessary to ensure that differences between grain surface morphology were not produced as a result of mineralogical differences. Thirty grains were examined and counted for each sample. The samples analysed were coded so that there was no _a priori_ knowledge of sample origins. They were decoded only after preparation of the final results diagram. This is necessary to enable confident unbiased interpretation of the results.

Grain analysis involved scanning each grain for a number of individual features. Eighteen categories of surface morphological features were noted for each grain. The surface features analysed have been adapted from those used by Goudie and Bull (1984). In addition to the features recommended by Goudie and Bull (1984) a further group was required for the analysis of the grains in this study. This was a category for amorphous and microcrystalline precipitations or coatings of iron and aluminium (categories 26-30 below). The features analysed comprised 41 categories of five groups. These are listed in Table 3.14.

<u>Table 3.14</u>

<u>Surface Feature Categories</u>

1 Large conchoidal fractures

2 Small conchoidal fractures

3 Large breakage blocks

4	Small breakage blocks	
5	Arc-shaped steps	
6	Random scratches and grooves	
7	Orientated scratches and grooves	Mechanically
8	Parallel steps	derived features
9	Non-orientated V pits	
10	Meandering ridges	
11	Dish-shaped concavities	
12	Upturned plates	
13	Large micro-blocks	
14	Small micro-blocks	
15	Very rounded	
16		
17		Roundness
18		
25	Very angular	
26	Low incidence	
27		
28		Coating
29		
30	High incidence	
31	Low degree	
32		
33		Solution
34		
35		
36	High degree	
37	Low	

38

39 Relief

40

41 High

The 5 categories of surface features used are designed to represent features produced by mechanical modification (mechanical features categories 1-14); chemical solution (solution features categories 31-36); precipitation (coating features categories 26-30); and derived external morphological attributes of the individual grains which included degree of roundness (categories 15-25) and surface relief (categories 37-41). In an attempt to assess the proportion of each grain characterised by each type of texture an estimate of the proportion of the visual grain surface covered by that feature was allotted to one of five abundance categories: < 10%; 10-35%; 35-60%; 60%-85%; 85%-100%. The analysis of the 30 grains provided values for the average incidence of each textural category for each of the thirteen samples. The results of the surface texture analysis are presented in Figure 3.8. The symbols in Figure 3.8 similarly represent five abundance levels for the average incidence of each of the 41 categories for each sample. This method of presentation follows that outlined by Goudie and Bull (1984).

Results

Considering the mechanical features, categories 1-14 the overall pattern which emerges is one of similarity for all the samples apart from samples H and J. The latter two samples are the B horizon samples from the group II and I soils respectively. The features identified on the grains for all samples suggest that

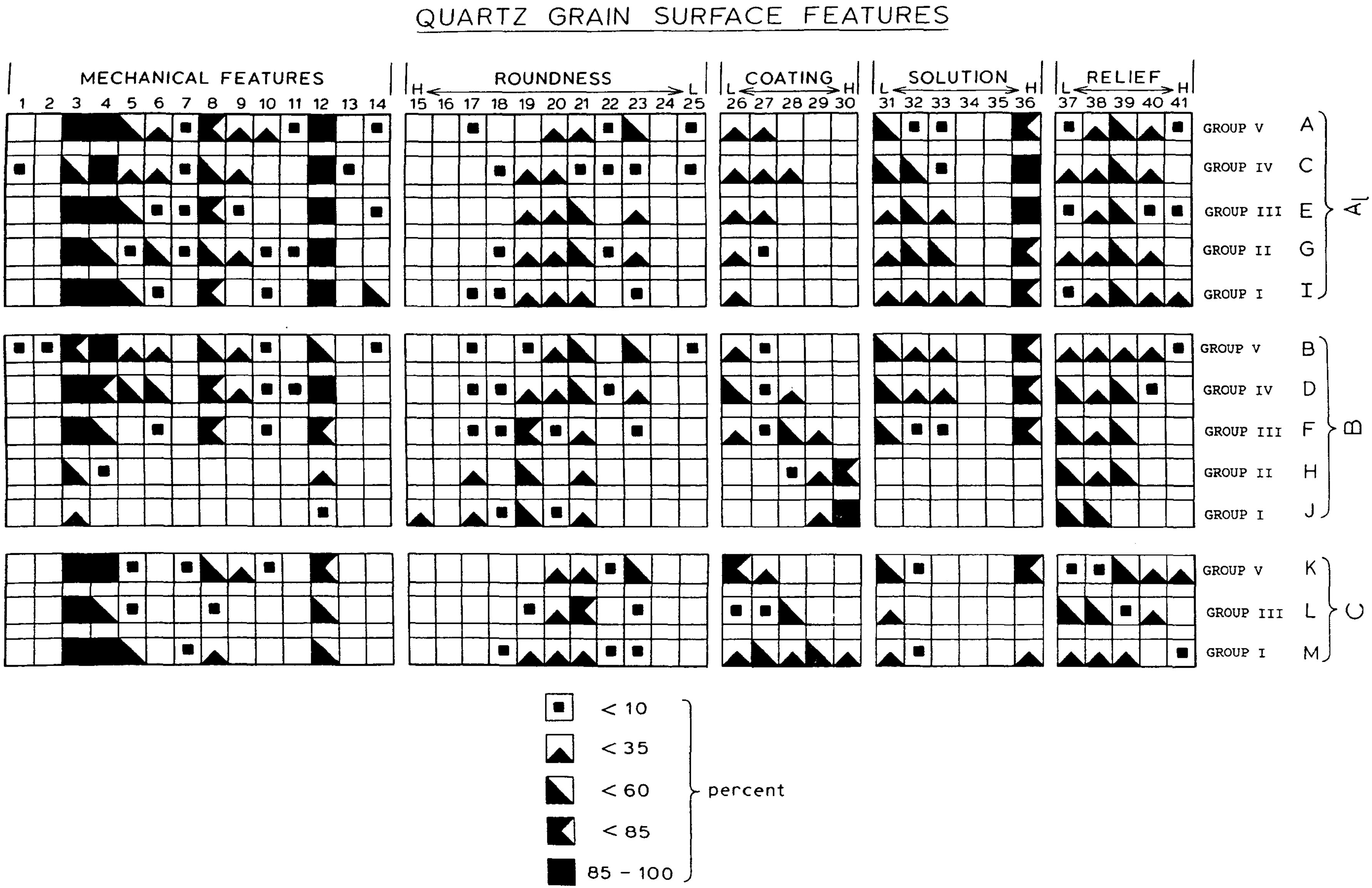

QUARTZ GRAIN SURFACE FEATURES
MECHANICAL FEATURES
ROUNDNESS
H L
COATING
L H
SOLUTION
L H
RELIEF
L H
1 2 3 4 5 6 7 8 9 10 11 12 13 14
15 16 17 18 19 20 21 22 23 24 25
26 27 28 29 30
31 32 33 34 35 36
37 38 39 40 41
GROUP V A
GROUP IV C
GROUP III E
GROUP II G
GROUP I I
A
GROUP V B
GROUP IV D
GROUP III F
GROUP II H
GROUP I J
B
GROUP V K
GROUP III L
GROUP I M
C
< 10
< 35
< 60
< 85
85 - 100
percent

Figure 3.8

they are unlikely to have been subjected significantly to glacial comminution processes and have probably not been transported long distances. This is indicated by the preservation of primary crystal forms on some of the grains. The suite of mechanical features most commonly observed on the grains from all samples included breakage blocks, scratches, step features indicative of lower grade breakages, and the presence, in some samples, of impact V pits, possibly caused by grain to grain collision in a relatively high energy subaqueous transport. These features, viewed together, may be indicative of a glaciofluvial depositional environment (Bull, 1978), thus suggesting the possibility of reworking of the valley fill deposits to provide the sediment source for the lower terraces.

The samples from the lower A horizons (samples A,C,E,G,I) do not show a marked between-sample difference for the roundness and relief categories. However, there is a gradual increase in the abundance of solution features from the group III to the group I samples. Such features include solution pits and hollows, crevassing and etching of the grain. Most of the lower A horizons have some grains which exhibit grain coatings. Sample C, belonging to the group IV soils exhibits the greatest number of grains with coatings. From the peak in the group IV samples there is a steady decline in abundance of coatings around the grains to the group I samples.

The trends shown in Figure 3.8 for the B horizon samples (samples B,D,F,H,and J) are considerably more marked than the rather gradual trends in coating and solution features shown by the lower A horizon samples. There are several obvious trends in the data which are related to one major process, that is the accumulation of grain coatings around the mineral grains in the B horizons of the soils. The group V sample (B) shows very little coating around the mineral grains whilst the B horizon grains for the group IV sample (D) similarly exhibit only relatively small amounts of coating. However, there is a marked increase in percentage of coating with the group III B horizon sample (F). This trend continues through the group II sample grains (H) to the group I samples (J) where there is almost a 100% incidence of grain coating. Roundness and relief similarly show marked trends in the B horizon samples. There is a decline in the relief of the grains from the group V to the group I samples with the grains also becoming considerably more rounded. This trend occurs as grain surface irregularities are effectively obfuscated by the accumulating coatings. The presence of coatings around the grains from the group II and group I B horizons is also likely to account for the absence of both mechanical and solution features for samples H and J. The obfuscation of such features as a result of the translocation of sesquioxides and the development of cutans around the mineral grains in the B horizons of podzols has been observed by Pye (1981) for podzolic soils developed in Queensland beach dune sands. Farmer et al. (1985) also note the presence of cutans around the mineral grains in a podzol developed in Late Devensian glaciofluvial deposits in north-east Scotland.

Analysis of the C horizon samples revealed little difference between the C horizon grains and most of the lower A horizon samples. There are two notable exceptions to this general trend:-

(1) There is a decrease in solution features in the C horizon samples relative to the lower A horizon samples. This suggests that, for samples K and L at least, the grains

have not been subject to the same processes that have produced the solution features in the lower A horizon grains;

(2) Some of the grains from sample M, the C horizon sample of the group I soils, have grain coatings. The origin of these coatings is discussed below and is suggested to be different from the coatings observed around the B horizon grains.

These observations from the C horizon samples suggest that the features observed on the grains in the lower A horizons and the B horizons are not inherited from the parent material grains. Rather they have developed since onset of the present cycle of pedogenesis.

Interpretation

The lower A horizon samples for the five soil groups show a gradual increase in degree of surface dissolution and etching, relative to the degree of etching exhibited by the C horizon samples. This was found to be the case particularly for the group III through to the group I soils. Quartz grains undergo quite rapid dissolution under extreme basic conditions (>pH 10), but are usually stable over the range of commonly occurring values in natural environments. However, dissolution does also occur at very low pHs, such as can occur in bleached horizons of podzols, but at much slower rates than under high pH. Dissolution and etching of grains in the pedogenic environment has been reported by several workers who have examined feldspar, amphibole and pyroxene grains (Dearman and Baynes, 1979; Berner et al. 1980; Farmer et al. 1984). Dissolution by organic acids is generally considered to be the prime genetic process. Farmer et al. (1985) noted that intensity of mineral weathering as indicated by degree of etching of feldspar grains was much more intense in the lower A, or Ea horizon of the podzol studied than in the Bs and C horizons. They concluded that intensity of mineral weathering is at a maximum in the upper horizons of the soil.

Cleary and Conolly (1972), conducted a scanning electron microscopy study of thin sections of soils of varying maturity sampled from a transect across the South Carolina coastal plain. They reported that the upper horizons of the red-yellow podzolic soils contained the most highly dissoluted quartz grains. They concluded that the organic acids produced in the rooting zone affect silica dissolution by complexing H_4SiO_4, thus resulting in greater rates of dissolution in the upper horizons than in the lower soil horizons.

As quartz is generally relatively resistant to weathering, chemical weathering of quartz grains will proceed at a slower rate than that for feldspars, pyroxenes and amphiboles. Nevertheless, considering the samples from the lower A horizons of the five Glen Feshie soil groups, a trend of increasing degree of grain surface etching is evident. Plates 3.9 a-e are scanning electron photomicrographs showing quartz grain surface etching on grains from the group V to the group I soils respectively. This trend may be due to dissolution of the grains by organic acids produced as podzolisation of the soils proceeds and as the lower A horizons become progressively more bleached from the group V to the group I soils.

Coatings around mineral grains in the B horizons of podzolic soils have been observed using scanning electron microscopy on

Photomicrograph of lower A horizon grains of the group V soil

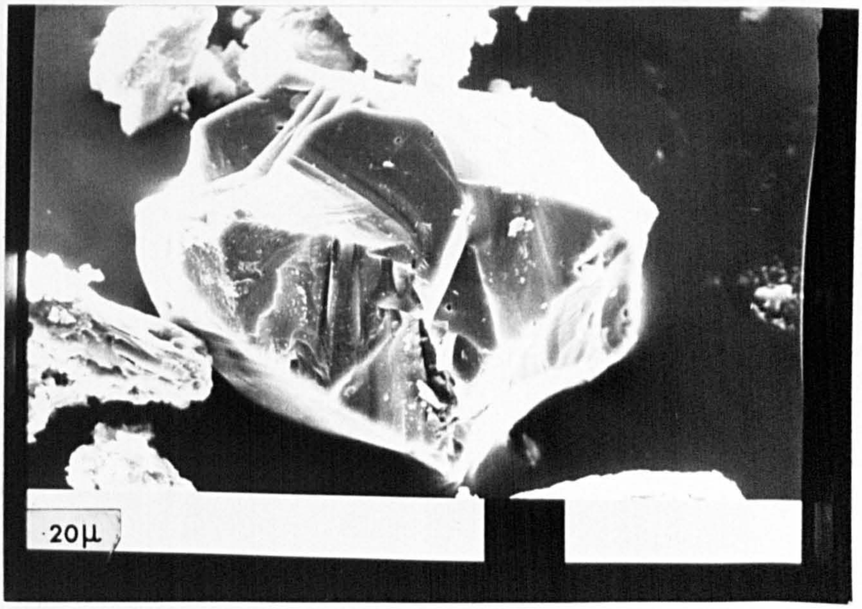


Plate 3.9a

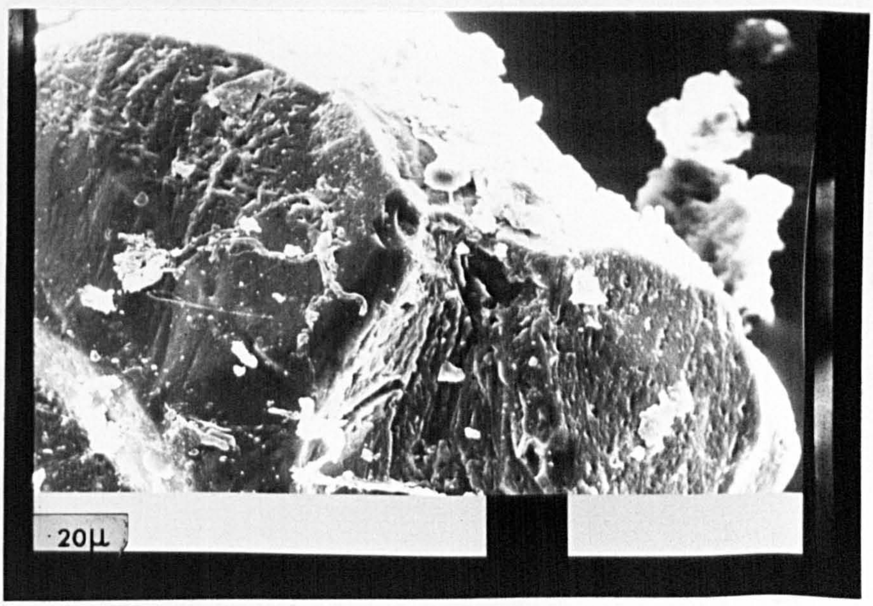


Plate 3.9b

Photomicrograph of lower A horizon grains of the group IV soil

Photomicrograph of lower A horizon grains of the group III soil

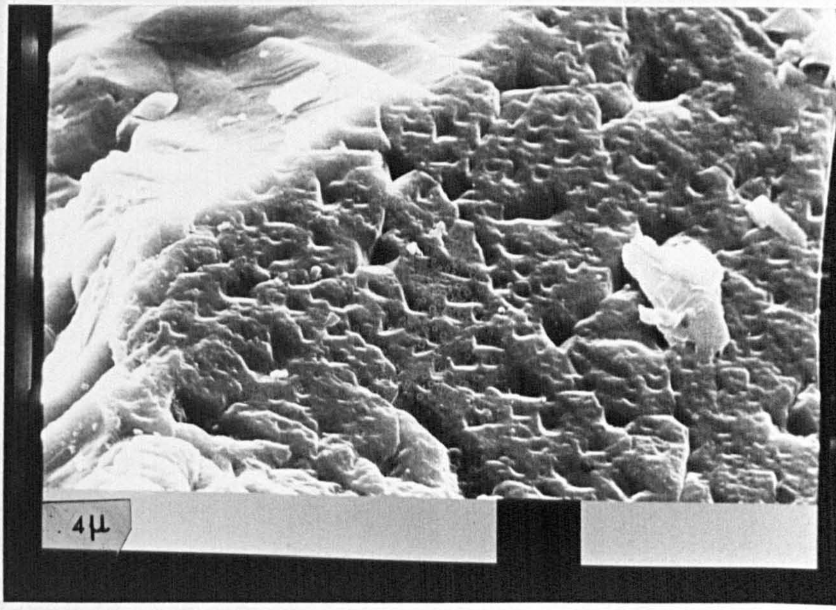


Plate 3.9c

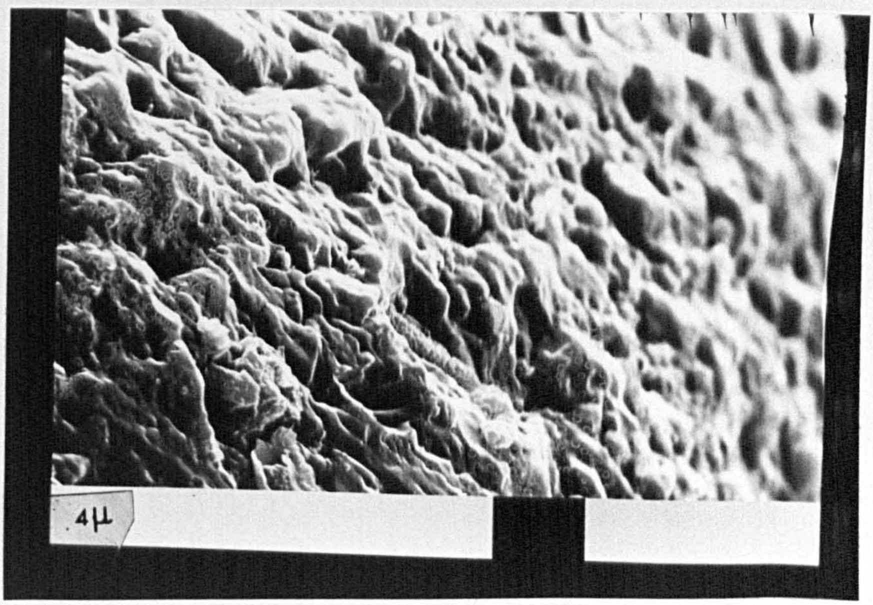


Plate 3.9d

Photomicrograph of lower A horizon grains of the group II soil

Plate 3.9e

Photomicrograph of lower A horizon grains of the group I soil.

either thin sections or individual mineral grains by a number of authors (De Coninck, 1980; McKeague and Wang, 1979; McKeague and Protz, 1980; McKeague, 1981; Pye, 1981; Farmer et al., 1985). However, these observations have all been made on well-developed, mature podzols. No previous study has examined the development of coatings around mineral grains in podzols which show differing intensities of podzolisation, both visually and in terms of bulk chemical analyses.

Plates 3.10 a-e show the progressive development of coatings around individual quartz grains from the group V to the group I soils. Reference to Figure 3.8 and the photo-micrographs shows that from the group V to the group I soils the individual quartz grains develop increasingly more extensive grain coatings. The group V soils (sample B) have only small numbers of grains which display very patchy, thin coatings. This group has a concentration of coated grains in the lower A horizon (sample A). Such a maximum is also displayed by the group IV soils (sample C) although there is some increase in the percentage of partially coated grains in the B horizon of the group IV sample (sample D) relative to the group V sample. This would accord with the bulk chemical analyses (Table 3.15 below) which shows that for the group V and group IV soils the maximum zone of accumulation of iron is near the surface.

Table 3.15

Site	Horizon	pH	% OC	%Fep	%Fed
Group V	Ah(A)	4.6	5.9	0.29	0.33
	BC(B)	5.0	0.89	0.10	0.21
	C (K)	5.2	0.04	0.04	0.17

Photomicrograph of coatings around
B horizons grains from the group V soil

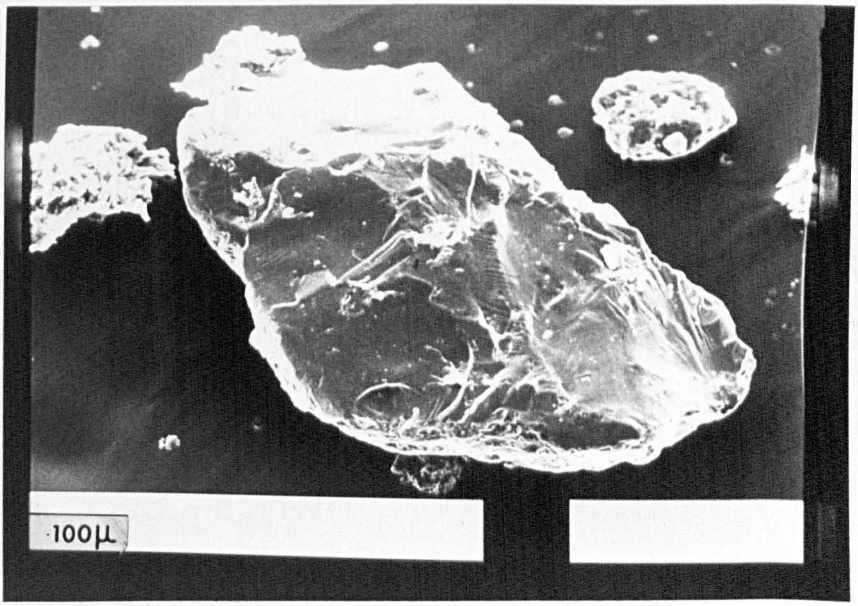


Plate 3.10a

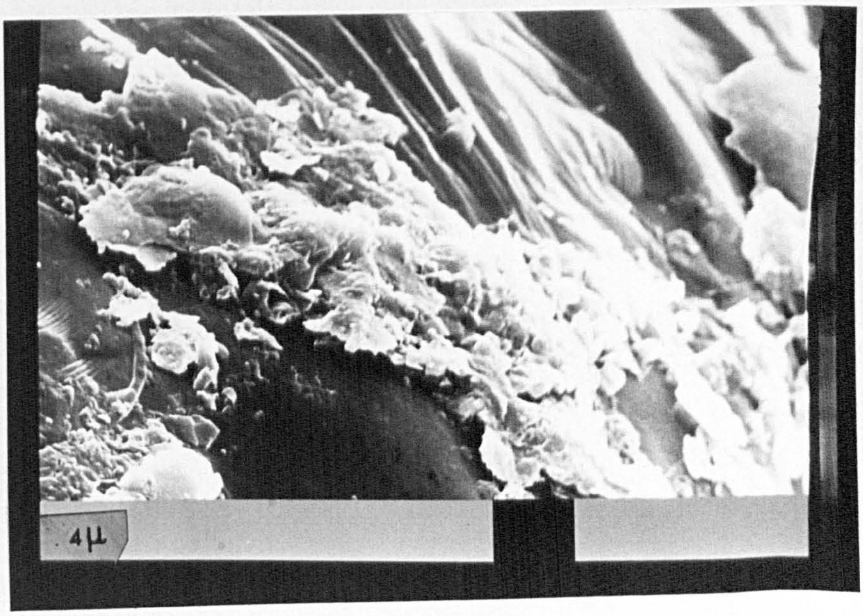


Plate 3.10b

Photomicrograph of coatings around
B horizons grains from the group IV soil

Photomicrograph of coatings around

B horizons grains from the group III soil

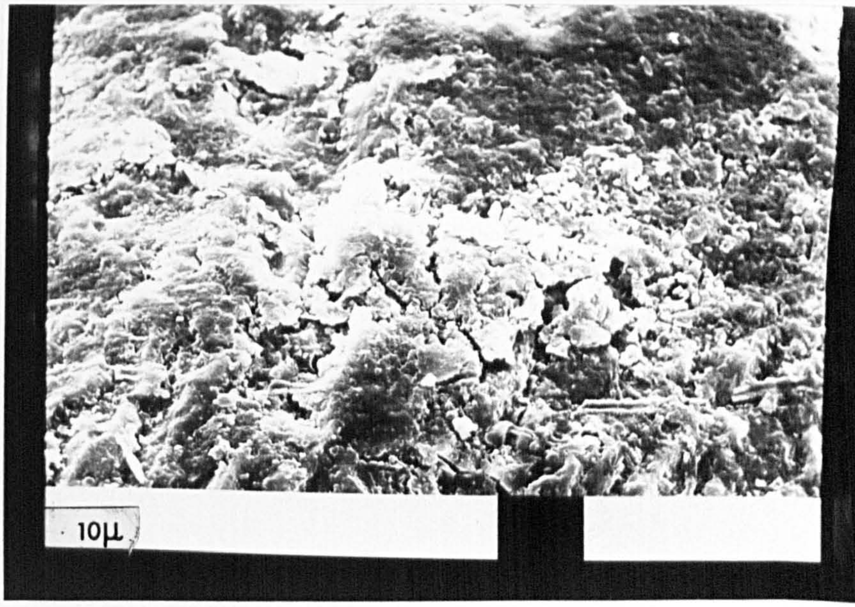


Plate 3.10c

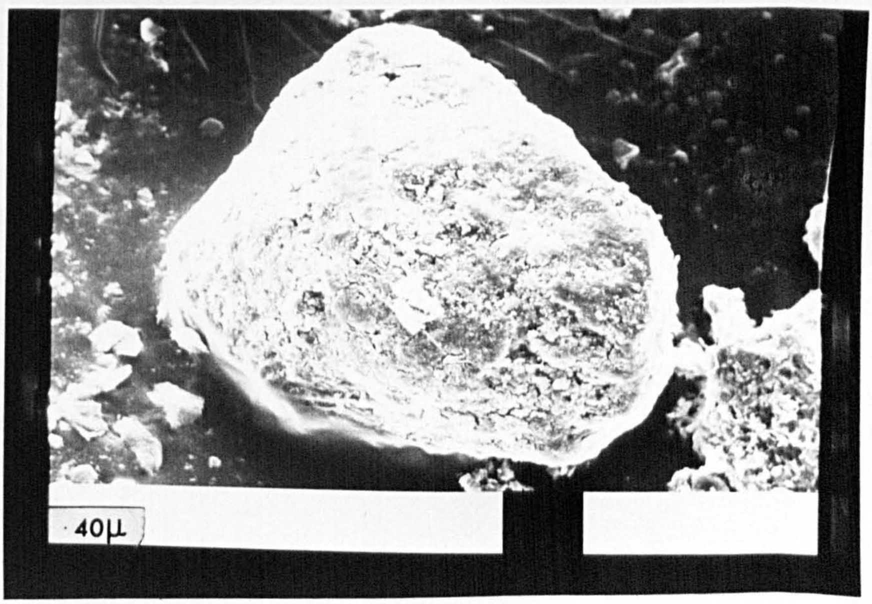


Plate 3.10d

Photomicrograph of coatings around

B horizons grains from the group II soil

Photomicrograph of coatings around
B horizons grains from the group I soil

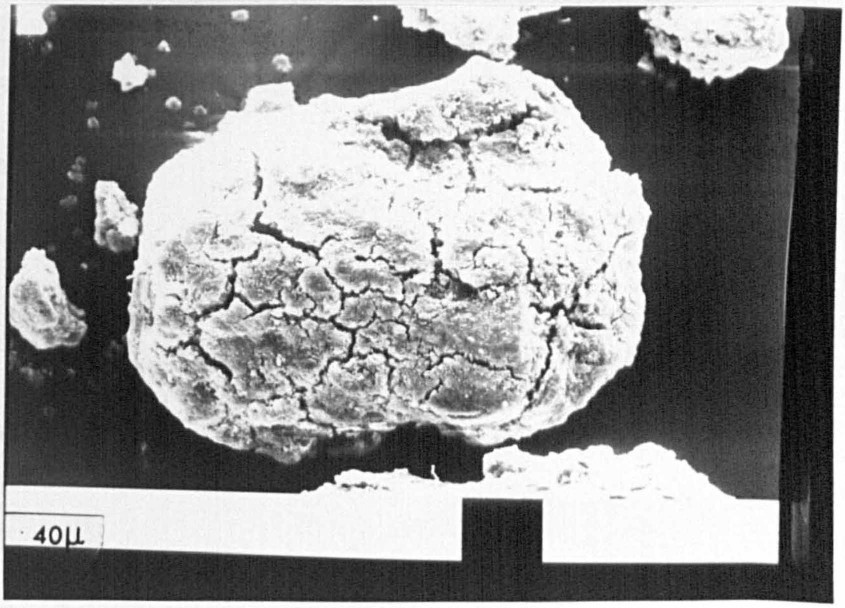


Plate 3.10e

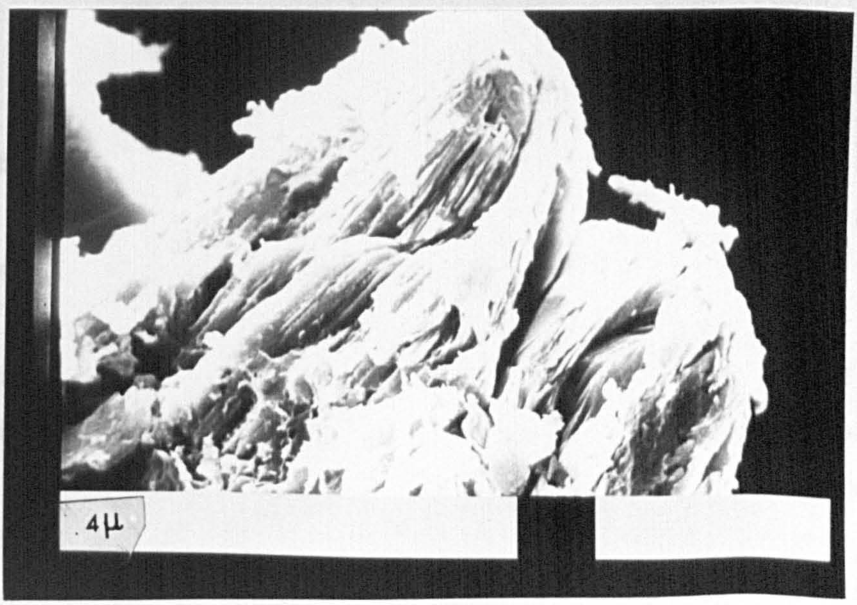


Plate 3.11

Photomicrograph of grain coating showing
fibrous appearance

Group IV	HA(C)	4.68	6.7	0.33	0.35
	B (D)	4.92	4.2	0.25	0.37
	C	5.18	0.05	0.09	0.23
Group III	H	4.25	10.7	-	-
	A (E)	4.31	2.1	0.44	0.17
	Bh(F)	4.95	3.3	1.07	0.40
	C (L)	5.29	0.06	0.24	0.22
Group II	H	4.38	18.4	-	-
	AEh(G)	4.00	2.1	0.14	0.21
	Bs (H)	4.62	5.5	0.03	1.31
	C	4.95	0.04	0.01	0.31
Group I	H	4.3	25.3	-	-
	Aeh(I)	4.41	4.16	0.25	0.37
	Bs (J)	5.2	9.7	0.34	1.91
	C (M)	4.7	0.02	0.01	0.31

Again with the group IV sample grains are only partially coated although with slightly more of the visible area of the grains being covered than was typical of the group V grains.

The group III soils show a marked rise in the presence of coatings around the mineral grains in the B horizons (sample F), both in numbers of grains coated and the extent of coverage of the grain surface.

The amount of the individual grain covered by coatings increases from the group III to the group I soils in the B horizons (from sample F to J). For the group II soils every grain examined was coated, and over 85% of the grains exhibiting an almost complete coating or cutan. For the group I soils most of the grains exhibited a complete coating. Again this accords with the bulk chemical analyses. There is a rise in the absolute amounts of iron extracted from the B horizons from the group III to the group I soils (Table 3.15).

The disposition of the cutans on the group IV and III grains suggest that coatings initially infill depressions on the grain surfaces (Plate 3.10 b) and are of a fibrous nature (Plate 3.11). From the group III sample to the group II sample the coatings progressively extend over the whole grain surface. Coincident with the increase in extent of the coatings is a gradual alteration in the physical appearance of the coatings. In the group II and especially the group I soils (Plates 3.12 a-b), the coatings display marked polygonal cracking. The group III coatings have an appearance intermediate between the group IV and the group II soils, with some observable cracking (Plate 3.10 c). The cracked coatings may develop in mature, older soils when the translocated sesquioxides are transformed from a gel to a solid state (De Coninck, 1980).

The C horizon sample for the group V soils shows evidence of some coating on the C horizon grains. Some translocation of material is thus occurring in what is therefore a slightly altered C or Cox horizon. Similarly the group III soils show some coating in the C horizons although to a lesser extent than was observed for the B horizon. This again suggests that the C horizon is slightly altered by pedogenic processes.

Energy dispersive spectrometry was used to examine the composition of the coatings around the grains for all samples. Selective spot energy-dispersive X-ray analyses of small areas

Photomicrograph of group II coating

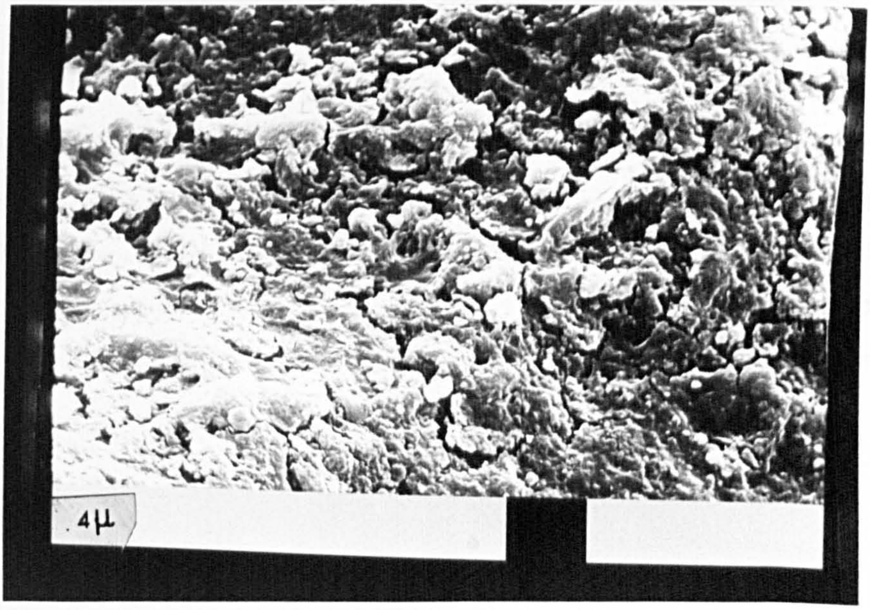


Plate 3.12a

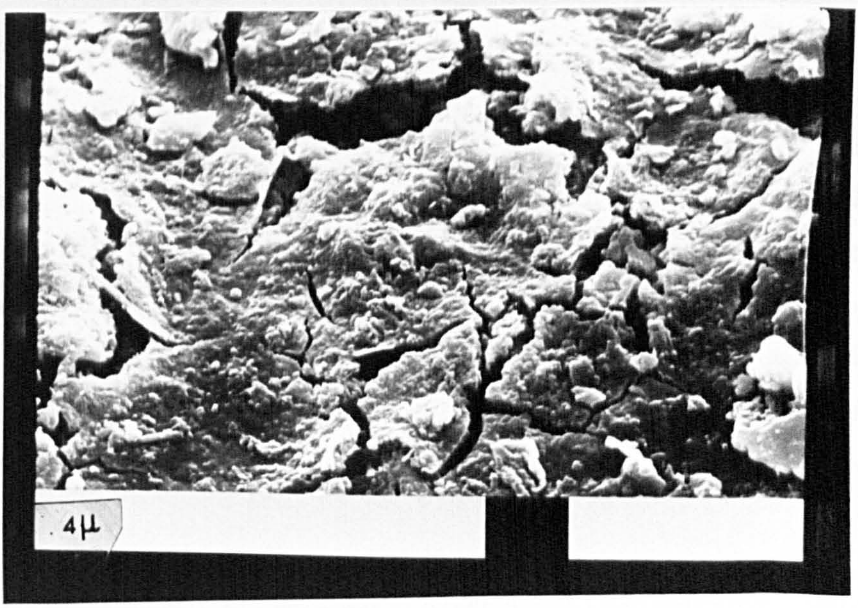


Plate 3.12b

Photomicrograph of group I coating

of coating on grain surfaces were made. Used in this way energy dispersive spectrometry may facilitate elucidation of soil genesis by permitting determination of the approximate chemical composition of the coatings identified from the scanning electron microscopy (McKeague, 1981). X-ray spectra from the coatings on the grain surfaces from the group V to the group I soils are shown in Figures 3.9 - 3.13. For comparison Figure 3.14 is an X-ray spectra from an uncoated surface of a mineral grain. The single peak in silica on the trace shows that the grain is quartz and has no adhering particles or coating.

The overall impression that emerges from these x-ray spectra is one of chemical similarity. For example, a spot analysis on a small area of coating for the group IV sample is shown in Figure 3.10. This exhibits a high peak in silica which is probably the underlying surface of the grain, the coating on the grain being very thin. The peaks in the Fe and Al are slightly higher than those for the group V soils (Figure 3.9). The spectra for the group IV soils exhibit peaks in K. The background counts for the 1-4 KeV region are slightly higher than those for the group V soils, and are higher relative to the 6-8 KeV region. The group III soils (Figure 3.11) exhibit a very high Fe peak and an Al peak, with minor peaks in S and K; there are also minor peaks in Mg and Ti. The background counts for the 1-4 KeV region are quite high. In contrast to Figure 3.11, Figure 3.15 is the X-ray spectra for the Fe coating shown in Plate 3.11 from the group III soils. This spectra has a very low background count in the 1-4 KeV region, a small peak in silica and Al and a dominating peak in the Fe. The group I soils show a dominating Fe peak in the spectra. There is a smaller Al peak and a peak in silica which is smaller than that for Fe. As with the spectra from Figure 3.15 there are low

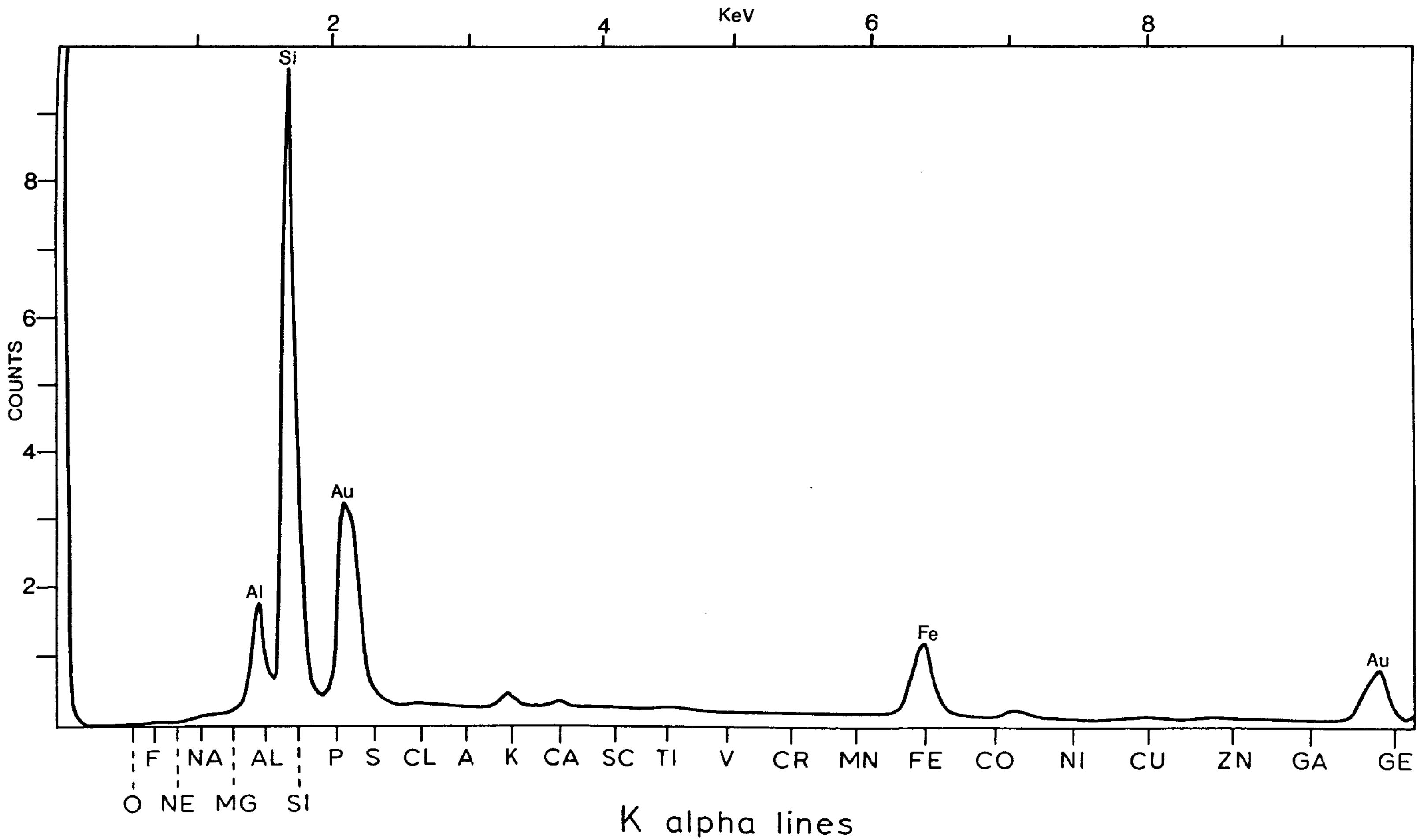


Figure 3.9 EDS Spectra of Group V soil grain

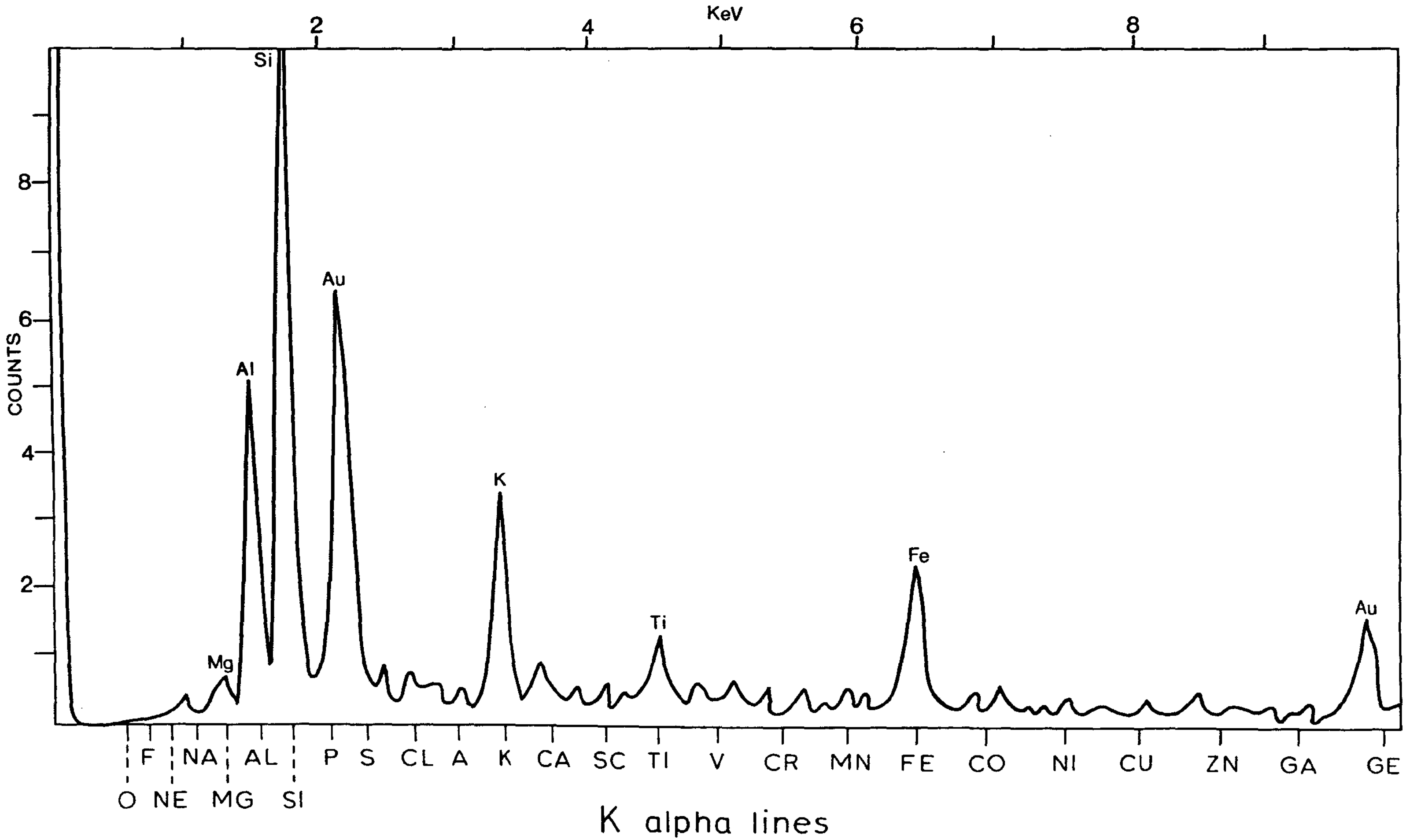


Figure 3.10 EDS Spectra of Group IV soil grain

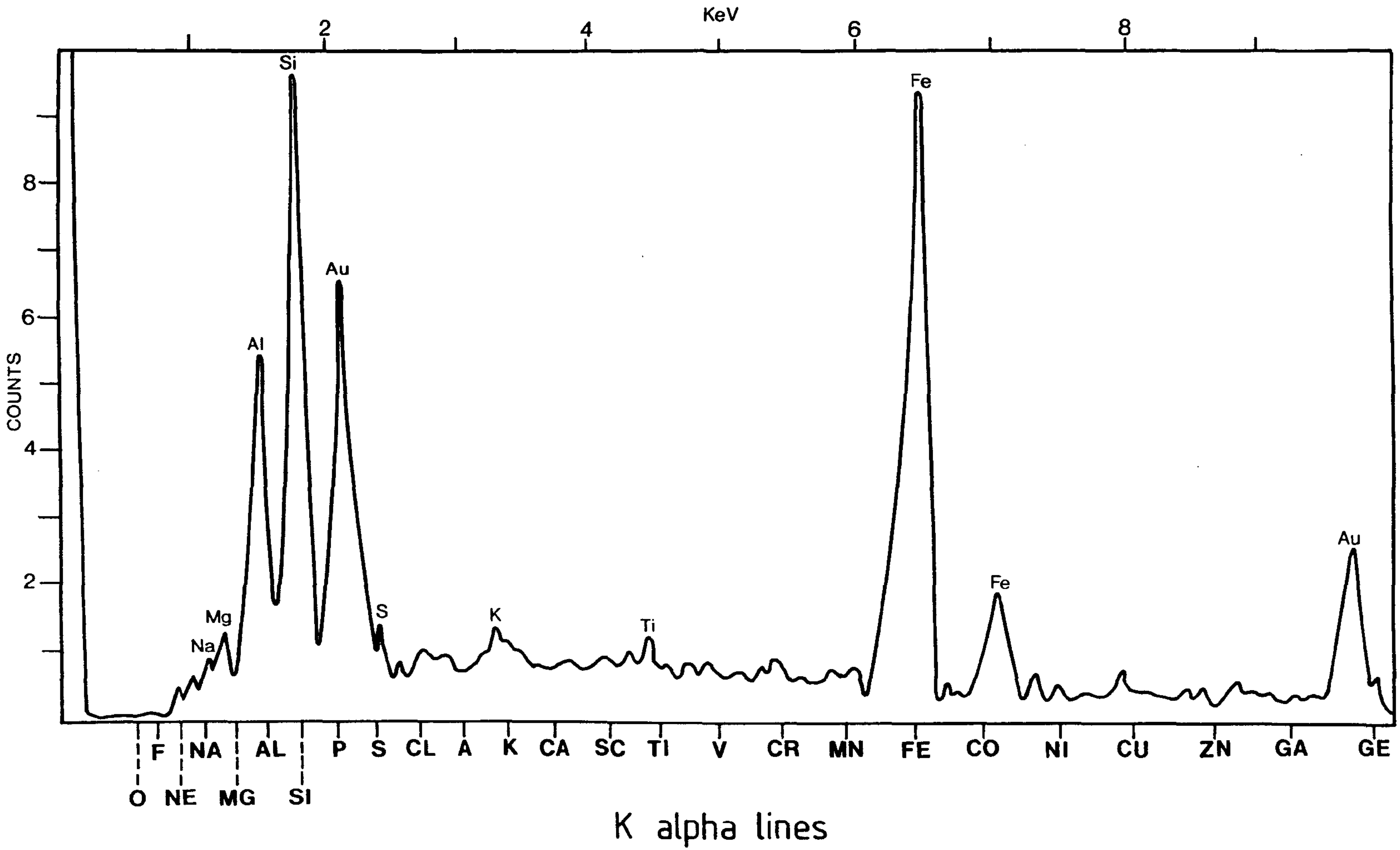


Figure 3.11 EDS Spectra of Group III soil grain

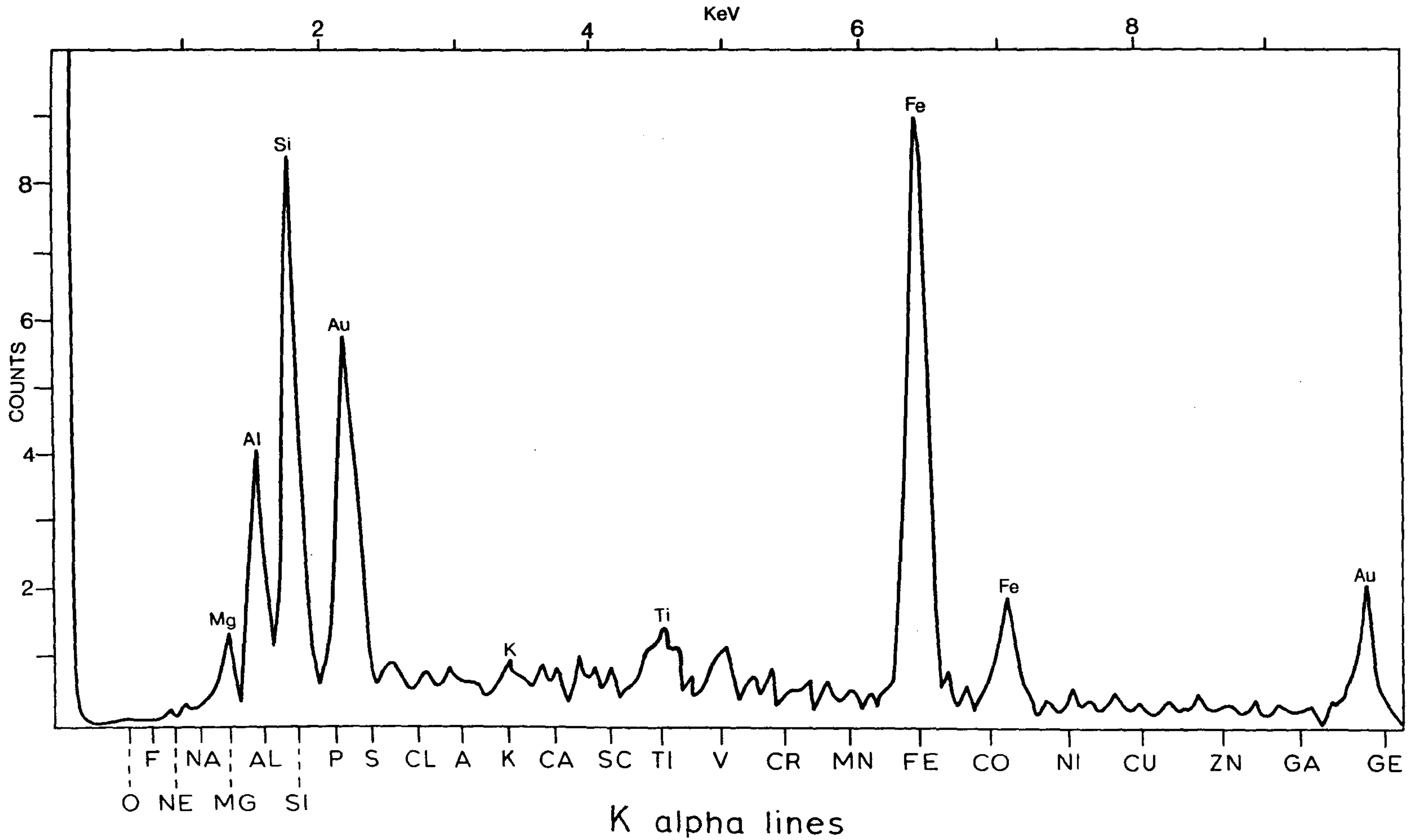


Figure 3.12 EDS Spectra of Group II soil grain

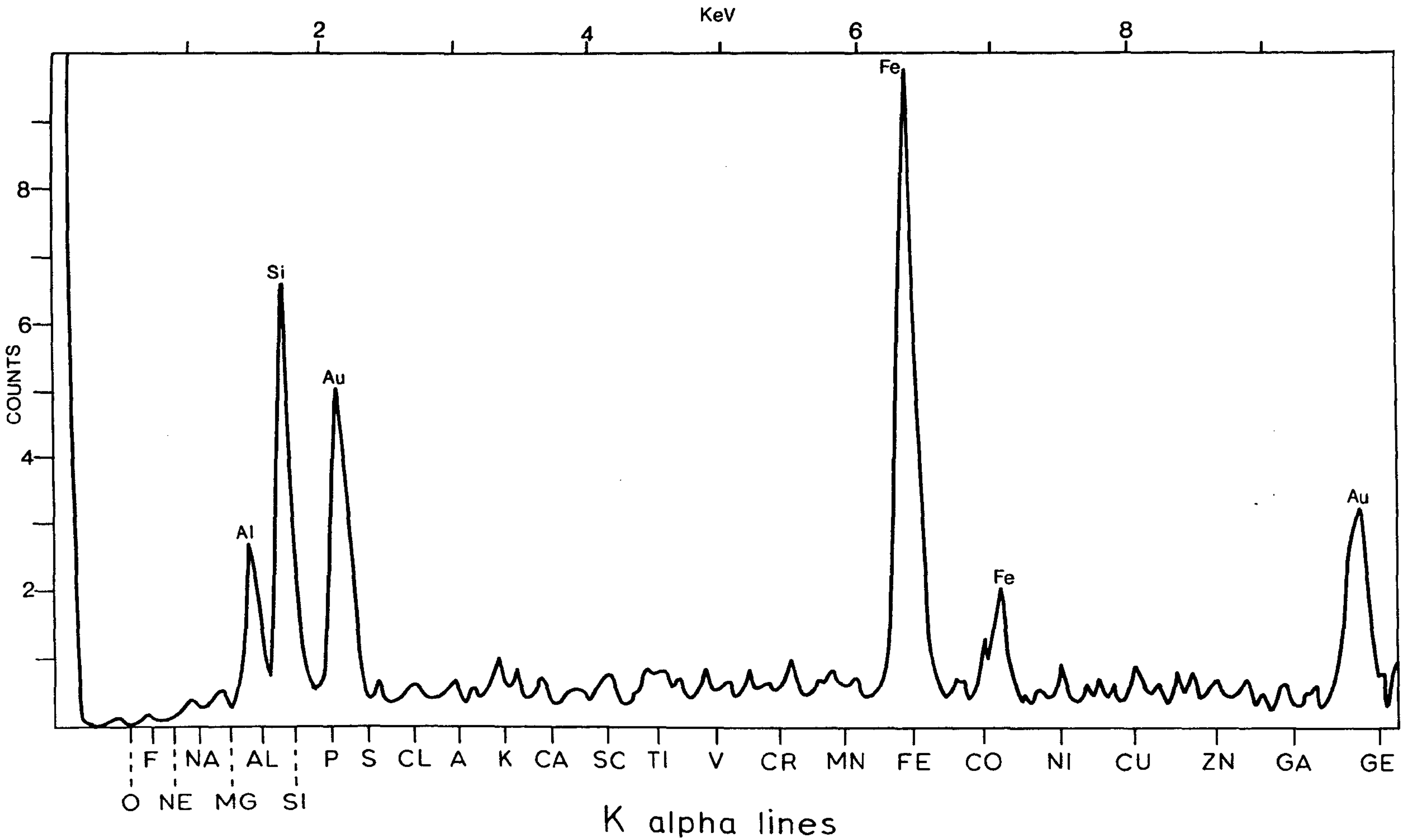


Figure 3.13 EDS Spectra of Group I soil grain

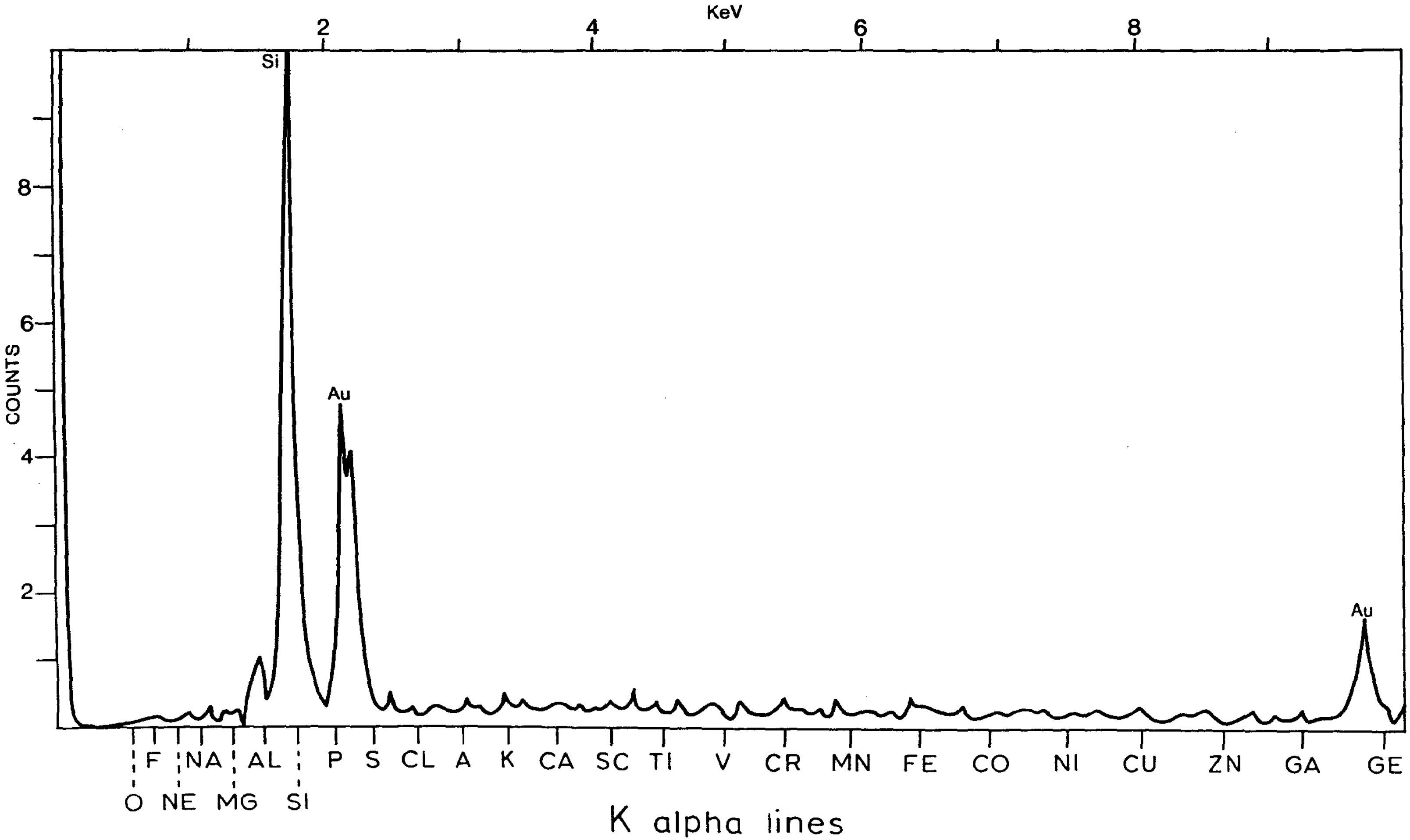


Figure 3.14 EDS Spectra of uncoated Quartz grain

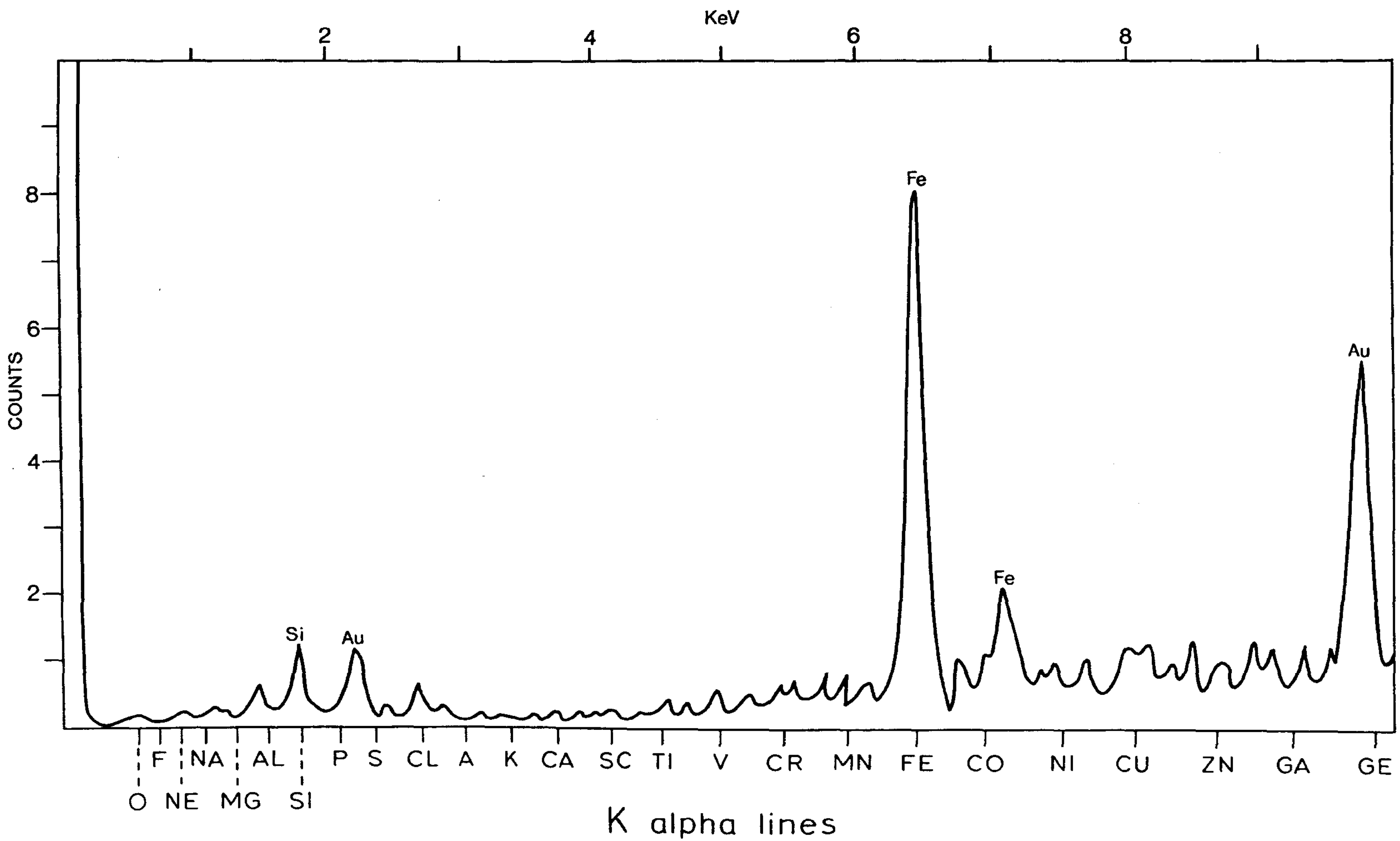


Figure 3.15 EDS Spectra of Group III soil coating fragment

background counts in the 1-4Kev region relative to the 6-8KeV region.

The evidence from the X-ray spectra suggests that the main constituents of the coatings on the grains are iron and aluminium. This finding is consistent with the bulk chemical analyses carried out on the Glen Feshie soils. Most spectra also show that fragments of silt and clay-sized fragments are incorporated into the coatings around the grain. Evidence for this comes from the X-ray link plots which show small peaks of titanium and magnesium. These are elements present in the micas and various feldspars of the Dalradian mica-schists which form the soil parent materials.

X-ray spectra from coatings around mineral grains in podzolic soils have been examined by Bisdom et al. (1976), McKeague and Wang (1980) and McKeague and Protz (1980). These studies show that X-ray spectra on plant remains always give high background counts in the 1-4 KeV region, relative to the 6-8 KeV region, and an S, K and usually a minor P peak. It is suggested that X-ray spectra from cutan material surrounding mineral grains which possess high background counts and peaks in K, P and S are indicative of coatings containing organic matter. For a podzol in the Netherlands Bisdom et al. (1976) show X-ray spectra which exhibit peaks in Fe, Al and K, as well as possessing relatively high background counts in the 1-4 KeV region. High peaks are also seen for silica which is probably the underlying surface of the grain, where the coating on the grain is thin. A number of X-ray spectra from horizons of podzols and podzolic B horizons in Canada are presented by McKeague and Wang (1980) and McKeague and Protz (1980). Silica peaks are much smaller than those from the spectra of Bisdom et al. (1976), but apart from this the spectra exhibit a marked degree of similarity with peaks in Fe, Al, and smaller peaks in S, P and K. High background counts are observed in the 1-4 KeV region. McKeague and Wang (1980) conclude that the cutans or coatings around the mineral grains are composed of silt and clay impregnated with Al and iron-organic material. For the podzolic B horizons McKeague and Protz (1980) suggest that mixed Al, Fe and silica hydrous oxides form the cementing material which coats and bridges the mineral grains. Subsequent work has shown that these podzolic horizons commonly contain imogolite and proto-imogolite, as suggested for the Scottish podzols by Farmer et al. (1985).

The X-ray spectra from the Glen Feshie soils are directly comparable with the spectra from the podzolic soils of Bisdom et al (1976), McKeague and Wang (1980) and McKeague and Protz (1980). The Fe peaks for the Glen Feshie soils are generally much higher than the peaks for Al. It would appear that Fe is thus the most important coating and cementing agent in the Glen Feshie podzolic horizons, with the peak in the Fe in the spectra increasing from the group V to the group I soils. This trend in iron content is also shown by bulk chemical analyses for the five soils examined (Table 3.15).

The relatively high background counts for the group IV and group III coatings may be indicative of iron-organic material coating the grains. However, Figure 3.15 for the group III sample also showed a high iron peak with very low background counts and was comparable with the spectra from the group I soils. This may indicate that these coatings are formed from inorganic translocaton of iron, that is allophane deposits incorporating clays and iron oxides as suggested for Scottish podzols (Farmer et al. 1985). These findings suggest that sesquioxide coatings around the grains, which comprise the cementing material for the podzolic B horizons for the Glen Feshie soils, is therefore likely to have been translocated from the upper soil horizons in both inorganic and organic form. This conclusion is consistent with the deductions based upon the bulk chemical analyses (Table 3.15).

The group I soils also possess coatings around the C horizon samples. Plates 3.13 a-b show the general appearance of the C horizon grains from the group I soils. From these micrographs it can be seen that there is a marked difference in the physical appearance of the coatings in the C horizon sample from those in the Bs horizon. The coatings are very thin and do not display the marked polygonal cracking of the mature Bs horizon. Small adhering particles are also evident and in places, disk-shaped secondary silica was observed to cover the grain surface (Plate 3.14). The latter have also been observed in the lower horizons of a podzolic soil in the Netherlands (Bisdom <u>et al</u>, 1975). Figure 3.16 is an X-ray spectra from these coatings on the Glen Feshie grains. A marked Al peak as well as an extremely high silica peak can be seen, thus suggesting that these coatings comprise mainly Al and silica. It would therefore apppear that the main cementing agents for this horizon are secondary silica precipitations and aluminium. This has also been suggested for a similar duric horizon in South Wales (Bull and Bridges, 1978)

In Section 3.8 (a) above, the existence of a Bx horizon in the group I soils was discussed. The boundary between this horizon and the C horizon is extremely difficult to detect in the field, and in fact characteristics of the Bx horizon are likely to persit to some depth in the profile. Fitzpatrick (1969) suggests that one of the characteristics of these indurated Bx

Photomicrograph of C horizon coating,

group I soils

Plate 3.13a

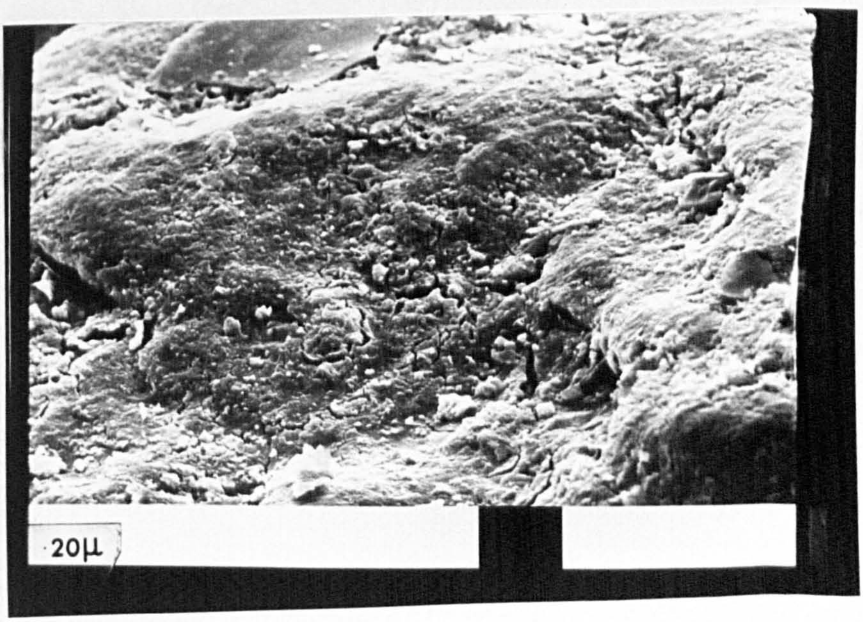


Plate 3.13b

Photomicrograph of C horizon coating,

group I soils

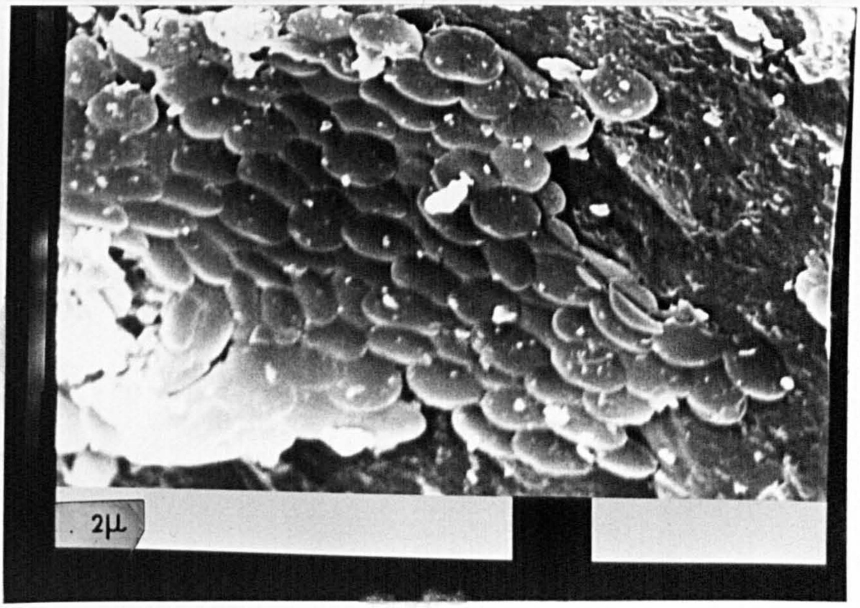


Plate 3.14

Photomicrograph of secondary silica
precipitation on the surface of
group I C horizon grains

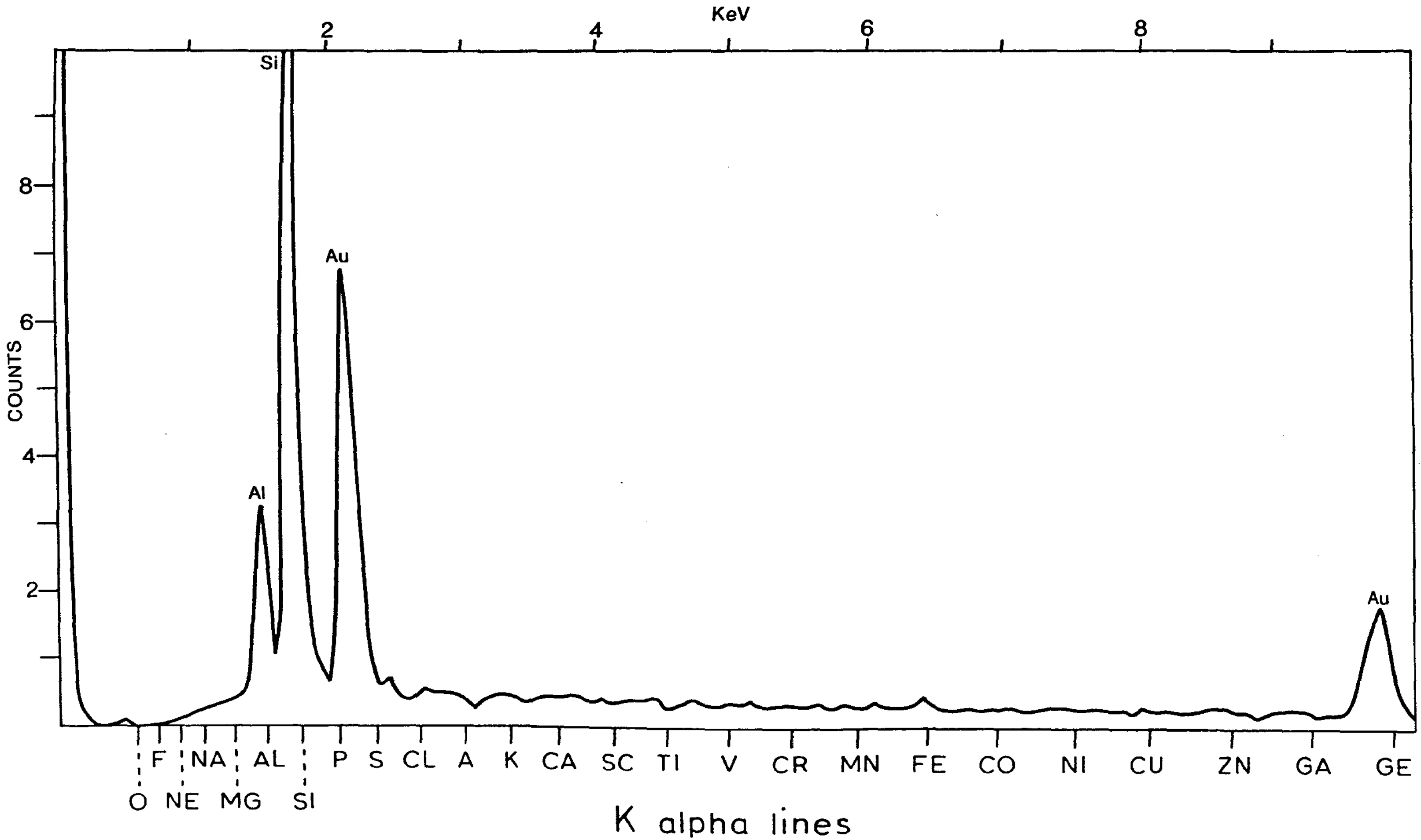


Figure 3.16 EDS Spectra of Group I soil grain

horizons is the presence of Al and silica as cementing agents. It is therefore likely that the C horizon sample from the group I soil is in fact the lower Bx or Cx horizon of the group I podzol.

The combination of scanning electron microscopy and energy-dispersive X-ray analysis has shown that the five Glen Feshie soil groups are genetically related but are all at varying stages of podzolisation. They are all subject to the processes of translocation of sesquioxides and probably also silt and clay particles which are incorporated into the coatings around the mineral grains. Translocation of sesquioxides and their subsequent precipitation in the lower horizons of the soils results in the formation of coatings or cutans around the mineral grains. In the group V and group IV soils these coatings cover only very small areas of the grains and are predominant in the lower A horizons of the soils. The coatings are more extensive in the group III soils with the maximum amount of coating occurring in the B horizon. Such maxima also occur in the B horizons of the group II and group I soils in which there are complete coatings around most of the grains. The group V and group IV soils show evidence of only incipient podzolisation which is consistent with the suggestion of young soils. The group III soils exhibit clear evidence of podzolic B horizons and the beginnings of intense weathering in the upper soil horizons which is associated with the development of bleached horizons in podzols. The group II and group I soils show well developed Bs horizons and evidence of intense weathering in the lower A horizons, as indicated by a rise in the degree of surface etching, on the quartz grains. The latter are clearly very well developed podzols which have been developing for a considerable length of time. The group I soils are even more mature than the group II soils.

(d) Dating control

Five soil groups have been objectively identified using a combination of Principal Components Analysis and cluster analysis from the data for 40 soil profiles developed on the terrace surfaces of the Glen Feshie terrace fragments. Analysis of variance was performed on a number of the physical and chemical soil properties. The clustered soils were shown to possess marked between cluster variation in both physical and chemical properties at greater than the 0.01 level of significance. Projection of the component one scores for each soil profile onto the first principal axis (Figure 3.2) showed the clusters of soils to be reasonably compact with analysis of variance showing the derived groups to be statistically significant at greater than the 0.01 rejection level.

If the Glen Feshie terraces were predominately outwash terraces related to decay of the Late Devensian ice sheet, or terraces related to high discharges consequent upon the deglaciation of the Zone III ice cap, as suggested by Young (1976), then terrace formation in Glen Feshie would have taken place over a time period of about 3,000 years, from 13,000 radiocarbon years BP to 10,000 radiocarbon years BP. Such an hypothesis would require stabilisation of the terrace surfaces shortly after the latter date of 10,000 radiocarbon years BP. Evidence for this would be revealed by the surface soil development on the surfaces of the terrace fragments. The soils would show a high degree of similarity and be at a similar stage in the podzolisation process.

The evidence presented in this chapter demonstrates at least five different stages of soil profile development for the Glen Feshie terrace soil sequence. The soil profiles from each of the statistically-derived groups have been demonstrated to form together a suite of genetically related soil profiles in various stages of podzolisation. It was argued above that the arrangement of the clusters along the first principal axis is chronological. The five soil groups which comprise the clusters form a relative age sequence from the incipient podzols of the group V soils to the well developed iron humus podzols of the group II and group I soils. As the soils progress from the group V soils to the group I soils the scores on the first component advance from the high negative scores of the incipient podzols to the high positive scores of the iron humus podzols.

The evidence from the soil profile development suggests that the terraces have developed episodically in response to five phases in the evolution of the Glen Feshie terrace sequence. The availability of some dating control allows the statistically-derived soil groups to be placed tentatively on an absolute time scale and some preliminary conclusions reached with regard to the age of the terrace surfaces.

The group I soils

All of the terrace fragments on which the group I soils are developed comprise the dissected deposits of the Glen Feshie valley palaeosandur (Chapter 2). Figure 2.2 shows that the pitted outwash in Glen Feshie terminates up-valley in kame and kettle deposits (dead ice topography). Kame and kettle deposits are ice-contact deposits which mark the terminal zones of valley glaciers from which the outwash rivers issued

(Sissons, 1967). The gradation of pitted outwash deposits into ice-contact features verifies the formation of the outwash as a proglacial sandur immediately downstream of decaying ice (Sissons, 1967, 1976; Kirby, 1969).

There is no evidence of a Loch Lomond Readvance ice limit in Glen Feshie. The limit for the Readvance in the central Grampians, the Gaick Plateau ice cap, has been placed in the upper valley of the Allt Lorgaidh (Sissons, 1974) where it is marked by abrupt termination of hummocky moraines (Plate 3.19). The firn line for the Gaick Plateau ice cap in the north-west of the Gaick Plateau is about 750-800m (Sissons, 1974) and for the Loch Lomond corrie glaciers in the Cairngorms is about 800-1000m (Sissons, 1979). It is therefore highly unlikely that a Loch Lomond valley glacier could have been sustained in Glen Feshie with a valley floor at an altitude of 380-300 metres. The palaeosandur and the kame and kettle or dead ice topography in Glen Feshie must therefore have formed during decay of the late Devensian ice sheet, as previously suggested by Young (1975).

Sissons and Walker (1974) suggest that total deglaciation of Scotland by 12,500 radiocarbon years BP is a conservative suggestion; deglaciation was certainly well underway by 13,000BP. These dates seem probable in view of the conclusion of Bishop and Coope (1977) that at about 13,000 radiocarbon years BP, temperatures rose very rapidly, with the mean July temperatures in south-west Scotland reaching about 15 degrees C. In addition, sea floor sediments indicate deglacial warming in the North Atlantic adjacent to the British Isles at about 13,500 radiocarbon years BP (Ruddiman and McIntyre, 1973). Radiocarbon dates from a number of Late glacial interstadial sites in Scotland suggest that many of the highland glens were deglaciated by, or shortly after, 13,000 radiocarbon years BP (Kirk and Godwin, 1963; Sissons and Walker, 1974; Pennington, 1975; Lowe and Walker, 1977; Vasari, 1977; Birks and Mathewes, 1978). In particular, radiocarbon dates of 13,125 radiocarbon years BP from Loch Etteridge 15km SW of Glen Feshie, and 12,700 radiocarbon years BP from the Abernethy Forest 20km NE of Glen Feshie strongly support this contention for the central Grampians. It therefore seems that considerable downwastage of the Late Devensian ice sheet in the central Grampians area must have taken place before 13, 000 radiocarbon years BP. Taking this evidence as a whole it is difficult to envisage survival of ice in Glen Feshie much after 13,000 radiocarbon years BP. Deposition of the gravels which constitute the group I terrace fragments therefore probably occurred about 13,000 radiocarbon years BP. This date then provides a maximum date for the onset of pedogenesis for the group I soils which subsequently experienced permafrost conditions.

Data from a pit dug into the surface of the Allt Garbhlach fan were inserted into the original soil data set and a new Principal Components Analysis carried out. The fan soil profile was typical of those for the outwash surfaces and possessed a well developed suite of permafrost features in the Bx horizon. Reclustering the scores on the first component clustered the Allt Garbhlach profile with the outwash surface soils, thus supporting Young's hypothesis of a deglacial age for the fan. The insertion of these data into the original data set did not change the composition of the soil groups.

The Group III soils

During the field examination of the soils in the Glen Feshie area a buried podzol was found in the terrace sequence of a tributary valley to the River Feshie, the Allt Lorgaidh (Figure 2.2). A ^{14}C date of 3620+/- 50 radiocarbon years BP (Har-4535) has been obtained on charcoal found in the organic-rich layer of a buried podzol (Robertson-Rintoul, in press). This podzol is buried beneath fluvial gravels in a low angle fan which forms the upper terrace surface in the tributary valley (Plate 3.15). The buried soil is traceable for some distance upstream in the Allt Lorgaidh and occurs in exposures on both banks of the tributary stream. The buried soil is overlain by about 10cm of sand and then 90-100cm of imbricated fluvial gravels (Figure 3.17). The alternation of sand and gravel and the depth of the overlying deposit varies between exposures. The fluvial gravels deposited over the buried podzol form the parent material for the modern surface soil. The buried podzol is developed in imbricated sub-angular fluvial gravels. The organic horizon of the buried soil which contains the charcoal is about 3cm thick with the charcoal occurring over the whole depth of the horizon. The particle sizes of the charcoal, with many pieces larger than 2cm, suggests deposition from local burning, rather than after reworking of an older deposit (Blong and Gillespie, 1978). Only one date was obtained on the charcoal so conclusions drawn must remain tentative. Clark's calibration curve (Clark, 1975) converts this date to a calender age of 4326 - 3715 calender years BP.

This date of 3,600BP gives an estimated age for the initiation of a phase of late Holocene sediment aggradation in the tributary valley. This phase of aggradation buried the podzol which marked the the former land surface of the upper terrace in the Allt Lorgaidh. The date gives a maximum age for the deposition of the gravels which comprise the parent material of

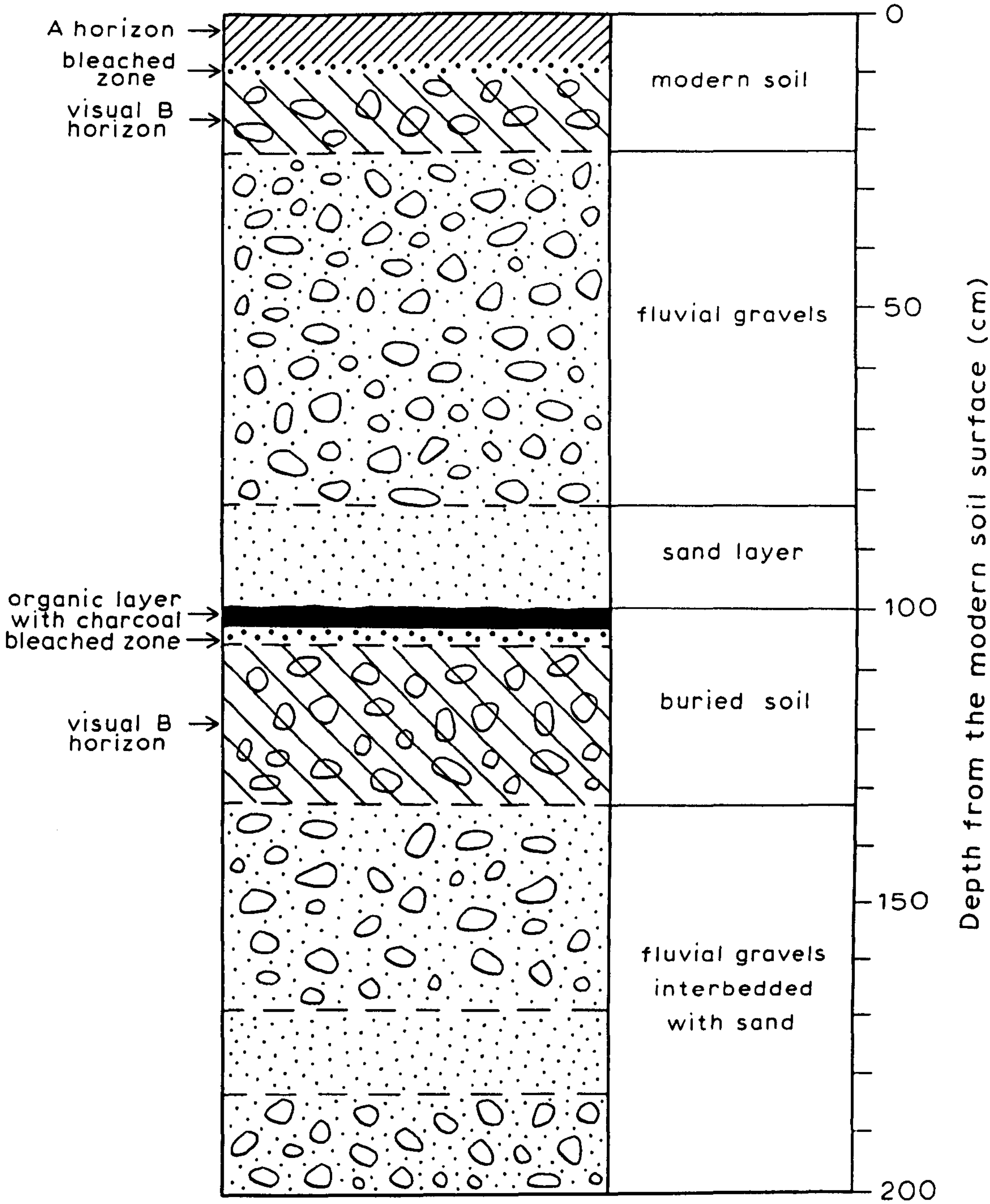


Figure 3.17 Stratigraphy of Buried Soil site, Allt Lorgaidh

the modern soil which has subsequently developed into the aggraded sediment (Plate 3.16). It therefore gives a maximum age for the onset of pedogenesis at this site. This presupposes that the sediment which buried the podzol was aggraded extremely rapidly. Several lines of evidence suggest that this is likely to have occurred. First, there is no evidence of buried A horizons which could suggest that the gravel and sand accumulated episodically and therefore over a long time period. Second, such rapid accumulation of a unit of sand and gravel in the River Feshie is possible if present day rates of aggradation can be extrapolated into the past. In the tributary valleys of the River Feshie over 50-75cm of sediment may be deposited over floodplain elements following a single flood (Plate 3.17). In the main River Feshie single floods may deposit up to 5cm of sand on the floodplain and 20cm gravel sheets over the floodplain elements. Further, Werritty and Ferguson (1980) report local incision and aggradation of active zone sediments of up to one metre in vertical extent during floods of about 100 $m^3 s^{-1}$ in the upper braided reach of the River Feshie.

Two soil profiles were analysed from the modern soil in the Allt Lorgaidh, one from the carbon date site (Plate 3.15 and site 37 in Figure 3.2) and one from an exposure upstream where the buried podzol was well exhibited (site 38 in Figure 3.2). The soil profile description for the modern surface soil, site 37, is given in Table 3.16.

<u>Table 3.16</u>

Profile name:	Allt Lorgaidh, ^{14}C site, modern soil
Slope:	0.5 degrees

The Allt Lorgaidh buried podzol

Plate 3.15

Plate 3.16

The modern soil above the Allt Lorgaidh buried soil site

Tributary overbank sedimentation in the Feshie catchment

Plate 3.17

Plate 3.18

1899 bar-channel complex, upper Feshie

Vegetation:	Vaccinio-Ericetum cinereae: lichen-rich heather moor
Soil drainage:	Free
Series:	Alluvial soil, Dryburn
Parent material:	River gravels
Rock type:	Mica schist and granite
Major soil subgroup:	brown podzolic soil

H : 0-10cm; very dark brown 10YR 2/2 no mineral content; semi-fibrous; moist; very weak fine granular structure; abundant fibrous roots; no stones; sharp smooth boundary.

AEh : 10-15cm; dark brown 10YR 3/3 fine sand; no mottles; very weak medium granular structure; moist friable; many fine fibrous roots; abundant medium sub-angular stones; gradual smooth boundary.

B sh : 15-30cm; dark brown 7.5 YR 4/4 loamy fine sand; no mottles; single grain structure; moist; friable; many fine fibrous roots; abundant medium sub-angular stones; gradual smooth boundary.

BC : 30-90cm; yellowish brown 10 YR 5/6 coarse loamy sand; no mottles; single grain structure; moist; very friable few fine fibrous roots; abundant medium sub-rounded stones; sharp smooth boundary.

The profile description for the modern soil above the ^{14}C site is thus typical of those of the group III soils. These two soil profiles from the tributary valley upper surface were included in the initial cluster analysis. Reference to Figure 3.2 shows that the two profiles have clustered with the group III Glen Feshie soils, thus suggesting that they are at a similar stage of podzolisation to the group III Glen Feshie soil profiles.

Pedogenesis must therefore have been initiated at approximately the same time for all the profiles. Clustering of the two modern soil profiles with the group III soils therefore enables an estimate to be made of the time of deposition of the gravels which form the parent materials of the group III soils in Glen Feshie. This close grouping of the tributary valley soils and the main valley group III soils suggests that the phase of gravel aggradation in the tributary occurred at a similar time period to the phase of braiding and gravel accumulation which was responsible for the building of the floodplain of the group III main valley terraces.

One date for the group III soils can only give an indication of the age of the deposition and stabilisation of the gravels which form the parent materials of the group III soils. It does not allow limits to be set on the temporal resolution of the diagnostic soil properties. Additional absolute dates from further sites would enable the age range of the terrace fragments to be established. However, the ^{14}C date obtained for this study does allow the group III soils, which have already been placed on a relative time scale using soil stratigraphic data, to be placed tentatively on an absolute time scale.

The Group V soils

Dating control for the group V soils comes from the 1869 and 1899 1st and 2nd edition 1:10 560 County Series Ordnance Survey maps, from 1946 1:10 000 and 1967 1:7 500 vertical air photographs. The first two editions can be directly compared and therefore used to plot the locus of most intense braiding in the River Feshie. The location and area of active gravel bars and vegetated bars may also be plotted (Werritty and Ferguson, 1980).

The channel systems, active gravel areas, and vegetated bar areas were mapped for this study from the 1st and 2nd edition 6" maps. Maps were also prepared from the 1946 1:10 000 vertical air photographs using a Bausch and Lomb zoom transfer scope. The 1:5 000 maps prepared from the 1967 1:7 500 vertical air photography were also used. In order to standardise the scale of all the maps, the original maps were traced onto clear film and the final planimetric maps produced using a Bausch and Lomb zoom transfer scope. The zoom transfer scope is a graphical data transfer instrument that will take output from conventional air photography, or from maps, compensate for various distortions of the aerial photograph, and superimpose the original map onto a scaled data base provided by a reference map (Theis, 1979). Controls allow the accurate matching of the photo scale, or map scale, to the data base scale and also provide optical corrections for tilt and relief of the photography. In this way the information from the photograph, or original map, may be plotted onto the new map scale. The 1971 1:10 000 National Grid edition Ordnance Survey Map was used as a base map. In this way the planimetric maps in Figures 3.18 a-c were produced.

Analysis of these data sources has revealed significant channel planform changes in three reaches of the River Feshie since the 1st edition map in 1869. Reach 1 (Figure 3.18 a) exhibits the most dramatic planform changes of the three reaches. Between the 1869 and 1899 maps there was a change in channel planform from a singlethread channel to a multithread channel. This change was accompanied by the development of large lobate unvegetated bar surfaces. The appearance of the bar surfaces

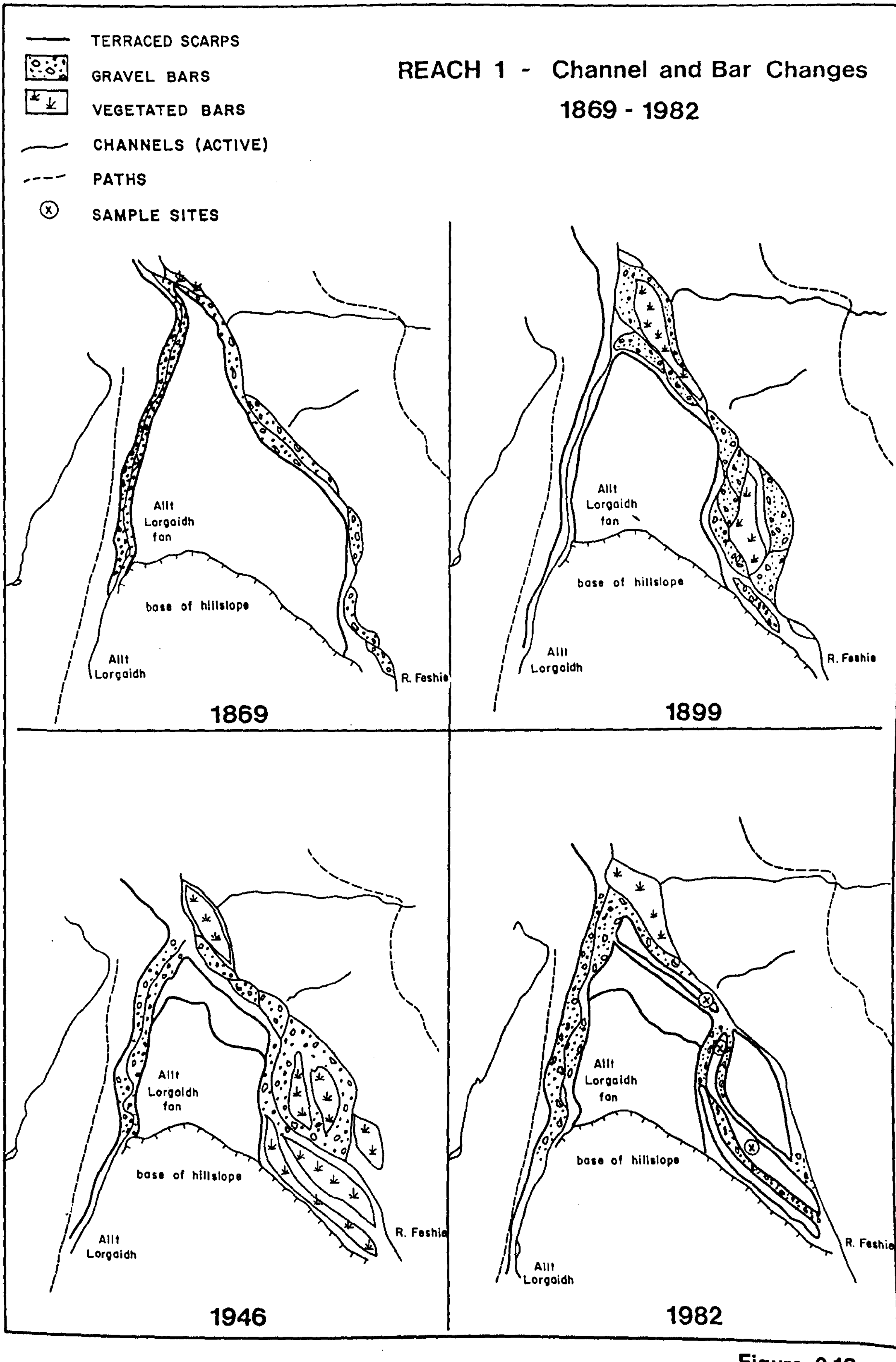
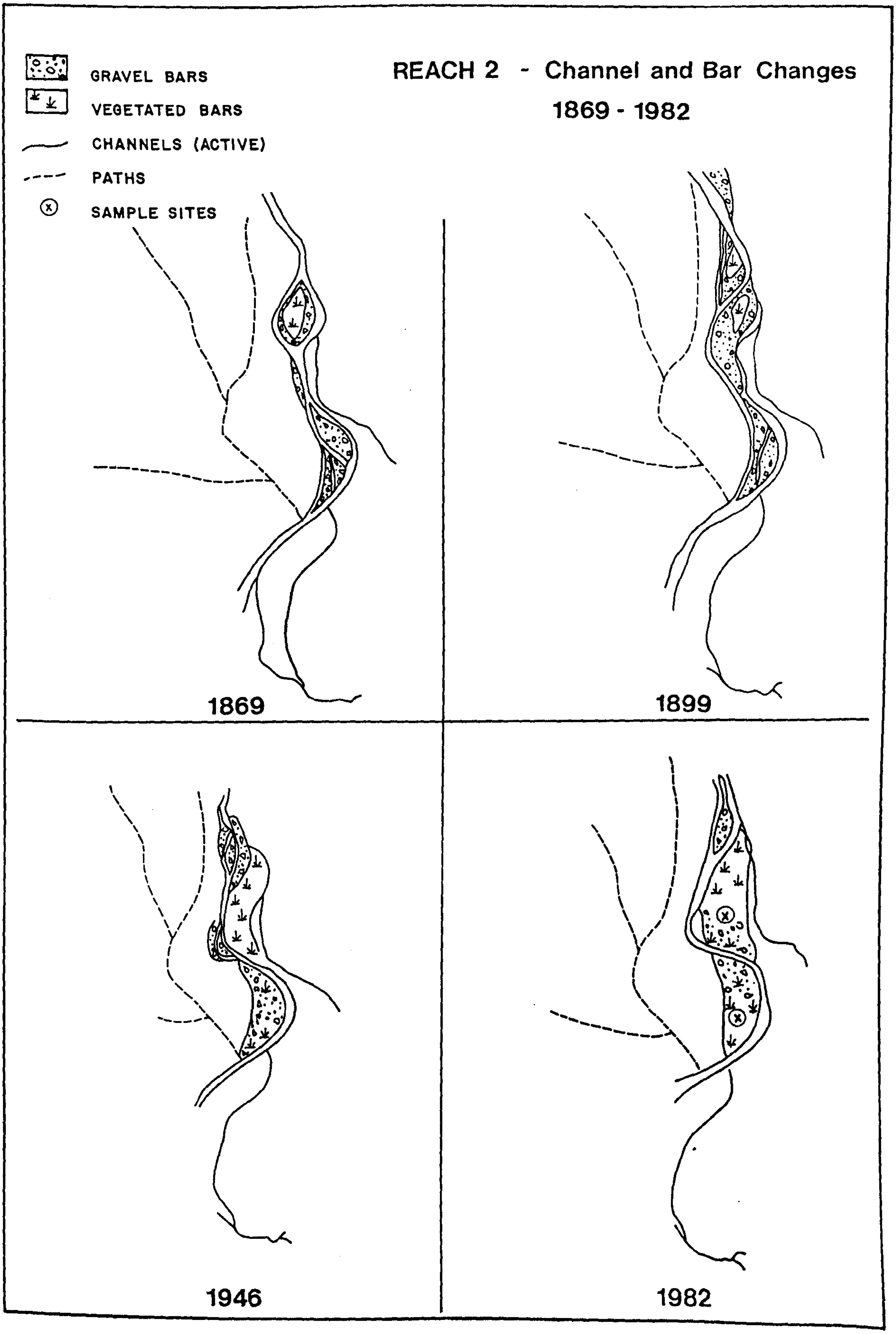

TERRACED SCARPS
GRAVEL BARS
VEGETATED BARS
CHANNELS (ACTIVE)
PATHS
SAMPLE SITES
REACH 1 - Channel and Bar Changes
1869 - 1982
Allt Lorgaidh fan
base of hillslope
Allt Lorgaidh
R. Feshie
1869
1899
1946
1982

Figure 3.18a



GRAVEL BARS
VEGETATED BARS
CHANNELS (ACTIVE)
PATHS
SAMPLE SITES
REACH 2 - Channel and Bar Changes
1869 - 1982
1869
1899
1946
1982

Figure 3.18b

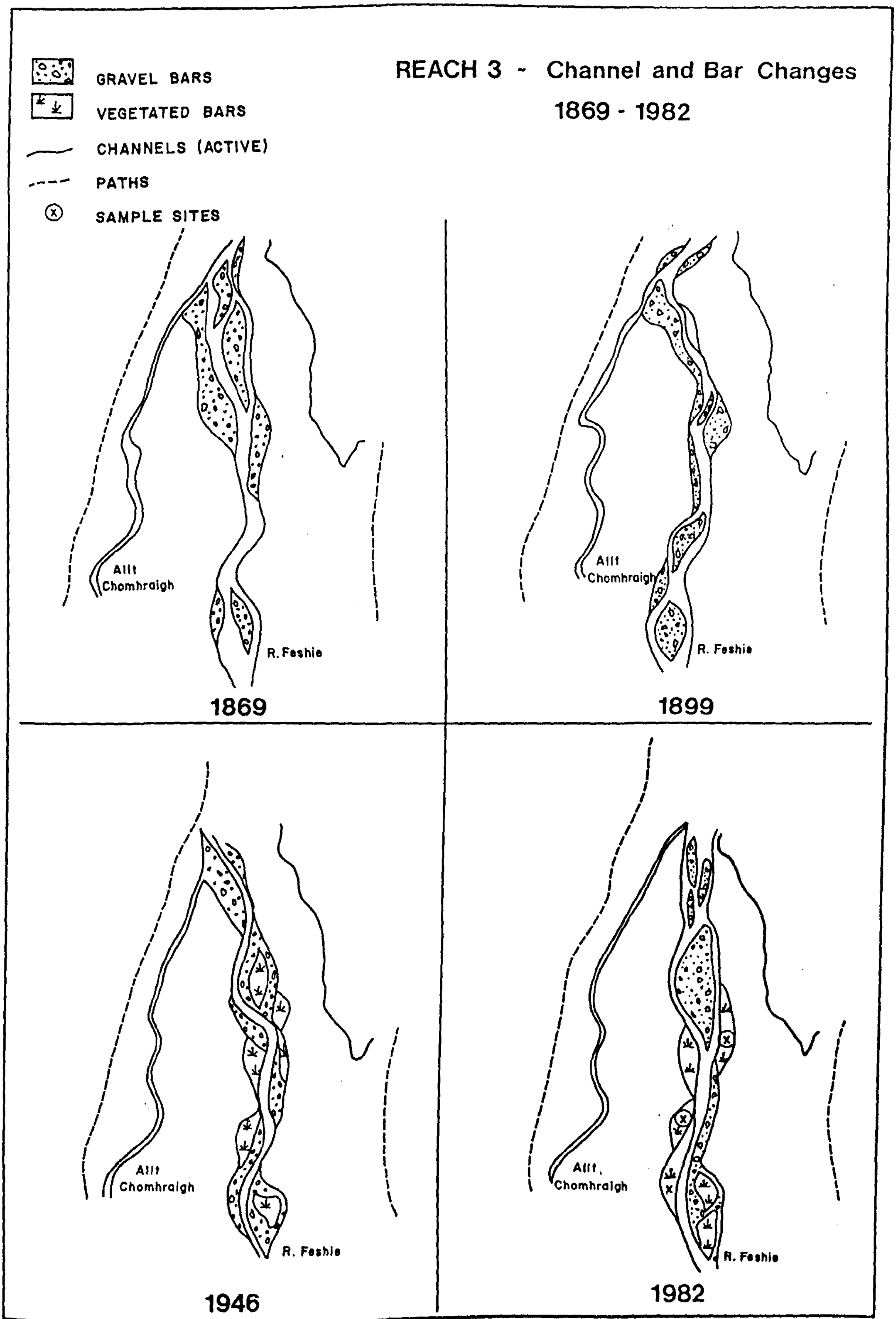

GRAVEL BARS
VEGETATED BARS
CHANNELS (ACTIVE)
PATHS
SAMPLE SITES
REACH 3 - Channel and Bar Changes
1869 - 1982
Allt Chomhraigh
R. Feshie
1869
Allt Chomhraigh
R. Feshie
1899
Allt Chomhraigh
R. Feshie
1946
Allt. Chomhraigh
R. Feshie
1982

Figure 3.18c

was concomitant with a reduction in the total area of the terraces and floodplain between the upper Feshie and the Allt Lorgaidh. Digitising the area between the two streams showed that the area was reduced by 36% of the original 1869 area. The change from a singlethread to a multithread channel appears to have been accompanied by bank erosion and reworking of the higher terraces. This reworking of the higher terraces is the most likely sediment source for the new gravel bars.

Between 1899 and 1946 there was a further enlargement of the active gravel area with a further reduction in the total area of the terraces, a further 15% of the original 1869 area. Between 1946 and 1982 the channels formed between 1869 and 1899 were gradually abandoned until the Feshie reverted to a singlethread channel and abandoned the newly created bar area. The present River Feshie is now incised about 1 metre below this new bar area. The gravel bars were fully vegetated by 1967. The large lobate vegetated bars produced as a result of the increase in braiding thus represent the most recently abandoned valley bottom surfaces in the reach (Plate 3.18) These surfaces were in existence as gravel bars by 1899, but post-date 1869. This suggests that deposition of the gravel bars occurred about 80 radiocarbon years BP, according to Clark's calibration curve. The date of 80BP for the deposition of the gravels gives a maximum date for onset of soil formation for these most recent surfaces, and is thus consistent with the maximum dates for the onset of soil formation for the group III and group I soils. The sample sites for the group V soils in this reach are marked in Figure 3.18 a.

The maps for reaches 2 and 3 (Figures 3.18 b-c) show a similar increase in the area of active gravel bars, although the planform changes for these two reaches are not as dramatic as those recorded for reach 1. Nevertheless, both reaches experienced an increase in the extent of the active gravel bar area between 1869 and 1899, followed by a gradual decrease and stabilisation of the gravel bars between 1899 and 1982. Again, as with the Reach 1 bars, in these reaches the vegetated surfaces represent the most recently abandoned surfaces and are no longer flooded during the mean annual flood. The sample sites for the group V soils are marked on Figures 3.18 b and 3.18 c.

The Group II and Group IV soils

Group II and group IV lack definite absolute dates. However, it is possible that the group II soils may be related to deglaciation of the Zone III Gaick Plateau ice cap which was situated immediately to the south of Glen Feshie. Sissons (1974) has suggested that the Allt Lorgaidh fan (Figure 2.2) immediately downstream of the hummocky moraine limit, (Plate 3.19) was produced as an outwash fan at the end of Zone III by wastage of the tongue of ice from the Gaick Plateau ice cap.

The soil profile developed on the upper unit of the fan (Plate 3.20) shows that it is a well developed iron humus podzol comparable to the group II Glen Feshie soils. Soil-stratigraphic data from a pit dug into the surface of the upper unit of the low angle fan was inserted together with the Allt Garbhlach fan soil data into the original data set for the second Principal Components Analysis. Reclustering the scores on the first component clusters the Allt Lorgaidh fan soil with fragments 4 and 23 which constitute the group II soils.

Plate 3.19

Loch Lomond hummocky moraine, Allt Lorgaidh

Plate 3.20

Loch Lomond aged soil from the upper unit of the Allt Lorgaidh fan

Pollen spectra from the basal sediments of a core taken from the Pass of Dromochter on the south-west side of the Gaick Plateau ice cap indicate that the Gaick Plateau area became ice free soon after the close of the Late glacial stadial, probably around 10,000 radiocarbon years BP (Walker, 1975). Deglaciation of the Gaick Plateau area must have occurred by at least this date. It is therefore possible that a tentative date of about 10,000 radiocarbon years BP could be placed on deposition of the parent materials of the group II soils.

Supportive evidence for the suggested age for the deposition of the Allt Lorgaidh fan gravels and the initiation of pedogenesis on the fan surface comes from pollen analysis of the H horizon from the soil pit on the upper Allt Lorgaidh fan. Two samples for pollen analysis were taken from the deep H horizon of the iron humus podzol, one from the base of the horizon, and one from half way up the horizon. A skeleton count of 200 land pollen were counted from each sample in an attempt to correlate the pollen found in the soil to regional pollen diagrams. The first sample is characterised by very high Coryloid values, in the region of 50-70%. Betula peaked at 30%. Small amounts of Pinus were also present. There are only low percentages of herbaceous pollen. In the second sample Coryloid pollen falls to 15-20% and Pinus rises to 89% of the tree pollen, with Alnus comprising 11% of the tree pollen. Calluna comprises 43% of the shrub pollen. The pollen data from the first sample with its very low arboreal pollen records and domination of the pollen sum by hazel indicates that the landscape was predominately covered by shrubs. The pollen records from the second sample suggest a landscape dominated by pine and a heather vegetation with little hazel scrub. The pollen extracted from the lower sample suggest that it may be correlated with early Flandrian pollen assemblages. Pollen assemblages containing Corylus and Betula pollen were widespread in the Grampians in the early Flandrian (Walker, 1975; Birks and Mathewes, 1978). Dates on the Corylus rise range from 9,700BP to 8,400BP (Birks and Mathewes, 1978; Rapson, 1984). The small amount of herbaceous pollen would also indicate an early postglacial assemblage (Rapson, 1984). It is possible that the pollen in the lower sample dates from 8,400BP at the latest. The pollen assemblage from the upper sample may possibly be correlated with the mid-postglacial Pinus-Betula pollen zone (Birks and Mathewes, 1978; Walker, 1984). The beginning of the Pinus dominance in the central Grampians began about 7,225 radiocarbon years BP.

This pollen evidence suggests that soil formation and accumulation of an organic horizon at the Allt Lorgaidh upper unit fan site was probably well underway by the early Flandrian. The upper unit of the fan is therefore not likely to be younger than early Holocene. Absence of the permafrost features in the lower horizons of the Allt Lorgaidh fan suggests that the upper unit fan soil has not experienced a period of intense periglacial activity either following the deglaciation of the Late Devensian ice cap or during the Loch Lomond Stadial. These two lines of evidence would suggest that a very early Holocene date for the onset of pedogensis is indicated. It is therefore possible that the gravels constituting the upper fan unit aggraded during deglaciation of the Gaick Plateau ice cap, as suggested by Sissons (1974). A maximum date of 10,000 radiocarbon years BP for deposition of the gravels of the fan is therefore probably realistic. As the group II Glen Feshie soil profiles were grouped with the Allt Lorgaidh fan soil it is possible to suggest that the onset of pedogenesis on these surfaces was contemporaneous with that on the upper unit of the fan. A phase of gravel accumulation and floodplain building may therefore have occurred in the main valley at the same time as fan building in the tributary valley.

The soil data for the group IV soils indicate that they are more fully developed than the group V soils. The former are podzolic in character, whereas the latter are only in the incipient stages of podzolisation. However, the group IV soils are less well developed than the group III soils. An age intermediate between the group III and group V soils is also clearly indicated by the clustering of the component scores on the first principal axis.

An attempt was made to estimate the age of the group IV soils. The mean component scores for each of the four dated soil groups was calculated and a constant of 2 added to the mean score before log transformation of the negative values. Values were log transformed to facilitate fitting the trendline. The component scores were then plotted against age of deposition of the parent material gravels and a trendline fitted by eye (Figure 3.19). Plotting the mean component scores for the group IV soils onto the component score/soil age plane gave an estimated age range of about 800-1000 radiocarbon years BP for the deposition of the parent material gravels of the group IV soils.

3.9 The river terrace fragments grouped into terrace surfaces

The various methods of absolute age determination discussed above permit a tentative assignment of absolute ages for the accumulation of the gravels which have formed the parent materials for the terrace surface soils. They also provide

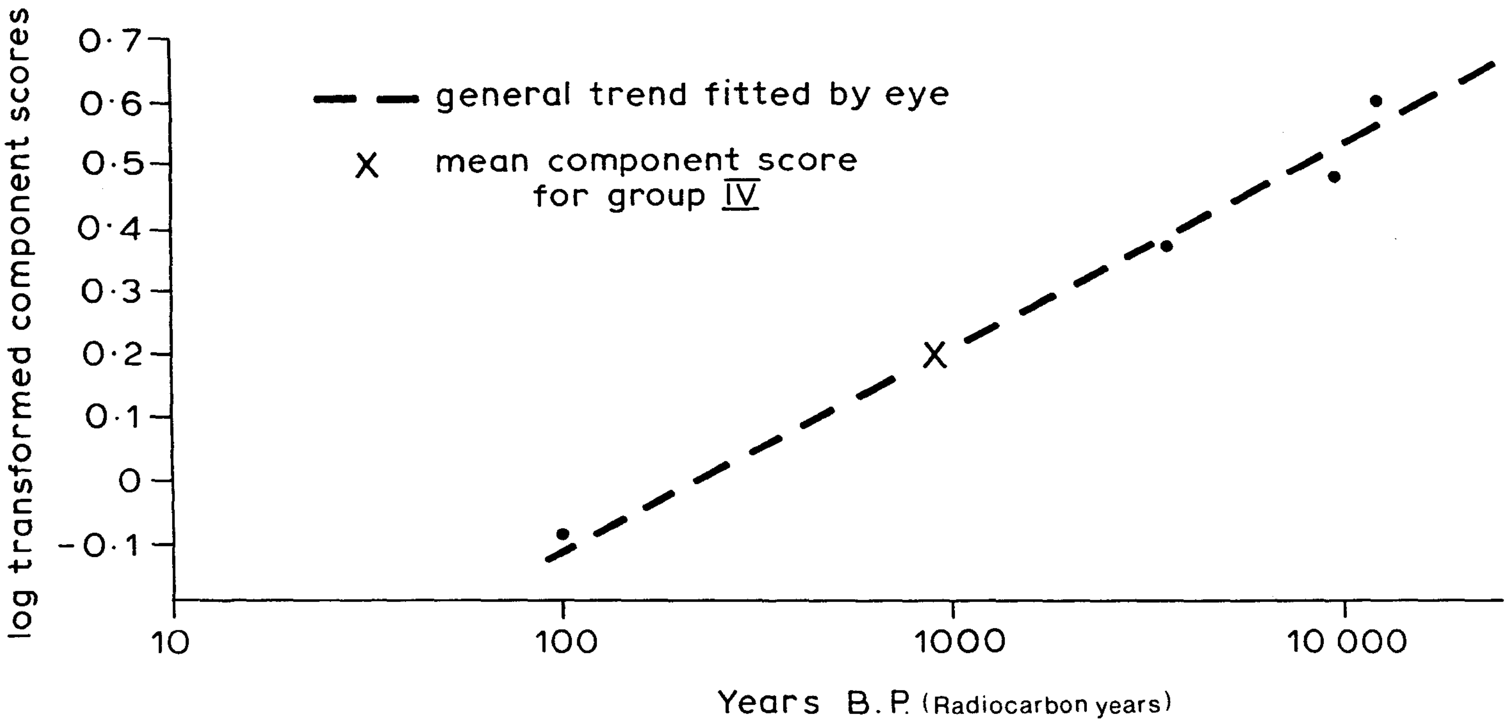


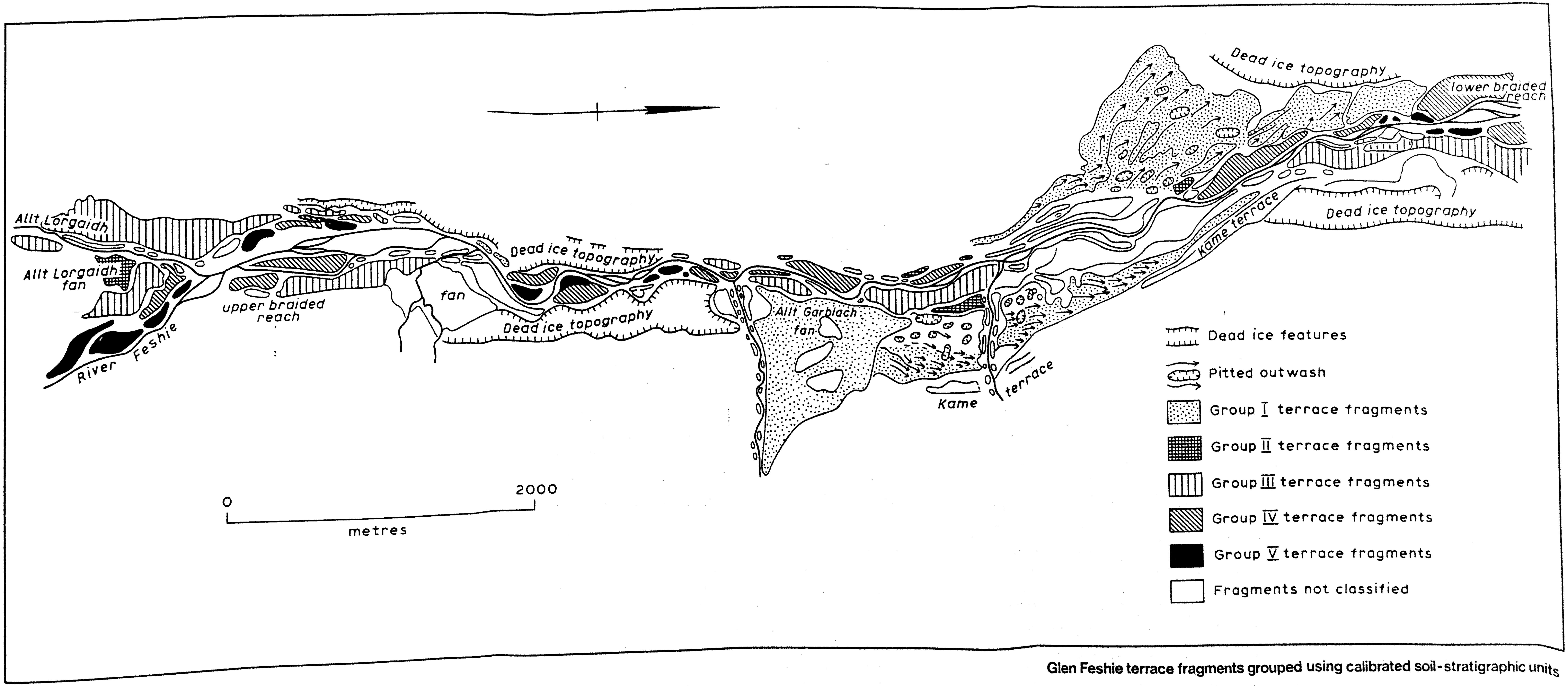
Figure 3.19 Trendline of Age against mean Component 1 score for each Soil Group

maximum dates for the onset of pedogenesis for the dated surfaces. Cluster analysis was used to objectively group the surface soils developed on the Glen Feshie terrace surfaces into soil units of differing degrees of podzolisation. As podzolisation is a time-dependent process this also grouped the soils into soil units of differing relative age. As the soil sites within one cluster have been shown to have distinctive physical and chemical soil properties which permit their recognition, and as each cluster has wide geographical distribution, each cluster of soil profiles comprises a soil-stratigraphic unit (Finkl, 1980). The grouping procedure thus allowed the derivation of five soil-stratigraphic units in Glen Feshie. The maximum dates for the onset of the phases of soil formation which produced the units have been placed at 13,000, 10,000, 3,600, 1,000 and 80 radiocarbon years BP.

As a soil-stratigraphic unit is, by definition, found only at one stratigraphic interval, on a deposit of one age (Birkeland, 1974), then each unit should also consist of terrace deposits which belong to one phase of gravel accumulation. Two lines of evidence may be advanced to support this suggestion for the Glen Feshie terraces. First, for both group I and group V, the majority of the terrace fragments can be independently dated and can be shown to be a response to one episode of gravel accumulation for each group. For the group I soils this phase of sedimentation was the deposition of the gravel deposits comprising the valley sandur which occurred during deglaciation of the late Devensian ice-sheet. The phase of accumulation which allowed the development of the group V soils was due to a period of gravel deposition consequent upon increased braiding intensity, bank erosion and valley fill reworking between 1869 and 1899.

Second, examination of the spread of the component scores along the first principal axis (Figure 3.2) for the group I and group V soils is indicative of some lateral variation in the soil properties. Some lateral variation in the physical and chemical properties of sites in a soil-stratigraphic unit is to be expected (Birkeland, 1974; Finkl, 1980). The spread of the points along the principal axis for the group III and group IV soils is no greater than that for the group I and group V soils, all of which belong to one episode of gravel accumulation. It therefore seems feasible to suggest that the group II, III and group IV fragments also belong to one period of gravel accumulation and subsequent stabilisation.

Each terrace fragment may therefore be assigned to an age group on the basis of the soil profile developed into its upper gravel units. The terrace fragments correlated into surfaces of the same age-range on the basis of the soil-stratigraphic data are shown in Figure 3.20. The dates that are available with which to calibrate the soil-stratigraphic units allow tentative age assignments for broad scale intervals of terrace development in Glen Feshie. The oldest fluvial terrace fragments are the dissected remnants of a valley sandur produced by meltwaters from the ice wastage of the Late Devensian ice sheet, about 13,000 radiocarbon years BP. The second, an areally restricted group, are suggested to be related to deglaciation of the Loch Lomond ice cap and are probably about 10,000 radiocarbon years in age. Three groups of late Holocene terraces have been distinguished. A date of 3,600 radiocarbon years BP provides a maximum date for the earliest late Holocene group of terraces. Historical data have provided a date of 80 radiocarbon years BP for the most recently abandoned fluvial surfaces in Glen Feshie. A date of about 800-1,000 radiocarbon



Glen Feshie terrace fragments grouped using calibrated soil-stratigraphic units

Figure 5.20

years BP has been estimated for the middle group of late Holocene surfaces. Although these dates must remain tentative they do provide convincing evidence of at least three phases of late Holocene valley floor instability and stabilisation.

CHAPTER 4

MULTIVARIATE MORPHOMETRY AND THE CORRELATION OF TERRACE FRAGMENTS

4.1 Introduction

Comparison of fluvial surfaces between successive reaches within a fluvial system, and, ultimately, comparison of fluvial surfaces between basins, directly raises the issue of correlation of alluvial deposits. The poor preservation of datable materials, particularly in gravelly river deposits, means that it is unrealistic to suggest that terrace fragments can be correlated solely on the basis of radiometric dates. Rather, the age assigned to a stream terrace should be based on the stratigraphic relation of the terrace fragment to dated deposits, or on the degree of post-depositional weathering of the surficial deposits (Born and Ritter, 1970). Correlation of terraces between reaches should therefore be made on the basis of variables which can be demonstrated to be time-dependent. The Glen Feshie terraces were correlated in Chapter 3 on the basis of age-calibrated soil-stratigraphic data derived from a podzolic chronosequence of soils. These data were derived from a number of parameters representing the physical and chemical properties of podzols that are time-dependent 'slowly adjusting' soil properties which may not reach a steady state in up to 10^4 years (Birkeland, 1974).

However, methods for both correlating terrace fragments between successive cross-sections in one river system and establishing their relative position in time have traditionally been morphological. The primary criterion for correlation has been longitudinal continuity of terrace remnants and an associated tendency for remnants of one terrace level to occur at a similar relative height above present river level, the latter usually established through a qualitative assessment of a height-range diagram (Leopold, Wolman and Miller, 1964; Kirby, 1969). This criterion for correlating terrace remnants is based on the premise that during a period of relative stability the stream constructed a floodplain that was more or less continuous along the valley (Leopold, Wolman and Miller, 1964).

The interpretation of the development and environmental context of terrace sequences in upland Scotland has been based on morphological data derived from levelling of terrace fragments and the construction of height-range diagrams (Kirby, 1969; Young, 1976, 1978; Sissons, 1979, 1982, Sissons and Cornish, 1982). As a result of the lack of datable reference levels, the establishment of a chronological framework within which to interpret the terraces has been based on the physiographic association of some terrace fragments with moraines and ice-contact deposits from the Dimlington and Loch Lomond Stadials, and the assumption that higher terraces are older.

It is now evident however, that the complex relationships between phases of cutting and filling in different parts of a drainage basin may mean that a height-range approach to terrace fragment correlation is inadequate. In recognition of the limitations of an approach to terrace remnant correlation based solely on morphological continuity and relative height above present river level, additional morphological parameters have been used to correlate and relatively date river terrace fragments. Coleman (1983) has described indices of terrace preservation and terrace scarp morphology which may provide additional morphological criteria by which to establish relative age. Mills and Wagner (1985) performed a statistical analysis of terrace elevations of discontinous, unpaired terraces producing a frequency distribution of terrace elevations. This distribution was used to group terrace fragments into surfaces of different age on the basis of ranges of heights above modern river level. Using these data Mills and Wagner established preferred elevations of terraces which were suggested to reflect changes in the rate of downcutting of the New River in southwest Virginia. It is notable, however, that the marked bimodality of the terrace heights and ages was confirmed using weathering indices.

The purpose of this chapter is threefold. First, it is proposed to discuss the problems involved in correlating terrace fragments and grouping them into surfaces of the same age range using the criteria of morphological continuity and relative height above present river level. Second, it is proposed to investigate the possibility of terrace correlation and relative dating using a statistical analysis of morphometric data. This will be achieved by using a number of morphometric parameters to group the terrace fragments in Glen Feshie statistically, employing the same multivariate techniques used to group the terrace remnants with age-related soil-stratigraphic data in Chapter 3. Third, the grouping of the Glen Feshie terrace fragments based on height-range methods, proposed by Young (1976), will be compared with the correlations made on the basis of the statistical analyses of the morphometric and soil-stratigraphic data. The differences between the grouping schemes will be discussed and evaluated.

4.2 The problems of terrace correlation using traditional height-based methods with particular reference to Holocene terraces

Morphological continuity of alluvial surfaces may produce misleading correlations of surfaces of different ages, particularly for low-level Holocene fluvial surfaces. Terrace and alluvial fan sequences in upland valleys in Britain are frequently characterised by low-level surfaces which exhibit minimal height variation and which appear, superficially, to be continuous but which are nevertheless diachronous. Some examples illustrate this point. First, at the upstream end of Jugger Howe Beck in the North York Moors, a compound, low angle fan occurs at the junction of a tributary stream, Hollin Gill, with the main stream, Jugger Howe Beck. The younger, lower part of the fan grades onto a well-defined, incised floodplain that is present along both banks of the study reach. The youngest unit of the fan has been dated radiometrically using organic material from an organic-rich layer in the lower part of the younger fan unit. This has yielded a date of 900+/-50BP for an alder fragment and 1,150+/-40BP for fine organic material (Richards et al., in press). Approximately 200m downstream from this site, lateral migration of the stream has exposed a cut bank in a small bog on the floodplain surface. Several radiocarbon dates ranging in age from 260BP to 200 BP show the floodplain itself to be extremely youthful, yet it shows a more or less continuous surface elevation from the lower part of the fan upstream. A morphological examination of these features would have placed them in terms of continuity and height above present river level, in the same relative age group. However the physical properties of the surface soils developed on the fan and floodplain surfaces reveal significant differences in total solum depth, depth of the B-horizon, and soil colour of the surface organic horizons and the B horizons (N. Peters, personal communication). These two surfaces would not be grouped in the same age range if correlation were based on soil-stratigraphic data.

A similar example may be cited from the Allt Gharbh Ghaig in the Gaick Plateau (M. Robertson-Rintoul and V. Brazier, in prep.). Here an alluvial fan has been divided into three different age units on the basis of the surface soil stratigraphy. There is a progression of relative age units from the apex of the fan to the toe, although each older age unit grades into the neighbouring younger age unit to produce a continuous down-fan gradient. A soil-stratigraphic analysis of the surface soils clearly differentiates between the different ages of the elements comprising the fan, but a morphological analysis based on continuity of surface would produce a misleading representation of this complex fan surface.

The complexity of the relationship between channel process and floodplain form, as well as cutting and filling in different parts of a drainage basin, is such that terrace morphology alone may be an inappropiate criterion by which to correlate and relatively date terrace fragments. The relative height of terrace surfaces above the present floodplain may not be uniform because the floodplain gradient of the stream may vary in response to changes in discharge, sediment calibre, stream sinuosity and structural discontinuities. The gradients of former floodplains may not be parallel so that terraces may converge or even cross. In this situation older terrace deposits may be buried by younger deposits, so that the criterion of relative height will produce misleading correlations. For example, Haible (1980) has described the changing gradient of Walker Creek, northern California as a result of upstream controls. Walker Creek in its upstream reaches has two terrace levels, the second of which has been created over the past 60 years as the channel has incised through its alluvial fill evacuating sediment from the upstream reaches. In the downstream reaches aggradation of the evacuated sediment has caused a flattening of the floodplain and river gradient. In consequence the modern floodplain has crossed the older terrace, burying the old landsurface with younger deposits. Downstream controls, in the form of base-level changes may result in variation of height above present river level for one terrace surface. Upstream valley lowering may continue as a knickpoint gradually recedes upstream, so that terrace remnants successively further upstream are increasingly lower relative to the original valley floor (Richards, 1982).

The criteria of relative height above present river level and longitudinal continuity used for correlating terrace remnants is based on the premise that during a period of relative stability the stream constructed a floodplain that was more or less continuous along the valley (Leopold, Wolman and Miller, 1964). However, investigations of the development of the active channel zone in braided streams suggests that floodplain evolution may be highly variable and lacking in spatial unity (Bluck, 1976, 1979). The result is a floodplain topography that is relatively high, as described by Lewin and Manton (1980). Thus although high-level terraces that are widely spaced in height terms can be easily distinguished inspite of this variation, low-level terraces are not so easily distinguished on a height-range diagram. Further, the channel zone and its associated floodplain is created during the lateral migration of the stream, the latter occurring as bars are stacked preferentially to one side of the channel zone. The pattern and locus of accretion in the channel zone may vary with bar type, sediment availability, which may change downstream (Church and Jones, 1982), and the process of channel avulsion. Isolation of discontinous, morphologically unpaired terraces may therefore be a likely adjunct to channel incision in braided streams. Such is the possible origin of the discontinuous, unpaired low-level terraces described from the Rakaia River, New Zealand (Rundle, 1985). Small (1973) has discussed examples in the Val d'Herens, Switzerland. The terrace sequence described by Born and Ritter (1970) exemplifies the development of terraces in braided channels that are shallow, and in which terraces are formed in thin sediment accumulations in response to high magnitude floods and downstream base-level control. Along the Truckee River, Nevada, which drains into Pyramid Lake, floodplains are aggraded during periods of very high flow. During such high magnitude events the entire braided zone in the valley floor is inundated resulting in considerable lateral erosion of the higher terraces. The sediment derived from erosion and reworking of the higher terraces provides the sediment for the aggradation of the new floodplain. Terracing of the floodplain occurs when the river returns to low flow and incision occurs in response to base-level change as the lake level is reduced. The extent of terrace preservation is determined by the magnitude of the lateral erosion during future high flows. Born and Ritter suggest the likelihood of the survival only of discontinuous terrace fragments because of the shifting nature of braided streams.

Multiple levels of discontinuous terraces which cannot be correlated morphologically may also be produced by a complex response of damped erosional and depositional phases triggered by one initial stimulus and which reflects coupling of main and tributary valleys (Schumm and Parker, 1973). Thus in Douglas Creek, western Colorado, a series of unpaired discontinous terraces have been produced in response to land use changes initiated in 1880's (Womack and Schumm, 1977).

Furthermore, the river response to changing environmental conditions may be spatially discontinuous, varying in a downstream direction. This may arise as a result of spatial variation in the depth of the valley fill/bedrock floor interface. Thus Harvey et al. (1984) describe the contrast between the upstream and downstream reaches of Carlingill in the Howgill Fells. Prior to 1900 AD the upsteam reach was unstable, migrating, and adding sedimentary units to the valley floor sediment surface. However, this reach is now stable and locked into bedrock. Conversely, the downstream reach, fed by sediment from the valley side scars, is unstable and has been adding units to the valley floor since 1900 AD. Similarly in Wales, Lewin et al. (1983) describe the discontinuous nature of the alluvial valley floor of the Afon Ystwyth. Above their study reach the river is confined by bedrock. Within the study reach the floodplain widens to about 250m and has been subjected to recent variation in sediment supply. This has resulted in the development of a series of low-level terraces which are not present in the upstream reach.

Spatial variation in the development of temporary sediment stores such as floodplain elements, may be caused by the alternation of transport reaches and sedimentation zones and their subsequent migration along gravel-bed rivers (Church and Jones, 1982; Church, 1983). Changes in bed elevation accompany accumulation and removal of sediment as migrating sedimentation zones along the river produce a stepped river profile. The location of the sedimentation zones and transport reaches may vary through time in association with periods of enhanced but localised sediment input. If river incision follows, morphological continuity of such surfaces may be difficult to trace along the stream. This effect is likely to be most pronounced in shallow gravel bed streams with relatively thin accumulations of sediment.

A further problem in correlating terraces from traditional morphological data relates to the actual construction of the final height-range diagram and is likely to be one that affects low-level terraces in particular. This is illustrated by Kirby (1969) who carried out a morphometric analysis of the glacio-fluvial terraces in the Esk Basin, Midlothian. A height-range diagram involves the selection of a projection line onto which the terrace elevations are transferred; problems will arise however in the choice of a single projection line :-

(1) Unless the line of levelling of the terraces is exactly parallel to the projection line, the projected points will give a series of points on the height-range diagram for which the best fit line is steeper than that measured in the field. For small differences in direction the projection error will not make a large difference to the slope. However, for a deviation of 30 degrees the projected slope angle would be increased by one seventh. Thus, a slope of 0.012 would be increased to 0.0137. For greater deviations the slope angle is increasingly in error. The chosen base line may not be the direction of the maximum, or valley slope preserved by the terrace surface, so that variation in the direction of valley slope for individual fragments in a reach may considerably alter the correlation of low-level surfaces which have height separations of less than a

few metres.

(2) Choosing a straight line as a projection line also compresses the extent of the terrace and the spacing of one fragment in relation to its neighbours. As Kirby points out this is likely to be less of a problem for outwash surfaces but is likely to present serious correlation problems for low-level terraces. The projection line chosen by Kirby gave a good representation of the high, glaciofluvial terraces, but gave relatively poor separation of the low-level, postglacial surfaces.

Problems such as those discussed above are likely to be exacerbated in upland valleys in Britain where braided or split gravel-bed channels are shallow, where terraces are frequently formed in relatively thin sediment accumulations and where recent terraces may be interrupted by rock outcrops. Under these circumstances it may be difficult to distinguish one terrace level from another even although there may be a time lag of several hundreds or thousands of years between the deposition of the terrace deposits of the varying levels. Furthermore, as incision occurs rivers migrate laterally so that remnants of former floodplains are often very fragmentary, or the shallow terraces formed during the migration of the river may be 'lost' in the older record. Morphological correlation of fragments may thus be of limited value when some terraces may be missing at some locations, and different terrace fragments may have been deposited at slightly different slopes.

4.3 A quantitative analysis of morphometric variables as a means of correlating and relatively dating river terraces

Five morphometric variables were used to group the Glen Feshie terrace fragments statistically; these were terrace gradient, height above present river level, maximum terrace scarp angle, mean terrace scarp angle, and total sinuosity.

(a) Terrace gradient

The gradient preserved by the terrace fragment is the valley gradient of the reach of the former floodplain preserved by the fragment (Richards, 1982). In this study an estimate for the gradient of each terrace fragment was required for two purposes:-

(1) For the comparative morphometric analysis to be carried out in this chapter. For this purpose it is essential that the slopes calculated for each terrace fragment are directly comparable. Most studies derive the terrace gradient by running a line of levels down the long axis of each terrace remnant assuming that this will give a representative value of the downstream, and therefore, valley gradient preserved by the terrace fragment (for example, Kirby, 1969; Young, 1976, 1978; Sissons, 1979, 1982; Sissons and Cornish, 1982). However, as terrace remnants possess lateral as well as downstream gradients, placing of the line of levels so that it is along the direction of the maximum gradient is extremely difficult to achieve in the field. It cannot be known therefore, if the gradient being computed is directly comparable for all terrace fragments, a problem which will be magnified if the terrace fragments are extensive both laterally and in a downstream direction. As many of the Holocene terrace fragments in the Glen Feshie study reach are over 500m and up to 750m long and up to 150-200 wide this was certainly a problem in the study area.

(2) For the palaeohydrological calculations to be made in Chapter 8, using the methodology outlined in Chapters 6 and 7. This methodolgy utilises a stream power term in the derived equations. The stream power used in the analysis is derived from the product of valley gradient and discharge. As the calculation of stream power is extremely sensitive to slope variation it was important to determine as accurately as possible for each fragment the terrace slope which best approximated the local valley gradient.

In practice the terrace fragment may be regarded as a plane surface, at least for short distances downstream (Culling, 1957). The slope of this plane surface will approximate to the maximum slope or valley slope preserved by the terrace fragment. The most appropriate method for calculating this former valley gradient is therefore to compute the slope of the best fit plane fitted over the surface using least squares regression analysis (Kirby, 1969). As the fitting of a best fit plane requires the spatial x and y, and elevation z, coordinates for a large number of points, preferably regularly spaced over the terrace surface, this meant the retrieval of a large amount of planimetric and height data for each terrace fragment. It was considered impractical to retrieve the required data from a field survey programme. The availability of high quality, large-scale aerial photography for Glen Feshie, and of a Kern PG2 stereoplotter (Chapter 2), meant that the most efficient and practical method of gathering the data required was to derive the terrace gradients from altitude matrices of terrace fragment heights generated from photogrammetric plotting.

(b) <u>Photogrammetric plotting of the Glen Feshie terrace fragments</u>

In order to derive the altitude matrices of terrace elevation points a net of grid lines was drawn onto plastic film overlays and placed onto the 1:5 000 scale planimetric maps prepared as described in Chapter 2. The altitude values were arranged on a rectangular grid, with a spacing of 50x25m on the ground. A 50x25m rectangular grid was selected to allow sufficient height points to be plotted in the x or cross valley direction, as well as in the y or down valley direction. Height points were therefore plotted at an interval of 25m across the terrace fragments and at 50m intervals in a downstream direction. All ground points were orientated relative to the National Grid. The three x, y and z coordinates in metres were obtained for each point. The x and y coordinates for each terrace fragment, although tied into the National Grid, were calculated independently for each individual terrace fragment for the purposes of the regression analysis. The origin, with x and y coordinates (0, 0), for the altitude matrix of each individual terrace fragment was taken as the most southwesterly point on the fragment. Each x and y coordinate was then calculated in metres with reference to the origin. The altitude matrices for each stereomodel are given in Appendix 3.

The accuracy of the spot heights obtained from the photography is a function of the flying height at which the photography was taken. The standard error of the spot heights may be approximated by calculating 0.02% of the flying height. For the Glen Feshie photographs this gives a standard error of +/-25cm-30cm. Field checking of spot height data indicated that the spot heights were accurate to within at least +/-20cm. As the amplitude of the surface roughness elements of the terrace surfaces, the braid bars and channels, was about 1m, a spot height accuracy of +/-20cm is more than adequate for the calculation of the slope of the best fit plane for each fragment.

(c) Statistical analysis of the output data from the photogrammetric plotting

The x, y, z coordinate data were used as input in fitting the best fit planes to each individual terrace fragment. This was carried out using Trend Surface Analysis. The trend surface model is a special case of the General Linear Model which also includes multiple linear and polynomial regression. However, there are several differences between Trend Surface Analysis and multiple linear regression which facilitate its use in analysing spatial data. One of the most important of these is the generation of powers and cross-products of the coordinates (the x_i, y_i) of the sampling points in their correct order. The trend surface procedure defines a spatial series in which the z observations are ordered with respect to the two spatial coordinates (Unwin, 1975). The equation

$$z_i = \alpha_o + \alpha_1 x_i + \alpha_2 y_i \qquad 4.1$$

where

z_i = height of the trend component at the i^{th} point

x_i = the x coordinate of the i^{th} data point

y_i = the y coordinate of the i^{th} data point

is similar to the standard multiple regression equation; the α coefficients are analogous to the β coefficients and are computed by the method of least squares (Mather, 1976). Fitting the best fit plane to the spatial and height data from the terrace fragments, means that the α coefficients are interpretable in physical terms; $\alpha 1$ is the slope, or rate of change of surface height, of the terrace in the x direction, and $\alpha 2$ is the slope, or rate of change of surface height, in the y or downstream direction.

For a terrace fragment whose three-dimensional shape is adequately summarised by the plane defined by Equation 4.1 above, it is necessary to define the <u>maximum</u> slope rather than the slopes parallel to either the x or y directions; the problem is analagous to that of finding the true dip of a stratum. The direction of the steepest slope on the surface is given by the angle θ with respect to the y direction,

$$\tan \theta = \alpha_1 / \alpha_2 \qquad 4.2$$

where α_1 and α_2 are the x and y coefficients from the regression. The slope itself is derived from Pythagoras theorem

$$\alpha_{max} = (\alpha_1{}^2 + \alpha_2{}^2)^{\frac{1}{2}} \qquad 4.3$$

The object of the trend surface analysis in this study was to obtain the steepest gradient of the best fit plane for each terrace fragment. The null hypothesis to be postulated in the context of the trend surface analysis is therefore that the variance regarded as being due to the trend in the sample data is the result of chance alone, or, a variance reduction as great or greater than that actually achieved, could have been obtained by random sampling of a population in which there was no trend (Unwin, 1975; Mather, 1976). In effect this gives a test of the hypothesis that all the population parameters are zero. It is tested by separating out the two sources of variation - systematic spatial trend and random local trend - into which the spatial series has been decomposed, to form a variance ratio, F, according to the relation

$$F = \frac{\%RSS.df_2}{(100-\%RSS).df_1} \qquad 4.4$$

where

%RSS = percentage of sum of squares accounted for by the linear surface

df_1 = degrees of freedom associated with the surface

df_2 = degrees of freedom associated with the residuals

Appendix 2 gives the results of fitting planes to all the terrace fragments considered. The equations describing the planes are given, as well as the valley slopes, the direction of the valley slopes and the results of the F ratio tests. The results of the F ratio test for all the terrace fragments allows the rejection of the null hypothesis that the planar shape of the terrace fragments may be due to chance variation in the z values of the data points. The analyses show that all the terrace fragments are predominately planar in shape, since the R^2 values (the percentage of variance explained by the linear surface) range from 85% to 97%.

The gradients of the 13,000BP outwash surface are, as would be expected, the steepest terrace gradients ranging from 0.02 immediately downstream from the Allt Garbhlach fan to 0.0172 at the lower braided reach. The gradients for the 10,000BP surface could not be measured because of the limited areal extent of the fragments. The gradients of the individual fragments comprising the 3,600BP-80BP surfaces are variable and are markedly lower than those for the outwash surface. The gradients of the

3,600BP fragments range from 0.012 at the upper braided reach to 0.015 -0.013 at Allt Garblach, increasing to 0.017 at the Achleum reach and declining to 0.013- 0.011 at the lower braided reach. As a result of the reduction in the valley slope between the outwash surface and the 3,600BP surface below it, the 13,000BP surface converges towards the 3,600BP surface at the lower braided reach. The 1,000BP surface has gradients of 0.011 at the upper braided reach, 0.013-0.009 at the Allt Garblach reach, 0.017 at Achleum and 0.013-0.012 at the lower braided reach. The 80BP surfaces have gradients of 0.014-0.012 at the upper braided reach and 0.014-0.013 at the lower braided reach. The gradients for the individual fragments are locally variable, but overall the gradients of these lower surfaces are roughly parallel. The average present valley gradient in the study reach is 0.012. As a consequence the terraces do not exhibit overall convergence towards the floodplain, although this may occur locally for individual terrace fragments.

(d) Height above present river level

The height above present river level for each terrace fragment was measured during the photogrammetric plotting programme carried out to calculate the terrace gradients. Subsequent field checking of these height data revealed that the plotted heights were accurate to between 10-20cm. Heights were calculated from the centre of the terrace fragments every 50m downstream and an average height used in the subsequent analysis.

(e) Terrace scarp angle

A number of studies have demonstrated that certain morphological variables may be potentially useful for estimating relative time differences between landforms of varying ages (for example, Welch, 1970; Brunsden and Kesel, 1973; Wallace, 1977). These studies have shown that for moraine slopes, wave-cut bluffs and slopes of fault scarps, the slopes change systematically with time as a result of erosion on the upper part of the slope and deposition on the lower part. Scarp slope was shown to be a function of length of time available for the operation of these proceses, although it was also found to be a function of scarp height. More recently, Coleman (1983) has argued that the slope of a terrace scarp may be used to estimate the age of a terrace. Coleman was considering the terraces of Rappahannock River in Virginia, which range in age from early Quaternary to about 187,000 years ago. Overall, the average maximum slope of the scarps decreases with age, although no significant differences were discerned between the three intermediate age terraces. It is possible that the extreme age of the terrace scarps and their advanced state of degradation may be responsible for lack of a more clearly defined trend in the data.

However, scarp angle may prove to be a successful discriminator of terrace age for terrace sequences where scarps are not in an advanced state of degradation. Accordingly, terrace scarp angles were measured for the Glen Feshie terrace fragments in the study reach. The slope profiles were measured using an Abney level secured onto a metre rule. A reading was taken every metre and at least three profiles were measured for each terrace fragment. The scarp associated with the development of the terrace is the scarp immediately above the terrace that was being undercut while the terrace surface was being formed, thus the terrace and the scarp above it are essentially of very similar age (Coleman, 1983). For each profile maximum slope angle and mean slope angle were obtained, and the average values of the three profiles for each terrace fragment were used in the analysis. The 80BP terrace scarps are those adjacent to the present river floodplain and have maximum scarp angles ranging from 59-42 degrees, and mean scarp angles ranging from 45-41 degrees; the 1,000BP surfaces have maximum angles ranging from 46-40 degrees and mean angles of 45-32 degrees; the 3,600BP surfaces have maximum scarp angles ranging from about 39-35 degrees and mean angles from about 32-29 degrees; the 10,000BP surfaces have maximum scarp angles of 34-31 degrees and mean angles from 27-29 degrees; the 13,000BP surface has maximum angles ranging from about 34-32 degrees and mean angles from about 29-24 degrees. These angles therefore appear to vary systemtically with age of the surface.

(f) Total Sinuosity

In Chapter 6 it is shown that the variable 'total sinuosity', may be used to quantify the degree of braiding of the active zone of a braided channel. This may be estimated for palaeo-channels preserved on terrace fragments using the methodology developed in Chapter 7. Measurements of total sinuosity of the palaeochannels on the terrace surfaces showed that degree of braiding of the River Feshie has varied through time. Total sinuosity was therefore used as an additional morphometric variable with which to discriminate between different terrace levels. Thus the 13,000BP terrace has total sinuosities ranging from 8.3-8.5; the 10,000BP terrace fragments are again too restricted to allow estimates to be made of this parameter. The 3,600BP terrace fragments have total sinuosities ranging from 4.5-5.83; the 1,000BP terrace from 2.6-4.9 and the 80BP surface has total sinuosities ranging from 3.3-4.7.

4.4 Numerical Analysis

(a) Principal Components Analysis

Principal Components Analysis was performed on the data matrix of morphometric variables. In this analysis it was not possible to include data from the 10,000BP surface. This is because the fragments of this surface are too restricted in areal extent to measure either a representative slope or total sinuosity. A 29x5 matrix was used. This matrix has a smaller number of data points for the terrace fragments than that used for the soil data. In the soil matrix two points were included from the tributary valley, the Allt Lorgaidh, points were included for the 10,000BP fragments and there were several duplicate soil pits on some of the terrace fragments.

The derived components, or eigenvectors, and the eigenvalues for the morphometric data set are presented in Table 4.1.

As with the statistical analysis of the soil-stratigraphic data in Chapter 3, Kaiser's rule was again used to determine the significant principal components. Only the first of the derived components had an eigenvalue of greater than one, this accounting for 74% of the original variance in the data.

Table 4.1

Eigenvalues and cumulative proportion of total variance for 2 principal components

1	3.7075	0.7415
2	0.7662	0.8947

Principal Components Loadings Matrix

1	Height	-0.9455	0.0322
2	Mean angle	0.9590	-0.1828
3	Max. angle	0.8066	-0.4709
4	Gradient	-0.6530	-0.6999
5	Tot. sin.	-0.9038	-0.1422

The loadings on the first component have strong negative correlations with relative height and total sinuosity, and strong positive loadings on mean terrace scarp angle and maximum scarp angle, with a weaker negative correlation with terrace gradient. Terrace gradient does change systematically from the higher, outwash surface to the lower, younger terraces as indicated by its negative loading onto the first component. Thus as relative height above present river level decreases total sinuosity and gradient of the terrace also decreases, but the angle of the terrace scarps increases. The variables which load onto this component are variables which would generally be expected to vary with the age of the terrace surface. Broadly, relative height above present river level would be expected to decrease with decreasing age of the terrace fragment. Scarp angle has been demonstrated to be a function of length of time available for modification of the slopes (Coleman, 1983). This first component may therefore be interpreted as having a temporal dimension. In this sense the first component is therefore similar to the first component in the soil-stratigraphic analysis, which, as a compound index of podzolisation, also had a temporal dimension.

The variable which loads most highly onto the second component is terrace gradient. Its loading onto the second component suggests that although terrace gradient does vary in a systematic way with terrace age, gradient must be influenced by factors, presumably spatial, which have not been represented in the first component. Thus component 2 isolates gradient variability that may relate to factors such as local variation of sediment supply and size, and valley confinement.

As with the analysis of the soil-stratigraphic data in Chapter 3, projecting the component scores of each terrace fragment on the first principal component enables the structure of the first component to be graphically illustrated. This is shown in Figure 4.1. The fragment scores are dispersed in discrete clusters along the axis representing the first component. With the morphometric data, five groupings of terrace fragments would not be expected as it was not possible to include data for the 10,000BP surface. However, if the multiparameter analysis of morphometric data is capable of achieving as fine a separation between the terrace fragments as the soil-stratigraphic data, there ought to be four groups of terrace fragments plotting onto the principal axis. However, on the principal axis plot there are only three discrete clusters of terrace fragments.

(b) <u>Cluster Analysis</u>

Cluster analysis was used to confirm statistically the identification of the clusters of points projected onto the first principal component.

The standardised scores on the first component were used as the input data for the clustering procedure, using the methods outlined in Chapter 3. The screen test was used to define a cutoff point for the acceptance of the number of groups in the

Projection of component 1 scores for morphometric data onto the first principal axis

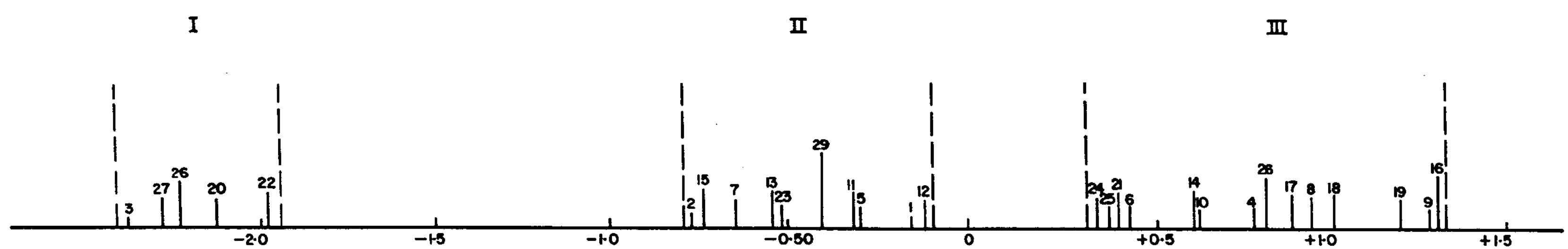


Figure 4.1

cluster analysis. Figure 4.2 shows the plot of within-group variance as a proportion of total variance against the number of groups for the morphometric data. This indicates that there is a discontinuity where the scores are clustered into 3 groups. Before the discontinuity the error sum of squares increases slowly. Beyond the discontinuity, from three clusters to two, there is a large increase in the error sum of squares. This would suggest that the level which best minimises the within-group variance and maximises the between-group differences is the three cluster grouping.

Analysis of variance was used to test the statistical significance of the grouping scheme. The resultant F ratio was 63.41. The 1% point for $F(2,26)$ is 5.53 so the null hypothesis of no significant difference between the groups can be rejected confidently. The three grouping scheme is significant at greater than the 0.01 level.

4.5 Comparison of the terrace grouping schemes using soil stratigraphic data, morphometric data and the grouping scheme derived from a published height-range diagram

Figure 4.3 shows the Glen Feshie terrace fragments within the study reach grouped into surfaces of the same age-range based on the soil-stratigraphic data. Figure 4.4 shows the grouping scheme derived from the morphometric data and Figure 4.5 illustrates the grouping of the Glen Feshie terraces based on the height-range diagram of Young (1976). The soil-stratigraphic grouping scheme gives five terrace levels, from Group I (the 13,000BP surface) through to group V (the 80BP surface). The morphometric data give three groups, groups I to group III.

Acceleration of information loss for cluster analysis of morphometric data

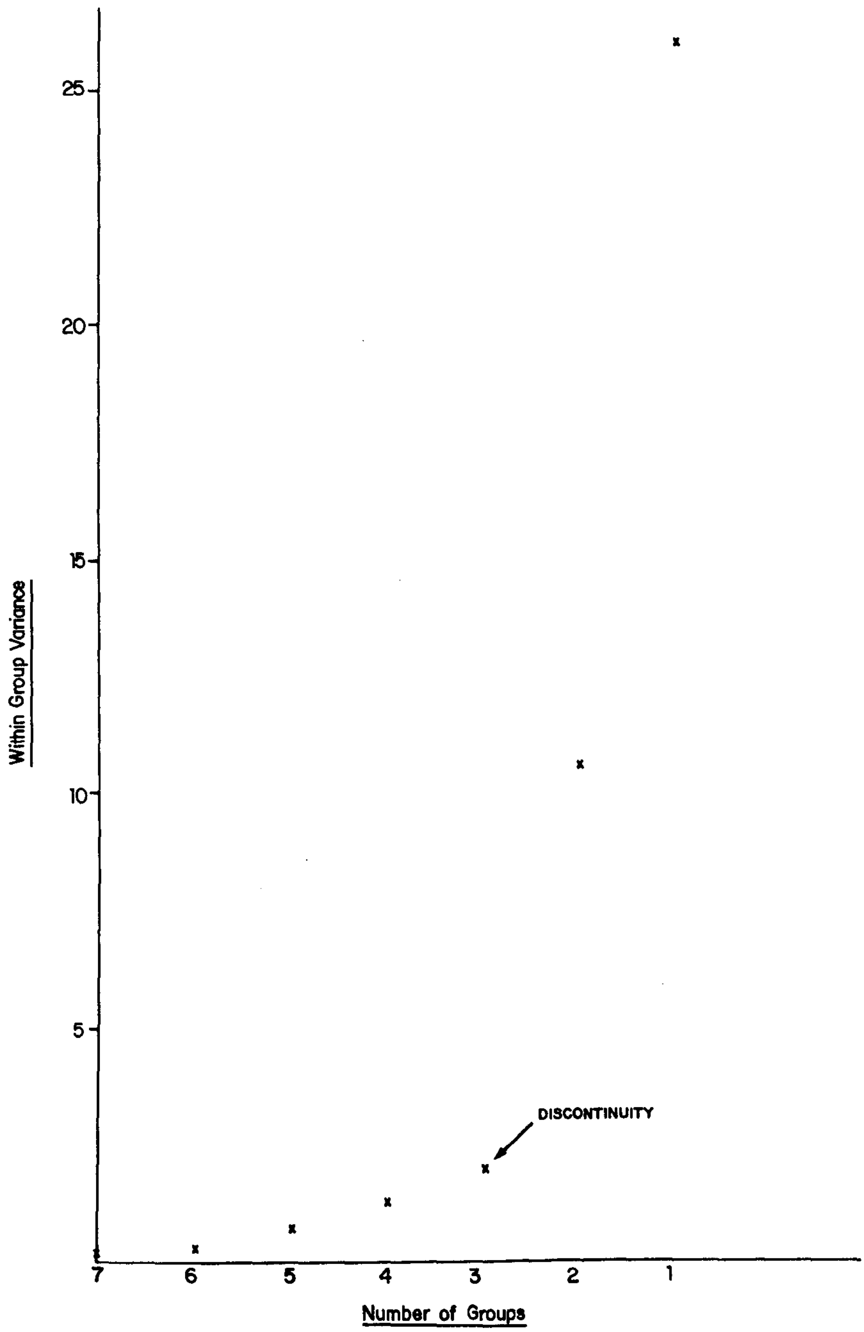
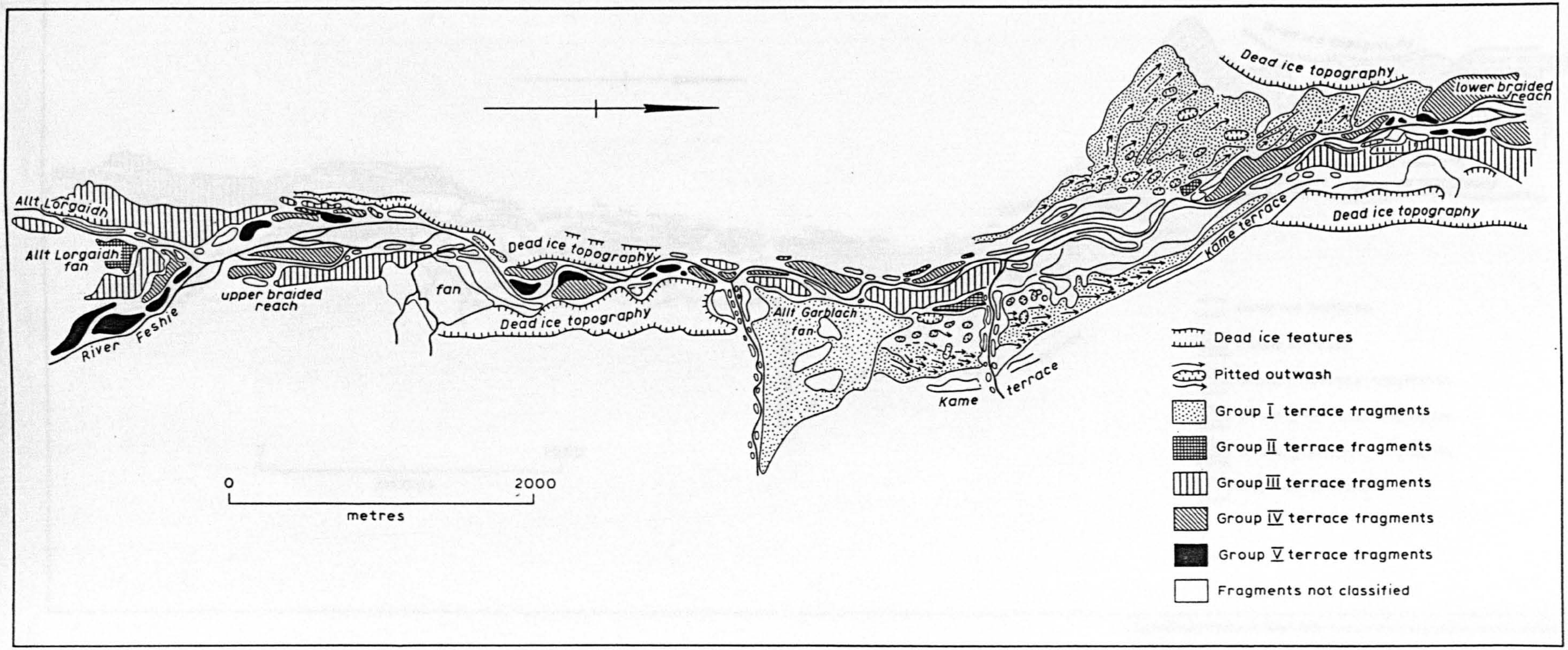
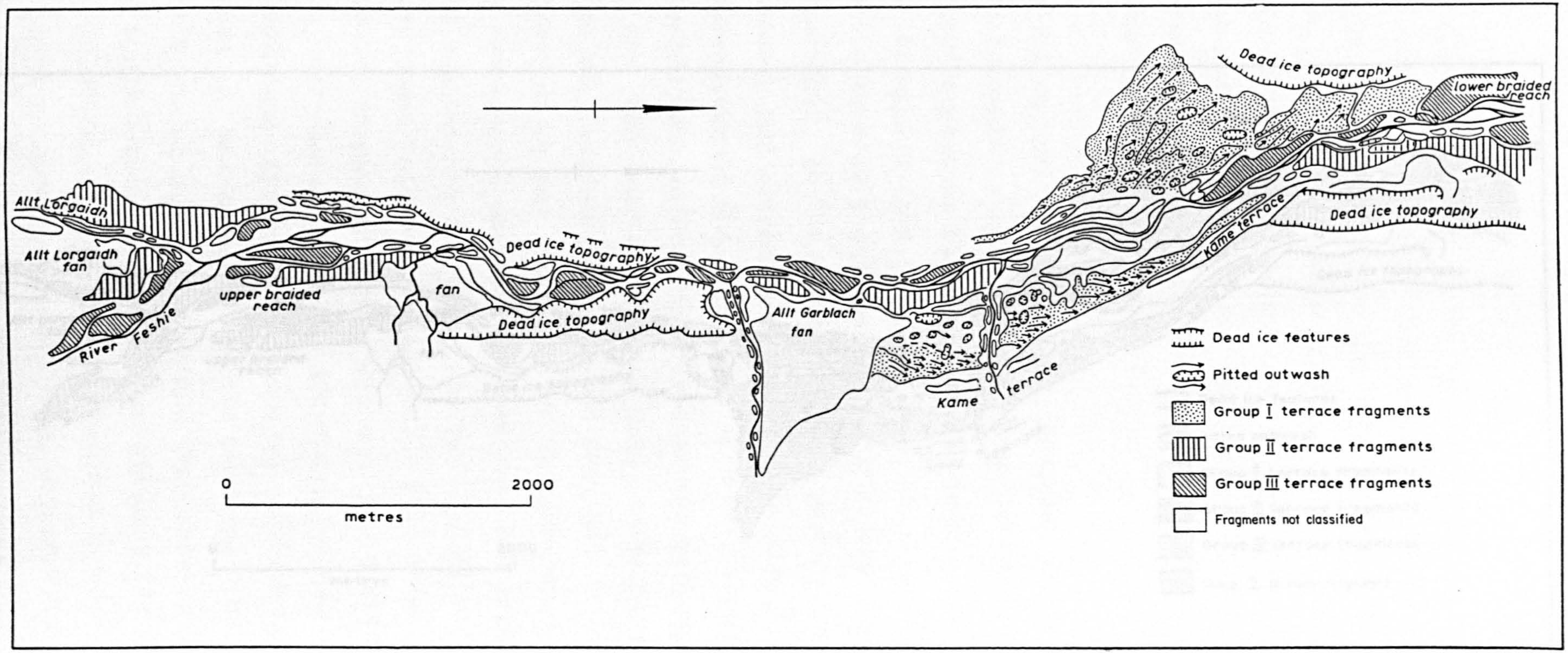


Figure 4.2



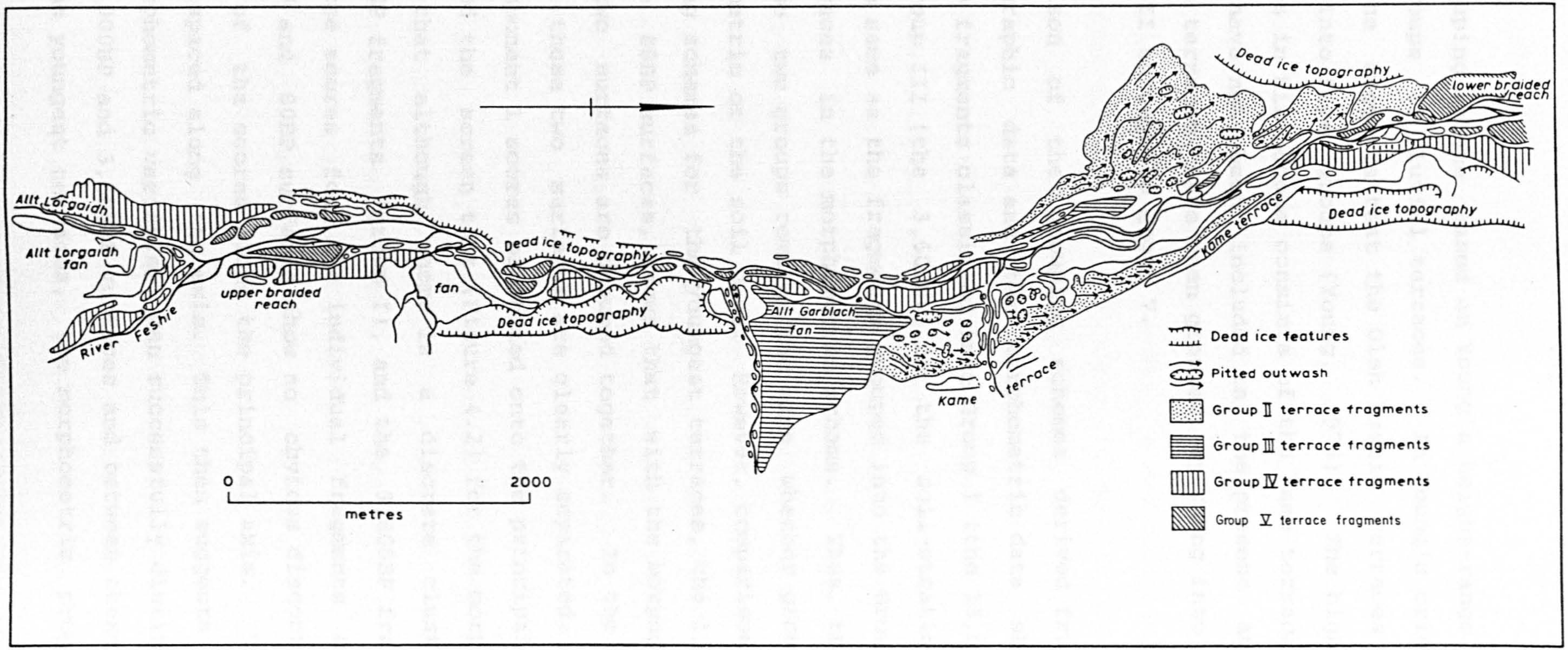
Glen Feshie terrace fragments grouped using calibrated soil-stratigraphic units

Figure 4.3



Glen Feshie terrace fragments grouped using morphometric variables

Figure 4.4



Glen Feshie terrace fragments grouped using height-range data (after Young, 1976)

Figure 4.5

The grouping scheme based on Young's height-range diagram gives four groups of fluvial terraces. In Young's original grouping scheme he suggests that the Glen Feshie terraces may be correlated into five groups (Young, 1976). The highest group of terraces in his scheme consists of the kame terraces (Chapter 2) which have not been included in the present analysis. The fluvial terraces have been grouped by Young into four groups, groups II through to group V.

Comparison of the grouping schemes derived from the soil-stratigraphic data and the morphometric data shows that the terrace fragments classified into Group I (the 13,000BP surface) and Group III (the 3,600BP) for the soil-stratigraphic scheme are the same as the fragments grouped into the Group I and Group II terraces in the morphometric scheme. Thus, the composition of these two groups remains the same whether grouped using the morphometric or the soil data. However, comparison of these two grouping schemes for the youngest terraces, the 1,000BP surface and the 80BP surfaces, shows that with the morphometric scheme these two surfaces are grouped together. In the soil grouping scheme, these two surfaces are clearly separated. Reference to the component 1 scores projected onto the principal axis (Figure 4.1) and the screen test (Figure 4.2) for the morphometric data shows that although there is a discrete clustering of the 13,000BP fragments (group I), and the 3,600BP fragments (group II), the scores for the individual fragments comprising the 1,000BP and 80BP surfaces show no obvious discontinuity in the spread of the scores along the principal axis. They are quite widely spaced along the axis. This then suggests that although the morphometric variables can successfully distinguish between the 13,000BP and 3,600BP surfaces and between these two surfaces and the youngest terraces, the morphometric properties cannot succcessfully differentiate between the most recent terrace surfaces.

However, comparison between the statistically derived grouping scheme using morphometric data and the grouping scheme derived from a height-range diagram suggests that, although a morphometric analysis may group together the younger Holocene surfaces, it is a method which will provide a more satisfactory approach to terrace correlation and relative dating than an analysis solely based on a height-range diagram.

Considering the terrace groups based on the soil-stratigraphic data and the height-range analysis (Figures 4.3 and 4.5) only one terrace surface contains fragments which are coincident between the two grouping schemes, whereas with the morphometric scheme there were two coincident surfaces. This is the group I outwash surface which is in fact defined by all schemes. Apart from this one group there are a number of fundamental differences between the fragments grouped using the soil-stratigraphic data and those grouped using the height-range methodology. These are :-

(1) Young's grouping scheme contains no comparable group to the 10,000BP (group II) surface from the soil-stratigraphic scheme. The fragments included in this soil-stratigraphic group have remained ungrouped in Young's scheme.

(2) Young's Group III fragments. This surface has no comparable group according to the soil-stratigraphic scheme, or the morphometric scheme. Young's Group III terraces are an areally restricted group which contains two extensive

elements and several small terrace fragments. These main elements are the Allt Garblach fan and a fragment on the west bank of the River Feshie. It is the gradient of the Allt Garbhlach fan which has been used as the basis for the derivation of this group. As such, the validity of the correlation of a group of mainstream terraces on this basis must be questioned.

(3) Young's Group V fragments consist of fragments which in the soil-stratigraphic scheme have been grouped separately as the 1,000BP and 80BP surfaces. Thus as with the morphometric scheme, the height-range method groups together the youngest of the late Holocene surfaces.

(4) Young's group IV terraces contain a large proportion of the Group III terraces (3,600BP) that have been classified using the soil-stratigraphic data and which were also comparably grouped using the morphometric data. However, Young's group IV also contains some fragments that are correlated as 1,000BP and 80BP surfaces in the soil-stratigraphic scheme (Figure 4.3 and Figure 4.5).

Some examples from the upper braided reach illustrate the differences between the two schemes for the low-level terraces. Between the upper Feshie and the Allt Lorgaidh soil-stratigraphic data have provided evidence for fluvial surfaces of four different ages (Figure 4.3). These are the Loch Lomond outwash fan which is dated at about 10,000BP, the 3,600BP surface (fragment 40 in Figure 4.6) the 1,000BP surface (fragment 39 in Figure 4.6) and the 80BP surface (fragment 36 in Figure 4.6). In Young's scheme the fragment 40 remains ungrouped, whilst fragments 39 and 36 are grouped together into the group IV surface

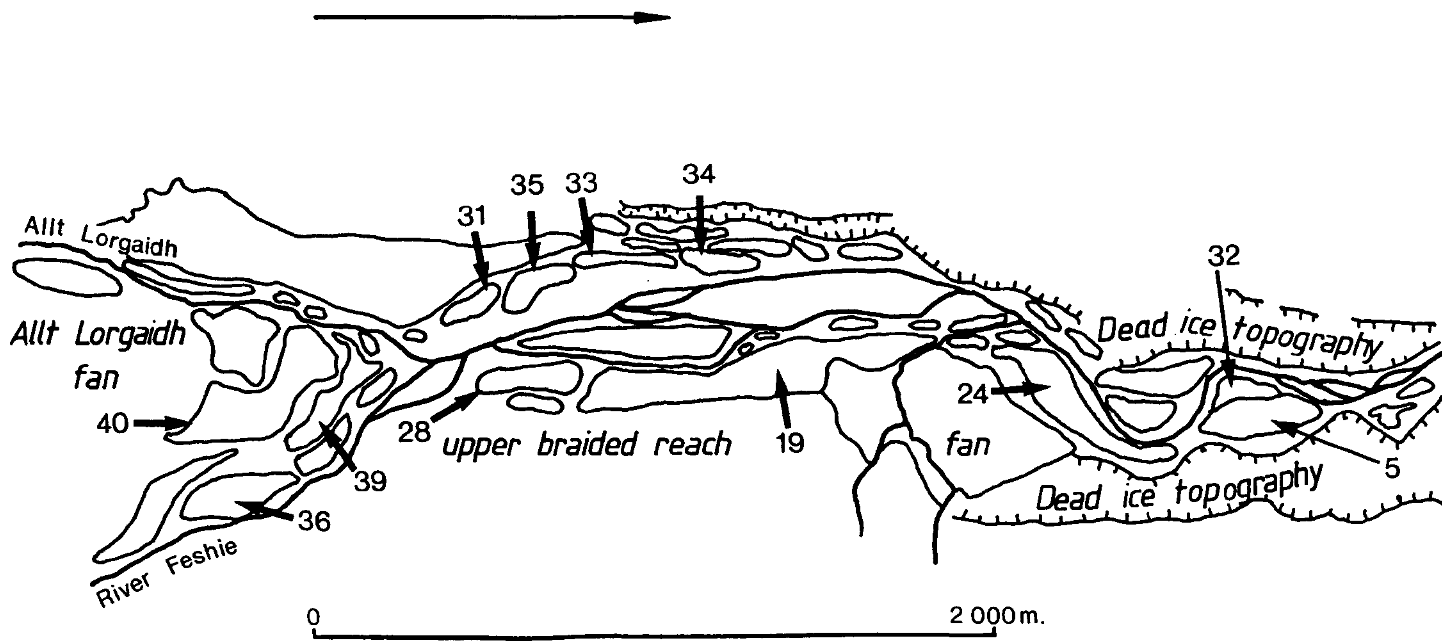
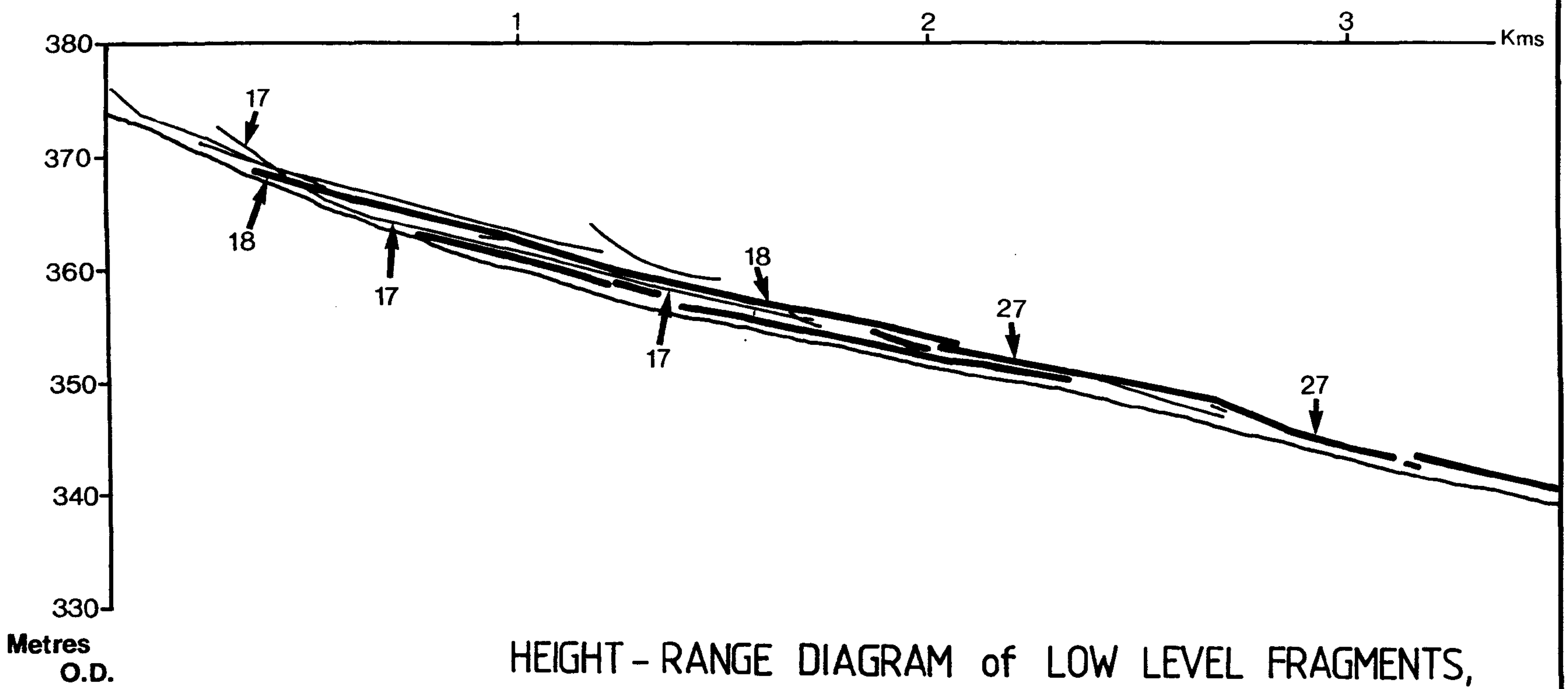


Figure 4.6 Location of Upper Braided Reach Terrace Fragments

in Young's scheme. This is the equivalent of the 3,600BP group in the soil-stratigraphic scheme. At the upstream end of the upper braided reach fragment 28 in Figure 4.6 has been correlated by Young with fragment 19 into his group IV surface. This fragment has been grouped as a 1,000BP surface under the soil-stratigraphic scheme. There is thus a mixture of 3,600BP, 1,000BP and 80BP fragments incorporated into Young's group IV fragments.

The inclusion of a range of terrace fragments of diverse age within one group may be the result of some of the problems associated with attempting to correlate low-level terraces on the basis of a height-range methodology. The direction of the projection line used by Young for the construction of the height range diagram for the mainstream Feshie terraces is N10°E/S10°W. It is suggested by Young that the long axis of most of the terraces is parallel with this line. However, for a number of the terrace fragments in this reach this is not the case. For example, the four fragments 31, 35, 33, and 34 (Figure 4.6) together comprise Young's fragment 17. Young's fragment 17 is shown in Figure 4.7 which is the height-range diagram for the low-level terraces from the upper braided reach. This fragment has been correlated by Young into his Group V terrace. The soil-stratigraphic data and historical evidence indicates that these four fragments comprise elements which are 1 000BP and 80BP years in age. However the height difference between the fragments is only about 0.5m. The long axes of the two most upstream fragments, 31 and 35, are about 50 degrees north of west, and therefore 60 degrees off the projection line. This factor will cause an oversteepening of the slopes of the terrace fragments by almost a third of their true value. However, the long axis of the downstream fragments, 33 and 34 is almost north/



HEIGHT - RANGE DIAGRAM of LOW LEVEL FRAGMENTS, FESHIE UPPER BRAIDED REACH, (after Young 1976)

Figure 4.7

south, and therefore experiences no oversteepening of the projected gradient. As a result of these variations in direction of the long axis of the terrace fragments, the upper two fragments on the height-range diagram have very steep slopes which cross fragment 18 and converge towards present river level, rising again downstream as the fragment slope flattens. A similar problem arises with fragment 27 on the height-range diagram. This is the equivalent to fragments 24, 32 and 5 on Figure 4.6. The long axis of the lower fragment is about 40 degrees off the projection line and as a result the gradient of the surface plunges. This fragment is correlated by Young with fragment 19 (Figure 4.6) which has been grouped as a 3,600BP surface in the soil-stratigraphic scheme. In the soil-stratigraphic scheme Young's fragment 27 has been grouped as comprising 1,000BP and 80BP units.

These examples show that the small height difference between the three terrace levels of the upper braided reach which range in age from 3,600BP to 80BP makes the correlation of fragments into surfaces on the basis of a height-range diagram extremely difficult. This difficulty is compounded by the nature of floodplain deposits in braided stream environments. The lateral migration of the stream across the valley floor and the tendency for bars to be stacked preferentially to one and then the other side of the active zone means that the long axis of individual terrace fragments is likely to vary considerably. A single projection line for all terraces is therefore likely to be inadequate, and to result in the projection of terrace slopes which artifically converge, diverge and cross. This difficulty may be compounded with slight slope differences for floodplain elements so that differentiation of fragments into surfaces on the basis of a height-range diagram becomes extremely difficult.

Finally, Young's two groups of low-level terraces, the group IV and group V terraces, are attributed by him to formation by meltwaters from the deglaciation following the Loch Lomond Stadial. They are thus suggested to have formed about 10,000BP. This inference is made from the height differences between the toe of the Allt Lorgaidh outwash fan and the low-level terraces. Young suggests that the low-level terraces merge with the toe of the fan. However, soil-stratigraphic data reveal that the Allt Lorgaidh fan is a complex fan, comparable with that in the Allt Gharbh Ghaig noted above. Three different age units may be clearly identified in both of these fans. The oldest unit in the Allt Lorgaidh fan is the upper unit which is the outwash fan dated at approximately 10,000BP (Chapter 3). The middle unit has a surface soil which correlates with the soils dated at 3,600BP, and the lower unit has surface soils which correlate with the soils estimated to have been developing for about 1,000 years (M.Robertson-Rintoul and V. Brazier in prep.). The low-level terraces discussed by Young in fact merge with the youngest unit of the fan.

Thus, although individual terrace surfaces may differ significantly in age, they may not differ significantly in height above present river level. Under these circumstances, where a few metres height difference may encompass several thousand years of incision, it may not be justifiable to correlate terrace fragments even approximately with dated land surfaces using the criteria of morphological continuity and height above present river level.

Comparison of the three different data bases for correlating terrace fragments, that is the soil-stratigraphic data, the morphometric data and the height-range data, suggests that a multivariate analysis of morphometric data will produce correlations of terrace fragments that are more reliable than those produced from a height-range analysis. However, it was not possible to differentiate between very young Holocene surfaces. Soil-stratigraphic data may, however, provide the necessary resolution to discriminate between surfaces of different age.

CHAPTER 5

A CRITICAL ASSESSMENT OF CURRENT PALAEOHYDROLOGICAL TECHNIQUES

5.1 Introduction

Fossil stream channels are frequently traceable over appreciable distances on the terraced surfaces of late Pleistocene and Holocene aggradation deposits. These channel traces have been used to assess qualitatively changes in the fluvial environment through time. For example, the existence of generations of palaeochannels on the terrace surfaces of valley fills in Poland (Froehlich _et al_., 1977; Kozarski and Rotnicki, 1983), and the south-west USA (Baker and Penteado-Orellana, 1973) have made it possible to determine some characteristics of the former river channels which in the past constructed the fossil floodplains which are the terraces. Qualitative evidence of periods of relatively higher and lower energy fluvial systems has been gained from the stream patterns on the terrace surfaces. Extensive braided palaeochannel systems have been associated with high stream powers, unstable sediments and cold, unvegetated environments. Changing channel configurations from multithread to singlethread traces are suggested to have accompanied the transition from a cold to a temperate climate (for example, Briggs and Gilbertson, 1980). These channel changes have been attributed to gradual reduction in surface runoff and river discharge. A late Holocene return to braiding for some channels in Poland has been attributed to increased mobilisation of perimeter sediments as a consequence of deforestation induced by developing industrialisation in the eightee nth century (Mycielska-Dowgiallo, 1977). A similar argument has been advanced for the channel metamorphosis on the

Afon Ystwyth over the past 200 years. Here braiding occurred as the stream adjusted to increases in sediment load as a result of mining activities in the catchment (Lewin et al., 1983).

Whilst such research permits qualitative comparison of changes in rates of river processes and hydrology between terraces, more precise comparison of the fluvial systems requires quantification of channel form. The development of statistical relationships between quantitative measures of channel planform and the controlling variables of channel pattern development may allow more precise statements to be made about the past hydrological regimes of the prior river systems.

In this chapter relationships currently used to reconstruct the past discharges of both meandering and braided channels are reviewed. The current palaeohydraulic method for reconstucting discharges for braided streams is then tested using field data from the upper braided reach of the River Feshie. The results are evaluated in the light of recent research concerning resistance to erosion of gravel clasts. The discriminant function approach to paleohydrology is then discussed and the method again tested against field data from the River Feshie.

5.2 The palaeohydrology of meandering channel traces

A large body of morphometric research has been concerned with the statistical correlation of parameters describing meander channel cross-sectional form and pattern at the between river scale with the prevailing flow conditions for contemporary channels assumed to be in equilbrium with the process regime. These relationships have been developed with a view to correlating channel pattern with the possible environmental controls of channel pattern, both as a means of investigating the active processes of meandering and as an aid to palaeoenvironmental reconstruction.

(a) Cross-sectional form

Quantitative measurement of cross-sectional channel form as well as of the sedimentary properties of palaeochannel infills, has been used to reconstruct palaeodischarges in fine-grained palaeomeandering channels. Schumm's study of two generations of well preserved palaeochannels of the Murrumbidgee River, New South Wales, is a classic illustration of the use of channel morphology to predict parameters of the flow which created the channels (Schumm, 1968). Schumm's method was based on geomorphic relations that he established between channel morphology, perimeter sediments, and velocity and discharge measures derived from present day streams. Schumm suggests that the morphology of the two sets of palaeochannels reflects changing hydrological regimes at the time when the respective sets of channels were formed. The older channel is wide, relatively shallow and filled with cross-bedded sands. The younger set of channels is narrower, deeper and possesses channel fill deposits of silt and clay. Estimates of velocity of the palaeoflows were obtained from the Manning equation

$$V = \frac{R^{0.66} \quad S^{0.5}}{n} \qquad 5.1$$

where

V = velocity

R = hydraulic radius

S = water surface slope

n = Manning's roughness coefficient

Values for depth, slope and roughness were all obtained from field data. Palaeodischarges were computed from the product of velocity and cross-sectional area.

More recently Knox (1985) has developed a set of empirical relationships from present day channels in the upper Mississippi Valley which relate bankfull cross-sectional area of meandering channels to the magnitude of the 1.58 year flood. Application of the relationship

$$\ln Q = 0.882 + 0.724\ (\ln CSA) \qquad r^2 = 0.89 \qquad 5.2$$

where

Q = magnitude of the 1.58 year flood

CSA = bankfull cross-sectional area of the channel

to channel capacities of relict channels enabled estimates to be made of the magnitude of 1.58 year floods of relict Holocene channels. The bankfull cross-sections of the channels were determined using morphological criteria originally developed by Knox for modern channels.

However, the preservation of complete, well defined channel sections and their subsequent exposure for palaeohydrological analysis appears to be relatively uncommon, not only at the scale of the individual cross-section but also in terms of observable continuity of the channel in a downstream direction (Starkel and Thornes, 1981). In a single random exposure it will be impossible to determine how representative the observed section is of longitudinal variability in cross-section form.

(b) Meandering channel pattern morphology

As an alternative, the preservation of fragmentary palaeomeander traces on terrace surfaces may permit quantification of changes in degree of meandering between palaeochannels preserved on terraces of differing age, and thus allow the prediction of past river discharges. This may be effected by using some of the relationships which have been developed between meander geometry and discharge. In reality, a brief review of the palaeohydrological literature which has used quantitative estimates of meander dimensions to predict palaeodischarges, illustrates the difficulties of using such relationships for fossil streams.

Dury (1965), concerned with questions of meandering valleys and underfit streams, related meander wavelength to bankfull discharge as

$$L = 54.3\,Q_b^{\,0.5} \qquad 5.3$$

where

L = meander wavelength

Q_b = bankfull discharge

and used this relationship to reconstruct the valley formative discharges from valley meanders in central and southern England. However, several problems in attempting to estimate discharges from palaeochannel form arise with the application of this equation. First, Schumm has demonstrated that meander wavelength is dependent on perimeter sedimentology as well as discharge, as

$$L = 618Q_b^{\,0.43}\,M^{-0.74} \qquad r^2 = 0.88 \qquad 5.4$$

where

M = weighted silt-clay index.

Longer bends occur in wide, shallow channels with a sandy load. Hack (1965), in a study of postglacial streams on the Ontonagon Plain, Michigan, showed that meander wavelengths are longer in bedrock reaches than in alluvial ones. Channel boundary materials are therefore likely to be an important parameter controlling meander dimensions. The situation is further complicated by the assessment of magnitude and frequency relationships. Tinkler (1971) argued that whilst alluvial channel meanders may be adjusted to discharges with a return interval of about 1.5 years, valley meanders cut in bedrock may be adjusted to flows with a return period of 10 to 50 years. Baker et al. (1975) also concluded that valley meanders may be adjusted to high-magnitude, low frequency floods. The morphological and sedimentological roles of different event magnitudes need to be considered if relationships between process and form can be used for palaeohydrological analysis.

This point is also made by Baker and Penteado-Orellana (1977). They use several planimetric properties of river meanders, but particularly meander wavelength and stream sinuosity, as scale variables of the past river systems of the Colorado River in central Texas. They have shown that each of 8 stages in the stepped evolution of the Colorado has a channel pattern characterised by a particular meander wavelength and sinuosity. Comparison of these variables for the 8 stages shows a complex pattern of changes in the fluvial system with large meander wavelengths in stages 1, 2, 3 and 5 and relatively smaller wavelengths in stages 4, 6, 6A and 6B. Baker and Penteado-Orellana suggest that sufficient morphology is preserved on the various terrace surfaces to utilise empirical correlations of meander wavelength and river discharge to predict past discharges. However, they note that attempts to correlate progressive changes in meander wavelength with changes in river discharge are complicated in two respects. First, by the changes in the amount and calibre of sediment delivered to the rivers in association with changes in surface run-off. Second, by flood-magnitude frequency changes through time. Such changes may remove the study cases too far from the class of rivers for which the empirical equations were originally derived.

More recently Williams (1983) has developed a set of empirical hydraulic geometry relationships relating the radius of curvature of meanders to discharge. Williams suggests that the use of radius of curvature as opposed to meander wavelength is preferable as a palaeohydrological tool because complete meander wavelengths are rarely preserved in the fossil landscape. The two relationships

$$Q_{max} = 0.28\ r^{1.38} \qquad r^2 = 0.73 \qquad 5.5$$

$$Q_m = 0.025\ r^{1.58} \qquad r^2 = 0.66 \qquad 5.6$$

where

r = average radius of curvature of a group of meander arcs

Q_{max} = average annual peak discharge

Q_m = mean annual discharge

are used to reconstruct Q_{max} and Q_m for 9 terrace elevations of the River Mora in Sweden. This analysis suggests that the oldest and youngest palaeorivers had relatively higher discharges than the systems which produced the intermediate level terraces.

However, meander wavelength is a function of both river discharge and the calibre of the perimeter sediments. Meander wavelength and radius of curvature are related as (Richards, 1982)

$$L = 4.59\ rc^{0.98} \qquad 5.7$$

where

rc = radius of curvature.

Radius of curvature is approximately equal to 2-3 times channel width, and is a channel dimension strongly related to perimeter sediments. Therefore again it would have been appropriate to add a sediment term to the relationship, despite Williams' contention that water discharge is the most important variable controlling the development of meanders.

Williams' arguments for employing radius of curvature highlight a further problem with attempting to use hydraulic geometry relationships between meander bend statistics and discharges to predict past river discharges. Estimates of meander dimensions are frequently made from single meander cut-offs or averages of a number of cut-offs (Rotnicki, 1983; Williams, 1983). This is because, by the nature of the process of cut-off development, these features tend to be well preserved in the fossil fluvial landscape. However, meander cut-offs may be indicative of increasingly unstable meander development (Quick, 1974). Where flow and bank material conditions are conducive to an increase in amplitude and tightness of bends, a threshold sinuosity may be reached which the river can no longer maintain, so a meander cut-off develops. Cut-offs may be a symptom of instability, a response to excessive sinuosity which lowers the channel gradient to a state where the stream cannot transport its load

(Knighton, 1984). Application of relationships between bend statistics and discharge developed for equilbrium streams from data at the between-river scale may therefore give spurious estimates of palaeodischarges if applied to isolated meander bends or to cut-offs. Rather, reconstructions of palaeo-discharges should be made from geometrical parameters of the channel pattern which are derived from channel traces which incorporate a reach of river spanning at least several meanders. Furthermore, because empirical relationships developed between bend statistics and discharge show scatter, estimates of palaeodischarges should be assigned confidence limits.

An alternative approach to the analysis of meandering palaeo-channel traces has been offered by Ferguson (1977a). Ferguson (1975, 1979) considers that meandering, an environmentally modified deterministic process, can be characterised by meander wavelength, sinuosity and degree of irregularity. All three characteristics can be estimated from the analysis of direction or curvature series. The variance of the direction series may be used to estimate channel sinuosity since a close theoretical relationship can be demonstrated between these measures. The value of the approach is that direction variance can be calculated even for a small and scattered sample of palaeochannel remnants as long as they are representative of a sufficiently long reach to enable a value to be estimated for the average or quasi-equilbrium channel. From the direction variance the sinuosity of a meandering palaeochannel may thus be predicted even from discontinuous channel traces. Confidence limits may be attached to the estimated sinuosity by assuming that the measured directions from the palaeochannel traces are a random sample from a normal population. However, this model has not been expanded to enable parameters of the palaeoflow to be estimated from the predicted sinuosity.

5.3 The palaeohydraulic method of discharge reconstruction for gravel-bed channels

The palaeohydrological methods described above are not applicable to coarse-grained braided streams and no methods currently exist which are based on gravel-bed braided stream morphometric variables. Several workers have therefore attempted to develop an alternative methodology for reconstructing parameters of the palaeoflows of gravel-bed braided streams. These methods involve the estimation, from clast dimensions, of local depths and flow velocities assumed to be responsible for clast deposition. The method essentially represents a transformed measurement of particle size (Church, 1972). These methods have also been evaluated in the context of bedrock canyon flashflood deposits (Bradley and Mears, 1980; Costa, 1983). Palaeoflow estimates of past braided streams employing this methodology have been widely used for a large number of braided terrace systems in southern England (Cheetham, 1976, 1980; Clarke and Dixon, 1981; Briggs, 1983), the River North Esk in Scotland (Maizels, 1983a, 1983b), Greenland (Maizels, 1983b), Southern Norway (Maizels, 1983b), New Zealand (Maizels, pers. comm), and Baffin Island, Canada (Church, 1978). There have been several reviews of the methodology and some discussion of the assumptions underlying the use of the hydraulic formulae used for the computations (for example, Church, 1978; Maizels, 1983c). It is notable however that little attempt has been made to test the formulae used in this method with data from modern gravel-bed braided streams.

The success of predicting discharge using an approach based on grain size measurement depends on the ability to relate initiation of motion of gravel particles on the bed to the characteristics of the flow responsible for gravel motion and on an estimate of prior channel width. The method is justified only if there exists a well defined relationship between stream competence and shear stress or velocity, if the particle size measured sufficiently determines flow resistance (Church, 1978), and if former channel width is known. Established equations which predict the threshold value of flow intensity for particle movement may then be applied to the surface layer of clasts on terrace deposits to reconstruct the flow intensity necessary to initiate their motion. Estimates of flow characterisics derived from palaeohydaulic reconstruction are meant, therefore, to apply to flow conditions at the threshold of sediment transport, so that the flood estimates made should relate to the onset of gravel movement for particular clast sizes.

The initiation of particle motion depends on the supply of different size grades and on the hydraulics of flow within the channel. Bed velocity, mean velocity, bed shear stress and stream power have all been used as the flow intensity parameters for predicting initiation of motion of gravel particles. Predictions from coarse-grained particle size data of the velocity needed to initiate motion of particles of a given size have been made from both theoretically derived and empirically derived relationships. Thus, Helley's (1969) deterministic model, which included the effects of particle size, shape and orientation, predicts incipient motion of ellipsoidal pebbles using the principle of moments to define drag force. Solving for bed velocity Helley's formula has been used by both Bradley and Mears (1980) and Costa (1983) to predict bed velocity at the stage of incipient motion of boulders in bedrock gorge flashfloods. Empirical relationships have been derived from regression analyses of observed velocities for particle entrainment and from data on riprap stability (Novak, 1973; Bradley and Mears, 1980; Costa, 1983; Williams, 1983). Costa (1983) has applied the relationship

$$v = 0.20d^{0.455} \qquad r^2 = 0.71 \qquad 5.8$$

where

v = mean velocity in m s^{-1}

d = particle diameter (D_{90} , D_{95} , or the single largest moved under given flow conditions)

which predicts the mean velocity required to initiate motion of large bed particles. The relationship in equation 5.8 was derived from observations of particle movement and associated measurement of velocity. Costa also used an equation for estimating the limiting size needed for riprap stability. This is given by

$$v_b = 5.9\ d^{0.5} \qquad 5.9$$

where

v_b = bed velocity

d = particle diameter.

Bed velocity is multiplied by 1.2 to obtain an estimate of average velocity.

Both Bradley and Mears (1980) and Costa (1983), have used a range of such theoretical and empirical relationships to estimate the velocities needed to move boulder deposits in flash floods. Bradley and Mears found that, for all equations used and between extreme values, there exists a broad band of prediction overlap. For example, velocities needed to initiate motion of a particle 188cm in diameter predicted from six equations were estimated to range from 4.6 to 7.6 m s^{-1}. Similar results were achieved by Costa in predicting erosion velocities for particles in flashfloods in Colorado. Costa agrees with Bradley and Mears in considering that the use of several equations provides results with sufficient consistency to be useful.

Techniques for predicting discharge in gravel-bed braided streams are generally based on an estimate of critical bed shear stress, that is the shear stress required to initiate motion of a given clast size. Four basic steps are involved in arriving at an estimate of total discharge for gravel-bed braided palaeostreams. These are :-

(1) To establish the hydraulic properties of the flow (shear stress) at the threshold of erosion for a given clast size;
(2) To estimate the depth of flow given a predicted value for shear stress;
(3) To estimate the mean velocity of flow for the predicted flow depth;
(4) To predict a total discharge by multiplying the depth-velocity product, the specific discharge, by an estimate of flow width.

Shields' criterion has been used in most palaeohydrological studies of gravel-bed braided streams either for estimating critical bed shear stress or for predicting the flow depth at which initiation of particle motion should take place (Baker, 1974; Cheetham, 1976; Church, 1978; Maizels, 1983a, 1983b; Briggs, 1983). In the Shields relationship a dimensionless critical shear stress is related to the ratio of the critical mean bed shear stress to the weight per unit area of a single layer of submerged grains; this may be interpreted as the ratio between two forces per unit area, the drag force of the flow acting acting on a unit area of bed and the gravity forces that are opposed to the drag forces acting on the surface layer of grains. The Shields function relates the dimensionless critical shear stress Θ, to the particle shear velocity Reynolds number, $11.6\ D/\delta$, where D is depth and δ is depth of the laminar sub-layer. The relationship between Θ and particle Reynolds number yields the Shields entrainment function (Figure 5.1). The value for the latter has been considered constant for coarse bed particles at high particle Reynolds numbers, with Θ assuming a value of 0.056. Rearrangement of the Shields relationship allows prediction of mean critical bed shear stress from

$$\tau_c = 0.056\ (\gamma_s - \gamma_f)\ D \tag{5.10}$$

where

τ_c = mean critical bed shear stress

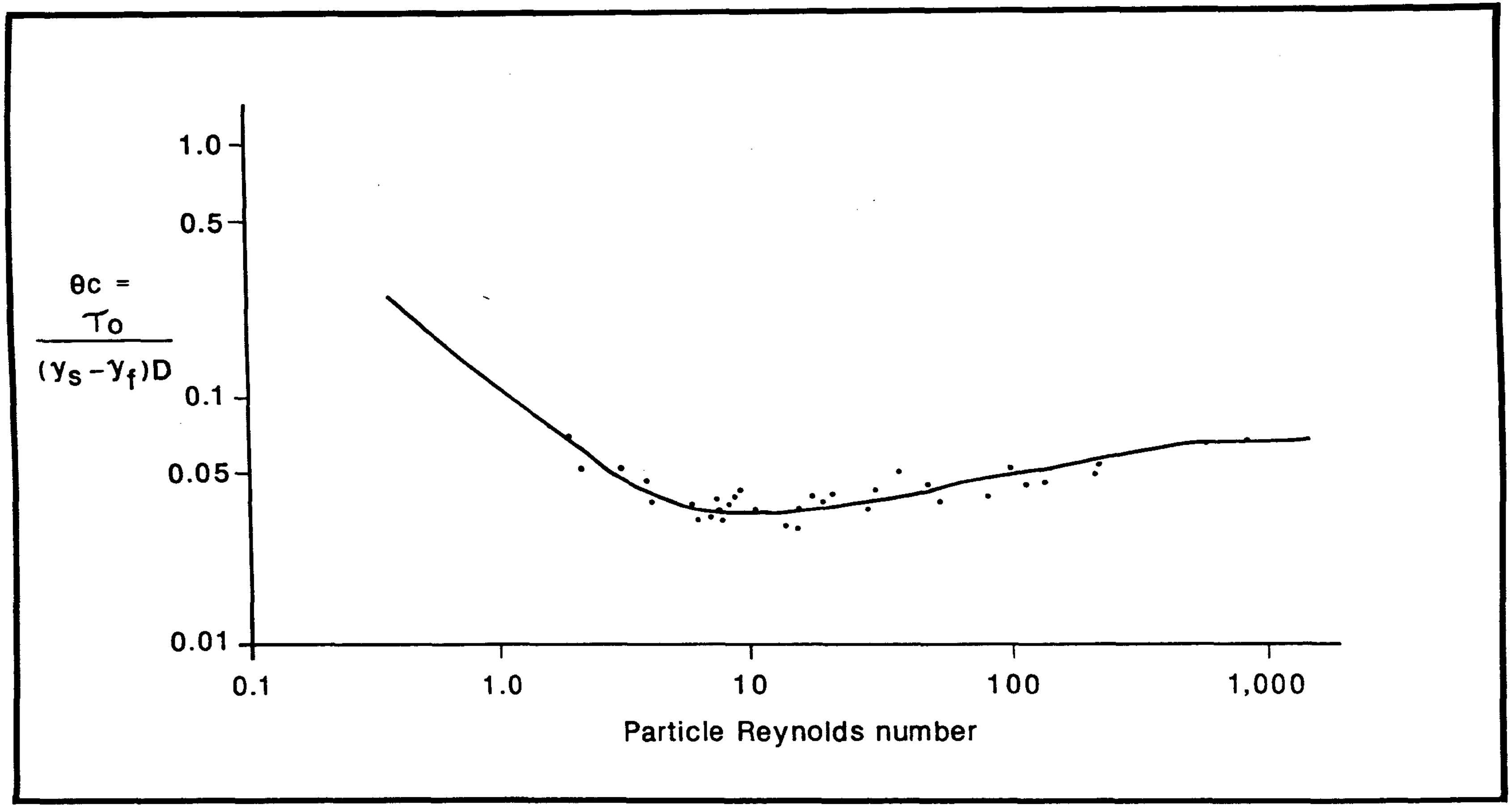
D = particle diameter

γ_s = unit weight of sediment

γ_f = unit weight of fluid.

For terrace deposits, D in equation 5.10 is obtained from particle size analysis of the deposits comprising the fossil braided channel systems under consideration. An estimate of D was made by Church (1978) by taking the mean of the five largest particles in the gravel unit under consideration. Baker used the mean of the largest 10 particles. Briggs (1983) and Maizels (1983a, 1983b) used D_{84}.

However, the value of Shields Θ may vary with grain geometry. Church (1978) has shown that Shields Θ may vary between the



The Shields diagram (after Vanoni, 1964)

Figure 5.1

limits of 0.02 and 0.10 for varying grain geometries in natural gravel-bed rivers. Values of 0.02 are appropriate for overloose grain geometries where packing is completely open. Conversely, where clasts are imbricated or underloose, Shields Θ may increase up to 0.10. Values of 0.056 may be appropriate for 'normal' sediments sitting in a non-dilated state, without surface imbrication.

Once the critical bed shear stress necessary to initiate motion of a particular clast size has been calculated, the flow depth associated with that shear stress is estimated by substituting the equation for mean bed shear stress

$$\tau_c = \gamma_f Rs \qquad 5.11$$

where

τ_c = mean critical bed shear stress

γ_f = unit weight of fluid

R = hydraulic radius

s = energy slope

into equation 5.10. This then becomes

$$\frac{\tau_c}{(\gamma_s - \gamma_f)\, D} = \frac{\gamma_f\, ds}{(\gamma_s - \gamma_f)\, D} = 0.056$$

$$\Rightarrow \quad d = \frac{0.056 \times 1.65\, D}{s} \qquad 5.12$$

where

d = depth of flow.

The value for the slope term is generally estimated from the slope of the terrace surface. Mean velocity of the flow at the estimated depth may be calculated from either the Manning equation, (equation 5.1), or by rearranging the Darcy-Weisbach equation as

$$v = 8.86(f)^{\frac{1}{2}} (ds)^{\frac{1}{2}} \qquad 5.13$$

where

v = velocity in m s^{-1}

d = depth in metres

s = bed slope

f = Darcy-Weisbach friction factor

The coefficient n for equation 5.1 may be calculated from the Limerinos equation

$$n = \frac{0.113\ R^{0.166}}{1.16 + 2.0 \log (R/D_{84})} \qquad 5.14$$

where

R = hydraulic radius

D_{84} = 84th. percentile (% finer than)

The friction factor in equation 5.13 is estimated from the Limerinos relationship (Richards, 1982)

$$\frac{1}{(f^{\frac{1}{2}})} = 1.16 + 2.0 \log (R/D_{84}) \qquad 5.15$$

Specific discharge is calculated from the depth-velocity product.

If some means of assessing stream width exists then total discharge across the section may be estimated. Church (1978) suggests two methods for estimating the width of the former active zone of a braided reach. First, by measuring the width of the extant channel traces on the terrace surfaces. Second, by assuming that the scaling of the total width with discharge of the terrace channels was comparable to that of the present day streams. The relationship between present day width and discharge can then be used to predict a width term for the past stream. Either way total discharge is calculated by multiplying specific discharge with total estimated width. Maizels (1983a, 1983b) suggests an approach similar to the first approach of Church, that the total number and mean width of the braided palaeochannels may be predicted from remnants of the original surface.

However, even if well defined braided palaeochannels have been preserved on terrace surfaces it is extremely uncertain as to the extent that the original active channel zone has been preserved by the terrace remnant. As Maizels (1983a) notes, the sandur deposits of the River North Esk have been reworked to form four successively lower terrace systems. Church (1972) also notes that reworking of higher terraces in Baffin Island sandurs supplies much of the sediment for successively lower terraces. Similarly, the River Feshie is actively reworking both glaciofluvial sedimentary landforms and the lower Holocene terrace deposits (Chapter 2). Mapping in the upper Feshie (Chapter 3) has demonstrated that in under 100 years the area of the Holocene terrace fragments between the upper Feshie and the Allt Lorgaidh has been reduced by over 50% in comparisonto the original area.

Unless under exceptional circumstances (for example, Claque, 1975), the total width of the former active channel zone of a braided network is therefore unlikely to be represented by the width of the terrace remnant. Lateral erosion and reworking of older, higher deposits by successively younger braided systems as the channel incises through valley fill deposits may result in an unknown area of each preceding older terrace level being removed. Recognising the problem of the unknown width of the active channel zone, Briggs (1983) attempted to predict total channel width of the braided system which built the Floodplain terrace deposits of the upper Thames using the relation developed by Schumm (1960)

$$w/d = 255M^{-1.08} \qquad 5.16$$

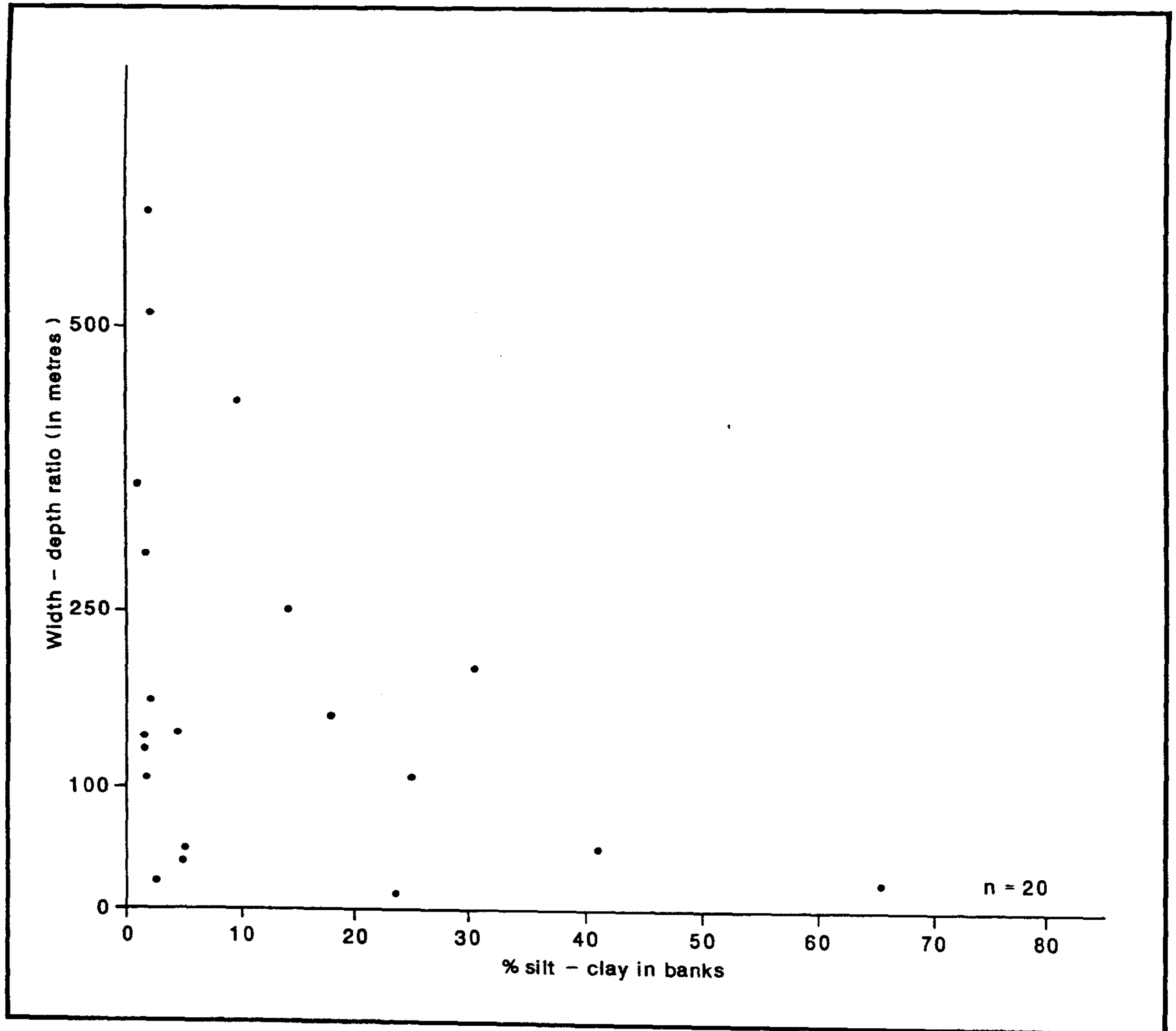
where

w/d = width:depth ratio

M = % silt-clay in the perimeter sediments

Depth was previously estimated from an inversion of the Manning equation. However, Schumm's empirical relationship was developed for 90 river cross-sections of streams flowing in fine-grained alluvium in a semi-arid region of the U.S.A. It is therefore unwise to apply such a relationship to periglacial, gravelly braided streams. To illustrate this point, a data set of width:depth ratios and % bank silt-clay content was compiled from 20 gravelly braided and meandering streams in New Zealand. Figure 5.2 shows a wide scatter of points over the silt-clay/width-depth ratio plane. Thus a bank silt-clay content of 2% is associated with width:depth ratios which range from 112 to 596, whilst width-depth ratios of 14 range in bank silt-clay content from 2.9% to 65%. The estimation of a reasonable channel width for the prior river channel under consideration remains one of the most serious problems facing the current method for palaeohydraulic reconstruction of past discharges.

5.4 A field test of the currently used palaeohydraulic approach to discharge estimation



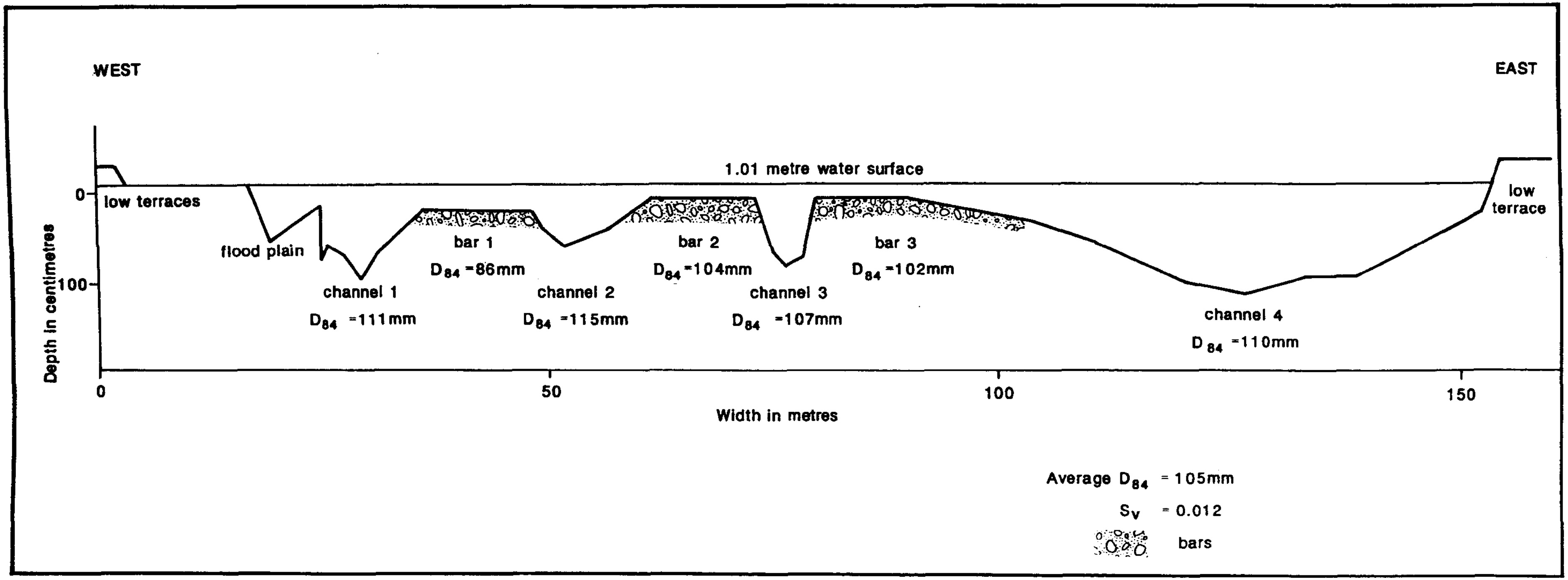
Plot of % silt - clay in banks against width - depth ratio for a sample of 20 gravel-bed rivers - data from Mosley,unpublished.

Figure 5.2

A comparison of field measurements with calculations based on existing equations for predicting the threshold of erosion, flow depths and velocities may be used to appraise existing palaeo-hydrological methods based on grain size transformations. This was made by attempting to estimate the discharge at which initiation of motion of gravel particles will take place in the upper braided reach of the River Feshie and comparing the results with field data of present discharges and sediment transport.

The cross-section of the upper braided reach used for the analysis (Plate 2.2) was surveyed as accurately as possible using a quick-set level and staff. The section was divided into bar and channel segments (Figure 5.3). The b-axis of 100 clasts was sampled from the surface layer of particles of each unit using the Wolman method (Wolman, 1954). From the gravel count data cumulative percentages of bed material sizes were calculated. Bed material size is expressed as D_{84} (% finer) in millimetres. The average D_{84} for the 8 units was 105mm (Figure 5.3). The channel beds and bar surfaces comprised a cobble and gravel-sized clast supported population with a smaller secondary population of finer material infilling the void spaces of the framework on the bars. Surface imbrication was universal on both channel beds and bar tops. Following the terminology of Church (1978) the sediments of both channel and bars would be termed underloose. The valley slope of the reach was 0.012. Bedslope was surveyed for each of the 4 channels.

The critical shear stress needed to initiate motion of particles of D_{84} size was estimated from equation 5.10. As the River Feshie channel and bar gravel is underloose a value for Shields' Θ from the range 0.07 - 0.10 should be used for equation 5.10



Cross - profile of active channel zone - upper braided reach

Figure 5.3

(Church, 1978). Calculations were made using a value of 0.07 for Shields' Θ. Substituting the appropriate values for Θ into equation 5.10 and using 2650 kgf m^{-2} (kgf m^{-2} = Kilogrammes of force per square metre) for grain density, 1000 kgf m^{-2} for fluid density, and 9.81 m s^{-1} for acceleration due to gravity the Shields' relation simplifies to

$$\tau_c = 115.5D \text{ (kgf m}^{-2}\text{)} \qquad 5.17$$

where

τ_c = mean critical bed shear stress

D = D_{84} in metres

For a clast size of 105mm this gives a predicted shear stress required for the threshold of motion of 12.13 kgf m^{-2}. Having calculated the critical bed shear stress required to initiate motion of this clast size, the flow depth associated with that shear stress is estimated from equation 5.12. The majority of palaeohydrological studies of braided streams (for example, Church 1978, Baker 1974) calculate only one flow depth from one measurement of clast size, so the same procedure was followed in this text. This gives a predicted flow depth for the River Feshie of 1.01m. Mean velocity is predicted from equation 5.13 with the friction factor estimated from equation 5.15. The predicted mean velocity for a depth of 1.01m and a slope of 0.012 is 3.1 m s^{-1}. The specific discharge, the depth-velocity product, is 3.13 m^3 s^{-1}.

To complete the estimates of discharge at the stage of incipient motion in the upper braided reach of the Feshie, an estimate of channel width was made by taking the depth estimate from the previous calculations, 1.01m, finding the minimum or deepest point in the main channels, drawing a perpendicular 1.01m in length from the points and drawing a water surface orthogonal to the perpendicular, across the section (see Figure 5.3). This procedure was adopted because it is likely to give the best estimate for channel width at the predicted flow depth, therefore giving the palaeohydrological method the fairest possible test. This method gives a flow width of 135m. The final discharge estimate was calculated from the product of specific discharge and total channel width. The predicted discharge is 422.5 m^3s^{-1} for the threshold of erosion of gravel particles of 105mm in size in an imbricated state.

Stage and discharge data and information on incipient motion of gravel and sediment transport for the upper braided reach of the River Feshie have been provided by Dr. A. Werritty and Dr. R. Ferguson (personal communication). The stage data were obtained by the University of St. Andrews water level recorder, and converted to discharge by a stage-discharge relationship calibrated by current meter measurements. The recorder was installed in a partially confined section of the stream immediately downstream of the upper braided reach. The period of gauging has been from 1978-1985. Although the gauging period has not been long enough to produce a set of representative flood magnitude statistics, estimates for the 1.5 year flood and the mean annual flood are available. The flood with a recurrence interval of 2-2.3 years, the mean annual flood, is estimated to be 80-90 m^3s^{-1}. The 1.5 year flood probably lies between 40-60 m^3s^{-1}.

The flood record for the River Feshie may be extended back to 1951 by considering the gauging data from the station operated from 1951-1974 by the Dept. of Agriculture and Fisheries for Scotland at Feshiebridge, near the mouth of the Feshie catchment. In September, 1961, a peak discharge of 200 m^3s^{-1} was recorded. This represented the largest event over the whole gauging record. A second peak discharge of 181 m^3s^{-1} was recorded at Feshiebridge in 1969. Approximate values for these peak events at the study cross-section can be estimated by calculating the specific discharges and multiplying this by the catchment area of the upper braided reach. This estimate will give only approximate values for specific discharge as specific discharge declines with increasing basin area. Upstream specific discharge would be higher and flood peaks would probably be larger. Specific discharges of 0.83 m^3 sec^{-1} km^{-2} for the 200 m^3s^{-1} flood and 0.75 m^3 sec^{-1} km^{-2} for the 181 m^3s^{-1} flood were calculated giving minimum flood peaks of 92 m^3s^{-1} and 83 m^3s^{-1} at the upper braided reach.

It has been observed that at discharges of about 14 m^3s^{-1} selective sediment transport occurs through the upper braided reach. At discharges of 20 m^3s^{-1} appreciable transport of gravel takes place. At discharges equivalent to the 1.5 year flood the threshold of gravel motion is approached over much of the bed, although there is spatial variation in competence. This spatial variation is probably linked to areas of converging and diverging flow due to the presence of large-scale bedforms (Ferguson and Werritty, 1983).

Indirect evidence of substantial sediment transport during floods of 70-100 m^3s^{-1} is also available (Werritty and Ferguson, 1980; Ferguson and Werritty, 1983). Detailed surveying of closely spaced cross-sections in the upper braided reach before and after these major flood events has revealed channel incision and infill that can locally exceed 1m in depth in an active channel zone whose total bed relief is about 1.5m.

Overbank deposition of extensive cobble sheets on segments of floodplain also occurs after high magnitude 100 m^3s^{-1} floods. During flood events approximately equal to the mean annual flood, the progressive aggradation and migration of major bars, usually occurring in association with bank erosion often exceeding several metres, further demonstrates the efficacy of sediment transport.

Substantial sediment transport in the present day upper braided reach of the River Feshie thus takes place during the flood which is approximately equal to the mean annual flood. Incipient motion of gravel takes place for selected particle sizes at discharges not much more than 14 m^3 s^{-1}, whilst the threshold for erosion over much of the bed occurs with floods of 40-60 m^3s^{-1}. At this discharge most size ranges of particles are moved.

The application of the palaeohydraulic method to the River Feshie cross-section was conservative both in the estimation of the flow depth over the bars and channels and in using a known value for the width of flow for the defined stage. However, the palaeohydraulic method applied to the present day channel gives an order of magnitude over-estimate for the discharge at which initiation of motion of gravel particles of a D_{84} size takes place. The palaeohydrological method based on a transformed grain size is meant to predict discharges at the stage of incipient gravel motion for a specified clast size. The palaeohydraulic method of discharge reconstruction estimated a discharge of 422m^3s^{-1} for incipient motion of clasts of D_{84} size in the upper braided reach. This discharge estimate is far in excess of the largest flood on record since 1951 and does not have a known recurrence interval. In the upper braided reach

D_{84} sized particles are moved at discharges of about 40-60m^3 s^{-1}. Church (1978) and Maizels (1983) quote likely errors of up to an order of magnitude for the final discharge estimates for palaeostreams using the palaeohydraulic methods described above. However, the assessment of order of magnitude errors is based mainly on variation of Shields'θ. In reality, when predicting discharges for palaeostreams, errors may be larger than an order of magnitude.

5.5 Discussion

Application of the currently used palaeohydrological methodology to a cross-section of the upper braided reach of the River Feshie has resulted in an over-estimate of the discharge required for incipient motion of gravel of D_{84} size. Several factors have probably contributed to this over-estimate. First, the critical shear stresses needed to initiate motion in coarse gravel may diverge considerably from values predicted by the Shields' criterion (Baker and Ritter, 1975). Plotting the Shields' criterion on a regression of shear stress against particle size Baker and Ritter showed that data points representing deep flows plot to the right of the Shields' criterion whilst those data points representing shallow, extremely turbulent flows plot to the left. Particle motion was predicted at much lower shear stresses than would be predicted by the Shields equation for the latter type of flows. Carling (1983), examining the threshold of coarse sediment transport in Great Egglesthope Beck, a broad, shallow relatively staight cobble-bed stream in the north Pennines, has plotted τ_oc against D_5. This relationship

$$D5 = 18.28\ \tau_oc^{\ 1.58} \qquad r^2 = 0.57 \quad 5.18$$

where

τ_{oc} = total shear stress

D_5 = the mean of the five largest clasts measured

differs from the Baker and Ritter relationship

$$D_{max} = 65T_c 0.54 \qquad r^2 = 0.85 \quad 5.19$$

where

D_{max} = D_{90}

T_c = critical bed shear stress

in both slope and intercept coefficient. Carling's data agrees with that of Baker and Ritter in that for shallow flows the shear stress required to initiate motion of any grain size is less than that predicted by the Shields' criterion. However, differences in the slopes of the two regressions indicate that considerably higher values of shear stress are required to initiate motion in Carling's stream than in the sample of Baker and Ritter. Grain geometry may be an important factor in contributing to variation in the parameters of the two equations (Carling, 1983). The bed sediments of Great Egglesthorpe Beck are compacted. This will affect both grain geometry and the relative protrusion of individual particles from the stream bed into the flow. Recent experimental work has demonstrated the importance of grain geometry and relative protrusion of individual grains from the bed on particle mobility. Fenton and Abbot (1977) observed a decline in Shields' θ as the relative protrusion of particles increased in fully turbulent flows of constant depth, θ being reduced from 0.04 to 0.01 as relative protrusion increased. Particle mobility is also influenced by factors such as bed relief and sediment packing (Laronne and Carson, 1976).

The effect of the variables encompassed in grain geometry on particle mobility will be compounded when attempts are made to apply threshold criterion to coarse, heterogenous sediments, and particularly when extended to the highly complex bed of a braided stream. Difficulties of assessing a representative value for Θ over a naturally sorted gravel-bed river may mean that estimates for Θ may be either over-estimated or under-estimated depending on the bar/channel unit of the cross-section under consideration. Andrews (1983), for example, considering the effect of nonuniform bed material on the value of Θ, suggests that most palaeohydraulic studies over-estimate the value of Θ. This means that flow depths will be overestimated thus leading to an even greater error in the estimated flood discharge as discharge is approximately proportional to the 2.5 power of flow depth.

Second, particle entrainment in gravel-bed streams is a stochastic process involving the random effects of turbulent fluctuations on a spatially varied, non-cohesive stream bed (Baker and Ritter, 1975). Kalinske (1947) estimated that normally distributed instantaneous velocities may be up to 2-3 times the mean bed velocity for about 5% of the time. Instantaneous drag and lift forces were shown to be 4-9 times the mean value. Cheetham (1979) devised a technique for measuring point bed-shear stresses in a small gravel-bed braided stream in Norway. He demonstrated that instantaneous critical bed shear stresses are up to 10 times greater than the unit bed shear stresses that would be predicted from equation 5.11 for mean bed shear stress. This is because of the combined effects of lift and drag forces that are induced by instantaneous turbulent velocity fluctuations. Mean drag and lift forces are thus less than instantaneous peak values. Entrainment at the bed depends not on mean values for drag and lift but on instantaneous values which cannot be predicted from equations such as Shields' or that for mean bed shear stress.

Third, there is growing literature recognising the existence of small and medium scale cluster bedforms, and their importance in controlling the mobility and availability for erosion of individual clasts (Brayshaw, 1985). Field experiments have demonstrated that 87% of particles are eroded from a plane bed during flood flows, in comparison to 46% from a bed characterised by cluster bedforms.

The ability of threshold equations to predict the threshold of erosion for coarse particles on natural stream beds may therefore be extremely limited whilst the application of empirical equations such as that of Baker and Ritter (1975) may be as difficult in a palaeohydrological context as is the application of the Shields' criterion. This is because the parameters of the equations are likely to vary between data sets, with variation in grain geometry, and flow conditions.

The experimental work of Cheetham (1979) highlights a further difficulty in attempting to apply competence relationships such as Shields' criterion to gravel-bed rivers, and more particularly to braided streams. Cheetham suggests that a competence threshold exists which is defined by channel geometry and stage. Beyond a limiting width at constant discharge, competence decreases because of increased frictional losses as relative roughness increases towards the edges of the wide, shallow channels. In channels with high width-depth ratios competence is stage-dependent, loss of competence occurring as excessively wide cross-sections develop on rising stage. In channels with a low width-depth ratio, that is, narrow gravel-bed streams, the effects of wall friction causes threshold values of boundary shear stress to rise in comparison to broad gravel-bed streams (Carling, 1983). Spatial variation of competence over a cross-section and between successive cross-sections is therefore likely to be appreciable. A single estimate of critical shear stress over a braided stream cross-section is not likely to be representative of the average conditions over the whole bed.

Palaeohydraulic reconstruction such as that made for the upper braided reach of the River Feshie is based on estimates of depth, friction factors and velocities which essentially represent a transformed grain size. This methodology assumes that skin resistance is the dominant form of resistance in the braided reach (Church, 1978; Maizels, 1983; Briggs, 1983), and that form resistance may be neglected. However, this assumption may not be reasonable. Church and Jones (1982) suggest that barforms in gravel-bed braided streams constitute important form resistance elements interacting with and conditioning the flow pattern. Such barforms have heights comparable to bankfull channel depth (Bridge, 1985) and are likely to persist even in high magnitude floods. Thus, the gravel barforms of the Knik River, Alaska, are not drowned out at high peak floods (Bradley <u>et. al.</u>, 1972). Prestagaard (1983) has demonstrated empirically that both particle size and bedform roughness constitutes the total channel resistance for a sample of straight and braided gravel-bed channels at bankfull stage. Bedforms were shown to interact with the flow pattern at the reach level, thus supporting the suggestion of Church and Jones that bedforms are important flow resistance elements. This was also demonstrated through the work of Cheetham (1979) who showed a marked spatial variation of competence as a result of the presence of gravel barforms.

For the River Feshie cross-section used in the field test the presence of large-scale bedforms causes significant variation in channel roughness, flow depth and channel cross-sectional area through the section. Relative roughness increases with increasing bar size, and decreasing depth of flow as bar tops are inundated during floods. Significant changes in relative roughness and depth across the section will produce corresponding changes in shear stresses so that a single estimate of shear stress, and consequently depth, for one cross-section, is inappropriate.

Maizels (1983a, 1983b) has attempted to overcome the latter source of error by distinguishing between in-channel flow and bar top flow, suggesting that a component method for a slope/area estimate of discharge should be used. Estimates of relative roughness, friction factors, critical mean depth, velocity and discharge should be made for each bar and channel unit. Total discharge is then determined from either the product of mean discharge of each channel and bar multiplied by the total number of bars and channels, or from the product of specific discharge per unit width and the total width for each type of flow.

Maizels makes separate calculations of shear stress, water surface elevation, relative roughness and velocity for each set of bars and channels in isolation. The result is a stepped water surface profile across the section with bar top discharges which exceed those of channel flow discharges, largely because of high flow depths reconstructed across the steeply graded bar-tops. This is clearly an unrealistic basis for estimation of total section discharge. Discharge estimates could be better made by defining a series of water surface elevations for the cross-section on morphological grounds. The discharge through each bar/channel segment could then be calculated for each water surface elevation and component discharges summed across the section (Chapter 8).

Finally, maximum sediment availability is assumed in all palaeohydrological calculations, that is, it is assumed that all sizes of sediment are available for transport. The maximum particle sizes present in the deposit are assumed to represent stream competence rather than the lack of any larger particles in the source region (Church, 1978; Koster, 1978). If larger sizes could have been transported but were not available estimates of flow intensity may represent some unknown minimum value. Maizels (1983c) suggests that this assumption of maximum sediment availability may be acceptable in most glacial and periglacial environments. However, maximum particle size will depend on the availability and size range of the particles from the source areas. Only a certain size range of particles may be available for transport, so that particle size may not necessarily be successful in differentiating between different degrees of high energy environments transporting gravel and boulder-sized bedload. Rather maximum particle size measured may represent spatial variation in particle sizes available for transport due to the heterogenous nature of the valley fill deposits.

Testing the current methodology for palaeohydrological reconstructions based on a transformed grain size has given an order of magnitude error for the discharge prediction at the stage of incipient gravel motion for a cross-section of the upper braided reach of the River Feshie. It was suggested earlier that this palaeohydraulic methodology is justified only if there exists a well-defined relationship between stream competence and shear stress, and if the measured particle size sufficiently determines flow resistance. These two assumptions however, may not be justified. Significant sources of error in prediction may arise from:

(1) entrainment at the bed depends not on mean values for drag and lift but on instantaneous values which cannot be predicted from equations such as Shields' or that for mean bed shear stress;

(2) the highly complex bed of a gravel bed stream interacts with the flow and causes marked spatial variation in competence which cannot be accounted for using the current method of palaeohydrological reconstruction;

(3) the current method of investigation assumes the dominance of skin resistance which may not be reasonable in wide shallow cross-sections of braided streams.

Current methods of palaeohydraulic reconstruction cannot identify the frequency of predicted discharge. Furthermore, the method predicts only local depths and velocities. The extrapolation of these point estimates to averages for wide, shallow, and morphologically complex cross-sections such as are found in braided streams is uncertain. These difficulties coupled with the uncertainity of the prior channel system width introduce serious doubts concerning the efficacy of this methodology for estimation of palaeoflow data for past gravel-bed braided streams.

5.6 The discriminant function approach to palaeohydrology

An alternative approach to the estimation of a minimum "bankfull" discharge for braided palaeochannel networks on terrace fragments has been used by several workers (Cant and Walker, 1976; Cheetham, 1976, 1980; Clarke and Dixon, 1981; Briggs, 1983). This is the discriminant function approach to palaeohydrology of braided streams. In 1957, Leopold and Wolman (1957) demonstrated that meandering and braided streams could be separated by the discriminant function

$$s = 0.013 Q_b^{-0.44} \qquad 5.20$$

where

s = channel slope

Q_b = bankfull discharge in m^3s^{-1} .

For a given slope braided channels have a higher discharge than meandering channels. This relationship may then be used to estimate the minimum "bankfull" discharge needed for braiding given a value for the slope term. Taking the slope of the terrace, the Leopold and Wolman relationship has been used to predict minimum "bankfull" discharges for braided networks on terrace fragment surfaces.

Since slope and discharge are the two variable quantities of stream power, slope-discharge plots such as that of Leopold and Wolman (Figure 5.4) imply the exceedance of an energy threshold for braiding. Thus more powerful rivers braid and less powerful rivers meander. However, Ferguson (1984) has shown that the Leopold and Wolman line corresponds to a threshold stream power of about 50 Wm^{-2} , which is well within the stream power range commonly found in both meandering and braided rivers. Analysing a data set of gravel-bed rivers Ferguson (1984) demonstrated that the critical stream power needed for braiding is at least three times that implied by the Leopold and Wolman plot. A threshold equation of the same general form as the Leopold and Wolman function provides a better discrimination of meandering and braiding

$$s = 0.056Q_b^{-0.5} \tag{5.21}$$

However, since bed material transport and bar formation is necessary in both meander and braid development, the threshold between patterns should also relate to bedload calibre (Richards, 1982). Henderson (1961) re-analysed Leopold and Wolman's data to devise an expression of the form

$$s = 0.002D_{50}^{1.15} Q_b^{-0.46} \tag{5.22}$$

where

D_{50} = median grain size in mm

which shows that a higher threshold slope is necessary for braiding with coarser bed material. This is shown in Figure 5.4 where lines of constant grain size have been drawn on the Leopold and Wolman plot. More recently, Ferguson (1984) produced the multivariate discriminant function

$$s = 0.017Q_b^{-0.49} D_{90}^{0.27} \tag{5.23}$$

where

D_{90} = the 90th. percentile particle intermediate axis

which shows that a greater stream power is needed for braiding

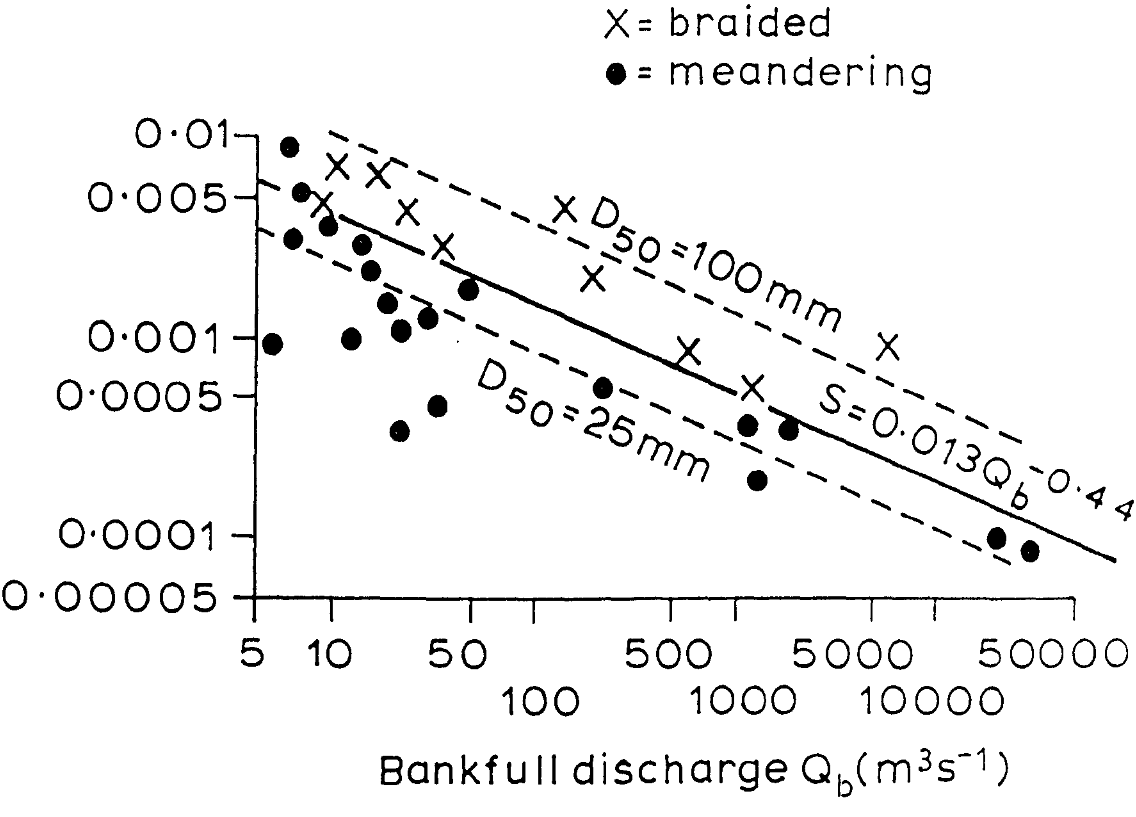


Figure 5.4 The Leopold & Wolman Discriminant Function

(after Leopold & Wolman, 1957)

as D_{90} increases in size. A family of curves is therefore neessary to define the threshold slope that is needed for braiding at a given discharge and grain size.

In order to test the discriminant function approach for predicting palaeodischarges a number of discriminant functions (Table 5.1) have been used to predict the minimum "bankfull" discharge needed for braiding in the present upper braided reach of the River Feshie given the channel slope of 0.0094. The results are shown in Table 5.1. Using the Leopold and Wolman function the minimum discharge needed for braiding is predicted to be 2.08 m^3s^{-1} . However, the Bray (1982) function, produced from coarse gravel-bed rivers in Alberta, predicts a discharge of 102 m^3s^{-1} . The predicted discharge can be seen to vary widely with the predictive function used.

TABLE 5.1

Leopold and Wolman (1957)	$s = 0.013Q_b^{-0.44}$
Predicted bankful discharge	= $\underline{2.08}$ m^3s^{-1}
Henderson (1961)	$s = 0.002D_{50}^{1.15}Q_b^{-0.46}$
Predicted bankfull discharge	= $\underline{49.0}$ m^3s^{-1}
Bray (1982) (2 year flood)	$s = 0.07Q_2^{-0.44}$
Predicted 2 year flood	= $\underline{102.0}$ m^3s^{-1}
Ferguson (1984)	$s = 0.056Q_b^{-0.5}$

Predicted discharge $= \underline{35.0}\ m^3s^{-1}$

Ferguson (1984) $s = 0.017Q_b^{-0.49} D_{90}^{0.27}$

Predicted discharge $= \underline{53.0}\ m^3s^{-1}$

In the palaeohydrological context, attempts to predict a minimum "bankfull" discharge needed for braiding using the Leopold and Wolman plot generally have been restricted to one estimate made for a single terrace surface from one river system. However, comparison of data from several terrace fragments from one river system suggests that this method is not a very useful one for estimating even a minimum bankfull discharge for a braided terrace system. Data from 7 terrace fragments in the Feshie ranging in age from 1,000 BP to 13,000 BP were used to predict the minimum "bankfull" discharges for braiding using the Bray (1982) and the Ferguson (1984) discriminant functions. The results (Table 5.2) indicate that as terrace slope increases the predicted discharge needed for braiding inevitably decreases. Using the Bray function the discharge needed for braiding, given the measured slopes, decreases from $102m^3s^{-1}$ for the modern stream, to 98 m^3s^{-1} for the 1000 BP terrace fragment to 17.6 m^3s^{-1} for the 13,000 BP outwash surface. This suggests that since 13,000BP the River Feshie has been moving closer to the meandering / braiding threshold.

Considering the 3,600 BP terrace fragments in a downstream direction, the Bray function predicts a discharge of 54 m^3s^{-1} for the upstream fragment, 48 m^3s^{-1} for the Allt Garbhlach reach fragment, 3km downstream, and 25 m^3s^{-1} for the Achleum fragment, 5km downstream from the upper braided reach. The Ferguson function shows the same trend in the predicted discharges.

TABLE 5.2

Bray Function	Ferguson Function
$s = 0.07Q_2^{-0.44}$	$s = 0.017Q_b^{-0.49} D_{90}^{0.27}$
$Q_2 = \left(\frac{s}{0.07}\right)^{-2.27}$	$Q_b = \left(\frac{s}{0.017 D_{90}^{0.27}}\right)^{-2.04}$

3500 BP fragment: Upper braided reach

$s = 0.0121$

Bray Function	Ferguson Function
$Q = 54\ m^3s^{-1}$	$Q = 27m^3s^{-1}$

1000 BP fragment: Upper braided reach

$s = 0.01$

Bray Function	Ferguson Function
$Q = 83\ m^3s^{-1}$	$Q = 41\ m^3s^{-1}$

3500 BP fragment: Allt Garblach reach

$s = 0.0127$

Bray Function	Ferguson Function
$Q = 48\ m^3s^{-1}$	$Q = 33.5\ m^3s^{-1}$

3500 BP fragment: Achleum reach

$s = 0.017$

Bray Function	Ferguson Function
$Q = 25\ m^3s^{-1}$	$Q = 18.5\ m^3s^{-1}$

13000 BP fragment: Outwash terrace

$s = 0.0198$

Bray Function	Ferguson Function
$Q = 17.6\ m^3s^{-1}$	$Q = 13.0\ m^3s^{-1}$

The application of the discriminant functions in palaeohydrology is to produce a minimum estimate for the "bankfull" discharge as a guide to the discharge of the past channel. With the discriminant functions, the steeper the slope, the smaller the discharge required for braiding. Thus, because there is an overall decline in slope with decreasing age of the terrace surface (Chapter 4), comparing predicted discharges for the different terrace levels of the River Feshie inevitably produces the wrong temporal trend. Similarly, when considering the downstream trend, local increases of slope in a downstream direction, produce discharge estimates which decline downstream. For comparative studies, the discriminant function approach to palaeohydrology is of little value, even as a check on the minimum "bankfull" discharge required for braiding. These functions can only predict the limiting conditions for the occurrence of pattern type. This means that although the prior braided systems may have plotted well above the discriminant function line it cannot be known exactly where on the plot the different generations of the Feshie would actually have plotted, so that different degrees of braiding cannot be differentiated.

CHAPTER 6

THE CONTINUUM OF CHANNEL PATTERN MORPHOLOGY

6.1 Introduction

It has been argued that reconstruction of palaeodischarges may be made more successfully from investigations of channel pattern and the morphological and geometric parameters of a river channel than from reconstructions made from individual clasts sampled from channel/bar deposits (Bridge, 1985; Miall, 1985; Rotnicki and Borowka, 1985). At present however, there does not exist a quantitative model which satisfactorily relates numerical parameters of braided streams in particular, or channel pattern morphology as a whole, to the possible controlling variables influencing stream planform. A considerable amount of attention has been devoted to investigating the relationships between the geometry of meandering streams at the between river scale and possible controlling factors. In contrast, the present state of knowledge regarding braided streams appears to lag behind that concerning meandering streams. For example, little comparable work has been conducted on either the development of morphometric parameters with which to describe braided stream planform at the between river scale, or the possible relationships of braided stream morphometry to the hydrological and sedimentological controls of braided planform development.

Much research concerning river planform development and its controlling variables has concentrated on the distinction between meandering and braiding streams. This research has focused attention on the development of threshold slope-discharge criteria which discriminate between the meandering and braiding channel states (for example, Leopold and Wolman, 1957; Henderson, 1961; Osterkamp, 1978; Bray, 1982; Ferguson, 1984). The sample of channels used for the derivation of the discriminant functions is usually drawn from a preexisting classification of channel planform. This treatment of channel pattern morphology as comprising discrete pattern states has encouraged the perpetuation in the literature of a classificatory approach to the analysis of channel pattern morphology and has inhibited the adoption of a unified approach to channel pattern morphology.

However, for channels in near uniform materials, there may be a complete gradation of channel geometry, flow, and sedimentary processes between channel patterns, depending on the imposed discharge and sediment load (Bridge, 1985). The identification of a continuum of channel pattern morphology, and its general relationship to the controlling variables of pattern morphology, requires the development of a quantitative index of form that can be applied to both meandering and braided streams. If channel pattern morphology is treated as a continuous quantifiable variable, then it becomes possible to test more rigorously the hypothesis that meandering and braiding streams may be related to a single general model in terms of a set of continually measurable variables. In this chapter, therefore, a common measurable parameter of channel pattern, total sinuosity, is proposed for free alluvial channels. The physical basis for this parameter is examined in the light of existing theoretical, analytical and empirical studies. The parameter is then statistically related to three possible controls of channel pattern morphology, discharge, valey slope and bed material size.

It may then be possible to apply any relationships established between channel pattern and its controlling variables in present day rivers to a palaeohydrological investigation of palaeo-channel traces on terrace fragments. Comparisons between palaeochannnel networks on a single numerical scale may be facilitated not only between varying terrace levels of one river system, but also between terraces of different river basins. Furthermore, the same palaeohydrological method could be used for single and multithread streams, as well as for braided streams exhibiting differing degrees of braiding.

6.2 The continuity of channel pattern morphology in free alluvial channels

The concept of a continuum of channel pattern is not new. Leopold and Wolman (1957) originally argued that a continuum of channel pattern should exist in natural streams. They suggested that each nominally classified pattern type, that is straight, meandering, or braided, should be associated with a particular combination of the controlling variables of pattern morphology operating at different intensities. As the same basic physical principles operate to produce all pattern types, channel pattern morphology should intergrade with meanders and braids forming end members of a sequence. This view has been restated more recently by Parker (1976), Chang (1979), Richards (1982) and Bridge (1985).

Free alluvial streams are generally defined as those that are free to adjust their planform within the constraints of the prevailing hydrological and sedimentological conditions. For such streams at the between river scale, it may be possible to justify a unified approach to channel pattern morphology in terms of: (1) the physical processes of bank erosion, bed material transport and channel/bar construction; (2) continuous changes in channel cross-sectional geometry; and (3) the creation of channel pattern morpholgy as a mechanism for dissipation of excess energy.

(a) <u>The continuum of macrobar forms</u>

Channel pattern adjustments at the between reach scale take place as a result of the common physical processes of bank erosion, sediment transport and bar formation common to all alluvial rivers (Parker, 1976). The available evidence suggests that the increasing bank erosion and concomitant rise in width-depth ratio that is associated with the sequence of straight, through meandering to braided streams is accompanied by a sequential change in the type and pattern of the macrobar forms within the channel zone. These changes in bar forms that occur with changes in pattern morphology have been identified through both flume and field studies.

The flume experiments of Schumm and Khan (1972) demonstrated that as the flume runs progressed several different bar types developed (Figure 6.1). These changes in bar configuration occurred as stream power, sediment transport rates and width-depth ratio all increased. As changes in bar configuration and increase in width-depth ratio occurred the channel progressed from a relatively straight channel with point bars to a multi-thread channel with point bar type bedforms and mid-channel bars.

Similar sequential changes in bar configuration have also been described from the flume experiments of Ashmore (1982). In

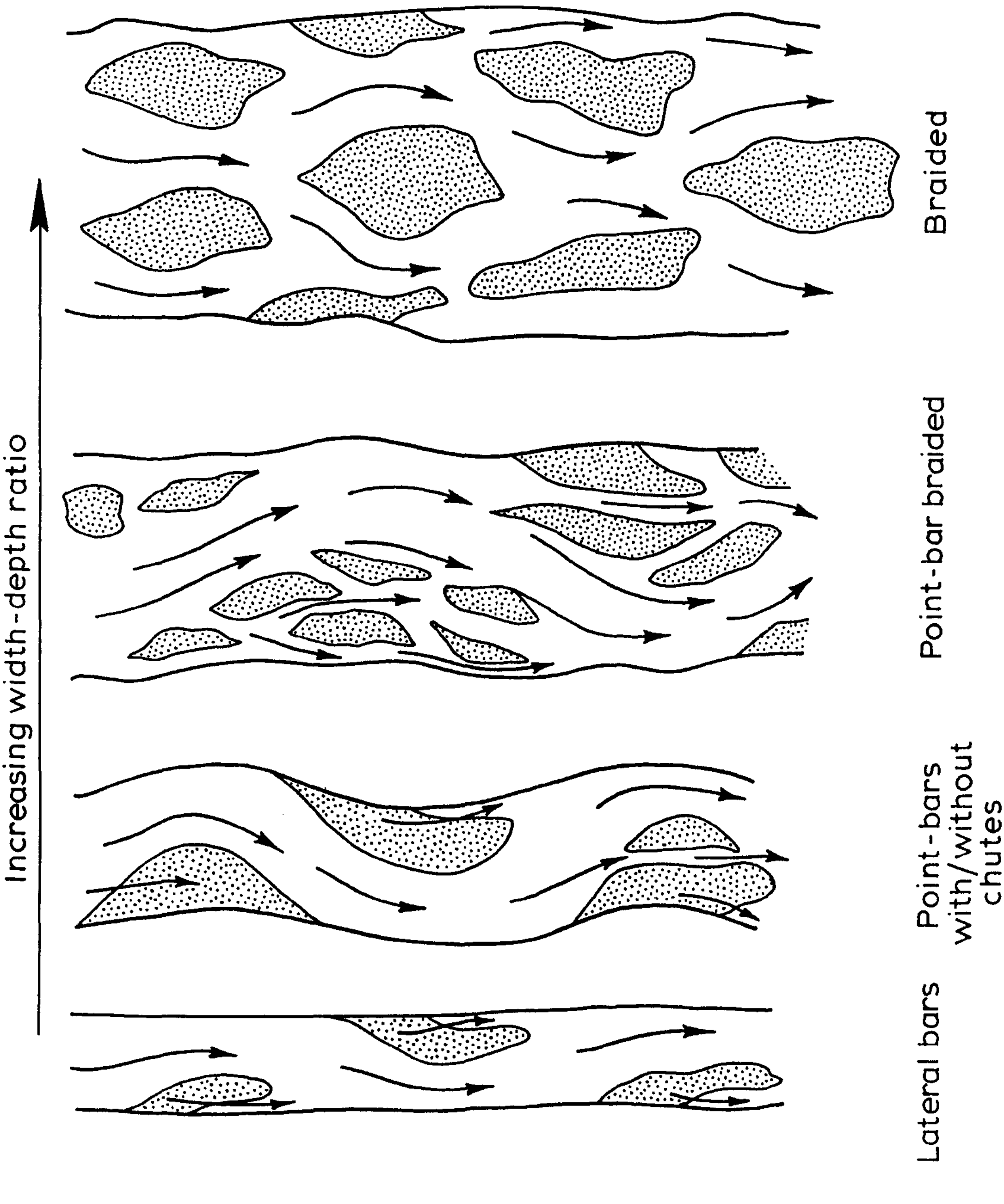


Figure 6.1 Continuum of bars in flume channels (after Schumm, 1972)

Ashmore's experiments, stream power remained constant, with sediment load being adjusted to prevent aggradation or incision of the bed. The development of the flume channel pattern occurred in sequence from alternating bars with a meandering thalweg in a straight channel, to a meandering channel as the flow deflected by the bars eroded the channel banks. As the sinuosity of the channel increased, bank erosion became more extensive, the width-depth ratio of the channel increased and the regular alternating bar pattern broke down as the flume channel began to braid. Ashmore's flume experiments support the theoretical work of others (Engelund and Skovgaard, 1973; Parker, 1976) who suggest that meandering and braiding may be treated as different degrees of the same instability phenomenon, with both pattern types developing from straight alluvial channels and alternate bar formation.

In natural gravel-bed rivers, the sequence of bedforms as channels change from meandering to braided has been illustrated by Bluck (1976) and Church and Jones (1982). As the pattern changes from meandering to braided, point bars are gradually replaced by bar forms that are progressively less stable and less attached to the bank. This sequence (Figure 6.2) progresses from

(1) point bar with a chute near the river bank,

(2) lateral bar that is bank attached with chutes,

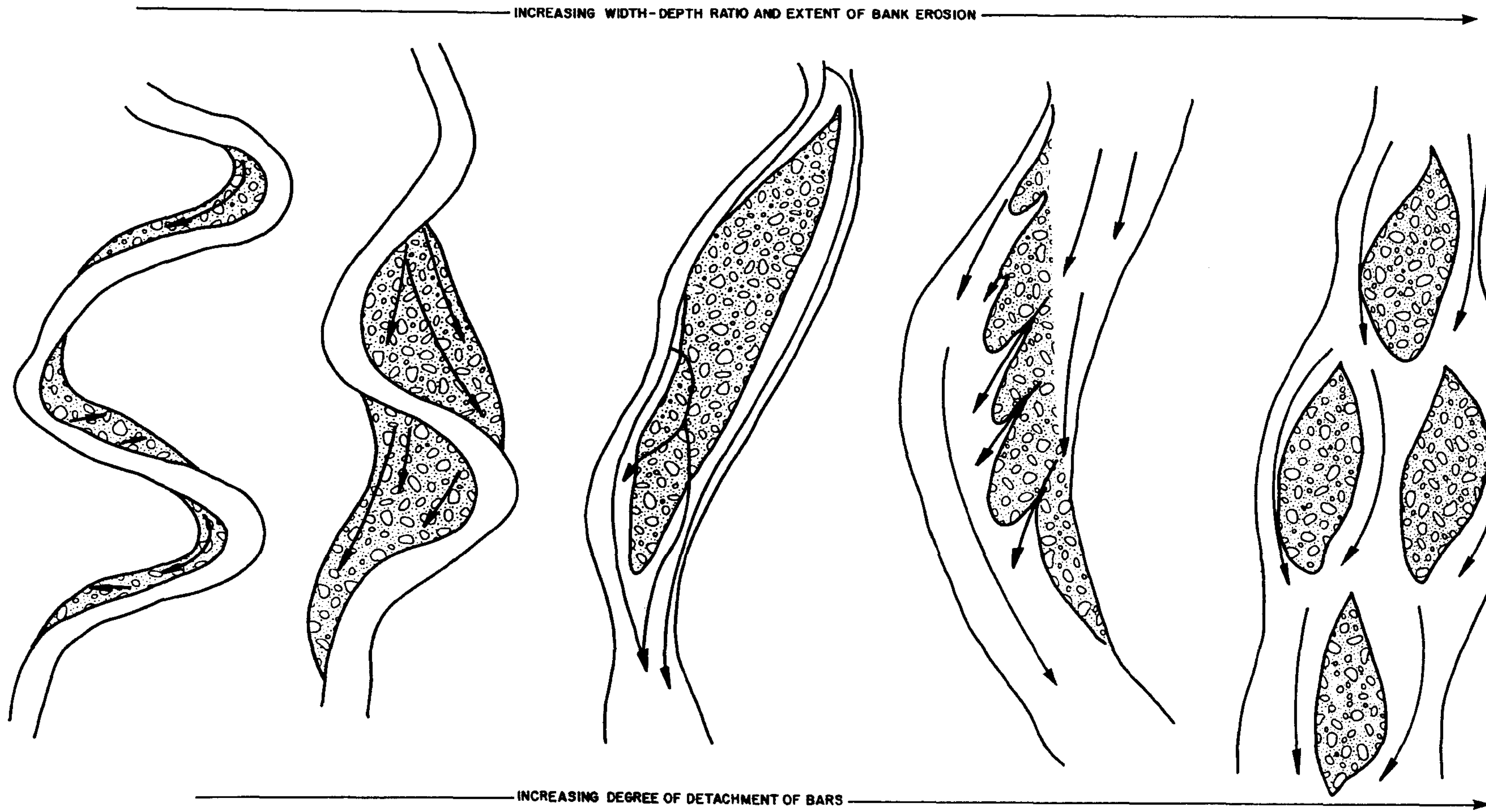
(3) lateral or diagonal bar crossed obliquely by the main flow and

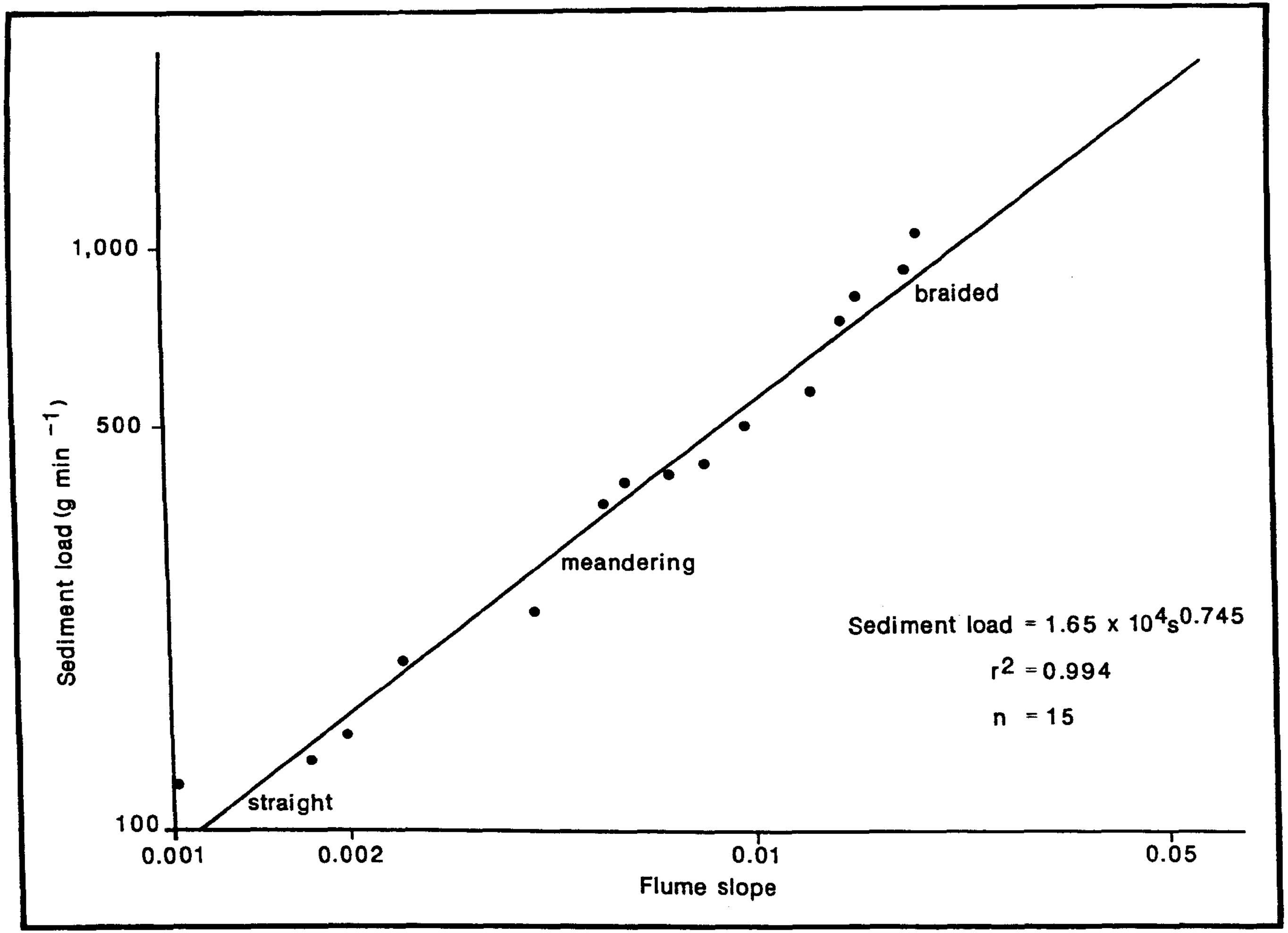
(4) lateral bars with river channels that are smaller than the main flow, and finally

(5) medial bars with approximately equal flows on either side of the bar.

Figure 6.2

Continuum of gravel bars in natural rivers





Sediment transport and channel pattern
in a flume channel at increasing slopes

Figure 6.3

(b) Continuous changes in cross-section geometry

Analytical and flume studies have shown that the progression of straight through meandering to braided streams is related to changes in the cross-sectional geometry of the channel (Ackers, 1964; Schumm and Khan, 1972; Engelund and Skovgaard, 1973; Parker, 1976; Chang, 1979), and more particularly to an increasing width:depth ratio. These changes in width:depth ratio of the channel occur in association with changes in the channel / bar configuration and channel pattern morphology (Figure 6.1 and 6.2).

A high width-depth ratio is associated with extensive bank erosion, high sediment transport rates and braiding (Ferguson, 1984), a view anticipated by the flume experiments of Schumm and Khan (1972). Schumm and Khan have suggested that their plot of sediment transport rate against slope demonstrates discontinuities or thresholds between the different channel pattern types. However, correlation and regression analysis of their data for sediment load against slope gives a correlation coefficient of 0.996 (Figure 6.3) for the relationship

$$Sed_1 = 1.65 \times 10^4 \, s^{0.745} \qquad r^2 = 0.994 \quad 6.1$$

where

Sed_1 = sediment load (g min^{-1})

s = flume slope (for constant discharge)

The flume data may therefore be interpreted as demonstrating a strong, continuous relationship between bedload transport rates and flume slope, which at a constant discharge rep resents a stream power index (Figure 6.3). Increasing width-depth ratio of the flume channel is also highly correlated with increasing bedload transport rates in the experiments. The flume studies suggest that changes in the cross-sectional channel pattern morphology and geometry of river channel occur as a continuous direct function of stream power and bedload transport rates. This hypothesis is supported by data from 28 meandering and braided rivers in New Zealand. A plot of width:depth ratio against an index of stream power, the discharge-valley slope product (Figure 6.4), suggests that the width-depth ratio does vary as a continous function of stream power. The relationship

$$w/d = 78.5\ \Omega^{0.69} \qquad r^2 = 0.77 \qquad 6.2$$

where

w/d = width:depth ratio

Ω = stream power index

between stream power index and the width:depth ratio shows no discontinuity between the scatter of points for the meandering and braided streams.

(c) Creation of channel pattern morphology as a mechanism for the dissipation of excess energy

In free alluvial channels local excess energy availability may induce pattern adjustment which allows a more equal distribution of potential energy loss per unit length/area, while at the same time minimising rate of energy loss per unit length/ area of channel. If there is a high initial rate of energy dissipation per unit length/area of stream, the pattern morphology created by active transport processes either lengthens the river path and reduces the gradient (Yang, 1971), or increases bed area (Chang, 1979) so that power per unit/length is minimised. Both meandering and braided streams may be interpreted within this

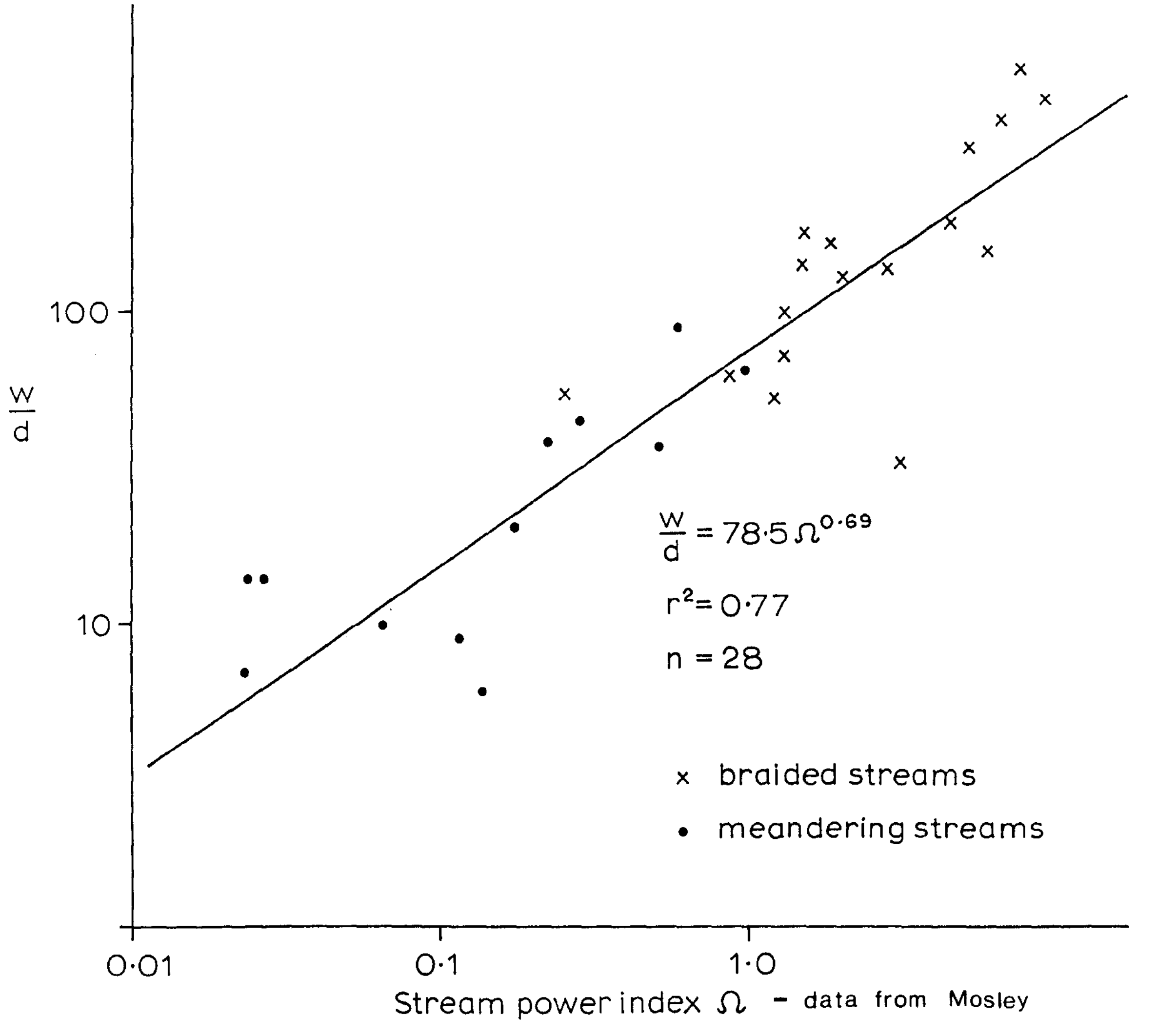


Figure 6.4 Plot of Width–Depth Ratio of Gravel–Bed Rivers against Increasing Stream Power

general context.

Stream sinuosity in singlethread streams depends on the initial rate of power expenditure per unit area (Ferguson 1973). In near uniform materials the highest sinuosities should develop in the channel reaches with the greatest initial power expenditure rate per unit area. Here, excess energy is available for bank erosion. Bank erosion will continue increasing stream sinuosity until the excess potential energy is minimised. An equilibrium sinuosity indicates the stage at which an initial power excess has been dissipated. Meandering is therefore a means by which a stream is able to minimise the rate of potential energy expenditure per unit length of channel. This is a view supported by the flume work of Schumm and Khan (1972). Here, continuity of sediment transport was maintained in the flume runs so that equilibrium conditions prevailed. Plotting sinuosity against power (Figure 6.5), Schumm and Khan found a continuous increase in sinuosity with stream power, until a threshold value was apparently reached when braiding began.

Braided streams minimise excess potential energy loss through the formation of braids until a quasi-stable state, the prevailing braid mode, is achieved (Parker 1976). The stable state of a channel is represented by the parameter

$$\varepsilon^* = \frac{S_e}{\pi\ (F.d/B)} \qquad 6.3$$

where

F = Froude number

B = channel width

d = channel depth

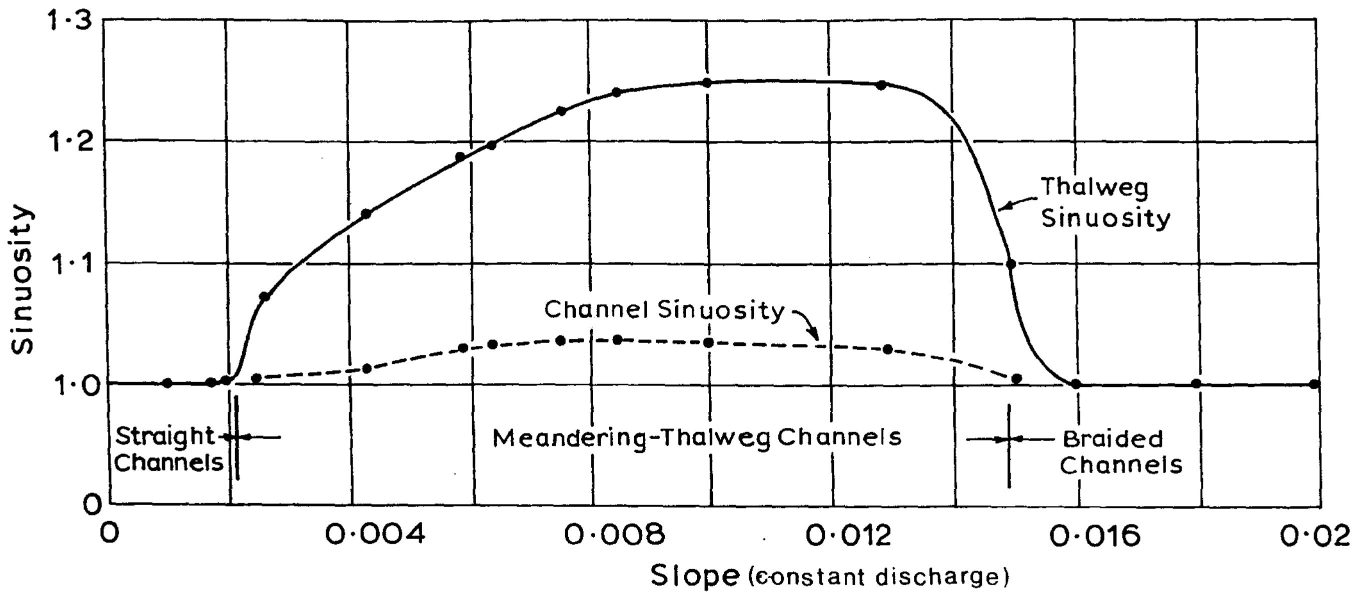


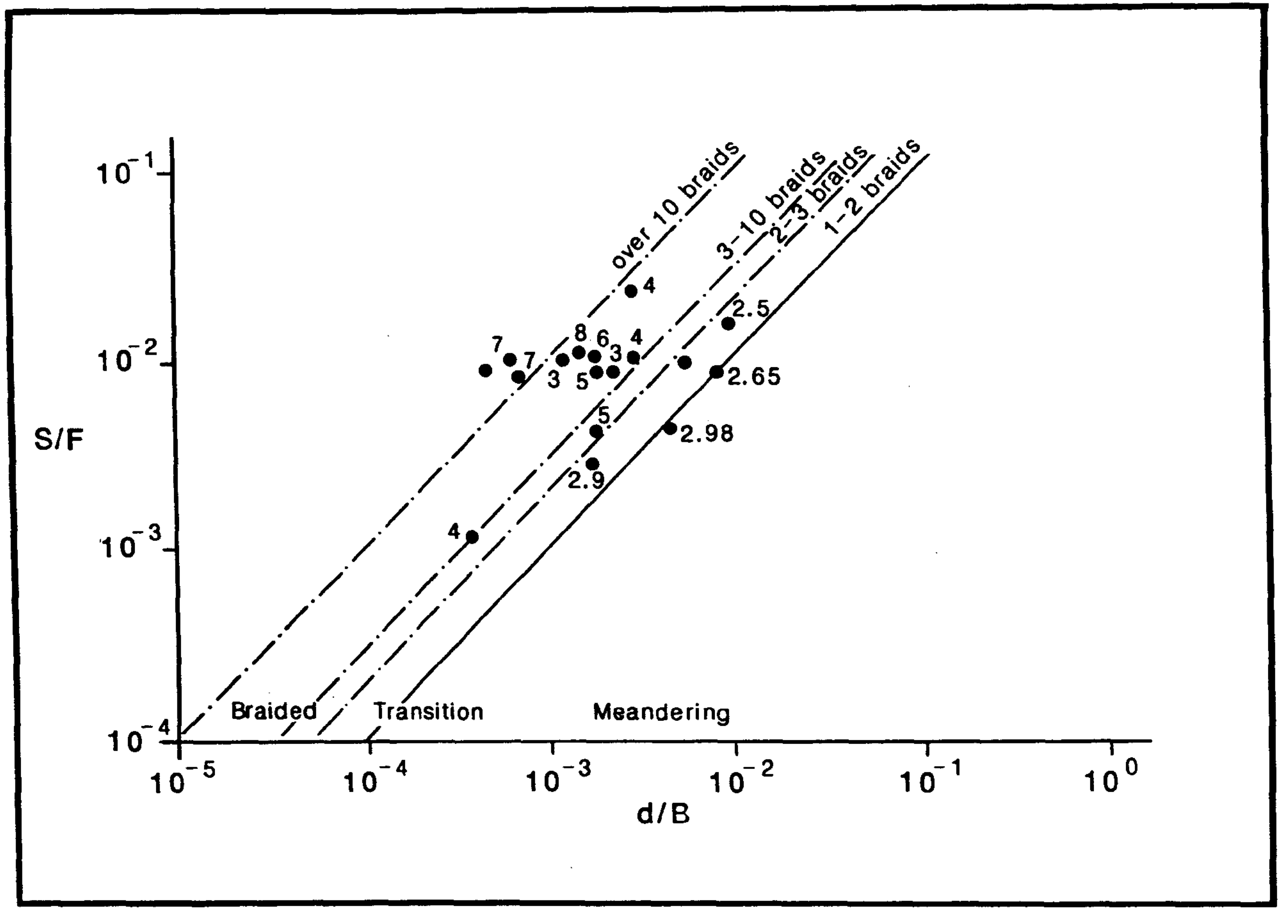
Figure 6.5 Changing Sinuosity in a flume channel at increasing slope (after Schumm & Khan, 1972)

Se = energy slope

which is a ratio of energy gradient to the energy dissipation characteristics of the channel (Church and Jones 1982). When ε^* is greater than one the number of braids at a cross-section in a channel is estimated by m, the braid mode, which is approximately equal to ε^*. Data for 15 braided gravel-bed streams from New Zealand were plotted on the Parker braided-meandering regime diagram, together with the number of braids representative of each braided reach (Figure 6.6). The plot demonstrates that the parameter ε^* gives a generally accurate description of the number of braids in natural braided channels.

The braid bars represent energy dissipators that enable the channel to maintain a prevailing braid mode whilst still providing an energy gradient sufficient for periodic ~~sediment~~ sediment transport over the bed (Church and Jones, 1982). Highly braided channels, that is those with a high number of braids at a cross-section, represent channels with a high initial excess of potential energy. Whilst the actual location of the channels in a braided reach may vary, and are therefore unstable, the prevailing braid mode, approximated by the number of braids in a reach, may be relatively stable under the prevailing hydrological and sedimentological conditions of the catchment (Bradley et al., 1972; Church and Jones, 1982). Essentially, this is no different from the idea of a stable reach sinuosity for a meandering river in which meander bends tighten and eventually cutoff.

A braided stream may therefore dissipate excess potential energy by increasing its total path length, also therefore effectively increasing its total bed area, through the formation of braids. This process can be compared with that of single thread channels



Braided/meandering regime diagram
(after Parker, 1976)

Figure 6.6

which minimises stream power expenditure by increasing in sinuosity thereby increasing channel path length and again total bed area per unit valley length.

For singlethread channels in uniform materials sinuosity is proportional to initial power expenditure rate per unit area of channel. Mathematical modelling has demonstrated that with an increase in stream power highly sinuous singlethread channels become braided (Chang, 1979), whilst beyond this transition the braiding tendency, that is the number of braids at a cross-section, increases with increasing stream power. Thus the prevailing braid mode becomes larger as stream power increases. The progression of streams in the Parker regime diagram (Figure 6.6), therefore represents a progression from lower powered to high powered streams. The streams which plot in the 1-2 braid band will have a lower stream power than the streams with 8 braids at a cross-section. Chang (1979) is proposing a unified approach to channel pattern morphology within which, for uniform materials, channel pattern morphology is variable with spatial variation in stream power. These analytical conclusions of Chang are supported by the empirical work of Richards (1982), Begin and Schumm (1984) and Howard et.al. (1970).

Richards (1982), defining pattern morphology for single and multithread streams using a sinuosity index similar to that of Le Ba Hong and Davies (1979), related the sinuosity of 19 sand and gravel-bed channels to a stream power index, the discharge-channel slope product. The relationship

$$P = 2.64\Omega^{0.1} \qquad r^2 = 0.41 \qquad 6.4$$

where

P = sinuosity

Ω = stream power index

shows a statistically significant correlation between total sinuosity and stream power. However, the amount of the variance of the dependent variable, sinuosity, that is explained by the independent variable, stream power, is only 41%. This may be because of several factors. First, Richards used the variable channel slope to calculate the stream power index. However, channel slope is not a fully independent variable being partly dependent on channel pattern. The logically relevant control on channel pattern is the initial gradient before any pattern development takes place (Ferguson, 1984). This gradient is best approximated by the valley gradient. Second, a mixture of sand and gravel-bed streams was used to derive the relationship. However, because of the different sediment resistance characteristics of sand-bed channels the flow range in which sand beds are mobile is very large, whereas for gravel beds this flow range is much more limited. For sandy-bed channels braiding will occur at relatively low stream powers, whereas the power needed for braiding in gravel is much greater (Ferguson, 1984). The functional relationship between total sinuosity and stream power is therefore likely to vary with the nature of the perimeter sediments. Furthermore, the stream power needed for braiding in gravel-bed rivers increases with increasing D_{90}, (Ferguson, 1984) so that a multivariate expression relating total sinuosity, stream power and grainsize may be needed.

Begin and Schumm (1984) conclude that the probability of passing from one stream pattern type to another is a continuous function of relative shear stress, shear stress being power per unit bed-area divided by velocity. An increase in shear stress causes high sinuo sity singlethread streams to become braided. For streams that are already braided, the number of braids at a cross-section has been shown to increase with increasing stream power (Howard et al., 1970).

Up to a point therefore, and in streams with near uniform materials, a stream may dissipate its energy by increasing its path length, and bed area, by increasing its sinuosity. Eventually, the energy level may be sufficiently high for it to be necessary to braid and create extra channels and bars in order to generate sufficient bed area for the potential energy to be extended on periodic sediment transport.

6.3 The quantification of channel pattern morphology

(a) Total sinuosity

If meandering and braided streams are to be related to a single general model in which channel pattern morphology is represented as a continuum in terms of a set of continually measurable variables, then a parameter of channel pattern morphology is required which is capable of quantifying meandering and braided stream planform at the between-river scale.

Two factors need to be taken into account when attempting to quantify channel pattern using a single numerical parameter. First, a common measurable parameter of channel pattern must be dimensionless. Only then can comparisons between river systems of different sizes be made. Second, a parameter used to quantify channel pattern must be interpretable in terms of the physical adjustment of channel pattern to the possible controlling variables of river planform.

The problem is to find a parameter of channel pattern morphology that can describe both meandering and braided patterns. The sinuosity parameter has been used for meandering streams to quantify degree of meandering. Various braid parameters for braided stream morphology have been previously proposed. Brice's dimensionless braiding index (Brice, 1964) is a quantitative index of the degree of braiding of a stream, but has been applied to both singlethread and multithread channels. The index

$$BI = \frac{2\,[\text{sum of the lengths of islands and/or bars in reach}]}{\text{reach length at mid channel}} \qquad 6.5$$

where

BI = braiding index

constructed for sand-bed channels, is a quantitative measure of the extra bank length created by braiding. Bars are recognised as non-vegetated, and submerged at bankfull stage, whereas islands are vegetated and emergent at bankfull stage. The braiding index may therefore be expressed as "transient" for bars, "stabilised" for islands and "total" for the two combined. As Brice remarks, the transient index is heavily stage-dependent, but suggests that the stabilised braiding index is nearly constant regardless of river stage. However, the stabilised index is not of great value if it is considered desirable to quantify the channel morphology of the active braided areas of the channel. Brice himself notes that although the braiding index may be useful for the morphological description of changing bed area with rising stage, it has little hydraulic significance. Further, the index is not particularly useful as a quantitative index of pattern morphology as applied to both meandering and braiding, for meandering streams of different sinuosities cannot by successfully differentiated.

Rust (1978) has argued that the problem of stage dependence may be largely overcome by defining the perimeter of a braid as the mid-line of the channels surrounding each bar or island. The braid length is the straight line distance between the extremities of the braid. A measure of braiding intensity, the braiding parameter, is then the number of braids per mean meander wavelength, the latter approximately 1.25-1.5 times the mean braid length. The braiding parameter may be calculated for individual bar orders (see below, Williams and Rust, 1969). Whilst this method has the advantage of overcoming a large degree of stage dependence, provided bars of the same order are compared, the parameter is not dimensionless and therefore makes comparison of braiding intensity between rivers difficult. Also, as with the Brice index, it cannot sucessfully differentiate between varying degrees of meandering. However, both indices do show that the fundamental properties that characterise channel pattern morphology are sinuosity and degree of channel splitting, a view also expressed by Miall (1985).

In near uniform materials channel adjustment to changing hydraulic conditions involves changes in total path length relative to valley length, and therefore changes in total bed area. This has been shown to be the case for both meandering and braided streams. It follows, therefore, that a common measurable parameter of channel pattern which is both dimensionless and has meaning in terms of the physical processes of channel pattern adjustment could be the ratio of total active channel length to valley length. Traditionally this ratio, channel sinuosity, has been applied to singlethread channels.

However, the same ratio could be used to describe the sinuosity, or rather, the <u>total sinuosity</u>, of a braided channel. Braided and meandering streams could then be compared so that a braided stream is more sinuous or has a higher ratio of active channel length to valley length than a singlethread meandering channel. A continuum of channel pattern form therefore more clearly exists from meandering through to braided streams whilst different degrees of meandering and braiding are quantified on a single numerical scale.

Le Ba Hong and Davies (1979) have proposed a sinuosity index for braided streams which is comparable to that proposed for single-thread channels. They calculated the sum of the total channel length of all the channels in a braided reach which is then divided through by the reach length. The sinuosity of a braided reach is defined as

$$P = \frac{\Sigma L}{l_v} \qquad 6.6$$

where

P = sinuosity

ΣL = total channel length

l_v = reach length

that is, the total length of channel per unit length of river. This index is the same as that used by Richards (1982) and by Mosley (unpublished data set) who used sinuosity to quantify the degree of braiding of New Zealand multithread streams.

(b) An operational definition of total sinuosity for multithread channels

In this study, sinuosity was calculated for singlethread channels by digitising the mid-line of the channel and dividing this by valley length. For multithread streams the total active channel length was calculated by digitising the mid-line of the channels adjacent to the macrobar forms in the channels measured (Figure 6.7). Thus total sinuosity is

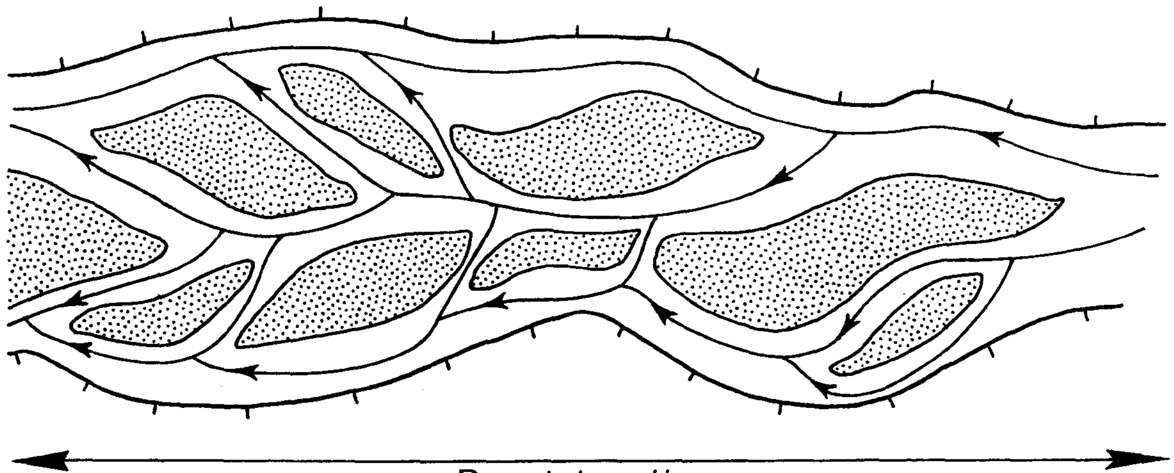
$$\Sigma P = \frac{\text{Total active channel length}}{\text{reach length}} \qquad 6.7$$

where

ΣP = total sinuosity

The total active channel length is the sum of all the lengths around the macrobars in the channel zones. As suggested by Rust (1978), digitising the mid-line of the channels helps to minimise the problem of stage dependence. The scale of measurement chosen in this study, the macrobar, is equivalent to the macrobar forms, the large bars or bar complexes, in the hierarchical classification of Church and Jones (1982). These bars are channel-scale sedimentary accumulations that are major features of gravel-bed river morphology. The latter scale is equivalent to the first order channels as defined by Williams and Rust (1969), that is the channels which delineate the main bar forms in the active channel zone. This scale of measurement is also consistent with the scale of the braid mode modelled by Parker (1976) in his stability analysis.

This scale of measurement was considered appropriate for an analysis of channel pattern morphology at the between river



$$\Sigma P = \frac{\text{Total active channel length}}{\text{Reach length}} = 4{\cdot}05$$

Bars - (macrobar forms)

Channels - (first order channels)

Figure 6.7 Definition of Total Sinuosity scale for several reasons. First, as Church and Jones (1982) point out, the macrobar forms in gravel-bed rivers not only act as major storage places for bedload sediments that are only moved sporadically, but they also form important resistance elements interacting with and conditioning the flow pattern. This scale of measurement can therefore be justified in terms of the hydraulics of the flow and is consistent with an analysis conducted at the between river scale. Second, whilst the channel configuration of gravel-bed braided streams may be highly unstable, the main channels and bar complexes, even in environments as active as valley trains, may remain relatively stable for a number of years (Church, 1972; Bradley et al., 1972; Church and Jones, 1982). It is the minor channels, the higher order channels in Williams and Rust's classification, which exist as ephemeral channels only. The first order channels are the channels which show the greatest fluvial activity and carry most of the discharge of the stream. These macrobars are at much lower topographic levels than the smaller second and third order bars which may be superimposed on them. As such photogrammetric or field recognition of these macrobar forms is facilitated. Digitising first order channels should therefore produce consistent measurements between rivers.

(c) The relationship between number of braids at a cross-section and total sinuosity

The analytical studies of Parker (1976) and Chang (1979), and the empirical study of Howard et al. (1970) has shown that the number of braids at a cross-section of an active braided reach is functionally related to stream power, although this relationship has not been quantified. Considering braided streams alone, if total sinuosity can be used to differentiate between different braid modes which are produced as a result of the spatial variation in stream power, then there should be a close correlation between number of braids and the total sinuosity of a braided reach.

The total sinuosity was calculated for 22 gravel-bed braided rivers. The number of braids at a cross-section was also calculated for each reach. Although the prevailing braid mode may be relatively stable the configuration of the channels may not be so. The number of channels may therefore vary between cross-sections of the reach. Therefore the mean number of braids in each reach was calculated by drawing lines in a downstream direction across the active channel zone orthogonal to the valley direction, and counting the number of first order channels bisecting the lines. The mean number of braids for each reach was defined as the total sum of the braids divided by the number of cross-sections. The number of braids, and the total sinuosity, was defined for each reach of constant discharge and valley floor width.

Total sinuosity was plotted against number of braids (Figure 6.8). The statistical correlation of the relationship

$$\Sigma P = 0.98\ Br + 0.45 \qquad r^2 = 0.96 \qquad 6.8$$

where

ΣP = total sinuosity

Br = number of braids

between the two parameters is high, with a linear correlation coefficient of 0.98. This result is as anticipated, for total sinuosity and mean number of braids are two dimensionless measures of the same property, that is the total length of

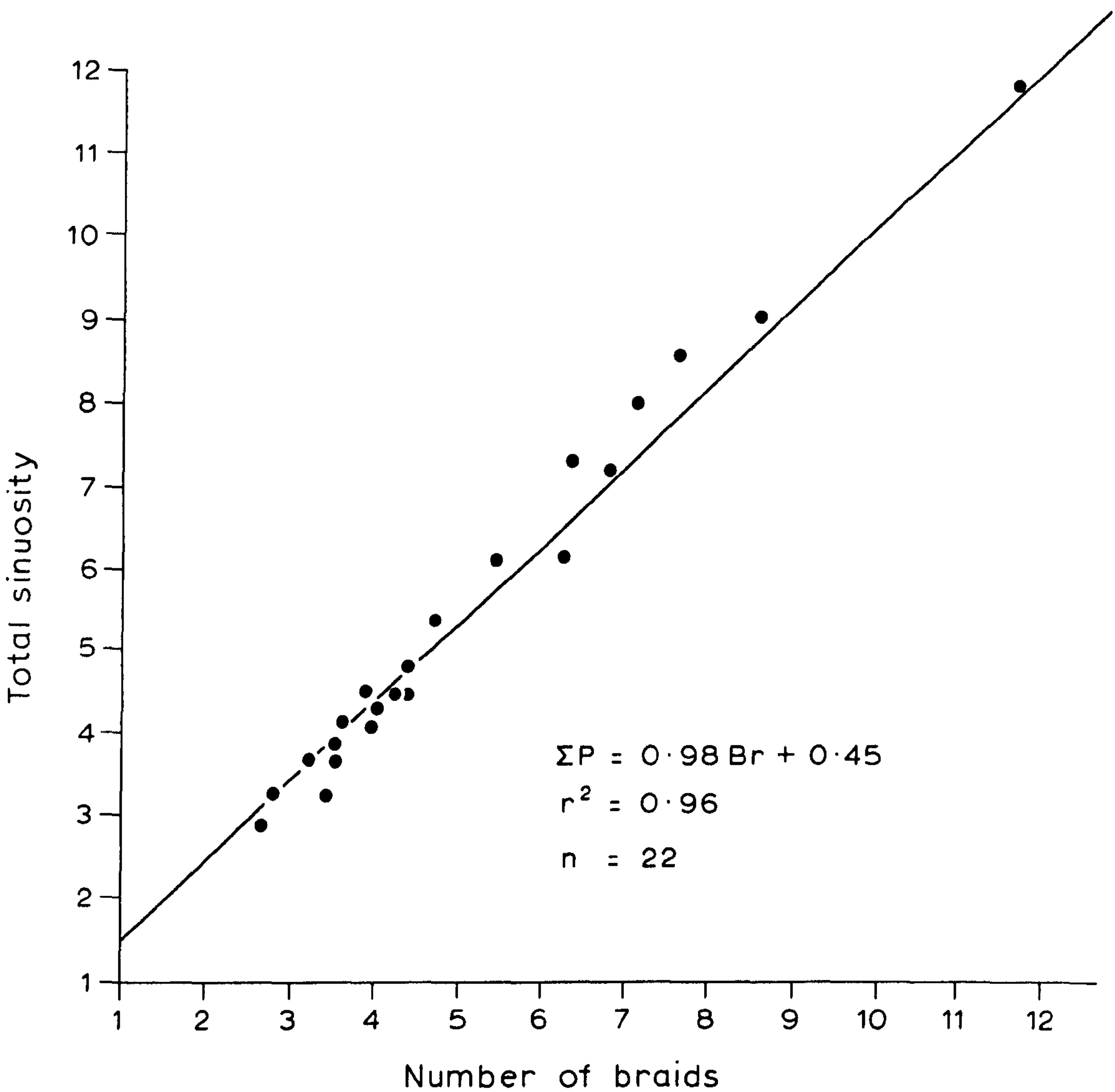


Figure 6.8 **Total Sinuosity against Number of Braids for Gravel-Bed Rivers**

channel per unit length of river. The total sinuosity of a braided stream may be approximated by counting the number of first order channels that bisect a line drawn across the reach. As the number of braids has been shown to be a function of stream power, and as there is a close and predictable relationship between number of braids and total sinuosity, then total sinuosity may also be regarded as a function of stream power.

(d) The question of stage dependence

Some field data suggest that in gravel-bed braided streams total sinuosity may be relatively constant over a range of discharges which include the 1.5 and 2.3 year floods.

Data relating the number of braids at a cross-section to discharge has been collected by Fahnestock (1963) for the active proglacial White River, Mount Rainier, Washington. As mentioned above number, direction and size of channels in a braided reach may vary over a given reach, so that the number of braids at a cross-section will vary both downstream and temporally, even for a reach of constant discharge and valley floor width. The mean number of braids for n cross-sections of a reach of constant discharge and width will give a more representative indication of the variation of the number of braids with discharge. The mean number of braids was therefore calculated from 3 cross-sections of a single reach of the White River using Fahnestock's 1958 and 1959 data sets. These data were then plotted against discharge. A number of salient points emerge from an examination of this plot (Figure 6.9). First, at low flow, (below about 6 m^3s^{-1}) only a few channels are occupied. Second, for flows over about 6 m^3s^{-1} and between 15 m^3s^{-1} the

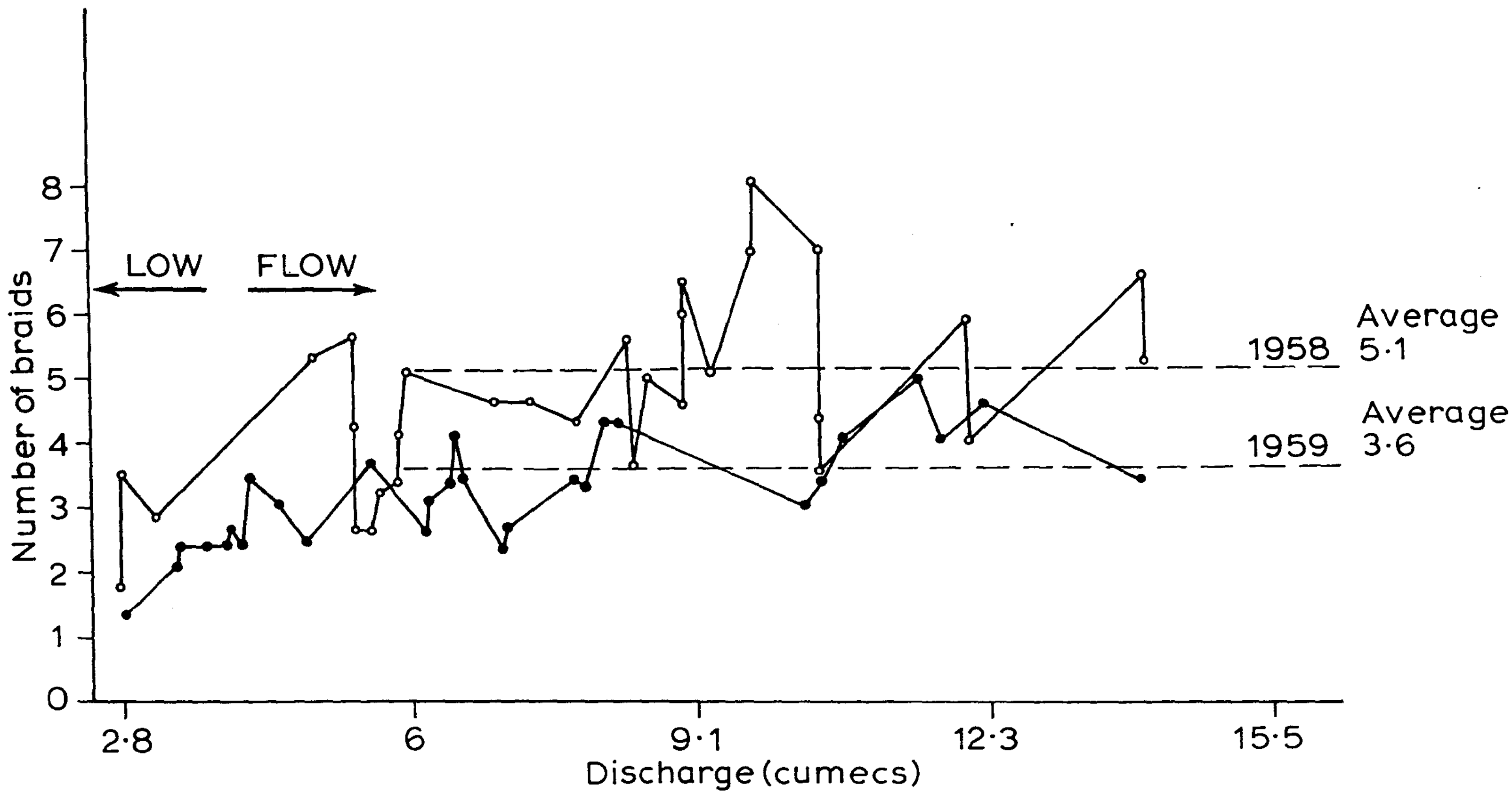


Figure 6.9 Number of Braids against Discharge for the White River

average number of channels tends towards a modal value of 5 channels for the 1958 data and 3.6 for the 1959 data. Over about 6 m^3s^{-1} the coefficient of variation of the number of braids for a given discharge is 22.5% for the 1958 data and 15% for the 1959 data. The relatively low values for the coefficients suggests that the number of braids does not vary greatly between discharges even for the flood discharges above about 11.5 m^3s^{-1} . This analysis is consistent with that made for the same raw data by Pickup and Higgins (1979). Pickup and Higgins calculated the frequency distribution of the number of braids over the same range of discharges that produced Figure 6.8. The distributions are all negatively skewed to the modal value which varies between 2-4 channels. Above 11.5 m^3s^{-1} the modal value increases to 4 channels. This modal value of 4 channels for the high magnitude flood flows represents the dominant braid mode of the low order braid channels. Fluctuations about these values are due to higher order channels becoming occupied as a result of slight increases in water surface elevation.

Both methods of analysis of the White River data show that the low order braided network does not exhibit a significant increase in the number of channels with increasing discharge. The dominant braid mode fluctuates about a value of 3.6-5.0. As number of braids is closely correlated to total sinuosity these data suggest that over a range of discharges above low flow, which include high magnitude flows, total sinuosity is likely to vary about an average value which represents the prevailing braid mode of low order channels. A similar conclusion has been reached by Thompson (in press) who was examining channel response to flood events in Langden Brook in the Bowland Fells, north Lancashire. Thompson suggests that continuous variations in sinuosity and degree of channel division, two variables which may be quantified by total sinuosity of the channel, represent a dynamic form of equilibrim which reflect the channel's sensitivity to short term changes in the balance between major and moderate flood events.

The relative stability of total sinuosity over a range of discharges which include flood flows is also supported by data from the gravelly River Ohau in New Zealand. Total sinuosity was calculated from the 5 published aerial photographs of Mosley (1983). The discharges represented in the photographs range from 25-500m^3s^{-1} . At very low flow as with the White River, the number of channels occupied is relatively low, total sinuosity having a value of 5.3. However for flows between 57-240m^3s^{-1} which includes the 1.5 and 2.3 year flood total sinuosity only varies between 7.3-8.0 reaching a maximum value of 8.0 at the mean annual flood of 240m^3s^{-1} . For flows of 500m^3s^{-1} which represents a very high magnitude, low frequency event, the total sinuosity declines to 5.0.

The data from both the White River and the Ohau River show that above low flow the value for total sinuosity remains relatively constant for flows which include relatively high magnitude flood flows up to and including the mean annual flood. These data, derived from two rivers of widely differing sizes and environments, show that total sinuosity is a much less sensitive index to variations in stage than indices based on either bar length or area.

6.4 Numerical analysis : total sinuosity as applied to singlethread and multithread streams

The discussion presented above proposes that total sinuosity, the ratio of total active channel length to reach length, is a measurable parameter of channel patten morphology that is both dimensionless and has physical meaning in terms of the physical adjustment of the channel to spatial variation in the controlling variables of river planform. It therefore fulfills the necessary requirements suggested for a common measurable parameter of channel planform at the between river scale. Using this parameter, channel pattern morphology of both meandering and braided streams could be quantified on a single scale and related to the possible controlling variables of planfrom for free alluvial streams. The analytical and empirical studies discussed above suggest that channel pattern morphology in free alluvial streams will be determined ultimately by river discharge, valley slope (which together are the variable quantities in stream power) and the nature of the perimeter sediments, which determines resistance to entrainment and transport.

These generalisations have not yet been tested with an appropriate range of field data. Two hypotheses are therefore considered in this study. First, having defined total sinuosity as a continuous measurable parameter of channel pattern, it is proposed that channel pattern varies as a continuous function of stream power and bed material size; total sinuosity should vary directly with stream power and inversely with bed material size. Second, the functional form of the relationship between total sinuosity and stream power should vary with the nature of the perimeter sediments, so that separate continua may be required for gravel-bed and sand-bed streams; this may reflect the different magnitude-frequency characteristics of bed material transport in gravel and sand-bed streams.

(a) <u>The variables</u>

Two data sets were compiled, one for gravel-bed rivers and one for sand-bed channels. The data sets used in the analysis were derived from three sources:

(1) the catalogue of alluvial river channel regime data compiled by Church, Moore and Rood (1981)

(2) an unpublished data set provided by Dr. P. Mosley (Mosley, MS Report)

(3) a selection of published research papers.

In all, 60 river systems were considered. The rivers were from widely differing sizes and discharges, and came from North America, New Zealand, Scandinavia and Britain. No data were included from plain sandurs. The river reaches included in the data set were restricted as far as possible to free alluvial streams which are actively migrating and reworking their valley deposits. Although several data sets of singlethread gravel-bed channels are available (Church <u>et al.</u>, 1981) the channels included in the present data set represent only a small selection of streams from these data sets. This is because a number of rivers from these data sets include powerful single-thread channels which are incised or possess confined meanders, or are relatively straight, powerful streams that are limited in sediment supply (Bray, 1982). The river reaches considered were all sufficiently long to encompass a representative selection of either meanders or number of braids for the river system as a whole. The data sets comprised 40 gravel-bed streams, (24 multithread and 16 singlethread channels), and 20 sand-bed channels, (6 multithread and 14 singlethread streams). The variables in the data set included total sinuosity, discharge, valley gradient, and bed material size.

Total sinuosity

Total sinuosity was calculated for most of the braided reaches using the method described above. First order channels were digitised from either published large-scale maps compiled from aerial photographs, or from planform maps of braided networks especially compiled for the present study from vertical air photographs. For several of the braided reaches of New Zeland streams the total sinuosity was taken from the data set of Mosley. Sinuosity for most of the singlethread rivers were derived from published data, and the data set of Church et al. (1981).

Taking a value for total sinuosity from one planform map or aerial photograph may not give a representative figure for the average channel system conditions. However, in comparing different rivers in order to infer general relationships between form and process this problem can be minimised by taking as large a sample as possible so that any deviations from equilbrium conditions may tend to cancel out (Ferguson, 1977b). Values of total sinuosity for the multithread reaches ranged from 2.05-9.0 and for the singlethread channels from 1.3-2.78.

Discharge

The question of the dominant discharge or most effective discharge, in terms of the active transport processes responsible for the morphology of gravel-bed braided streams has hardly been addressed. The dominant discharge, or the most effective discharge in terms of the morphology of the stream created by active sediment transport processes, may be defined with reference to the flow which performs most work, where work is considered with reference to sediment transport (Wolman and Miller, 1960). In gravel-bed rivers much of the material stored in bars remains immobile during most flows, the frequency with which material is moved being dependent on the character of the sediment and the hydrological regime. However, as bedload transport of gravel demands the exceedance of a threshold shear stress, the most effective discharges are likely to be higher flow events. Pickup and Warner (1976) calculated the total bedload transport associated with a range of discharges, using the Meyer-Peter and Muller transport equation for streams transporting fine gravel and coarse sand. In the sections investigated, the return period of the effective discharge ranged from 1.15 to 1.4 years - a flow less than the most probable annual flood and bankfull discharge. Flood flows greater than this may individually transport greater loads but occur too infrequently to have a cumulative effect on channel morphology. Investigating the effective discharge for 15 rivers of widely differing sizes and grain size distributions in Colorado and Wyoming, Andrews (1980) suggests that the effective discharge has a recurrence interval of 1.18 to 3.26 years in the annual flood series.

Considering event frequency and morphological adjustment of upland streams in England Harvey _et al_. (1979) and Thompson (in press) suggest that major events occurring from once or twice a year to events with a recurrence interval of 1.8 years on the partial duration series (2.3 years on the annual series) are the creating events for channel morphology. In the gravelly braided Langden Brook, major events (LC 2 events) cause substantial movement of sediment including clasts over 200 mm, formation of new gravel bars and bank erosion. It is the highest class of event, the LC 1 event, however that results in completely new channel alignments as a result of channel avulsion. On a larger scale, Werritty and Ferguson (1980) describe channel avulsion in the upper braided reach of the River Feshie after the largest flood on the gauging record at Feshiebridge (the 1961, 200 m^3s^{-1} flood). Subsequent gravel transport events have resulted in the simplification and rationalisation of the braided pattern, with floods equal to the 1.5 year flood and the mean annual flood causing substantial sediment transport, bank erosion and bar migration (Ferguson and Werritty, 1983).

These studies suggest that, for gravel-bed streams, substantial amounts of sediment transport and bar development take place during moderate floods. High magnitude events are often accompanied by channel avulsion and destruction of the previous channel configuration. Gravel-bed braid modes may experience short term fluctuations about a dynamic equlibrium form (Harvey, et al., 1979; Thompson, in press), as was suggested by the analysis of the White River data. Effective discharges for braided channels may possibly be represented by moderate to relatively high magnitude floods with recurrence intervals of about 1.0 to 3.0 years. The mean annual flood, with a recurrence interval of 2-2.3 years may therefore be reasonably representative of significant channel-forming discharge for gravel-bed braided streams.

Data for the mean annual flood, which may be indexed by events of return period of 2.0 - 2.3 years in the annual flood series (Church et al. 1981), were available for the majority of the rivers in the sample data set. Where these data were not available, values quoted for the 1.5 year flood or bankfull discharge were used. Discharges for the sample rivers ranged from 1.9 - 5500 m^3s^{-1}.

Valley slope

Total sinuosity should be related to the initial power expenditure rate per unit length of channel. This requires the use of the initial gradient before any channel sinuosity develops. This is best approximated by the use of valley gradient.

An index of stream power was calculated for each stream using the discharge-valley slope product.

Bed material size

For stability and resistance considerations some relatively large grainsize is usually used to represent the distribution of coarse bed material. The b-axis diameter of the clast sizes representing D_{90} and D_{84} have frequently been used and can be justified on the grounds that it is the coarsest elements of the bed material which provide the roughness elements for the river flow. B-axis diameter of D_{84} is therefore used as an index of resistance to erosion. For the gravel-bed rivers D_{84} varied from 36mm to 225mm.

(b) Multiple Regression Analysis

Multiple regression and partial correlation analyses offer a means of empirically testing the hypotheses proposed above. Multiple regression analysis of total sinuosity against stream power and grainsize for the 60 sand-bed and gravel-bed reaches in the data set shows that total sinuosity varies as a continuous function of stream power and bed material size as

$$\Sigma P = 4.4\ D_{84}{}^{-0.03}\ \Omega^{0.24} \qquad r^2 = 0.74 \qquad 6.9$$

where

ΣP = total sinuosity

D_{84} = the $84^{th.}$ percentile intermediate particle diameter mm

Ω = stream power index

The partial correlation coefficient for stream power is significant at greater than the 0.001 level and the partial correlation coefficent for grainsize is significant at the 0.05 level for 57 degrees of freedom for a two tailed test.

However, because total sinuosity must have a limiting value of 1, a functional relationship of the form

$$\Sigma P = 1+k\ D_{84}{}^{-n}\ \Omega^{p} \qquad 6.10$$

is required. The same data were therefore fitted to an equation of this form giving the relationship

$$\Sigma P = 1+ 3.42\ D_{84}{}^{-0.06}\ \Omega^{0.40} \qquad r^2 = 0.73 \quad 6.11$$

with the partial correlation coefficient for stream power significant at greater than the 0.001 level and that for the grain size term significant at the 0.05 level.

Plotting the data points on the total sinuosity-stream power plane with continuous curves for grainsizes of D_{84} from 10mm to 100mm shows that some data points have not fallen onto their correct curves of bed material size (Figure 6.10). There may be several reasons for this:-

(1) the functional form of the relationship between total

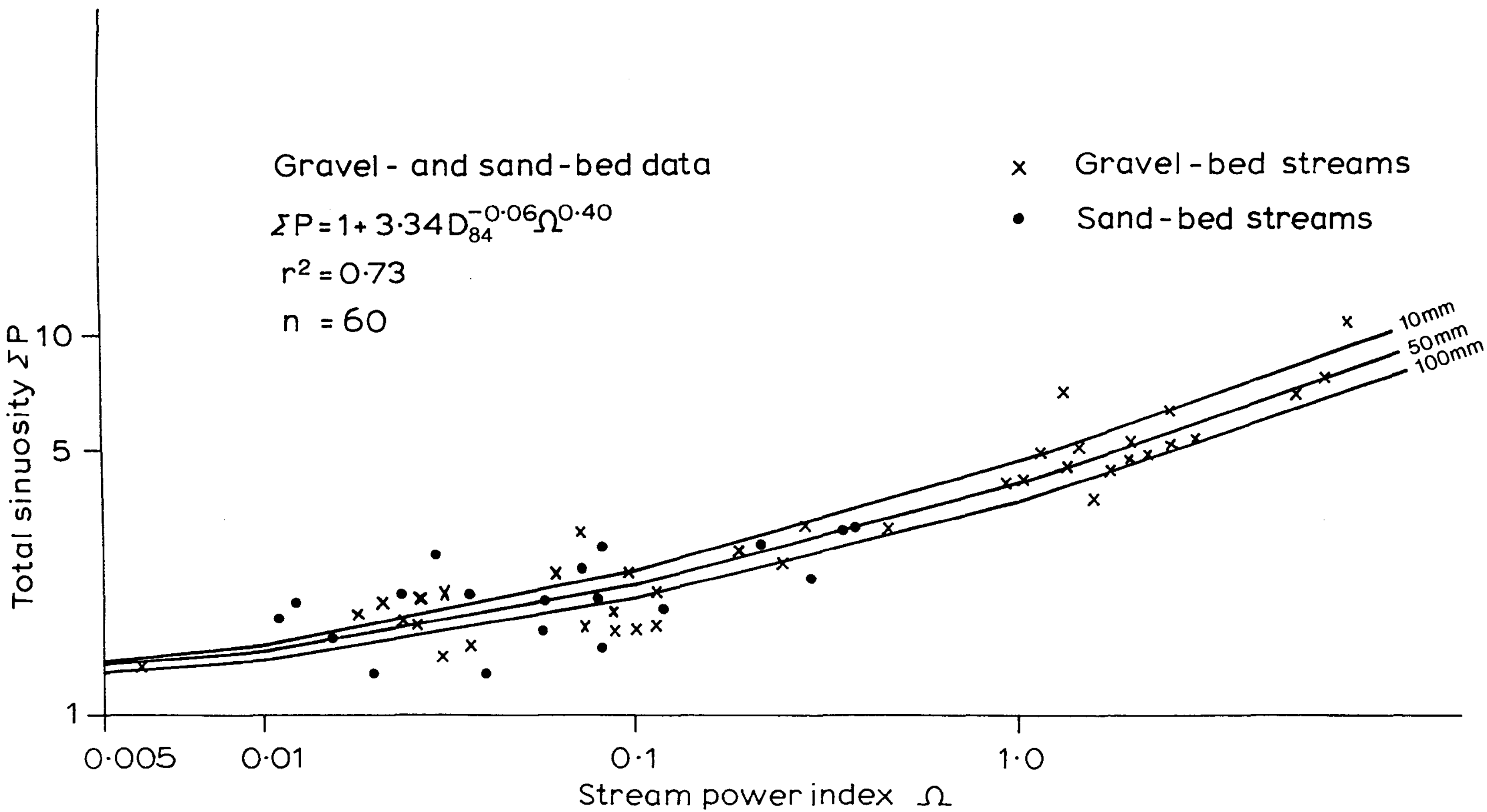


Figure 6.10 Multiple Regression of Total Sinuosity and Stream Power : Grain Size

sinuosity and the three independent variables, stream power (discharge x slope) and D_{84} may vary for sand-bed and gravel-bed rivers;

(2) one or more independent variables influencing total sinuosity have been omitted from the analysis;

(3) limitations of the data set.

In order to determine if separate continua would better describe the data, and hence test the hypothesis that the functional form of the relationship between total sinuosity and stream power will vary with the nature of the perimeter sediments, separate multiple regression analyses were carried out for the gravel and the sand-bed data sets. These analyses yielded the following relationship for the gravel-bed data set

$$\Sigma P = 1+5.52 D_{84}^{-0.138}\ \Omega^{0.403} \qquad r^2 = 0.86 \qquad 6.12$$

and

$$\Sigma P = 1+2.64 D_{84}^{-0.44}\ \Omega^{0.38} \qquad r^2 = 0.48 \qquad 6.13$$

for the sand-bed data set. The partial correlation coefficients for stream power for both data sets are significant at greater than the 0.001 level for stream power and 0.05 for D_{84} with 37 and 17 degrees of freedom respectively.

Formal tests of equality were carried out to test for homogeneity of the two relationships. Parameters which correspond to the differences in the intercepts, coefficients for stream power, and coefficients for D_{84}, were derived. Applying t-tests to the ratios of the parameter estimates and their standard errors, statistical differences in the intercept coefficients and the D_{84} exponents are significant. Only the coefficients for stream power are not statistically different. A single regression model does not provide an acceptable description of both the gravel-bed and sand-bed data sets, as they require different intercepts and coefficients for the grain size terms. Two separate continua are required to describe the sand-bed and gravel-bed channels.

6.5 Discussion

The differences in the parameters for the two equations describing the funtional form of the relationships between channel pattern morphology and the controlling variables of channel planform for the sand-bed and gravel-bed channels, emphasise the differences between the two sedimentological types of channels. Both intercepts as well as the coefficients for the grain size term differ between the two equations. The importance of the sedimentological differences between channel types has been noted by both Ferguson (1981) and Church and Jones (1982). One of the most notable differences between the two sedimentological types of channels is the mobility of the macrobar forms of the sand-bed channels in a wide range of flows. Gravel-bed channels on the other hand are immobile during most flows below a threshold flow for transporting gravel. Considering the sand-bed equation the coefficient of determination is low, only 48%, the variance of Y attributable to the regression therefore being slightly less than half. This indicates that over half of the variance in Y must be due to variables not included in the regression. This may reflect the different magnitude-frequency characteristics of gravel-bed and sand-bed rivers. It was discussed above that gravel requires the exceedance of a threshold before initiation of motion will occur. This is likely to occur at some flood flow with a recurrence interval of about 1.0 to 3.0 years. However, channel/bar complexes in sand-bed rivers are much more active for a wider range of flows, as sand moves in all but the lowest flows (Church and Gilbert, 1975; Parker and Petersen, 1980). The total sinuosity may not be as stable for sand-bed streams over a range of flows which includes 1.0 - 3.0 year floods. Floods as large as the 2.3 year flood may be inappropriate as the channel-forming for sand-bed channels. Further, variables not included in the analysis may have important controls on the development of channel pattern morphology in sand-bed channels. It is likely that bank material properties will have an important control on channel shape as bank retreat rates and the processes of bank erosion are strongly affected by the percentage of silt and clay in the perimeter sediments (Schumm, 1960). However, insufficient data were available to attempt further analysis.

For the gravel-bed data, equation 6.12 helps to verify the hypothesis that channel pattern morphology varies as a continuous function of stream power and bed material size. The independent variables stream power and D_{84} account for 86% of the variance in the dependent variable, total sinuosity. This lends support to the use of total sinuosity as a quantitative index describing channel pattern morphology for meandering and braided streams, at least for gravel-bed rivers. Braided streams may thus be quantified in terms of degree of channel splitting and sinuosity. Greater stream power is needed for braiding than meandering, and increasing stream power is required for progressively higher total sinuosities or degrees of braiding. Thus, the braided streams plotted on the Parker regime diagram (Figure 6.6) have progressively higher stream powers as they proceed from the 1-2 braid band to the plotting position for streams over 10 braids, so that total sinuosity varies as a direct function of stream power. Equation 6.12 therefore confirms the observations of Howard et al. (1970), Chang (1979) and Begin and Schumm (1984) that high sinuosity singlethread streams will braid at higher stream powers and that intensity of braiding increases with stream power. The high coefficient of determination for equation 6.12 suggests that the three independent variables used in the regression analysis, discharge, valley slope and D_{84}, are, as suggested by the studies of Leopold and Wolman (1957), Howard et al (1970), Ferguson (1973), Parker (1976), Chang (1979), and Richards (1982), important variables controlling the development of channel pattern morphology in free alluvial gravel-bed streams. The 14% unexplained variance is likely to be due to measurement errors, the exclusion of a bank material properties variable which accounts not only for the effects of percentage of fines but more particularly the binding effects of bank vegetation, and the difficulty of using D_{84} as a summary statistic for resistance to entrainment and transport.

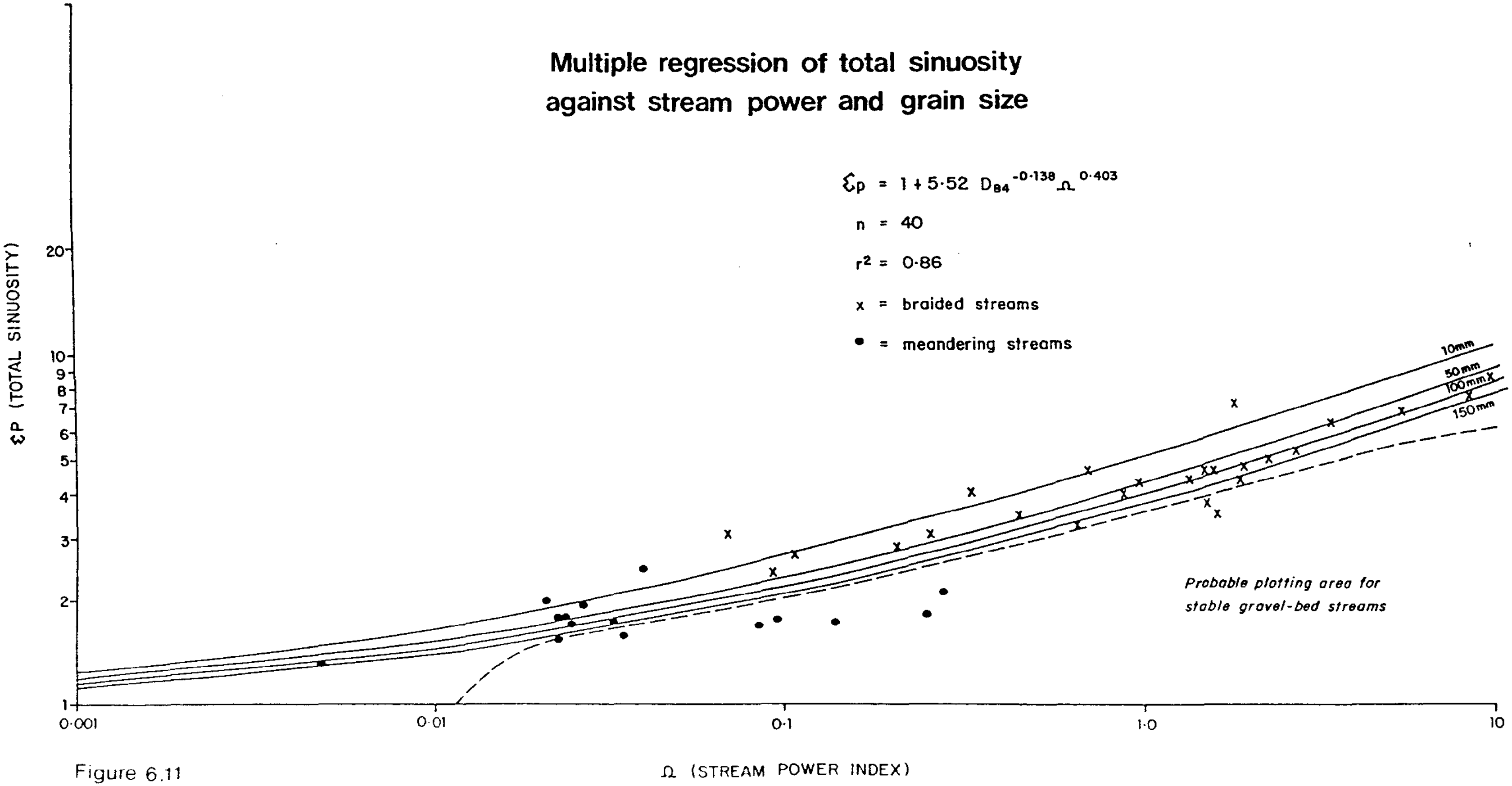
The multivariate expression, equation 6.12, for the gravel-bed data shows that sediment properties interact with stream power to determine channel pattern morphology in gravel-bed rivers. This was also demonstrated but in a different way by the multivariate discriminant function of Ferguson (1984)

$$s = 0.017 Q_b^{-0.49} D_{90}^{0.27} \qquad 6.14$$

who stressed the need for a family of discriminant functions to define the threshold slope for a given discharge and grain size needed for braiding. The advantage of quantifying channel pattern morphology on a single numerical scale is that, as opposed to the discriminant function approach to channel pattern development, channel pattern morphology is not initially assigned to a nominal scale pattern type according to some preexixting classification scheme. Channel pattern morphology is objectively defined and then related in a single multivariate model to the controls of planform development, discharge, valley slope and sediment characteristics.

This statistical model suggests that channel pattern morphology in free alluvial gravel-bed streams results from the interaction between two sets of variables measuring the force applied by the fluid and the resistance to erosion offered by the perimeter sediments. Temporal and spatial variation in one or more of the controlling variables will have a corresponding effect on the channel pattern as the channel adjusts its morphology to maintain a dynamic equilbrium with the new hydrological or sedimentological conditions. Greater stream power is required to maintain a given total sinuosity as the size of D_{84} increases; this is demonstrated from the continuous grain size curves plotted on the total sinuosity-stream power plane (Figure 6.11). For a stream power index of 1 total sinuosity may vary from 3.7 to 5.7 as D_{84} varies from 10mm to 150mm. As the continuous grain size curves diverge with increasing stream power, the range of total sinuosity for changing grain size (holding stream power constant), will increase with greater stream powers. For example, with a stream power index of 10 total sinuosity varies from 8.7 for 150mm grain size to 13.9 for a grain size of 10mm.

Streams that plot below the envelope (Figure 6.11) formed by the constant grainsize curves represent rivers which have a total sinuosity less than would be expected for their stream power.



Multiple regression of total sinuosity against stream power and grain size
$\Sigma p = 1 + 5{\cdot}52\ D_{84}^{-0{\cdot}138}\ \Omega^{0{\cdot}403}$
n = 40
r^2 = 0·86
x = braided streams
• = meandering streams
10mm
50mm
100mm
150mm
Probable plotting area for stable gravel-bed streams
ΣP (TOTAL SINUOSITY)
20
10
9
8
7
6
5
4
3
2
1
0·001
0·01
0·1
1·0
10
Ω (STREAM POWER INDEX)

Figure 6.11

These rivers are relatively inactive streams with an inherited perimeter sedimentology or streams with heavily vegetated banks. Some streams may be limited in sediment supply or may be incised. These rivers are not free to develop an equilibrium total sinuosity commensurate with the prevailing stream power. The probable plotting area for these relatively stable gravel-bed streams is illustrated in Figure 6.11.

This analysis has shown that discrete channel pattern states are less apparent when common indexes of pattern morphology are considered, although they may be more applicable to gravel-bed rivers than to sand-bed channels. The formation of different channel pattern types involves, to varying degrees of intensity, the processes of bank erosion, bed material transport, channel/bar construction, and the creation of pattern morphology as a mechanism for the dissipation of excess energy. Channel lengthening and increased bed area in both meandering and braided streams may be seen as morphological adjustments which minimise the rate potential energy loss per unit bed area.

For gravel-bed rivers different pattern morphologies in free alluvial streams are a response to spatial variation in stream power and bed material calibre. A strong relationship between a quantitative measure of channel pattern morphology, total sinuosity, and three controlling variables of channel pattern morphology has been established. This gives empirical support to already existing strong theoretical and physical arguments for a unified approach to channel pattern morphology. The results of the analysis suggest that, for gravel-bed rivers, the dimensionless variable, total sinuosity, is an appropriate parameter with which to quantify channel pattern morphology. Channel pattern morphology may be represented as a continuum in terms of a set of continually measurable variables.

CHAPTER 7

THE PREDICTION OF STREAM POWER AND THE MEAN ANNUAL FLOOD

7.1 Introduction

The concept of a channel pattern continuum related to flow and sedimentological characteristics of the stream offers the potential for the development of a new approach to palaeohydrology for gravel-bed rivers. This approach is in contrast to previous palaeohydraulic approaches which reconstruct flows for braided streams using individual clasts from former braid bar and channel deposits.

7.2 Palaeohydrological Equations : Stream power

As Richards (1977) and Williams (1985) have pointed out, relationships established by least squares regression techniques between variables describing river planform and parameters of river flow, should not simply be inverted to solve for the desired parameter of the flow regime; for example, inverting the linear equation

$$Y = a + bx \qquad 7.1$$

to solve for X as

$$X = (y - a)/b \qquad 7.2$$

introduces new constants which are the result of the mathematical manipulation which has produced equation 7.2. However, these new constants may be invalid and therefore introduce potentially large errors in the estimates of X. This rearranged equation also ignores the error distribution of Y in the original least squares regression so that confidence limits cannot then be applied to the manipulated equation.

If it is intended to predict parameters of the paleoflow regime from variables describing river planform, a new regression should be performed on the original data set (Richards, 1977; Williams, 1985). Confidence limits may then established for the error distribution of the new Y. This is the correct method to derive an equation that will best estimate the parameter of the flow that is being recovered.

As it was intended to predict stream power from total sinuosity and D_{84} the original data set used to derive equation 6.12 (Chapter 6) was fitted to an equation of the form

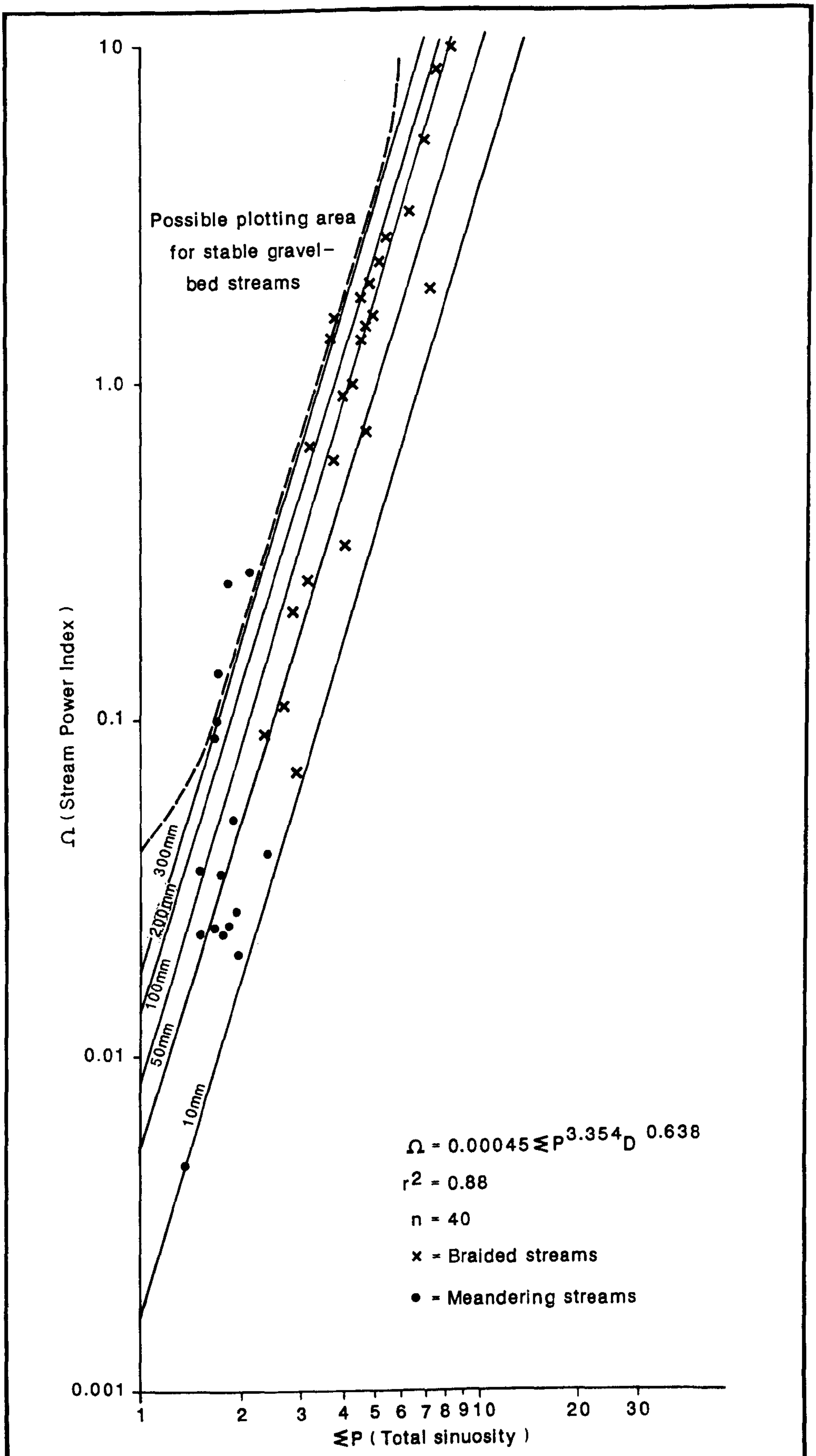
$$y = k\ a^{n}\ b^{p} \qquad 7.3$$

using least squares regression. The dependent variable, y, is the stream power index and the independent variables are total sinuosity and grain size. This analysis produced the relationship

$$\Omega = 0.00045\Sigma P^{\ 3.354}\ D_{84}^{\ 0.638} \qquad r^2=0.88 \qquad 7.4$$

where the partial correlation coefficient for D_{84} is significant at the 0.01 level with a two-tailed test and the partial correlation coefficient for ΣP significant at greater than the 0.001 level. The plotted points and constant grain size lines are shown in Figure 7.1.

The construction of confidence intervals for the multivariate expression in equation 7.4 has been made following the method



Stream power plotted as a function of stream total sinuosity and grain - size

Figure 7.1

given by Snedecor and Cochran (1967). The quantities Σx^2, Σx, x_2 and $\Sigma x_2 2$ used in the computations for the least squares regression form the matrix

$$A = \frac{(\Sigma x_1^{\;2} \quad \Sigma x_1 x_2)}{(\Sigma x_1 x_2 \quad \Sigma x_2 2\ )} \qquad 7.5$$

where

$\Sigma x_1 2$ = the sum of the squared deviations of the first independent variable

$\Sigma x_2 2$ = the sum of the squared deviations of the second independent variable

$\Sigma x_1 x_2$ = the sum of the products of the deviations from x_1 and x_2.

The inverse of this matrix is

$$A^{-1} = \begin{pmatrix} C_{11} & C_{12} \\ C_{21} & C_{22} \end{pmatrix} \qquad 7.6$$

The elements C_{ij} are the Gauss multipliers required for calculation of the confidence intervals. They are found by solving two sets of simultaneous equations

<u>First set</u>

$$C_{11}\ \Sigma x_1 2 + C_{12}\ \Sigma x_1 x_2 = 1 \qquad 7.7$$

$$C_{11}\ \Sigma x_1 x_2 + C_{12}\ \Sigma x_2 2 = 0 \qquad 7.8$$

<u>Second set</u>

$$C_{21}\ \Sigma x_1 2 + C_{22}\ \ \Sigma x_1 x_2 \qquad = 0 \qquad 7.9$$

$$C_{21}\ \Sigma x_1 x_2 + C_{22}\ \ \Sigma x_2 2 \qquad = 1 \qquad 7.10$$

From equation 7.4

$\Sigma x_1 2 \quad = 1.9818$

$\Sigma x_2 2 \quad = 1.9067$

$\Sigma x_1 x_2 \quad = 0.6259$

Substitution of these values into the first set of simultaneous equations gives

$$C_{11}\ 1.9818 + C_{12}\ 0.6259 \qquad = 1 \qquad 7.11$$

$$C_{11}\ 0.6259 + C_{12}\ 1.9067 \qquad = 0 \qquad 7.12$$

and solving gives

$C_{12} \quad = -0.1848$

$C_{11} \quad = \ 0.5629$

Substituting values for C_{12} into the second set of equations gives a value of C_{22}

$C_{22} \quad = 0.5851.$

The formula for the estimated standard error of Y is (Snedecor and Cochran, 1967)

$$S_y = s((^1/_n + C_{11}\ x_1 2 + C_{22}\ x_2 2 + 2C_{12}\ x_1 x_2)^{0.5})t_{0.05} \qquad 7.13$$

where

s = standard error of estimate

$x_1{}^2$ = squared deviation of X_1 from x_1

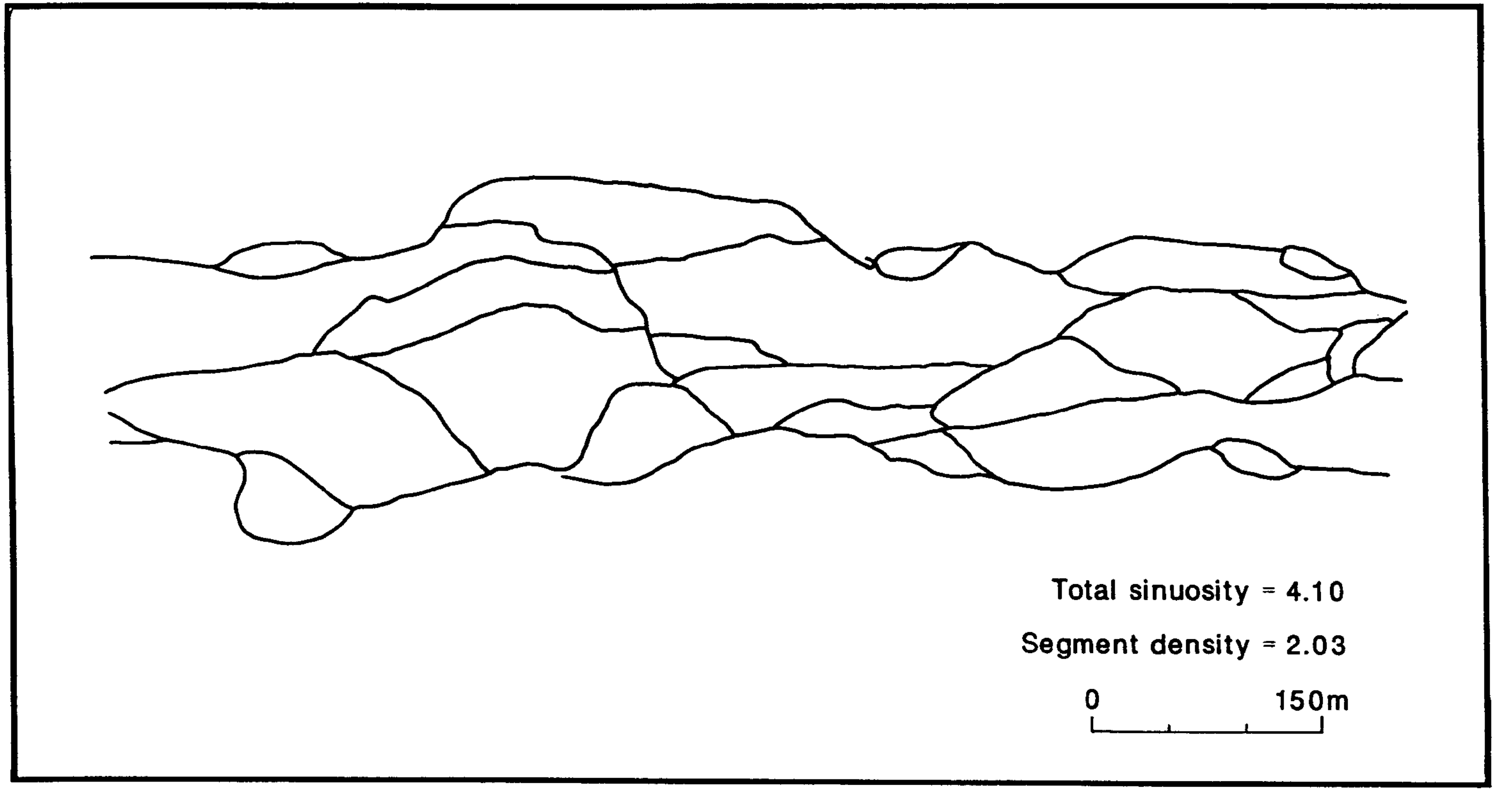
$x_2{}^2$ = squared deviation of x_2 from X_2

x_1x_2 = product of the deviations of x_1 and x_2 from the x_1 and x_2

Data for the upper braided reach of the River Feshie were not included in the regression analysis for either equation 6.12 or equation 7.4. The ability of equation 7.4 to predict stream power may thus be tested using data from the upper braided reach of the River Feshie for which the stream power index and mean annual flood is known. A planform map of the active area of the upper braided reach at a scale of 1:5000 was prepared from stereopairs of vertical air photographs. The midline of the channels between the first order bars was then plotted to give the planform of the first order channels (Figure 7.2). The total length of the first order bars was then digitised for the reach and the total sinuosity calculated from equation 6.7 (Chapter 6). The value for total sinuosity of the reach was calculated as 4.1. D_{84} for the upper braided reach averages 110mm. Inserting values of 4.1 for total sinuosity and 110mm for D_{84} into equation 7.4 gives a predicted stream power index of 1.025. Dividing this through by the valley slope of 0.012 gives a predicted mean annual flood of 85$m^3\ s^{-1}$. This prediction is at the mid-range for the 80-90 $m^3\ s^{-1}$ range for the mean annual flood estimated from the gauging data. Confidence limits may be calculated for this estimate as

$$S_y = s((^1/_{40} + 0.5629(0.0204) + 0.5851(0.0178) + 2(-0.1848\text{x}0.01914)^{0.5})t_{0.05} \qquad 7.14$$
$$= +/- \ 0.12$$

The interval estimate for the predicted stream power index of



Channel network of the upper braided reach of the River Feshie

Figure 7.2

the River Feshie is 0.91-1.135 and dividing these values through by slope gives a confidence interval of $76m^3\ s^{-1}$ $-94.5m^3\ s^{-1}$ for flood approximately equal to the 2-2.3 year flood. Equation 7.4 thus gives an accurate prediction of the mean annual flood for the upper braided reach of the River Feshie.

7.3 Palaeohydrological equations : Segment density index

(a) Introduction

Total sinuosity is derived from the quotient of the total channel length and reach length. This can only be measured over the complete active channel zone of a braided reach. However the total sinuosity of the active channel zone of the palaeo-river which constructed the terrace deposits cannot be estimated from the palaeochannels preserved on the surfaces of the terrace fragments. This is because of the uncertainty as to the extent of post-depositional reworking of the original floodplain (Chapter 5). A portion of the original active channel zone will be left preserved on the terrace surface, but it cannot be known how much of the original active zone this actually represents. A method is therefore required which will allow the estimate of a value for the total sinuosity of the whole active zone without reference to the original channel width.

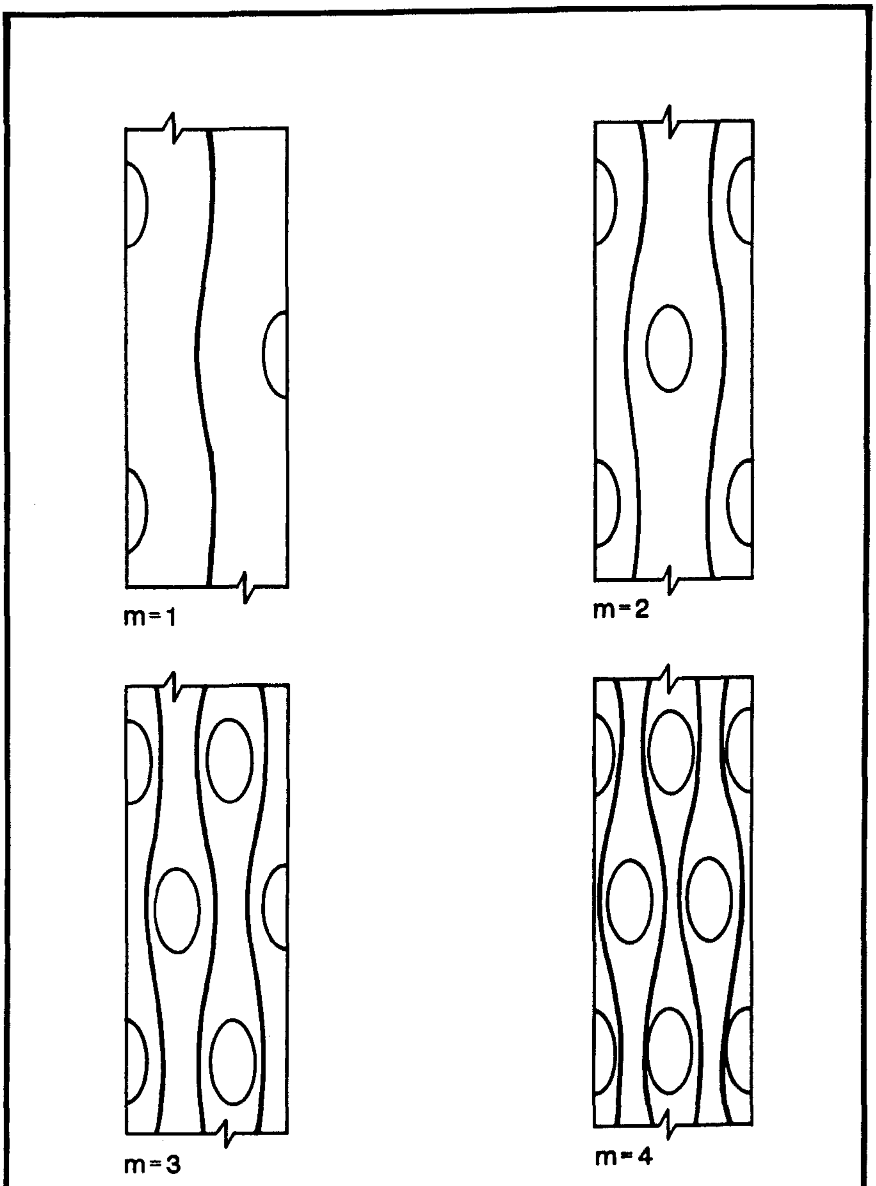
A solution to this problem may be found in an examination of the spatial arrangement of the channels within a braided channel system. This may be represented by several characteristics of the braided channel network.

(b) Channel Density

The prevailing braid mode or total sinuosity of a braided reach may be approximated by the number of channels at a cross-section (Chapter 6). Figure 7.3 shows the bed pattern associated with various braid modes in Parker's (1976) stability analysis of meandering and braiding. The diagram shows that as the braid mode or total sinuosity increases, the number of channels at a cross-section increases. Holding channel width constant, the density of the channels or total length of channel per unit area of active zone must also increase. Channel density of a braided network is therefore a network characteristic which is a measure of the increase in path length that is associated with the physical processes of braiding.

As total sinuosity and channel density are both related to the increase in path length that is associated with the creation of bars and channels in the braiding process, there should be a close and predictable relationship between channel density and total sinuosity. The channel density of a braided stream network is analogous to the drainage density of a drainage basin. As with drainage density, a measure of channel density should, by definition, be the same if calculated for a part of the braided reach or from the whole braided reach. The development of a relationship between channel density and total sinuosity would then allow the prediction of total sinuosity from the channel density calculated from a part of the network of the braided stream.

However, channel density has the dimensions of L^{-1} and is therefore not dimensionless. In order to compare braided streams of different sizes a dimensionless parameter is required.



The bed patterns associated with various braid modes (after Parker, 1976)

Figure 7.3

(c) Average Segment Length

A braided stream network may be characterised by channel segments (anabranches), and nodes where segments branch or join (Howard et al.,1970). The topological definition of these properties is shown in Figure 7.4. Deriving a number of dimensionless ratios which include measures of channel segments, Howard et al., (1970) have demonstrated that geometric similarity is preserved between braided streams with the same total sinuosity, but of differing sizes of active channel zone. It was also shown that segment length could be statistically correlated to a number of size parameters of the channel system of a braided reach.

Segment length could therefore act as a scaling variable for channel width and provide the additional length dimension required to develop a dimensionless relationship between channel density and total sinuosity. In order to investigate this further average segment length was calculated for 22 braided reaches by digitising individual segments between nodes and dividing the sum of all segment lengths by the number of segments in the reach. This was plotted against width of the active channel zone, measured for each of the 22 reaches (Figure 7.5). The derived relationship

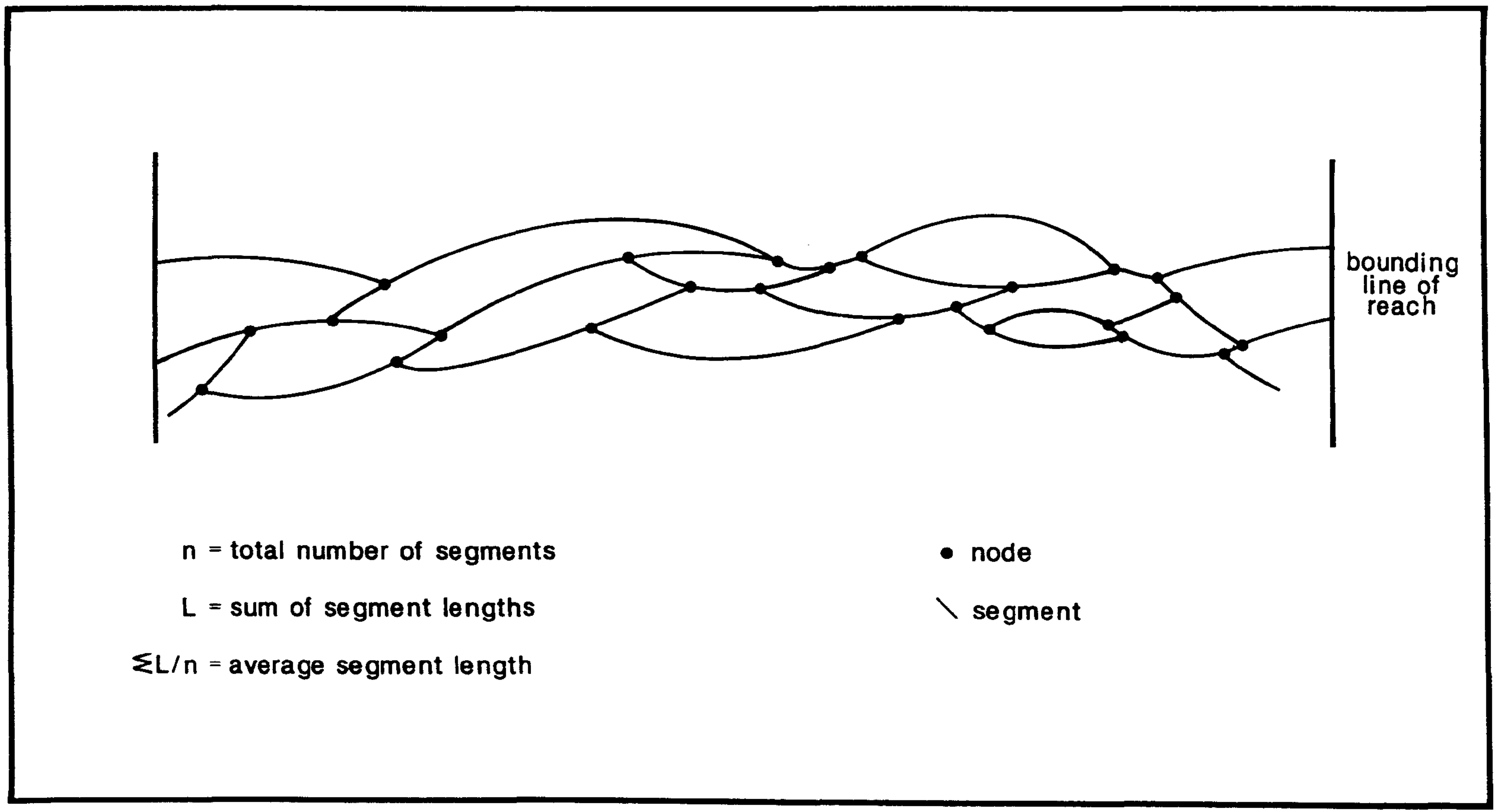
$$seg_1 = 2.96 + W0.514 \qquad r^2 = 0.92 \qquad 7.15$$

where

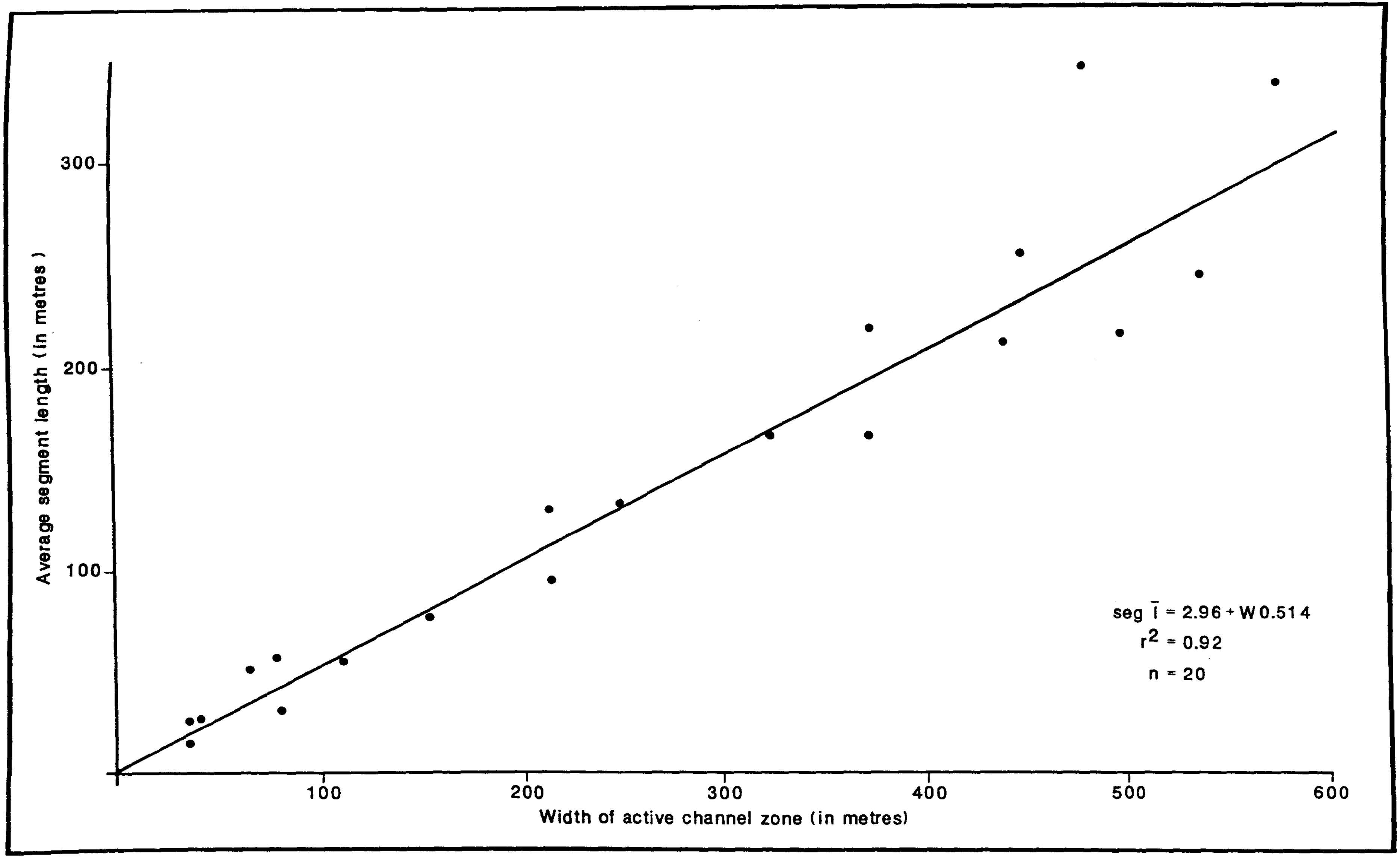
seg_1 = average segment length

W = width of the active channel zone in metres

shows a strong linear correlation between channel width and average segment length, with average segment length increasing



Topological variables of braided streams (after Howard et al, 1970) Figure 7.4



Plot of active channel zone width against average segment length

Figure 7.5

directly with channel width. Average segment length may therefore act as a scale variable for channel width.

The two network parameters, channel density and average segment length, may be combined to form a dimensionless variable, <u>the segment density index</u>. The segment density index is defined as

$$seg_d = (\Sigma L/A).\ seg_l \qquad 7.16$$

where

seg_d = segment density index

ΣL = total channel length

seg_l = average segment length

This is the product of braided channel density and average segment length. In dimensions this gives

$$\frac{L.L}{L^2} = 0 \qquad 7.17$$

This index is interpretable in terms of the physical processes of braiding, both the increase in path length with the creation of bars and channels and the increase in channel width as rates of sediment transport increase with higher degrees of braiding. As total sinuosity is also interpretable in terms of the same physical processes, the two indices of channel pattern morphology are physically related and may be seen as two indices measuring the same quantity, channel pattern morphology.

Streams of the same total sinuosity should therefore have the same segment density index regardless of size of the braided system. This is illustrated with the following example. Figure 7.6 shows two braided reaches of differing size. Measuring the first order channels, the braided reaches illustrated in Figure

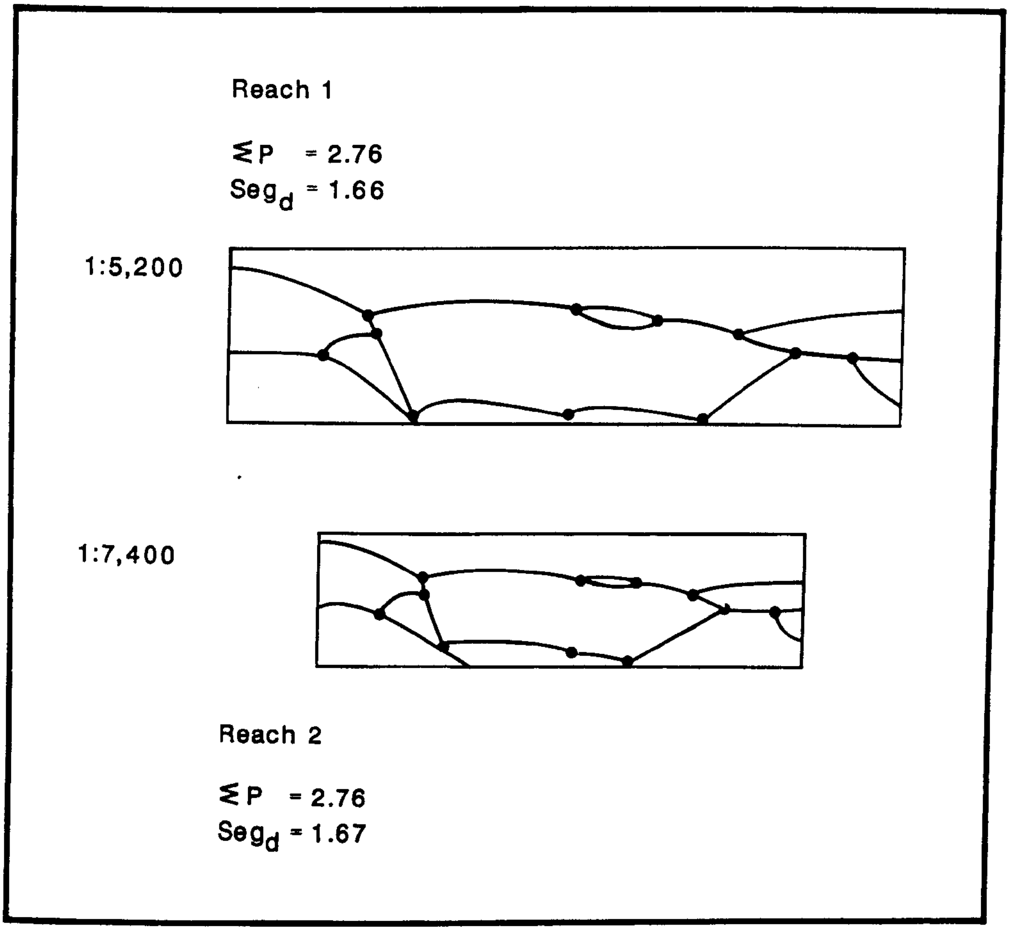


Figure 7.6

Diagram to show scale - independence of segment Density Index

7.6 have the same total sinuosity, 2.76. Reach 1 however has a larger active zone than Reach 2. The channel density for Reach 1 is 0.021 and that for Reach 2 is 0.031. The average segment length for Reach 1 is 78m whilst that for Reach 2 is 54.5m. Calculating the segment density indices for the two reaches gives 1.66 and 1.67 respectively.

A stream with a braid mode or total sinuosity of n should have a distinguishing set of network characteristics, represented by the two properties of channel density and average segment length which combine to form the segment density index. For each value of segment density there should be a unique value of total sinuosity. In order to test this hypothesis statistically, the segment density index and the total sinuosity was calculated for 22 braided stream reaches of widely differing sizes and total sinuosities. Plotting segment density index against total sinuosity for each reach demonstrates a high statistical correlation (r = 0.96) between the two variables.

Least squares regression analysis was used to establish the nature of the relationship between the two variables, segment density and total sinuosity. A least squares regression of Y on X assumes that X is the independent variable and Y is the dependent variable. In a given set of data there should be a clear dependence of one variable on the other (Till, 1973). In a situation when there is a problem in assigning dependency, the reduced major axis regression line is the correct type of linear fit to apply to the data set. The Y on X least squares regression minimises the sum of the squares of the residuals from the regression line using the vertical distance of the individual data points from the best-fit line; reduced major axis regression minimises the sums of the areas of triangles formed by projecting orthogonal lines from the data points to the regression line until it is bisected.

In the case of the relationship between the segment density index and total sinuosity, which are two variables measuring the same quantity of channel pattern morphology, strictly, reduced major axis regression would be the preferable method for fitting a regression line. However, as it was intended to calculate confidence bands for the regression, and as the direction of prediction is to be from the segment density index to total sinuosity, a Y on X regression was performed with segment density as the predictor (X) variable and total sinuosity as the dependent (Y) variable. The relationship between segment density and total sinuosity is illustrated in Figure 7.7 and has the form

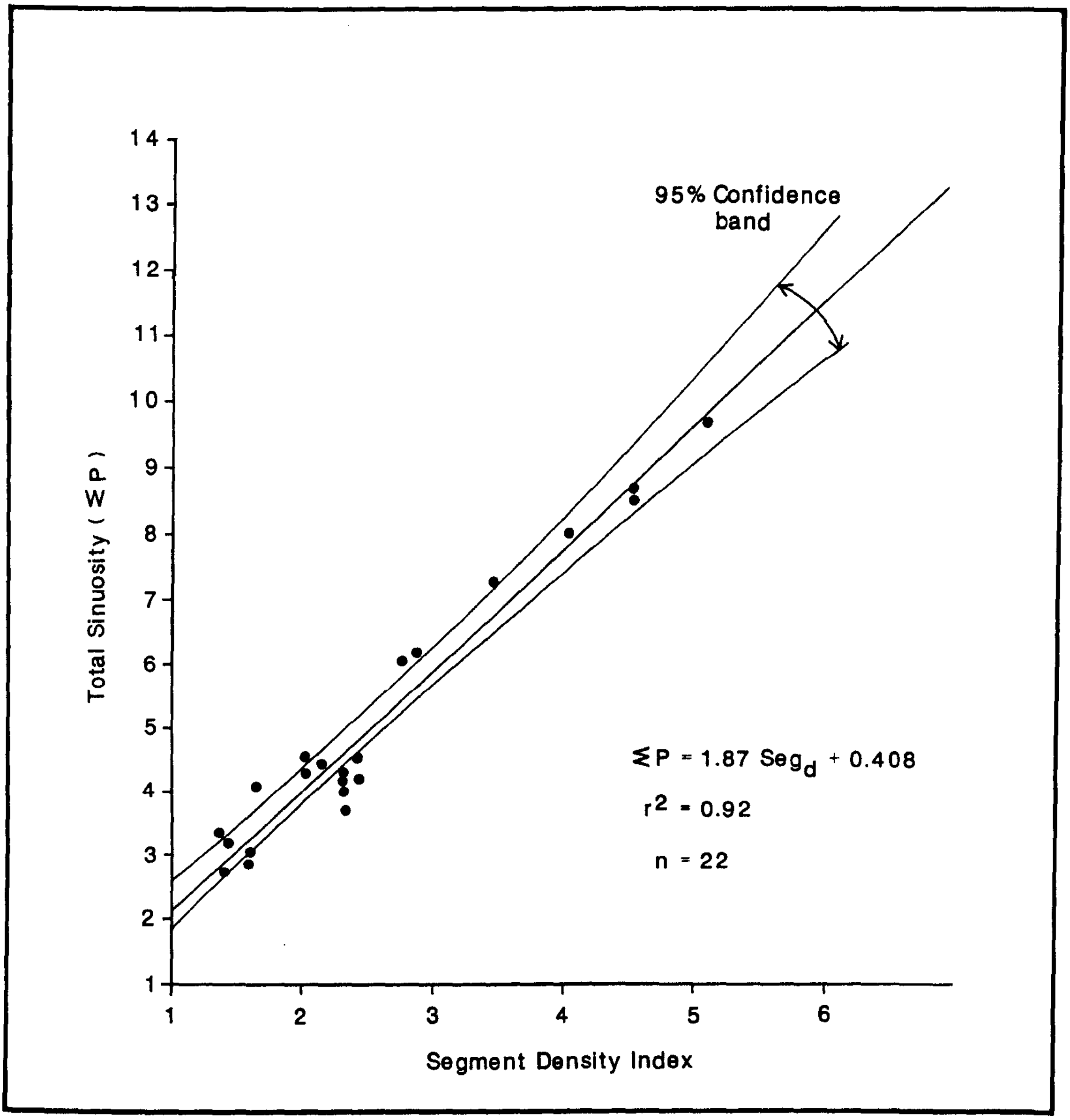
$$\Sigma P = 1.87\ Seg_d + 0.408 \qquad r^2 = 0.92 \qquad 7.18$$

where

ΣP = total sinuosity

Seg_d = segment density index.

There is a direct linear relationship between segment density and total sinuosity, so that as the segment density increases there is a corresponding increase in total sinuosity. The high coefficient of determination (0.92) supports the acceptance of the hypothesis that for each value of the segment density index there is a unique value for total sinuosity. Thus for example, a segment density index of 2.0 corresponds to a total sinuosity of 4.15, whilst a segment density index of 2.5 has a total sinuosity of 5.08. Prediction of total sinuosity from the measurement of the segment density index is therefore feasible within the range of values for the data set. This is as



Plot of Segment Density Index against Total Sinuosity

Figure 7.7

expected, because the two variables are both dimensionless, physically related quantities derived from parameters which can describe only one given channel pattern morphology. The 8% unexplained variance is accountable to measurement error.

An unbiased estimate of the true variance about the regression line is given by the residual mean square, with n-2 degrees of freedom for a bivariate regression. It is denoted by $S^2_{y.x}$. The square root of this quantity, $S_{y.x}$, is the standard error of the estimate or the standard deviation of Y holding X constant. A single standard deviation does not apply to all predicted vales of Y, but must depend on the X value that determines the Y population. For a sample of Y values with a fixed set of X values, then x (the mean of X values), is a constant while y (the mean of the Y values), and b (the slope of the regression line) are variables. Variation in y raises or lowers the regression line parallel to itself; the effect of this is to increase or decrease all estimates of means by a fixed value. Variation in b rotates the regression line about the point x;y and the effect on an estimate of Y depends on the magnitude of X - x. Variation in b has no effect on the estimate if X = x, but otherwise the effect of variation in b will increase in proportion to X - x.

A standard deviation applicable to an estimate must allow for variation in both y and b for the distance X-x. The estimated standard error of predicted Y is (Snedecor and Cochran, 1967)

$$S_y = S_{\hat{y}.x} \left({}^{1}/_{n} + x^2 / \Sigma X^2 \right)^{\frac{1}{2}} \qquad 7.19$$

with n-2 d.f.

where

$S_{y.x}$ = standard error of the estimate
$x^2 = (X-x)^2$
$\Sigma X^2 = \Sigma (X-x)^2$

Confidence limits can be attached to the estimates of predicted Y by multiplying the appropriate standard error by the tabulated t value for the 95% confidence level. Plotting these values in conjunction with the regression line forms the confidence band plotted in Figure 7.7. This band is therefore applicable point by point. The true value of Y estimated from the regression line is therefore 95% certain to lie within 2 standard errors either side of the estimated value.

For example, for a segment density index of 2.0 the best estimate of total sinuosity is given by

$$S_y = 0.466((0.045 + (2.0 - 2.308)^2)/14.81)^{\frac{1}{2}} \, x t_{0.05}$$
$$= +/-0.21 \qquad 7.20$$

A segment density index of 2.0 substituted into equation 7.18 predicts a total sinuosity of 4.14. The predicted value therefore has a 95% chance of lying between 3.92 - 4.35.

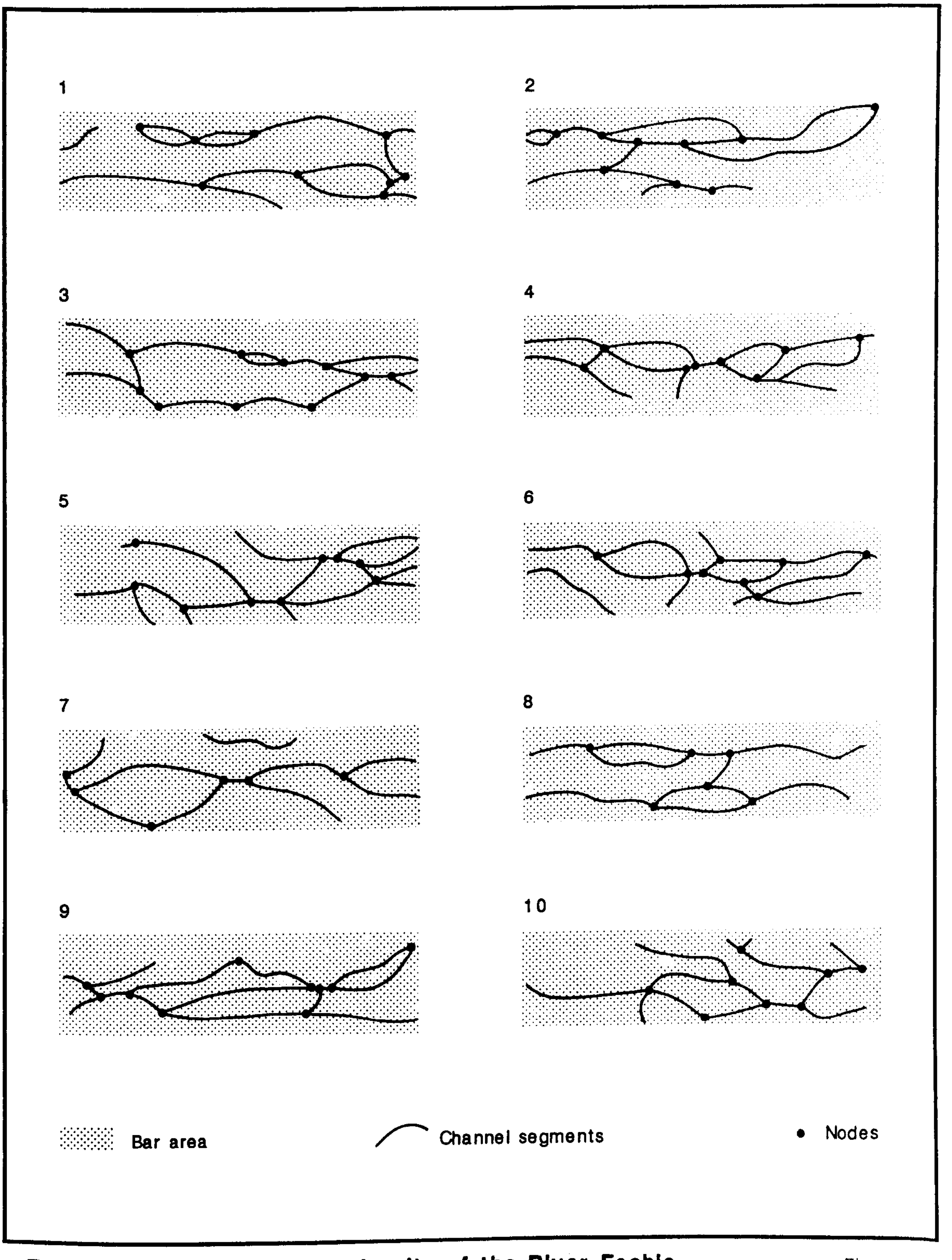
7.4 Prediction of total sinuosity from the segment density index calculated from an active braided reach

A stream with a braid mode or total sinuosity of n will have a distinguishing set of network characteristics. These are represented by the two properties of channel density and average segment length which combine to form the segment density index. In consequence measurements of the segment density index made from the whole braided reach and from a part of the braided reach should be equivalent as long as the section of the braided reach measured contains at least some complete segments connected by nodes. As with estimates of drainage density from quadrat analysis some variation of the segment density index between sections of the active zone is to be expected.

In order to test this with some field data, the segment density index was calculated for the upper braided reach of the River Feshie using the plot of the first order channels in Figure 7.2. The total active channel length of the first order bars was digitised for the reach, as well as the area of the active channel zone. The segment density for the upper braided reach of the River Feshie is 2.01 and the total sinuosity is 4.1.

Segment density indices were then calculated for 10 areal units of the active channel zone of the upper braided reach. The size of the areal units used to calculate the segment densities for parts of the upper braided reach was determined by the area of the low-level terraces bordering the floodplain of the river. Scaled rectangles representing the areas of the terrace fragments were located on the planform plot of the river using random numbers tables, having first superimposed a grid onto the planform map. The total active channel length and area of each sample was digitised, the average segment length calculated and the segment density index computed. The networks of each areal unit for the stream are shown in Figure 7.8.

If it is assumed that the areal units from which the segment density index is to be calculated represent a random sample from a normal population then it is possible to attach confidence limits to the predicted values of total sinuosity using the method given above. The confidence intervals for each segment



Ten randomly selected areal units of the River Feshie active channel zone

Figure 7.8

density measurement were calculated. Table 7.1 gives the results of the attempts to reconstruct the total sinuosity of the upper braided reach of the River Feshie from the segment density indices of ten randomly located sections of the active channel zone using equation 7.18.

TABLE 7.1

Seg_d	ΣP	Error	Confidence limits	
1.86	3.88	-5.2%	3.64	4.11
1.95	4.05	-1.3%	3.82	4.27
1.95	4.05	-1.3%	3.82	4.27
1.96	4.07	-0.65%	3.85	4.29
1.98	4.10	0.0%	3.89	4.43
2.00	4.14	0.97%	3.92	4.36
2.04	4.21	2.8%	3.99	4.43
2.10	4.32	5.5%	4.10	4.53
2.18	4.47	9.2%	4.26	4.68
2.20	4.51	10.0%	4.30	4.72

True Seg_d, 2.01; ΣP, 4.1

The mean segment density index for the 10 samples is 2.02. This represents a mean error of 0.47% from the segment density index calculated from the whole reach. The standard deviation of the ten samples is 0.102 and the coefficient of variation is 5.08%. The small coefficient of variation suggests that the spatial variation of the segment density index, as measured for first order bars, is relatively slight. The total sinuosity estimates using equation 7.18 are all within 10% of the true value of 4.1. However, two of the ten confidence intervals miss the true value for the total sinuosity, this occurring where the segment density index is higher due to a spatial concentration of channels in particular sections of the active zone. This is to be expected as there may be a random element in the distribution of channels in the network (Howard et al., 1970; Church, 1972). An estimate of the true segment density will therefore be achieved better by taking more than one sample for the reach in question and averaging the calculated segment densities.

This analysis shows that the segment density of part of a braided reach is not likely to vary significantly from a measure of segment density taken from the whole reach. More than one estimate for segment density is preferable; but, if a statistical approach to prediction of total sinuosity is necessary, measurement of the segment density index from only a part of the whole reach should produce a quite accurate prediction of the total sinuosity of the whole active zone. An estimate of channel pattern morphology can be derived from either of two physically-related indices, segment density and total sinuosity. The network parameters which define segment density must be unique to each value of the index. As the segment density index and total sinuosity are directly related, measurement of the segment density index from a set of braided palaeochannels preserved on a terrace surface should give a reliable estimate of the total sinuosity of the active channel zone that created the particular set of network characteristics measured by the segment density index.

7.5 The palaeohydrological application of the equations for estimating total sinuosity and stream power

The segment density index provides a statistical means by which total sinuosity of the whole active zone may be predicted from the segment density index measured from either a part of a braided network or from a complete active channel zone network. A statistical approach to the prediction of total sinuosity has been developed because the braided palaeochannel networks of a whole active reach are rarely preserved in the fossil fluvial landscape. Total sinuosity cannot be measured directly from the palaeochannels preserved on the terrace fragment surfaces because of the uncertainity of the scale of post-depositional reworking of the original surface by younger channel systems. If a part of the palaeochannel network is preserved complete with nodes and channel segments, then because the properties of the segment density index enable it to be calculated for a network of any size, regardless of the area of the original surface, the total sinuosity of the whole original network may be predicted. Confidence limits may be attached to the predicted values of total sinuosity for each palaeonetwork measured.

Once the total sinuosity of the palaeochannels has been predicted, prediction of the stream power index and ultimately the flood approximately equal to the 2-2.3 year flood is possible.

The methodology embraced in the total sinuosity approach to discharge reconstruction is outlined below, and is then tested with data from the modern upper braided reach of the River Feshie. It should be noted that equation 7.4 may also be used for reconstructing the stream power index of meandering palaeo-channels. This is provided that sufficient channel traces are available to acquire a reasonable estimate of channel sinuosity. Ferguson's (1977a) method for estimating channel sinuosity from the direction variance of the meandering palaeochannel traces should be used to estimate sinuosity.

The methodology

(1) The channel network, in which nodes and channel segments may be identified, is plotted as a scaled planform map from vertical air photographs.

(2) The total channel length of the network is digitised and the average segment length computed.

(3) The area of the channel network is digitised and the segment density index computed from equation 7.16

$$Seg_d = (\Sigma L/A).seg_l$$

(4) The total sinuosity is predicted from equation 7.18

$$\Sigma P = 1.87\ Seg_d + 0.408$$

(5) Stream power is predicted by inserting the appropriate values for total sinuosity and D_{84} into equation 7.4

$$\Omega = 0.00045\ \Sigma P^{3.354}\ D_{84}{}^{0.638}$$

where D_{84} has been previously estimated.

(6) The discharge approximately equal to the mean annual flood is calculated by dividing the stream power index through by valley gradient, estimated from the gradient of the terrace.

(7) Confidence limits may be attached to both the estimates of total sinuosity and stream power.

There is no standard method for combining the confidence limits from the two least squares regressions, the total sinuosity against segment density index and stream power against total sinuosity and grain size, to produce one final set of confidence limits for the prediction of the flood discharge. It is likely that simply combining the 95% confidence limits from both equations will produce limits that are approximately equal to the 98-99% level, although it is not possible to specify exactly at what level the limits are being set (Constable, personal communication). The final confidence limits for the discharge estimates were therefore calculated using the following procedure. The confidence limits are calculated for the estimated value of total sinuosity as outlined above using equation 7.19

$$S_Y = S_{\hat{Y}.X} \left({}^{1}/_{n} + x^2 / \Sigma X^2 \right)^{\frac{1}{2}}$$

with n-2 d.f.

The interval estimate for Y is made by multiplying Sy by a t value corresponding to the 0.05 level. The upper and lower limits for the predicted total sinuosity are then used to predict two values for the stream power index from equation 7.4. The confidence limits are then established for the two stream power indices using equation 7.13

$$S_Y = s\left(\left({}^{1}/_{n} + C_{11}\, x_1{}^2 + C_{22}\, x_2{}^2 + 2C_{12}\, x_1 x_2\right)^{0.5}\right) t_{0.05}$$

for n-3 degrees of freedom. The lower limit of the first estimate for stream power and the upper limit for the second estimate for stream power are then taken as the final confidence limits for the predicted stream power. This method gives the most conservative possible estimate for the confidence intervals for the final stream power and flood estimate.

The segment density indices calculated from the randomly located areal units of the upper braided reach of the River Feshie were used to test the estimates of a final stream power index and flood discharge using this palaeohydrology method against known values for these parameters of the flow. The results of these calculations are given in Table 7.2.

TABLE 7.2

Segment density = 1.86

Estimates		Confidence limits (95%)	
ΣP	3.88	3.64	4.11
Ω	0.85	0.58	1.13
Q	$71 m^3 s^{-1}$	$48.3 m^3 s^{-1}$	$94.5 m^3 s^{-1}$

Segment density = 1.95

Estimates		Confidence limits (95%)	
ΣP	4.05	3.82	4.27
Ω	0.98	0.708	1.304
Q	$82 m^3 s^{-1}$	$59 m^3 s^{-1}$	$108 m^3 s^{-1}$

Segment density = 1.98

Estimates		Confidence limits (95%)	
ΣP	4.10	3.91	4.35
Ω	1.025	0.74	1.38
Q	$85 m^3 s^{-1}$	$62 m^3 s^{-1}$	$115 m^3 s^{-1}$

Segment density = 2.04

Estimates		Confidence limits (95%)	
ΣP	4.2	3.98	4.38
Ω	1.11	0.808	1.449
Q	$92 m^3 s^{-1}$	$67.4 m^3 s^{-1}$	$120.7 m^3 s^{-1}$

Segment density = 2.1

Estimates		Confidence limits (95%)	
ΣP	4.32	4.11	4.52
Ω	1.22	0.908	1.556
Q	$100 m^3 s^{-1}$	$75.5 m^3 s^{-1}$	$129.6 m^3 s^{-1}$

Segment density = 2.2

Estimates		Confidence limits (95%)	
ΣP	4.51	4.30	4.72
Ω	1.41	1.27	1.77
Q	$117 m^3 s^{-1}$	$106 m^3 s^{-1}$	$148 m^3 s^{-1}$

This Table shows the results of attempts to reconstruct the flood approximately equal to the 2-2.3 year flood for the upper braided reach of the River Feshie from the ten random sample areas. For five of the samples the best estimates for the flood discharge are within the 80-90m^3s^{-1} range, whilst three are within +/- 10m^3s^{-1} of the range.

The confidence intervals for the predicted stream power index and flood discharge for eight out of ten of the samples all include the present stream power index of 0.96-1.08 and the present 2-2.3 year flood of 80-90m^3s^{-1}. Thus, for eight out of ten of the samples there is a 95% chance that the true value for the present mean annual flood lies with the discharge range given by the confidence intervals. For the two highest segment density indices of 2.18 and 2.2 the confidence intervals miss the true values. Table 7.2 shows that the total sinuosity approach to discharge estimation can give a reasonably accurate prediction of the true mean annual flood for the modern River Feshie. This then suggests that the method applied in a palaeohydrological context, to palaeochannels on a terrace surface, should be able to give useful estimates of the mean annual flood of the past rivers which created the bar and channel complexes being measured.

In Chapter 5 it was shown that the methodology currently used to reconstruct some of the parameters of the palaeoflows which constructed braided stream terrace deposits is unsatisfactory for a number of reasons. These are:-

(1) The method is based on a number of hydraulic relationships between parameters of the flow and the grain sizes present in a fluvial deposit. However, the use of some of these relationships embraces a number of unsupported assumptions.

(2) The methodology requires the extrapolation of parameter

estimates from at least four equations which means accumulating a hierarchy of errors. Confidence limits are not calculated and as each new estimated X will be subject to error the final error in the discharge estimate is likely to be high but unknown.

(3) The currently used equations can only predict local depths and velocities, that is, point estimates for various parameters of the flow. The extrapolation of these local estimates to the wide, morphologically complex cross-sections of gravel-bed braided streams is not sound particularly when coupled with the uncetainity of the prior channel width.

(4) The method cannot identify the frequency of the reconstructed discharge.

The advantages of a palaeohydrological methodology based on a ralationship between total sinuosity, stream power and grain size are :-

(1) It allows the retrodiction of past discharges directly from channel pattern morphology thus bypassing the need for a hierarchy of calculations based on uncertain deductions concerning the velocity or shear stress required for the incipient motion of gravel particles.

(2) The method allows the estimation of a discharge which is likely to approximate the mean annual flood.

(3) The method does not require an estimate of the total channel width of the prior active zone of the former

braided stream, therefore removing a serious potential source of error. One of the main sources of error in the palaeohydraulic method is likely to be the problem of estimating channel width. As the width of the surviving remnant is used in the palaeohydraulic reconstruction the resultant discharge is partly a function of the amount of the former floodplain that has survived reworking processes.

(4) Confidence limits may be set for the point estimates from the regression relationship.

The total sinuosity method of discharge reconstruction may therefore yield a more reliable pattern of results than the palaeohydraulic method.

CHAPTER 8

THE PALAEOHYDROLOGY AND PALAEOENVIRONMENTAL SIGNIFICANCE OF THE GLEN FESHIE RIVER TERRACES

8.1 Introduction

The present study has re-assessed the temporal significance of the terrace sequence in Glen Feshie and has provided evidence for major geomorphological development in the Holocene in the form of at least three phases of terrace development. This evidence questions the generality of the hypothesis that terrace formation was confined to periods of deglaciation in upland Scotland and necessitates a revision of the interpretation of the younger members of the terrace sequence in at least one major valley in upland Scotland.

Explanations for the development of individual terrace sequences in upland Britain have been given in terms of one or other of three main conceptual models. Each of these varies in the background environmental changes thought to induce terrace development. First, the concept of the proglacial development of terrace sequences in upland Scotland as propounded by Sissons arose from observations of upland valleys where terraces are associated with former ice-dammed lakes, such as those at Glen Doe, Glen Roy, Glen Spean and Achnasheen (Sissons, 1979, 1982). In these valleys terrace development has occurred in response to catastrophic draining of ice-dammed lakes at a time when there was an abundant supply of meltwaters from downwasting ice. In this situation Sissons' conclusions are ratified by tracing terrace fragments back to ice contact features and moraines. However, as the present study has attempted to show, the extrapolation of such a model of proglacial terrace formation to all terrace sequences in upland Scotland cannot be supported.

Second, Maizels (1983a, 1983b) elaborated the model of terrace formation in response to fluctuating discharges of proglacial rivers (1983). However, this model has been applied in several contexts to undated terrace deposits which have not been demonstrated to be palaeooutwash deposits, such as some of the lower terrace levels of the River North Esk, Scotland.

Third, late Holocene alluviation has frequently been interpreted within a conceptual framework encompassing time lags, thresholds and complex response together with the underlying background factor of increased sediment supply to the streams as a result of man-induced vegetation changes. For example, the terraces in Dovedale Griff in the North York Moors are suggested to be a result of complex response (Richards et al.,in press). However, problems may arise in relating alluvial response to general environmental conditions when dating control is minimal and when deposition and erosion can both arise because of extreme events. Although these may be more probable under some climatic/hydrological regimes, their random nature does not preclude their occurrence at any time, hence complicating environmental interpretation (Richards et al. in press; Harvey, 1986).

This chapter uses the total sinuosity approach to palaeodischarge estimation developed in Chapters 6 and 7 to predict the flood discharges of the prior braided streams of the River Feshie. An attempt is then made to place the findings of the palaeohydrology estimation and the soil stratigraphy within the broader context of Holocene environmental change and landform development.

The interpretation of the Glen Feshie terraces which is given below suggests that no single model of terrace development can adequately account for the formation of all terrace levels. This is because terrace development is likely to take place against a variety of background environmental conditions. Further, in highland valleys temporal variation in environmental conditions is likely to be combined with spatial variation in the sensitivity of the geomorphic landscape to erosion. This may occur particularly where development of the terrace sequence is complicated by spatial variation in the depth of the valley fill/bedrock interface.

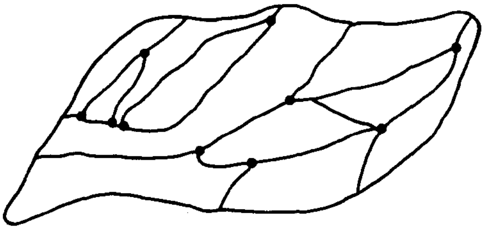
8.2 The River Feshie terraces : prediction of the flood approximately equivalent to the mean annual flood

The total sinuosity approach to predicting discharges was tested using data from the upper braided reach of the River Feshie (Chapter 7). The results showed that quite accurate estimates could be made of flood discharges approximately equal to the mean annual flood. It should be noted however, that it is difficult to be too precise about the return period involved because informaton is sparse concerning the relationship between braided channel morphology and formative or dominant discharge events. Nevertheless, the results suggest that if channel networks can be accurately plotted and the valley slope and D_{84} of a gravel-bed stream is known then the total sinuosity approach to prediction of past discharges can be applied with some confidence to braided palaeochannel networks on terrace fragments.

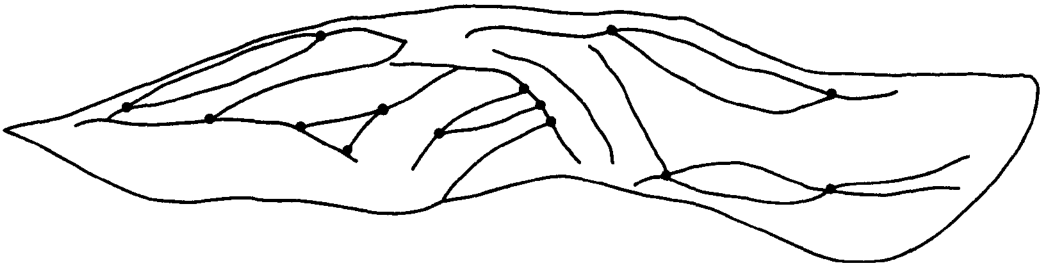
The River Feshie terrace fragments have been grouped into five terrace levels using age-dependent soil-stratigraphic data. Various methods of dating control have allowed the terrace levels to be assigned ages. Two groups of terraces are suggested to be of Late Devensian age, at 13,000BP and 10, 000BP respectively. The former is extremely extensive (Figure 3.20) whilst the latter is very limited in spatial extent. Three late Holocene terraces of 3,600BP, 1,000BP and 80BP age have been identified. Of these three surfaces, the 3,600BP is the most extensive with fragments of this surface being found at all five reaches studied.

Each terrace system, apart from the very restricted 10,000BP surface, exhibits terrace fragments with well preserved braided palaeochannel systems. In the River Feshie the possibly rapid abandonment of the former floodplains together with the lack of fines in the catchment has prevented the infilling of old channels. These therefore remain as distinct linear depressions in the former floodplain surfaces (Chapter 3). Some terrace fragments are small in areal extent and do not exhibit palaeochannels with a complete enough network for palaeohydrological analysis. Where the terrace fragments are areally more extensive distinct networks of braided palaeochannels inclusive of nodes and channel segments can be distinguished. These have been photogrammetrically plotted at a scale of 1:5000 and were subsequently checked in the field. The palaeochannels form pronounced topographic lows on the terrace surfaces and this factor coupled with the high optical resolution of the Kern P G 2 and the utilisation of high quality diapositives, meant that no difficulty was experienced in mapping the palaeochannels. The scale of the plotting and subsequent field checking make it clear that the majority of the channels mapped are first order channels. Examples of some of the plotted fossil channel networks preserved on the surfaces of the terraces are shown in Figures 8.1 to 8.3. The complete map of the Glen Feshie

Figure 8.1 **Glen Feshie terrace palaeochannels networks**



PALAEOCHANNELS - FRAGMENT 36



PALAEOCHANNELS - FRAGMENT 30

PALAEOCHANNELS - FRAGMENT 32

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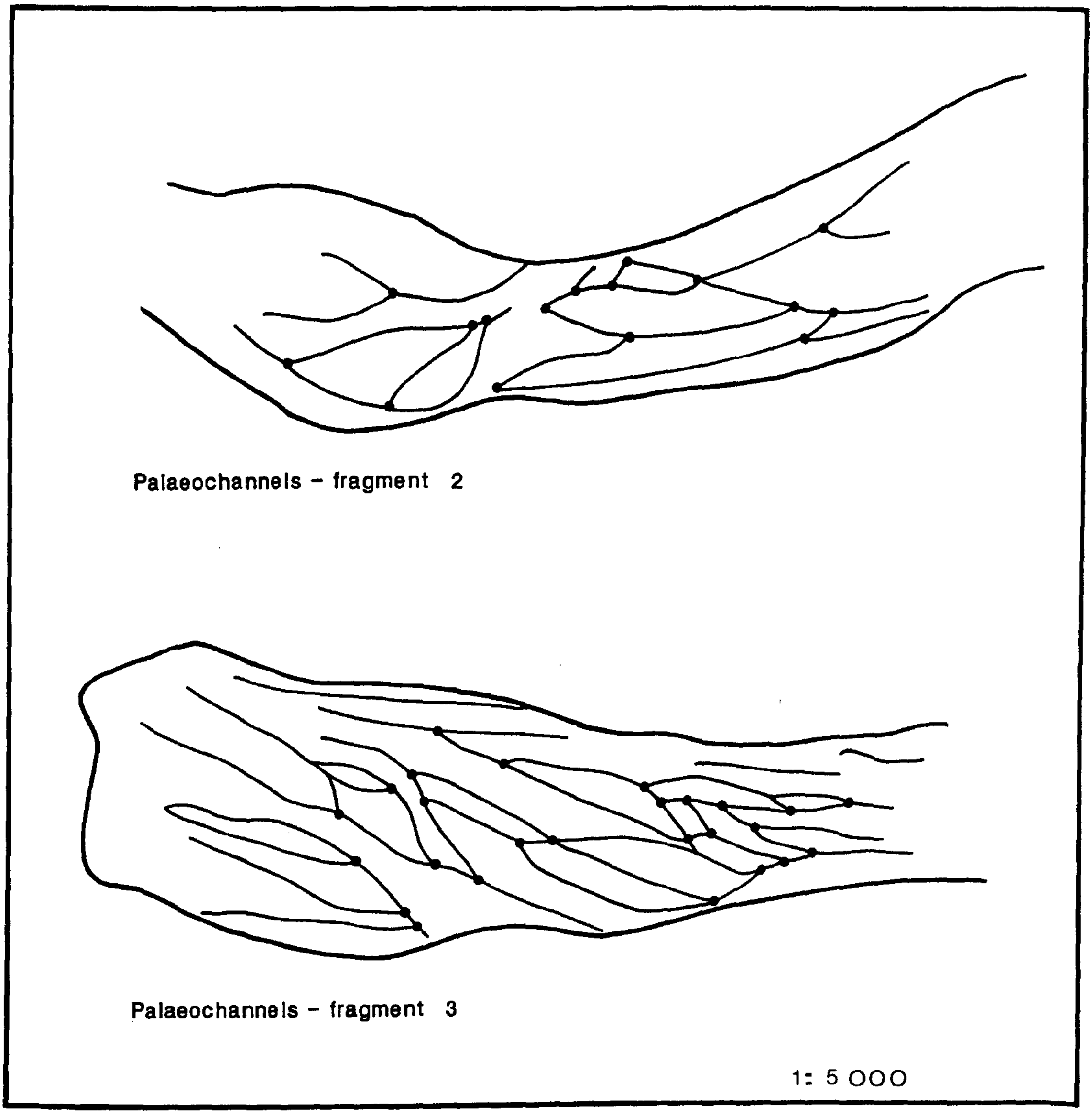
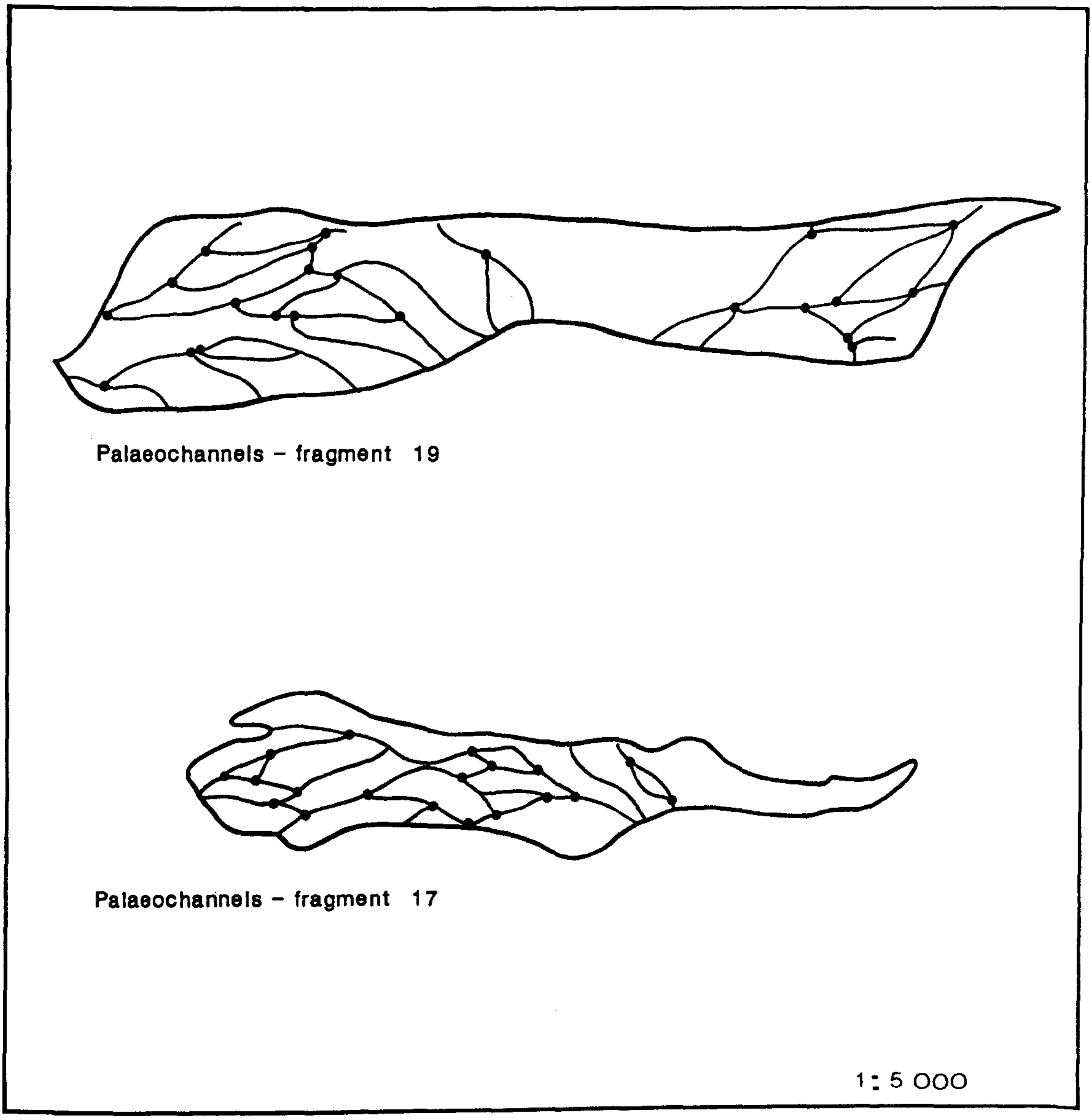


River Feshie braided palaeochannels

Figure 8.2

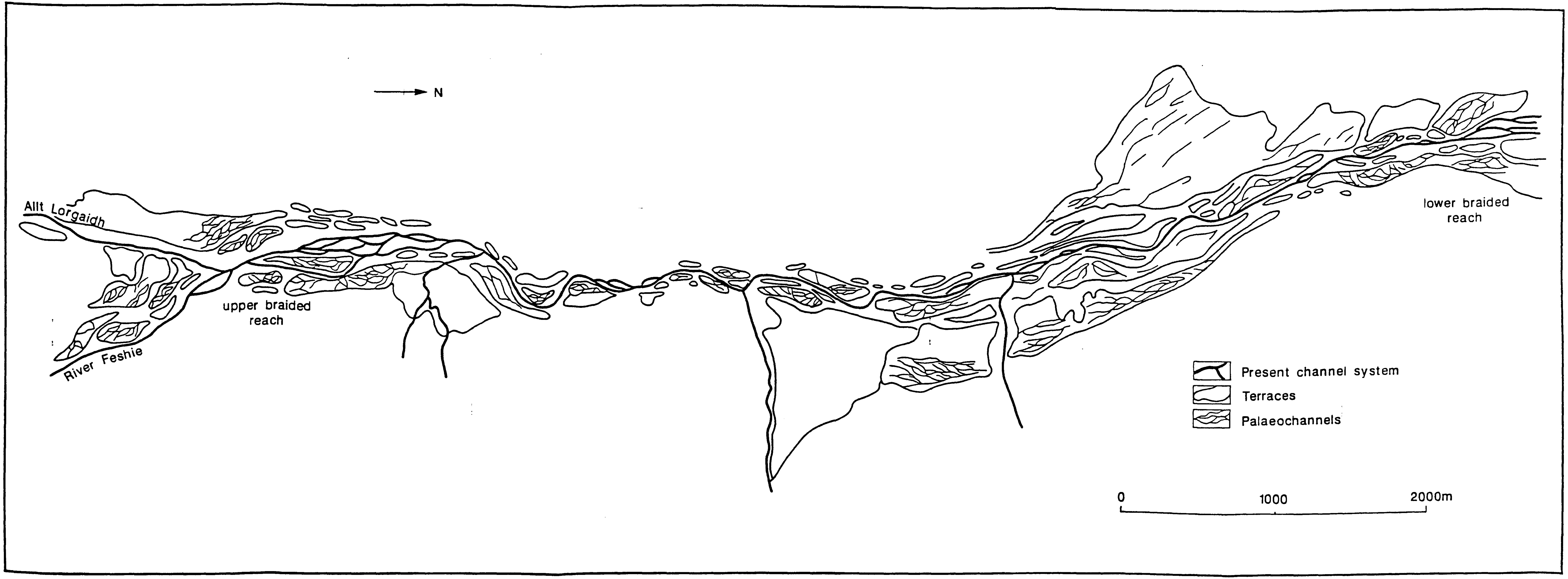


River Feshie braided palaeochannels

Figure 8.3

palaeochannels in the study reach is shown in Figure 8.4.

Sixteen of the terrace fragments were extensive enough and possessed sufficiently well preserved palaeochannel networks to enable 22 palaeodischarge estimates to be made. It was not possible to make any discharge estimates from the 10,000BP surface because the terrace fragments were too restricted in areal extent to enable a braided channel network to be mapped. Fragments of the 10,000BP surface do occur in several reaches along the River Feshie and include the Allt Lorgaidh upper alluvial fan unit, the Allt Fhearnagan fan, the terrace onto which this fan grades and a small fragment below the confluence of the River Feshie and the Allt Fhearnagan. The spread of the fragments along the length of much of the study reach suggests that it was probably a much more extensive surface, but limited preservation of terrace fragments has meant that palaeohydrological analysis could not be carried out for that surface. Palaeodischarge estimates were possible from the 13,000BP, 3,600BP, 1,000BP, and 80BP surfaces. In the case of the 13,000BP surface five individual networks comprising both west and east bank fragments were mapped. The 13,000BP surface does not extend upstream of the Allt Fhearnagan reach. At the level of this reach the outwash merges upstream into kame and kettle deposits. For the 3,600BP surface, terrace fragments have been preserved from all reaches studied so that palaeohydrological estimates of discharge changes for this surface could be made along the whole length of the study area. For the 1,000BP surface, fragments have again been preserved along the length of the study reach. Four estimates were made for discharges from this surface. Only two estimates were made for the 80BP surface.



Distribution of palaeochannel network of the River Feshie

Figure 8.4

The availability of a number of palaeonetworks from several terrace fragments of one terrace surface is valuable in three respects :-

(1) First, a number of discharge estimates are possible from one terrace level. This enables a check to be made on the consistency of the predicted palaeodischarges for an individual terrace level.

(2) Second, downstream changes in discharge for one terrace surface can be computed and compared with the downstream changes in discharge for the modern stream. Relative changes in discharge may then be calculated. Further, consistency of downstream increase in the predicted discharges helps to validate the method used.

(3) Third, because palaeodischarge estimates for the individual terrace levels can be made for a number of cross-sections along the study reach this enables meaningful between terrace discharge comparisons to be made. Generally, previous palaeohydrological estimates have not been controlled for downstream variation in discharge. Estimates have been made for successive terrace levels from palaeochannels which have been in varying locations downstream. The extensive, well preserved palaeochannels on the Glen Feshie terraces have enabled palaeodischarge estimates to be made for successive generations of the River Feshie at one cross-section.

Discharge estimates for the palaeochannel networks on the terrace surfaces were made according to the total sinuosity method of discharge estimation developed and field tested in

Chapter 7. Briefly, the total channel length for each network was digitised and the average segment length was calculated. The area of each terrace fragment covered by the network was also digitised and the segment density index calculated from

$$Seg_d = (\Sigma L/A) \cdot seg_1 \qquad 8.1$$

Where terrace fragments were very extensive and breaks in the network occurred, calculations were made for the individual networks. This was useful because it enabled a check to be made on the segment density indices from more than one part of the prior braided channel system. For example, two calculations were made for fragment 19 (Figure 8.5) giving segment density indices of 2.56 and 2.53. Three calculations were made for fragment 30 giving segment density indices of 2.75, 2.9, and 2.39. Two calculations were made for fragment 9 giving segment densities of 2.34 and 2.41; and three calculations were made for fragment 3 giving segment densities of 4.27, 4.2 and 4.3. The segment density indices calculated from the terrace fragments were used to predict the total sinuosities of the active zones of the former braided reaches using

$$\Sigma P = 1.87\ Seg_d + 0.408 \qquad 8.2$$

Confidence limits were calculated for each total sinuosity estimate using the method described in Chapter 7. The stream power index, the discharge-slope product, of each former braided channel was predicted by inserting the appropriate values for total sinuosity and D_{84} into the equation

$$\Omega = 0.00045\ \Sigma P^{3.354}\ D_{84}{}^{0.638} \qquad 8.3$$

The D_{84} data needed for insertion in equation 8.3 were obtained by the methods described in Chapter 2. The discharge approximately equal to the 2-2.3 year flood was computed by dividing the stream power by the valley gradient of the former braided reach. The valley gradient for each terrace fragment was obtained from the trend surface analysis of the spot height data on each terrace fragment (Chapter 4). Confidence limits for the final discharge estimates were made using the method described in Chapter 7. The data for each set of calculations are given in Appendix 4.

8.3 Calculation of present-day flood discharges for the five study reaches

Since the St. Andrews University gauging station is at the downstream end of the upper braided reach, present day discharge data are available only for the upper braided reach in the River Feshie. In order to compare palaeodischarges with the flood discharges of the modern Feshie, estimates of the flood discharges for the remaining four study reaches were calculated for 5 cross-sections of the River Feshie using a modified slope-area method of discharge estimation (Richards et al., in press). A well-defined trashline along both sides of the channel and deposited from the last high flood, 27 November, 1984, before the survey date in May 1985, enabled the cross-sectional area of flow to be calculated. Subsequent floods between November 1984 and May, 1985 only attained discharges of 20-30m^3s^{-1}, which are within channel flows. The existence of the trashline along the whole length of the reach for each cross-section enabled the water surface slope to be reconstructed. The flood which left the trashline was approximately equivalent to the 2-2.3 year flood, that is 80-90m^3s^{-1} at the upper braided reach gauging station. The location of the cross-sections for which the calculations were made is shown on Figure 8.5. In addition, two cross-sections of the upper braided reach were surveyed in order to compare the predicted discharge with the discharge recorded at the gauging station in attempt to assess the amount of error in predicting modern discharges using the slope-area method of calculation.

The method developed to estimate flood discharges for morphologically complex sections requires the accurate mapping of the form-sediment associations on the channel bed, and the identification of units within each cross-section within which dominant roughness influences can be identified (Richards _et al._, in press). Accordingly, each cross-section was surveyed with a quick-set level and staff. The b-axis of 100 clasts was sampled from the surface layer of particles using the Wolman method (Wolman, 1954). D_{84} for each cross-section was calculated from the cumulative frequency curves. The three braided cross-sections were divided into bar and channel segments and grain-size counts made for each unit.

Using the modified slope-area method, each cross-section was divided into segments, and calculations for roughness, velocity and cross-sectional area of flow made for each segment. At the stage of the mean annual flood, projection of large clasts is drowned out. Following Limerinos (1969) and Richards (1982), the ratio of flow depth to the 84th percentile B-axis measurement was used to calculate a roughness index from

$$1/(f^{\frac{1}{2}}) = 1.16 + 2.0 \log (d/D_{84}) \qquad 8.4$$

where

$(f^{\frac{1}{2}})$ = Darcy-Weisbach friction factor

d	= channel depth	
D_{84}	=the 84th percentile intermediate particle axis	

The Limerinos relationship as redefined by Richards (1982) employs a relative smoothness measure and is close to the theoretical relationship expected from a consideraton of the two dimensional velocity profile. The flood velocity for the segment was estimated from roughness, depth and trashline slope using

$$v = 8.86\ 1/(f)^{\frac{1}{2}}\ (ds)^{0.5} \qquad 8.5$$

where

f	= Darcy-Weisbach friction factor
v	= velocity in m s^{-1}
d	= depth in metres
s	= bed slope

The final flood discharge was calculated by multiplying the velocity estimate for each segment by the segment cross-sectional area. The component discharges were then summed to give the final discharge estimate. The final flood discharges calculated for each cross-section of the present River Feshie are shown on the unbroken line on Figure 8.5.

The flood discharges calculated using the component approach for the upper braided reach were $85m^3s^{-1}$ and $78.5\ m^3s^{-1}$. The close correspondence between the calculated discharges for the two cross-sections as well as the closeness of the predicted discharges to the actual value for the flood, $80\text{-}90m^3s^{-1}$, suggests that the modified slope-area or component method for discharge calculations can give fairly accurate results for discharge estimation of complex cross-sections as long as the water surface and roughness can be accurately defined. The predicted present day flood discharges should therefore provide a reasonable basis for comparison with the palaeodischarges estimated from the terraces.

8.4 Palaeodischarge estimates for the terrace fragments

Figure 8.5 (a) shows the between terrace changes in discharge at a cross-section for the five study reaches. The fragment number and estimated age of the terrace fragment are plotted as well as the estimated palaeodischarges for the floods approximately equal to the 2-2.3 year flood. Figure 8.5 (b) shows the variation in the magnitude of the flood approximately equal to the mean annual flood for each terrace fragment expressed as a percentage of the modern mean annual flood as

$$(Q_{past} - Q_{pres})/Q_{pres} \times 100 \qquad 8.6$$

The maximum number of terrace levels available for comparison at one cross-section is four. Three of these levels are of late Holocene age. It is not possible, therefore, to discuss a complete sequence of discharge changes through time since deglaciation. The most complete data are for the late Holocene terrace fragments, which do allow comparison of late Holocene flood discharges with those of the present day.

Examination of the two plots shown in Figure 8.5 reveals an overall consistent downstream pattern in the palaeodischarges for the terrace fragments relative to the discharges of the present day stream. The consistency helps to support the suggestion that estimation of discharge using the total sinuosity of the palaeochannels on terrace surfaces provides reasonable estimates of the palaeoflood discharges for the

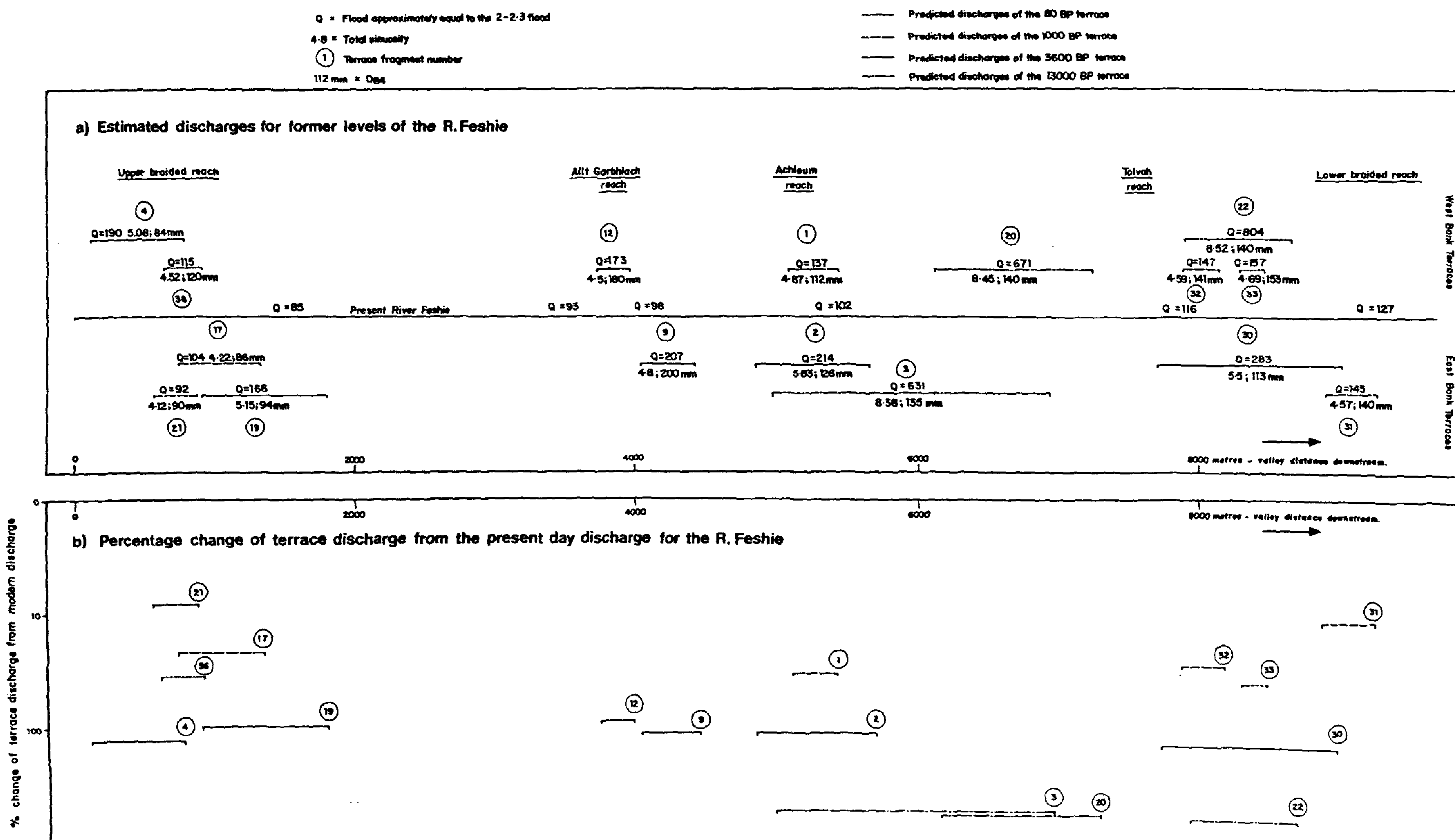

Q = Flood approximately equal to the 2-2.3 flood
4.8 = Total sinuosity
1 Terrace fragment number
112 mm = D84
Predicted discharges of the 80 BP terrace
Predicted discharges of the 1000 BP terrace
Predicted discharges of the 3600 BP terrace
Predicted discharges of the 13000 BP terrace
a) Estimated discharges for former levels of the R. Feshie
Upper braided reach
Allt Garbhlach reach
Achleum reach
Tolvah reach
Lower braided reach
West Bank Terraces
East Bank Terraces
Present River Feshie
Q=190 5.08; 84mm
Q=115 4.52; 120mm
Q=173 4.5; 180mm
Q=137 4.87; 112mm
Q=671 8.45; 140 mm
Q=804 8.52; 140mm
Q=147 4.59; 141mm
Q=157 4.69; 153mm
Q=85
Q=93
Q=98
Q=102
Q=116
Q=127
Q=104 4.22; 86mm
Q=92 4.12; 90mm
Q=166 5.15; 94mm
Q=207 4.8; 200mm
Q=214 5.63; 126mm
Q=631 8.38; 135 mm
Q=283 5.5; 113mm
Q=145 4.57; 140mm
0
2000
4000
6000
8000 metres - valley distance downstream.
b) Percentage change of terrace discharge from the present day discharge for the R. Feshie
% change of terrace discharge from modern discharge
10
100
1000

Figure 8

former braided channel systems of the River Feshie.

The most obvious overall trend in the data is a reduction in the total sinuosity of the preserved braided river palaeochannels from the 13,000BP surface to the present stream in its braided reaches. According to the thesis outlined in Chapter 6 this reduction in total sinuosity reflects a changing flood discharge from the 13,000BP surface to the present day river. Total sinuosities for the outwash terrace range from 8.3 at the Achleum reach to 8.52 at the lower braided reach. Discharges for the palaeooutwash in the Achleum, Tolvah and lower braided reaches were about 520% higher than those of the present day. These represent values of 650 m^3s^{-1} at the Achleum reach and $804m^3s^{-1}$ at the lower braided reach where the main Feshie outwash is joined by the Allt Chomhraig outwash on the western bank.

The total sinuosities of the 3,600BP surface range from 4.8 to about 5.5. The stream which built the former active zones represented by the 3,600BP terrace surfaces had discharges that were approximately 100-120% greater than present day discharges. The discharge for the 3,600 BP terrace fragments increases from about $166\text{-}190m^3s^{-1}$ at the upper braided reach to $214m^3s^{-1}$ at the Achleum reach to $283m^3s^{-1}$ at the lower braided reach.

By the level of the 1,000BP terrace total sinuosities at all reaches had declined substantially from the 3,600BP surface and were closer to present day values of 4.1 for the upper braided reach and 4.3 for the lower braided reach. Total sinuosities for this surface range from 4.22 to 4.87. The 1,000BP terrace discharges are thus considerably lower than those for the 3,600 BP surfaces. They were between 8-34% greater than present day discharges. The two 80BP fragments exhibit increases in total sinuosity from the 1,000BP surface. Total sinuosities were measured for two fragments only, one from the upper braided reach and one from the lower braided reach. Total sinuosities are 4.52 and 4.9 respectively. Associated discharges which were 35% higher than the discharges of the present day River Feshie, suggesting a rise in flood discharge from the 1,000BP fragments to the 80BP fragments followed by a decline to present day values.

Measurement of rates of bedload transport in modern gravel-bed braided streams is insufficient to relate quantitatively variation in bedload transport capacity to variation in total sinuosity and stream power. However, some indirect evidence may be advanced to suggest qualitatively that the extent of bank erosion and the capacity to transport sediment will increase with increasing total sinuosity. This may then be used to specify the direction of change in sediment transport rates in the Feshie during the late Holocene.

Changes in cross-sectional geometry and channel pattern morphology of free alluvial channels may occur as a continuous, direct function of stream power and bedload transport rates. Analytical and flume studies have shown that the progression of straight through meandering to braided streams is related to changes in the cross-sectional geometry of the channel (Ackers, 1964; Schumm and Khan, 1972; Engelund and Skovgaard, 1973; Parker, 1976; Chang, 1979), and more particularly to an increasing width-depth ratio (Chapter 6).

Bagnold (1977) has suggested that in equilibrium channels the width-depth ratio is adjusted, through bank erosion rates and bed deposition, so that the mean sediment transport rate is just equal to sediment supply from upstream. Bedload transport rates are directly related to stream power per unit bed area and to relative roughness, so that a wide, shallow channel will be a more efficient transmitter of sediment than a narrow, deeper channel. Thus the continuum of channel pattern can be associated with variation in sediment transport rates. A high width-depth ratio is associated with extensive bank erosion and high sediment transport rates. This association was anticipated by the flume experiments of Schumm and Khan (1972). These demonstrate that bedload transport is, as demonstrated by Bagnold, directly related to stream power. Increasing width-depth ratio of the flume channel is also highly correlated with increasing bedload transport rates in the experiments. In Chapter 6 a relationship between width-depth ratio and stream power was established for 28 gravel-bed rivers from New Zealand. This relationship suggested that the width-depth ratio varies as a continuous function of stream power. Regression of width-depth ratio of these same streams against total sinuosity gives a correlation coefficient of 0.92 for the relationship

$$w/d = 0.008\Sigma P^{3.184} \qquad r^2 = 0.85 \qquad 8.7$$

where

w/d = width-depth ratio

ΣP = total sinuosity

which is significant at greater than the 0.001 level of significance. Thus total sinuosity may vary concomitantly with a rise in stream power, bank erosion rates and increasing capacity to transport sediment.

During the late Holocene the River Feshie has experienced variation in total sinuosity and therefore also, probably, variation in stream power, sediment transport rates and extent of bank erosion. Considering the data from the late Holocene terraces this implies that the River Feshie at 3,600BP with a stream power index of 2.04-2.86 and a total sinuosity ranging from 4.8-5.5 had a greater capacity for sediment transport than at 1,000BP or 80BP. An increase in stream power index from 0.97 at 1 000BP to 1.5 for the 80BP surface with an associated rise in total sinuosity from 4.22 to 4.52 at the upper braided reach suggests an increase in sediment transport rates from 1,000BP to 80BP.

8.5 Vegetational, palaeolimnological and geomorphic evidence of Holocene environmental changes in upland Scotland

As far as is known there have been no published palaeohydrological estimates made for prior Holocene rivers in the Highlands. The data discussed above suggest phases of increased fluvial activity, relative to the present River Feshie, during the late Holocene. This is shown by higher values for the total sinuosity of the former active zones, and hence higher discharges and probably greater capacity for sediment transport. This evidence for these palaeohydrological changes needs to be evaluated in the context of available palaeoclimatic/palaeoenvironmental information for upland Scotland.

Stratigraphic analysis of pollen in bogs and in lake sediments has allowed the postglacial vegetation history of upland Scotland to be intensively documented and recent geochemical investigations of lake sediments have provided additional information for environmental reconstruction. Lake basins, as sediment sinks, contain a record of processes which have taken place within the lake catchment (Edwards and Rowntree, 1980). Integrating pollen stratigraphy with the analysis of the sedimentological and chemical properties of the sediments in lake cores is useful because it helps to provide an estimation of the history of landscape stability as it responds to climatic change, soil development and human disturbance (Engstrom and Wright, 1984).

In Scotland, the change to warmer conditions at the end of the Loch Lomond Stadial was very rapid. Minerogenic sedimentation was rapidly replaced by organic accumulation and by about 9,700-9,500BP summer temperatures in south-west Scotland were at least as warm as present (Bishop and Coope, 1977). Pollen analytical evidence suggests that vegetation responded rapidly to the climatic amelioration and by about 10,000BP birch-juniper scrub had replaced open tundra vegetation throughout much of the Scottish Highlands (Vasari, 1977; Lowe and Walker, 1977). A slightly later date of 9,673BP has been recorded for the _Juniperus_ expansion around the Abernethy Forest, the latter about 15km north-east of Glen Feshie (Birks and Mathewes, 1978). Between about 9,000-7,225BP the characteristic regional pollen zone is one dominated by _Betula_ and _Corylus_ (Vasari, 1977; Birks and Mathewes, 1978) with the birch-hazel forest forming a dense vegetation cover. The mid-postglacial was characterised by a phase of _Pinus-Betula_ domination which extended from about 8,000BP to 5,000BP. _Pinus_ probably began to migrate into the Grampians by about 8,000BP, with a date of 7,225 BP at Abernethy Forest (Birks and Mathewes, 1978). Immigration of _Pinus_ was probably complete by 6,600BP (Pears, 1968; O'Sullivan, 1976). By about 6,500BP _Alnus_ had expanded into Scotland and had begun to replace other tree species in the mid-Flandrian forests.

During this phase, which represents the climatic optimum, concentrations and totals of arboreal pollen were at their highest (Birks and Mathewes, 1978; Rapson, 1984) and tree lines in the Cairngorm Mountains reached their maximum elevation of about 790m. This is in comparison to the present natural treeline which probably averages 640m (Pears, 1968, 1969).

The geochemical stratigraphy of lake sediments in northern Scotland suggests that the early Holocene history of lake catchments involved rapid stabilisation of the landscape. A number of indices are used as indicators of periods of, respectively, stability and low erosion, and instability and high erosion. Mackereth (1966) proposes that the inorganic component of lake sediments calculated from loss-on-ignition data, and the proportions of sodium and potassium, reflect erosion intensity in the lake catchment. If the landscape is not stable erosion may cut through the soil cover and contribute relatively unweathered mineral material into the lake. This is recorded by relatively high contents of sodium and potassium and a higher ratio of mineral to organic matter. Variation of particle size may also be indicative of increased erosion levels in the catchment (Edwards and Rowntree, 1980) particularly if used in combination with other erosion indicators. During the first half of the Holocene lake sediments in northern Scotland were characterised by relatively low values of sodium and potassium and high values of organic matter. These trends are indicative of a stable landscape with limited erosion in the lake catchments (Pennington et al., 1972).

Similar trends in lake sediments have been observed from cores extracted from several Cairngorm corrie lakes (Rapson, 1984). Two of the Cairngorm corrie lakes, Coire an Lochain Uaine and

Coire an Lochain lie at altitudes between 910 - 1,010m and are therefore above the highest known tree lines at any time in the Holocene. As they lie above the tree line, the corrie lakes will record regional changes in pollen, whilst the erosion indicators may reflect changes in environmental conditions that are not due to human interference but may be climatically induced. The period about 8,000-4,000BP was characterised by a high ratio of organic matter to mineral matter in the core sediments and low values of sodium and potassium. Particle sizes of the mineral sediments indicated very fine material entering the corrie lakes. These trends are indicative of a period of low erosion in the catchments around the corrie lakes. This period is also associated with very high values of arboreal pollen suggestive of closed forests in the valleys.

Various palaeoenvironmental indicators thus suggest that the early-mid Holocene in upland Scotland was characterised by a stable forested landscape with low rates of slope erosion. This period represents a large part of the time when soil formation was occurring on the land surface comprising the 10,000BP terrace. The pollen analytical evidence derived from the organic horizon of the 10,000BP soil (Chapter 3), suggests that soil formation and accumulation of organic matter began on a stable surface very quickly in the early Holocene. Pollen extracted from the base of the organic layer is dominated by an early Holocene <u>Betula-Corylus</u> assemblage probably dating from about 9,000BP. The dominant tree pollen from the sample taken half way up the organic layer is <u>Pinus</u> with some alder and non-arboreal pollen of <u>Calluna</u> . This assemblage is probably a mid-Flandrian assemblage. These data suggest that soil formation during the early-mid Holocene was occurring at a time when the landscape was dominated first by the early Holocene shrub vegetation and then by the mid- Flandrian forests. The correlation of these pollen assemblages with the regional pollen zones suggests that the regional pollen diagrams may be applicable in Glen Feshie.

Palaeoenvironmental and geomorphic evidence suggest that the late Holocene stands in marked contrast to the relative stability of the early-mid Holocene. The late Holocene appears to have been characterised by a cooler, wetter climate, the widespread decline of the forest cover, the expansion of heathland vegetation and accelerated slope erosion and slope instability. The late Holocene is the period when man becomes increasingly important as a potent agent of environmental change (Goudie, 1977). However, differentiation between climatically and man-induced changes during the period dominated by anthropogenic activity is very difficult to achieve unequivocally (Edwards and Rowntree, 1980).

Pine stumps preserved in peat may be useful climatic indicators if wetness of the bog is directly related to increased wetness of climate. However, detailed studies of several sites are required to assess the local circumstances of the growth of blanket peats and the death of the pines before climatic change can be inferred. Birks (1975) has carried out a detailed study of pine stumps in Scottish blanket peats in order to evaluate their climatic significance. It is suggested that the synchroneity of stump ages and dates for the <u>Pinus</u> pollen decline from northern Scotland may be the result of increased climatic wetness from about 4,000BP. Pennington <u>et al.</u> (1972), commenting on the occurrence of pine stumps in northern Scotland, remark that widespread deforestation on both upland and peat bog locations was unlikely. They suggest that pine and birch grew on blanket peat as it was relatively dry, but that a climatic deterioration about 4,500BP led to an increased wetness, resulting in inhibited regeneration of pine, and swamping by the increased growth of blanket bog. Tree stumps in the vicinity of the Cairngorms show a much wider age range, but the existence of several dates around 4,000BP allows the tentative suggestion that increased wetness also affected the Cairngorms at this time (Birks, 1975). The onset of a period of accelerated erosion into the lakes appears to have accompanied this change in environmental conditions. This is inferred from the increase in sodium and potassium in lake sediments and decline in organic matter (Pennington et al., 1972; Engstrom and Wright, 1984).

Erosion indicators from the Cairngorms corries are suggested to be indicative of climatic deterioration at least from 3,300BP (Rapson, 1984). High values of sodium and potassium, low values of organic matter in proportion to mineral matter, and an influx of coarse sediments are indicative of accelerated erosion into the corrie lakes. Climatic deterioration in the form of cooling is also suggested by intermittent movement of solifluction lobes on the Cairngorm mountains. Radiocarbon dating of over-ridden organic material has yielded dates ranging from 5,440BP to 2,680BP (Ballantyne, 1984).

At Loch Garten in Strathspey, O'Sullivan (1974) has a radiocarbon date of 3,635+/-205BP for the onset of the late Holocene pollen zone characterised by a marked decline in the relative proportion of tree pollen and the expansion of taxa which are indicative of heathland, Calluna, Gramineae and Juniperus. Sporadic instances occur of weed pollen types such as Plantago lanceolata, Rumex, and Artemisia. At Loch Pit-youlish, Speyside, forest clearance appears to have been slightly later than is indicated for the area around Loch Garten. A radiocarbon date of 2,990+/-60BP marks the increase in the relative proportion of non-arboreal pollen and the onset of forest clearance (O'Sullivan, 1976). A decline in the organic content of the sediment in Loch Pityoulish together with the expansion of deteriorated pollen points to increased soil erosion within the catchment. This is further suggested by the post-2,990BP inversion of ^{14}C dates in the loch sediments. In the Cairngorm corries the onset of the pollen zone characterised by a decline in the relative proportion of arboreal pollen and an increase in non-arboreal pollen has been dated at 3,320BP- 3,150BP (Rapson, 1984, 1985).

In the vicinity of the Cairngorms dated pollen records from a number of sites indicate that deforestation began about 3,600-3,000BP leading to the widespread development of Calluna-dominated heath. This would suggest that the Group III soils may have been forming in an environment dominated by heathland vegetation. This is supported by pollen analytical data derived from samples taken from the base of two organic horizons of the Group III soils. Both samples indicate a dominance of Calluna with some Pinus. In Chapter 3 it was suggested that the pyrophosphate-extractable iron peak in the group III soils may be indicative of an overall change in environmental conditions from the onset of soil development of group I soils to the group III soils. A possible explanation for the higher pyrophosphate-extractable iron maxima in the group III soils may be that these soils began to form at a time in the Holocene when there was a change in environmental conditions to a cooler, wetter climate and with the development of heath-type vegetation. This explanation would be consistent with the evidence presented above of a possible climatic deterioration together with the development of heathland vegetation.

Although there is some evidence of a possible climatic change to cooler, wetter conditions it seems likely that the activity of man accelerated vegetation changes at about 3,600-3,100BP in the Cairngorms area. Generally, archeological evidence for settlement of the Cairngorms area is scarce. However, there may have been an influx into the Speyside area of a population about 3,600BP which was responsible for the construction of the Speyside Clava-type Chambered Tombs. The onset of forest clearance may therefore be associated with the influx of population into Speyside at about 3,600BP (O'Sullivan, 1974).

At Loch Pityoulish pollen analysis has provided evidence of a later, second phase of forest clearance. This is associated with pollen indicative of agricultural activity, including arable farming, and is attributed to the establishment of either Celtic or Pictish settlements in the Speyside area at about 1,000BP (O'Sullivan, 1975). Studies of the pollen content and the radiocarbon age of mor humus layers in the Forest of Abernethy have shown that a period of widespread forest clearance and heathland formation took place in that area about 1,500-1,000 years ago (O'Sullivan, 1973). This is consistent with the Loch Pityoulish evidence and is suggested to show that in the Spey valley a major period of agricultural settlement took place at this time (O'Sullivan, 1975).

Sedimentological and chemical data from the sediments in the corrie lakes of the Cairngorms show peak values of sodium and potassium, extremely low values of organic matter and increased proportion of coarser sediments entering the lakes from about 1000BP to the present day. This is suggestive of accelerated slope erosion in the corrie catchments in the very late Holocene. Evidence of widespread debris flow activity over the past 500 years in the Cairngorms (Innes 1982, 1983) also suggests that the slopes did not regain the apparent stability of the early-mid Flandrian. Radiocarbon dating of buried organic material in debris flows in the upper Feshie provides evidence of late Holocene slope instability in Glen Feshie. In the upper Feshie compound debris flows have been undercut by the river. This has exposed a number of buried organic horizons in the debris flow deposits thus indicating episodic renewal of debris flow activity at this site. A number of radiocarbon dates on the buried organic horizons suggest two phases of debris flow activity (V.Brazier, personal communication), an earlier,late Holocene phase of debris flow activity and a later historical phase of slope instability.

The earlier, late Holocene phase of debris flow activity is suggested by a date of 2090+/-50 radiocarbon years BP from the organic horizon of a soil buried by debris flow deposits. A renewed phase of activity in Glen Feshie probably began about 320BP. Innes (1983) suggests that historical debris flow activity in the Cairngorms area may be attributable to anthropogenic rather than climatic factors, that is to landscape instability initiated by burning and overgrazing. However, the date of about 300BP for the second phase of debris flow activity in upper Glen Feshie would place this in the Little Ice Age. Debris flow activity does continue to the present day and its initiation in Scotland is frequently reported to occur in association with intensive rainstorms (Common, 1954; Baird and Lewis, 1957; Innes, 1982). The role of extreme events may also be important in destabilisation of the landscape and in initiation of debris flow activity.

High magnitude events may also be responsible for widespread channel planform changes in catchments in upland Scotland. Changes in channel planform over the past 250 years, that is increases in sinuosity of meandering streams, increases in braiding intensity and local changes from meandering to braiding, in upland catchments in the Spey (including Glen Feshie), Dee and Tweed basins have been shown to be a response to relatively high random magnitude events (McEwan, 1985). Examination of the long term rainfall records for Braemar demonstrate high rainfall peaks in the period 1870-1900. Similarly, higher rainfall annual maxima are found during the 1870's to 1890's. Higher rainfall maxima are reflected in the reconstructed regional flood record with an increase in magnitude of the moderate regional floods. In the mainstream Spey there was an increase in incidence of moderate regional storm events with floods occurring in 1892, 1893 and 1894, these floods frequently having a major snowmelt contribution. Significant channel changes have been shown to have occurred in response to these events (McEwan, 1985).

The impact of extreme events on valley floor development has also been demonstrated from upland England. Evidence of geomorphic impact of intensive rainstorms in the initiation of shallow turf slides and debris cones has been reported by Harvey (1986) from the Howgill Fells, north-west England. Harvey also demonstrated that, for one valley in the Howgill Fells, an extreme, localised flood, generated by a high intensity convectional storm, caused a largely singlethread channel to jump to a braiding mode.

The palaeoenvironmental indicators discussed above suggest the hypothesis of a relatively stable early-mid Holocene followed by an unstable late Holocene. This evidence is thus comparable with the trends refered to above for changing environmental conditions during the Holocene for upland England. Here recent work has led to the development of the concept of an unstable late glacial-early Holocene, and a stable early-mid Holocene followed by an unstable late Holocene (Harvey <u>et al</u>., 1981; Richards, 1981; Harvey <u>et al</u>., 1984; Richards <u>et al</u>., in press). Thus, Carlingill in the Howgill Fells contains a high terrace which is probably late Pleistocene in age. Formation of this surface appears to have been followed by a long period of slope and valley floor stability which allowed the development of mature podzols on the high terrace. A period of slope and channel instability occurred about 2,500- 1,000BP which resulted in the formation of valley side gullies and a low terrace surface. In Jugger Howe Beck in the North York Moors, a low angle fan occurs at the junction of a tributary, Hollin Gill, and Jugger Howe Beck. The fan is a compound feature possessing two age units. The upper unit exhibits a well developed, mature podzol which correlates with similar podzols on high terrace fragments upstream in Jugger Howe Beck. Renewal of fan activity has resulted in the development of a younger fan unit, and a radiocarbon date of 1,150BP on fine organic material and 900BP on an alder fragment in an organic layer located towards the base of the younger fan unit deposits suggests a period of slope and valley instability in the late Holocene. In these valleys in upland England it is suggested that late Holocene destabilisaton of the landscape may have been consequent upon Viking-early mediaeval vegetation disturbance (Harvey <u>et al</u>., 1981; Richards <u>et al</u>., in press).

The data discussed above suggest that the later Holocene in upland Scotland was characterised by widespread environmental changes, both climatically-controlled and man-induced. These environmental changes have been reflected in geomorphic evidence of slope instability in the Cairngorms area. Evidence of recent and historical debris flow activity and channel pattern changes shows that late Holocene instability has continued to the present. However, evidence of historical activity also reflects the importance of extreme events in destabilisation of both the slopes and river channels in upland Scotland.

8.6 Fluvial response to environmental change in Glen Feshie

(a) The Temporal Response

Dating of the terraces (Chapter 3) in Glen Feshie suggests that there may have been a relatively long gap between the phase of floodplain development at 10,000BP and that which occurred at about 3,600BP. The occurrence of terrace surfaces dated to 3, 600BP, 1,000BP and 80BP suggests that floodplain formation and subsequent incision to produce terraces may be a recurrent feature of the late Holocene, and that palaeoenvironmental evidence of landscape instability, already indicated by slope instability in the Cairngorms area, is reflected in valley floor instability in Glen Feshie. Thus as in upland England, the late Holocene appears to be characterised by valley floor landform development. The late Holocene therefore presents a sharp contrast to the period between ca. 10,000BP and 3,600BP for which there appears to be no record in the stratigraphic sequence represented by the Holocene terraces. It may be that these two periods of the Holocene, the period from ca. 10,000-3 600BP, and ca. 3,600BP to the present, represent two periods with differences both in environmental conditions and fluvial activity.

Climatic change, vegetation change and high magnitude events are background environmental factors which are indirectly responsible for phases of aggradation and incision in fluvial systems. The physical processes responsible for terrace development are those of valley bottom sediment surface elevation changes which are a response to changes in sediment supply and/or runoff conditions within the catchment. As these elevation changes must be transmitted via the channel system they are likely to involve adjustments in channel pattern morphology.

Phases of aggradation and incision cannot be unequivocally related to particular environmental changes although there are several lines of evidence to indicate that phases of increased landscape instability have occurred at various times in the Cairngorms area. These may have contributed to the development of the Glen Feshie terraces. As discussed above, several lines of evidence indicate that the early-mid Holocene was one of relative landscape stability. Closed forests existed in most of the valleys and slope erosion into the lochs was at its lowest for the Holocene period. In Glen Feshie there is no evidence for the development of a terrace surface between the 10,000BP and 3,600BP surfaces. It is possible that the stratigraphic hiatus represented by the period ca. 10,000BP to 3,600BP may have been a period when the River Feshie was incising below the surface of the 10,000BP floodplain. The change from the depositional mode which produced the 10 000BP surface, to an incising mode which resulted in its abandonment may have been the result both of a reduction in sediment supply as recovery from the Loch Lomond Stadial took place, as well as the stabilisation of the valley fill deposits by a dense vegetation cover. During this first period of the Holocene the river may have been of relatively low total sinuosity with a relatively reduced sediment supply which would have encouraged incision through the fill deposits.

Incision, as a result of downstream base-level changes which would result in a wave of erosion proceeding upstream, is a relatively unlikely alternative for the development of the Holocene terraces in Glen Feshie. This is because of the presence of a 1km long meltwater rock gorge at Feshiebridge below the lower braided reach. The latter has formed the base-level for the postglacial stream (Young, 1975).

With the change in environmental conditions that occurred about 3,600BP in the Cairngorms area, destabilisation of the slopes and valley floor is likely to have been initiated, a trend which has continued through historical times. The onset of this phase of instability appears to have been marked in Glen Feshie by sediment aggradation and burial of the former, possibly Loch Lomond age landsurface in the tributaries, and by gravel aggradation and braiding in the main valley (Chapter 3). The correlation of the terrace fragments (Chapter 3) demonstrates the probable existence of a continuous surface which aggraded along the length of the study reach at about 3,600BP. The mapping of the palaeochannels on the 3,600BP surface shows that the River Feshie was braided along the full length of the study reach. This must therefore have been a basin-wide phase of increased sediment supply and gravel aggradation and possibly also of increased runoff. The increase in runoff may have been consequent upon the reduction of vegetation cover suggested by the regional pollen analysis as well as a possible increase in wetness. This phase of braiding involved the migration of the channel across the full width of the valley floor in the upper braided reach and between the confines of the sandur deposits in the reaches downstream of the Allt Garbhlach. This phase of extensive braiding by a powerful, high sinuosity stream would have removed any evidence of the progressive incision of the River Feshie through the 10,000BP surface, if such evidence existed in the form of fragmentary unpaired terrace remnants. The palaeohydrological reconstruction of discharges for this terrace level certainly support the concept of a powerful braided stream with flood discharges 2-2.2 times greater than present day discharges.

Subsequently, the River Feshie incised about 1-1.5 metres below the surface of the 3,600BP floodplain to leave it abandoned as a terrace. The onset of incision is unlikely to have been caused by an increase in discharge. An increase in discharge, assuming the valley slope remained constant, would lead to an increase in stream power and a rise in bank erosion rates. The combination of these factors would have resulted in an increase in total sinuosity. Such an increase in total sinuosity would have resulted in the reworking of the 3,600BP surface rather than its incision. It is therefore more likely that a decrease in discharge and some decrease in sediment supply followed the construction of the 3,600BP active braided zone thus resulting in its subsequent incision and abandonment as a terrace. A reduction in discharge from the 3,600BP surface onwards is supported by reference to the discharge data presented in Figure 8.5. A decrease in sediment supply is feasible, as it is with the development of the floodplain of the 3,600BP surface that the river/slope coupling in the River Feshie was weakened. The weakening of river/slope coupling in Scottish upland streams has been noted by Milne (1980) and Werritty (1982) and is attributed to the development of braided floodplains which separate the channel from the base of the slopes, thus leaving only small areas of exposed gravel adjacent to the river which can be readily reworked. With the development of the extensive 3,600BP braided zone and its associated floodplain the main channels would have become increasingly separated from the slopes in the upper braided reach and the glaciofluvial deposits which it had been undercutting in the middle and lower reaches of the river.

By the same argument, the development of the 1,000BP surface and the 80BP surface must have occurred as a result of an increase in sediment supply to the stream and possibly an increase in discharge. The role of extreme events may be important in contributing to the development of these younger terraces in the River Feshie. In Chapter 3 it was shown that the 80BP surface was the result of an increase in braiding intensity in three reaches of the River Feshie. This occurred between 1869 and 1899, with the River Feshie changing its channel planform from a low sinuosity channel to a multithread channel with a total sinuosity of 4.52. It was shown above that Speyside had experienced an increased rainfall intensity and flood peaks in the 19th century (McEwan, 1985). In the mainstream Spey there was an increase in incidence of moderate regional storm events with floods occurring in 1892, 1893 and 1894. Significant channel changes were shown to have occurred in response to these events. Glen Feshie was affected by these and other regional storms. For example, reworking of terrace surfaces, the latter used as agricultural land, was reported to have occurred at the downstream end of the River Feshie in 1862 (McEwan, 1985).

It is thus likely that an increase in the frequency of flood events in the 1870's-1890's was responsible for the increase in braiding intensity in the River Feshie. The increase in braiding intensity resulted in the development of a large area of active unvegetated gravel bars. The extent of the gravel bar area also increased between 1899 and 1946. The appearance of the bar surfaces was accompanied by a reduction in the total area of the terraces between the upper Feshie and the Allt Lorgaidh (Figure 3.18 a). Digitising the area of the terraces between the two streams showed that the higher terraces were reduced in total area by 36% compared to the original 1869 area. Between 1899 and 1946 there was a further enlargement of the active area with a concomitant reduction of the total area of the higher terraces. These were reduced by a further 15% of the original 1869 area. Thus in just over 70 years the total area of the higher terraces between the upper Feshie and the Allt Lorgaidh had been reduced by over 50% of the 1869 area. The change from a lower total sinuosity channel to a more intensively braided channel was accompanied by extensive bank erosion and reworking of the floodplain and bounding terraces to provide the sediment supply for the active braided zone which developed after 1869. The present River Feshie in this reach has once again reverted to a low sinuosity channel occupying its pre-1899 position and has incised over a metre below the new bar area to produce a low-level terrace. The increase in braiding intensity in these three reaches is likely to have been the result of a period of increased storminess and frequency of regional storm events during the later 19th century. This may have been exacerbated by destabilisation of the vegetation on the higher terraces as a result of tree felling and crofting practices (McEwan, 1985).

Several inferences concerning the development of low-level terraces in unstable reaches of gravel-bed rivers may be drawn from these findings:-

(1) A mechanism is suggested for the formation of low-level terraces in Glen Feshie. In Glen Feshie unstable reaches responded to moderately high flood events by extensive bank erosion, reworking of older terrace deposits and the formation of an active channel zone. The sediment supply for the gravel aggradation and braiding is derived from the reworked terrace deposits. Subsequent lower magnitude, more frequent events are responsible for the rationalisation of the gravel area into the first order bars and channels of a braided channel. These active areas are then abandoned, probably as a result of channel avulsion, and accretion onto older floodplain units may occur. The system readjusts after the perturbation of the high magnitude event, by abandoning the former active area. The channel has then abandoned its source of sediment and subsequent incision of the lower sinuosity main channel leaves these recently formed sedimentary units as low level terraces.

(2) These data imply rapid reworking, removal and accumulation of sediment. Average bank erosion rates of about 6.6m a year for this reach between 1869 and 1899 imply that a 250m wide active zone of a gravel-bed river could be reworked in about half a century. This therefore implies that previous terrace levels could have existed but have been reworked leaving no record of prior Holocene events. This may be particularly applicable to late Holocene low-level terraces which may have formed in a manner similar to those described above in response to regional flood events. These rates of erosion and sedimentation also imply that sedimentary units which may develop into low-level terraces bordering the modern river may accumulate very rapidly.

Although these rates of reworking and sedimentation are rapid they are comparable with the rates of erosion and reworking described by Ferguson and Werritty (1983) for the upper braided reach of the River Feshie and with the high rates of lateral channel shift noted for upland rivers in Wales (Lewin et al., 1977). They are also comparable with the rates of historical valley alluviation for the Afon Ystwyth (Lewin et al., 1983). Here series of low-level terraces have developed in response to increased sediment supply consequent upon mining activities since 1800 AD.

(3) This process is likely to be spatially variable. The 80BP terrace occurs in three reaches of the River Feshie, the upper Feshie, the upper braided reach and the lower braided reach. These three reaches are the reaches of the present day river where bank erosion, active gravel bar development and migration are taking place. There is no evidence of the 80BP surface in the middle reaches of the present River Feshie. In these reaches the river is either locally locked by bedrock or laterally confined by coarse valley fill deposits (Werritty and Ferguson, 1980). In the middle reaches the present stream is restricted to a single channel with occasional mid-channel bars. Reference to the 1869 and 1899 OS maps and the 1946 aerial photographs has shown that in these reaches the channel had already achieved its present position by 1869 and that lateral migration of the river since that time has been negligible.

(4) It is possible that localised terraces such as those that comprise the 80BP terrace will continue to develop in response to the processes currently operating in the

unconfined reaches of the River Feshie. However, upland streams in the highlands are susceptible to localised meteorological events (Werritty, 1982). This factor coupled with the spatial variability of sediment in upland valleys and the role of channel avulsion in producing these sedimentary units suggests the possibility that the resultant low-level terraces may be localised surfaces only and therefore vary in age not only between valleys but within valleys. Thus Gage (1970) has described the development of flights of terraces in valleys in New Zealand. Here during a single high intensity storm in 1965 the Waiho River produced a sequence of terraces. Elsewhere in the same region flights of terrraces developed and have been eliminated more than once over the 30 years preceding 1970 as a result of high intensity storms that may have a 10 year cycle.

The interpretation of the development of the Holocene terraces given above involves increased fluvial activity in the horizontal plane during phases of relative landscape instability, and, progressive incision, and therefore fluvial activity in the vertical plane, during phases of relative landscape stability. Phases of increased fluvial activity in the horizontal plane are likely to be accompanied by phases of braiding, whilst phases of fluvial activity in the vertical plane may be accompanied by a reduction in total sinuosity of the stream. Support for this interpretation of terracing in the environment of gravel-bed braided streams comes from terrace sequences in areas where sediment transfer by the fluvial system has been controlled by stadial and interstadial conditions (Church and Ryder, 1972; Mycielska-Dowgiallo, 1977; Briggs and Gilbertson 1980; Jackson et al., 1982).

In valleys where large volumes of sediment have been introduced to the fluvial systems as a result of glacial activity, heightened sediment movement will continue as long as sediment remains easily accessible to the fluvial system. The sediment transfer that occurs during this time results in the formation of fluvial deposits such as alluvial fans, debris cones and terraces that represent temporary storage areas for glacially-derived material. Such sediment transfer may dominate the operation of fluvial processes for thousands of years after deglaciation and has been termed paraglacial sedimentation. Changes in vegetation conditions may be crucial in determining whether such deposits may be regarded as unstable in the fluvial environment, or whether, as a result of stabilisation due to the spread of vegetation, such deposits become unavailable for erosion and redeposition elsewhere (Church and Ryder, 1972). Discussing the paraglacial origin of the terraced river sediments in the Bow Valley, Alberta, Jackson et al. (1982) describe phases of braiding and gravel aggradation during a period of sparse vegetation and cool climatic conditions following deglaciation at about 13,000BP. This was followed by climatic amelioration and the spread of vegetation which resulted in the decrease in availability of stored sediment. During this time the river assumed a low sinuosity, meandering habit displaying degradational behaviour. In Poland, extensive braided palaeochannels on terrace surfaces have been associated with unstable sediments and cold, unvegetated environments. The changing channel configuration from multithread to singlethread accompanied the transition from cold to temperate environment and resulted in degradation of the river to leave the former braided channel zones as terraces. Recent destabilsation of the slopes and valley floors as a consequence of deforestation has resulted in gravel aggradation and braiding (Mycielska-

Dowgiallo, 1977). These findings are consistent with those of Briggs and Gilbertson (1980) who interpret the development of the terraces of the upper Thames in terms of a model involving fluvial reworking of solifluction deposits, braiding and gravel aggradation during cold phases, and meandering and incision during warmer, more stable interglacial phases.

(b) The Spatial Response

The spatial response to episodes of aggradation and incision described above may however, be variable within a single river system. This variation is the result of spatial differences in the depth of the valley fill/bedrock floor interface, and in the nature of the valley fill deposits. The river response to spatial variation in these factors is manifested in the degree of activity of the river along a given reach and is reflected in the downstream variation in channel pattern morphology as measured by the total sinuosity of palaeochannels on one terrace surface.

The multivariate expression relating stream power, total sinuosity and grain size developed in Chapter 6 showed that sediment properties interact with stream power to determine channel pattern morphology. Temporal and spatial variation in one or more of the controlling variables will influence the total sinuosity of the stream as the channel adjusts its morphology to maintain a dynamic equilbrium with the new hydrological or sedimentological conditions. The Glen Feshie terraces exhibit a variation in total sinuosity in the spatial as well as the temporal dimension as a result of downstream variation in grain size for one terrace level (Figure 8.5). This trend is particularly evident with the most extensive Holocene surface, the 3,600BP surface.

Total sinuosity declines from the upper braided reach to the Allt Garbhlach reach although the discharge would be expected to increase downstream and with the entrance of two second order tributaries. Total sinuosity declines from 5.15 to 4.8 but grain size increases from 94mm to 200mm. As noted in Chapter 2, this increase in grain size is the result of coarse inputs of sediment from undercutting of coarse glaciofluvial deposits derived from the Allt Garbhlach and from the coarse deposits being evacuated from Allt Garbhlach tributary valley. Thus both controls vary spatially along the river as well as temporally. The addition of water and sediment from tributary sources of variable size produces discontinuous changes in the controls, with parallel discontinuities in channel pattern morphology (Knighton, 1984). At the Achleum reach D_{84} declines to 126mm and the total sinuosity increases to a value of 5.83.

The downstream decrease in total sinuosity with increasing grain size, exhibited by the palaeochannels on the 3,600BP terrace surface, shows the sensitivity of the total sinuosity parameter to changes in grain size. This is an encouraging result in support of the applicability of the concept of total sinuosity to gravel-bed braided streams, not only in the context of modern streams, but also in terms of its sensitivity in the palaeohydrological context.

Reference to Figure 8.5 shows that there is a spatial clustering of the 1,000BP and 80BP terrace fragments used for palaeohydrological analysis. Several of the terrace fragments of the 80BP and 1,000BP terrace fragments not used for the palaeohydrological analysis, possess palaeochannels with relatively low total sinuosities. For example, three 80BP fragments have total sinuosities of 3.3, 3.1 and 2.6 respectively. In the reaches where these fragments occur the stream which deposited the 80BP surface was confined by more resistant till deposits of the valley fill. These values may be compared with the total sinuosity of the 80BP fragments in the unconfined reaches that were used for the palaeohydrological analysis. These are 4.52 for the upper braided reach and 4.69 for the lower braided reach. In the Allt Garbhalch reach the 1000BP fragments have sinuosities of 2.6, 2.4 and 2.2. In the Allt Garbhlach reach the stream was confined both by coarse kame and kettle deposits and by bedrock. In reaches where the stream was unconfined the total sinuosities range from 4.1 to 4.59.

Where the stream is confined by glaciofluvial deposits, Ferguson (1981) suggests a mechanism by which pinning may take place. The stream begins to undercut the confining walls but is unable to remove the large quantities of coarse sediment released from the undercut bluffs. Milne (1983) has shown that this method of confinement occurs frequently in channels in upland Britain. He describes eleven streams which exhibit confined reaches which occur where the stream is incised into poorly sorted glacio-fluvial valley fill deposits.

The incision of the 3,600BP surface in Glen Feshie to the level of the 1,000BP surface reached the valley fill/bedrock floor interface in several localities. At these locations the bedrock outcrops are forming a series of local baselevels down-stream. These local base-levels encourage lateral migration of the stream upstream and thus removal of higher terrace levels. At these locations bedrock outcrops on the channel floor and banks. The 3,600BP surface has been deposited across the bed-rock and incision through the gravels has lead to its exposure. Upstream from such outcrops the 3, 600BP surface is virtually absent, with only the 1,000BP and 80BP surfaces being present.

This contrast in the spatial distribution of channel pattern character is likely to occur more generally in upland valleys in Britain and, as noted by Ferguson (1981), is likely controlled by availability of sediment and degree of confinment of the channel. McEwan (1985) found that the effect that random, high magnitude events had on the changing planform of upland streams was determined by the availability of sediment and the degree of confinement of the channel. Where the channel was confined little planform change was evident, the most marked channel changes occurring in reaches where the river was unstable and reworking of valley fill deposits was possible. Harvey et al. (1984) also suggest that as a stream progressively incises through the valley fill deposits it may eventually become locked by bedrock. However, this is likely to be a spatially variable process so that reaches which have a readily accessible supply of sediment will continue to modify their patterns in response to relatively frequent flood events.

CHAPTER 9

CONCLUSIONS

Detailed investigation of one major valley in the Scottish Highlands has provided the basis for an intergrated analysis of the temporal, palaeohydrological and palaeoenvironmental aspects of valley floor landform development. Evidence has been presented for the occurrence of major geomorphological development during the Holocene in the valley of the River Feshie, south-west Cairngorms. At least three phases of Holocene terrace development are suggested to have taken place. Significant changes in stream power, transporting capacity and channel pattern morphology appear to have accompanied this landform evolution. Assessment of these changes within the context of Holocene environmental fluctuations in the Cairngorms has suggested that the early-mid Holocene was a period of relative landscape stability while the late Holocene was characterised by increased landscape instability. This thesis therefore challenges, for at least one major site, the established concept that most of the Scottish river terraces in upland valleys are outwash terraces related to deglaciation at the end of the most recent cold periods. This reinterpretation of low-level terrace development for Glen Feshie was facilitated by the methodology developed in this thesis for rigorous correlation and relative dating of river terrace fragments and the palaeohydrological analysis of terraced gravel-bed braided deposits.

The geomorphological interpretation of upland valley terrace sequences is inhibited if correlations and reconstructed chronologies of evolution are misleading. The traditionally used morphological techniques for terrace correlation, based on analysis of height-range diagrams, may not always provide a sound basis for subsequent interpretation of the development and environmental context of terrace sequences. A soil-stratigraphic approach to terrace correlation has been developed using data from the surface soils developed on the Glen Feshie terraces. Measurements have been made of the physical and chemical properties of the soil profiles that are diagnostic of the soil forming processes and their temporal pattern. In Glen Feshie these processes are those which together contribute to podzolisation of the soils. The properties examined were therefore selected after a consideration of the processes of podzolisation. These data provided a strong quantitative basis for the correlation and relative dating of the Glen Feshie alluvial deposits. Principal Components Analysis was performed on the data matrix of soil variables. Projection of the component scores for each soil profile onto the first principal axis produced a point distribution of five discrete clusters. Cluster analysis confirmed that the five grouping scheme constituted an operational classification of the original data matrix. Interpretation of the derived components suggested that the first principal axis represented intensity of podzolisation. Intensity of podzolisation is a function of the length of time available for the operation of the soil forming processes. The first principal component can thus be interpreted as having a temporal dimension. The arrangment of the clusters along the first principal axis must be chronological with the five groups arranged on a relative time scale. Calibration of the sequence with some absolute dating control allowed the Glen Feshie terraces to be placed on a tentative absolute time scale.

The soil-stratigraphic data together with the availability of some absolute dates provide dates for the accumulation of the gravels of the various terrace surfaces. Five terrace surfaces spanning an age range of 13,000BP to 80BP have been identified for Glen Feshie. The low-level terraces at 80BP, 1,000BP and 3,600BP provide evidence for active terrace development during the late Holocene. The higher surfaces at 10,000BP and 13,000BP are suggested to have developed in relation to deglaciation at the end of the most recent cold periods.

The soil-stratigraphic method of terrace correlation and dating allowed each of the terrace surfaces to be interpreted in the context of a chronology of evolution. Comparison of the method with the traditional height-range techniques and with a multivariate morphometric correlation, suggests that the soil-stratigraphic data provides the necessary resolution to distinguish terrace surfaces which possess only slight morphological differentiation in terms of height above present river level.

The presence of soils on the terrace surfaces that are at very different stages of soil profile development suggests that the surfaces on which the profiles have evolved probably developed episodically. Comparison of networks of palaeochannels on the surfaces of the terrace fragments shows that this episodic development has also been associated with changes in channel pattern morphology. Comparison of channel pattern morphology on the terraces was promoted by the development of a unified approach to channel pattern morphology. In turn this allowed the development of a new approach to palaeohydrology which bypasses some of the problems attendent with currently used palaeo hydraulic techniques of discharge reconstruction.

A dimensionless variable, total sinuosity, has been developed in this study as an appropriate parameter with which to quantify channel pattern morphology for free gravel-bed alluvial channels. Total sinuosity is the ratio of total active channel length and reach length. Total sinuosity enables the range of channel pattern types from singlethread, meandering channels to multithread, braided channels to be quantified on a single numerical scale. The variable also permits quantification of degree of braiding on the same scale. This parameter has been shown to be a continuous function of stream power and bed material size. Channel pattern morphology is shown to vary directly with stream power and inversely with bed material size.

Having established the existence of a channel pattern continuum for free gravel-bed alluvial channels, a new approach to the palaeohydrology of gravel-bed channels in general, and braided channels in particular, was possible. Two new relationships have been defined which can be used to predict the stream power and flood discharge approximately equal to the 2-2.3 year flood.

Total sinuosity can only be measured over the complete active channel zone. However, the size of the active channel zone of a palaeoriver cannot be directly measured from terrace fragment palaeochannels because of the uncertainity as to the extent of the scale of post-depositional reworking of the original flood-plain. A new index, the segment density index, was therefore developed by which total sinuosity could be estimated from the portion of the original palaeochannel network preserved on the terrace surface. This index was tested with field data from the upper braided reach of the River Feshie and was shown to provide accurate estimates for the total sinuosity of the whole reach.

As relationships which have been developed between variables describing river planform and parameters of the flow should not be inverted to solve for parameters of the flow, a new regression relationship was developed which relates stream power to total sinuosity and bed material size. Once total sinuosity of the palaeostream has been estimated from the palaeochannel network preserved on the surface of the terrace, stream power may be predicted from this relationship. By assuming that the network is a random sample from a normal population, confidence limits may be set for the predicted value of stream power. An estimate of the flood approximately equal to the mean annual flood may be obtained by dividing stream power by the valley slope, the latter approximated by the slope of the terrace surface. Stream power and flood discharge can thus be predicted from a part of a braided network given an estimate of total sinuosity, particle size and a measure of terrace slope.

Application of these new relationships to the well preserved palaeochannel networks on the Glen Feshie terraces has allowed an assessment to be made of the palaeohydrological implications of the terrace deposits. The channel systems on the terrace surfaces show an overall trend of reduction in total sinuosity from the oldest surface to the younger surfaces. Total sinuosities vary from 8.3 to 8.52 for the 13,000BP surface; from 4.8 to 5.5 for the 3,600BP level; 4.1 to 4.3 for the 1,000BP surface; and 4.22 to 4.87 for the youngest terrace. This reduction in total sinuosity is shown to reflect changing flood discharges throughout the history of development of the valley floor sedimentary landforms. Discharges for the oldest surfaces represent an increment of 520% of present day discharges; 100-120% for the 3,600BP surface; 8-34% for the 1,000BP surface and 35% for the 80BP surface. The episodic development of the valley floor landforms indicated by the soil-stratigraphic data thus appears to have been accompanied by changes in channel pattern morphology, discharge and probably sediment transport capacity.

The analysis of the valley floor sediment surface elevation changes, in conjunction with the channel pattern morphology of the palaeochannels, against the background of known environmental changes in the Cairngorms, suggests that there was marked change in environmental conditions between the early-mid Holocene, which appears to have been relatively stable, and the late Holocene which appears to have been relatively unstable. These changes may have been associated with variation in river behaviour.

Phases of increased braiding seem to be coincident with periods of increased landscape instability. These phases represent periods of increased supply of sediment from the valley fill deposits, fluvial reworking of sediment, and gravel aggradation. Phases of incision appear to occur during periods when there is greater slope stability either because of the development of a vegetation cover, particularly trees, or because of a weakening in the river/slope coupling. During these periods, the stores of sediment in fill deposits become stable in the fluvial environment, there is a reduced sediment supply to the stream and a phase of reduced total sinuosity during which the channel progressively incises through the fill.

The 3,600BP surface is the most extensive of the Holocene surfaces. It represents a phase of instability that appears to have been catchment wide. This phase of instability caused aggradation of an older land surface in the tributary valley of the Allt Lorgaidh. In the main Feshie valley it led to the development of an active braided zone in all five study reaches. It is possible that during this phase of widespread braiding along the whole length of the study reach, the river may have removed much of the floodplain of a probable Loch Lomond age floodplain as the 3,600BP braided river migrated across the width of the valley floor reworking the valley fill deposits. This phase appears to have been a response to both climatic deterioration and changing vegetation conditions as the mid-Flandrian forests were replaced by widespread development of heathland.

The 1,000BP surface also occurs in all reaches of the River Feshie. However, in the Allt Garbhlach reach the river which built the 1,000BP channel zone deposits was locally confined by a high bluff comprising glaciofluvial deposits. This resulted in a constriction of the active channel zone and a reduction in total sinuosity. This surface is characterised by a reduction in discharge relative to the 3,600BP surface, probably as a result of decreases in both runoff and sediment supply. It is likely that this stream lacked the power, and possibly had insufficient time, to rework the full width of the valley floor. This has encouraged the preservation of extensive fragments of the older 3,600BP surface in most reaches other than those where bedrock outcrops have resulted in increased lateral migration upstream.

The areally restricted 80BP surface only occurs in unstable reaches where there is a readily available supply of sediment.

This surface developed in response to increased braiding intensity probably in association with a period of increased flood magnitude between 1870-1890.

The Holocene terraces in Glen Feshie have developed in response to a variety of background causes which have resulted in destabilisation of the landscape and reworking of the valley fill deposits. Spatial variation in the depth of the valley fill/bedrock interface and in the average grain size of the valley fill deposits has meant that there has been a spatial as well as temporal variation in both the degree of activity of the stream for a given surface downstream as well as in the degree of preservation of older terraces. Spatial variation in the depth of the fill/bedrock interface as wellaaverage grain size of the fill may therefore produce a discontinuous response to changing environmental conditions.

The identification, correlation and relative dating of the Glen Feshie terrace levels by the soil-stratigraphic methodolgy, the subsequent palaeohydrological interpretation of these surfaces and the provision of some absolute dating control has allowed a re-interpretation of the valley floor sedimentary landforms in one valley in upland Scotland. The methodology developed for this thesis has potential for application to other such sequences of landforms and would allow the correlation of terrace and alluvial fan sequences between mainstream and tributary valleys and between river basins.

APPENDIX 1

Laboratory Procedures

A. Organic carbon content

The method of determination is a modification of that of Tinsley (1950), developed by Chartres (quoted as personal communication in Ellis (1978a)). A 0.5g sample of <2mm soil, or 0.1g if the soil was very organic-rich, was weighed into a 500ml conical flask and 10ml N $K_2cr_2O_7$, were added, followed by 20ml conc. H_2SO_4. The mixture was shaken for 30 seconds and then allowed to cool for 30 minutes. 190ml water were then added, followed by 10ml 88% H_3PO_4, leaving the mixture for 10 minutes to cool. This was titrated against a solution of 0.5 N ammonium ferrous sulphate, freshly dissolved in 0.75 N H_2SO_4, using approximately 10 drops of diphenylamine indicator freshly prepared as a 0.5% solution in approximately N H_2SO_4. The colour change was from brown, through purple to green. A blank was prepared exactly as above, and this was also titrated against ammonium ferrous sulphate solution.

Ammonium ferrous sulphate solution is, however, rather unstable and its normality was checked daily. This was achieved by adding to 10ml N $K_2Cr_2O_7$, 50ml water followed by 10ml H_3PO_4, and titrating the resulting solution against ammonium ferrous sulphate using the indicator as above. The normality (N) of the solution was then calculated thus:

$$N = \frac{10}{\text{ml of titre of ammonium ferrous sulphate}}$$

To determine the percentage organic carbon by weight, the following equation was used:

$$\%C = \frac{0.3\ (\text{ml blank - ml test titre}) * N * 1.33}{\text{weight of soil in g}}$$

B. Pyrophosphate and Dithionite extractable Fe and Al sesquioxide content

The following methods (after Bascomb (1974)) use deionised water throughout.

B1. Pyrophosphate Extraction

0.5g <2mm oven-dry soil was placed in a 50ml polypropylene centrifuge tube to which was added 50ml 0.1M $K_4P_2O_7.3H_2O$ solution. The tubes were shaken overnight and then centrifuged at 2000rpm for 15 minutes. The supernatant (the pyrophosphate extract) was decanted into a polythene bottle for storage.

B2. Dithionite Extraction

0.5 <2mm oven-dry soil was placed in a 50ml polypropylene centrifuge tube to which was pipetted 50ml sodium acetate /glacial acetic acid buffer (pH 3.8) and 2.0g sodium dithionite ($Na_2S_2O_4$) powder. The tubes were shaken overnight and then centrifuged at 2000rpm for 15 minutes. Centrifuging must begin immediately as sodium dithionite decomposes rapidly in solution. The supernatant was decanted into a small beaker. A x10 dilution was made up by pipetting 10ml of the supernatant into a 100ml volumetric flask which was then topped up with deionized water. This dilution was transferred into a polythene bottle for storage.

The dithionite and pyrophosphate extracted iron and aluminium was determined in the Atomic Absoroption Spectrophotometer, and calculations of actual content were made applying the following equation:

$$\% \text{ extractable Fe, Al,} = \frac{\text{ppm Fe, Al} \times \text{dilution}}{100}$$

C. Particle size analysis

This method is a modification of that of Bascomb (1974). Whole bulk samples were weighed, ground with a rubber pestle and passed through a 2.0mm sieve. The weight of gravel was recorded and expressed as a percentage of the total sample weight.

Approximately 50g of the <2.0mm fraction was treated with 6% H_2O_2 in order to remove organic matter. The material was then oven-dried and weighed, the mass in g being denoted as Wp. The sample was then dispersed by warming for 10 minutes in 100ml sodium hexametaphosphate (Calgon) solution prepared at a concentration of $40gl^{-1}$. Meanwhile a blank 100ml sample of Calgon was oven-dried and the mass denoted as Wc.

After dispersal the sample was wet-seived on a 63µm sieve, the material passing through being transferred into a 500ml sedimentation tube. Material retained on the sieve was oven-dried, ground with a rubber pestle and shaken for 10 minutes on a nest of 590, 210 and 63µm sieves. Material passing through the 63µm sieve was added to that in the sedimentation tube which was then made up to 500ml with distilled water. The sieve residues were then weighed, thus enabling the percentages by weight of coarse sand (2000 - 63µm) to be calculated on a gravel-free basis.

The sedimentation tube was placed in a water bath overnight in order to attain constant temperature conditions during sedimentation. After shaking the tube for 1 minute, it was immediately replaced in the bath and the time noted. Sampling the suspension was conducted at 10cm depth with a 25ml pipette whose contents were then transferred to a weighed beaker, oven-dried and weighed (Wd). Sampling was carried out at intervals of 4 minutes 48 seconds, 1 hour and 8 hours after shaking. According to Stokes' Law, at a temperature of 20^{o}C, material of specific gravity 2.5 should settle out through a 10 cm column of water as follows: 20 µm (4 minutes 48 seconds), 6 µm (1 hour) and 2µm (8 hours). The specific gravity figure of 2.5 was considerd to be realisitic, since this value is generally accepted as being about the mean in relation to the range of rock types found in the earth's crust.

In order to calculate the percentage by weight of the various size fractions obtained during sedimentation, the following relationship was used:

$$\% \text{ material} < \text{diameter } d = \left(\frac{Wd\ V}{V} - Wc\right) \frac{100}{Wp}$$

where V = volume of sedimentation tube, and v = volume of sampling pipette. This methoid allows the percentage by weight of medium silt (20 - 6µm), fine silt (6 - 2µm) and clay (<2µm) to be determined. The percentage of coarse silt (60 - 20µm) was calculated by subtraction of the sum total of the remaining fractions from 100%.

D. <u>Preparation of samples for SEM - EDXRA analysis</u>

Samples of approximately 20g of <2mm material were washed in distilled water to remove macro-organic fragments and then dried on petrie dishes at 60 deg. C. No other preparation was undertaken so as to cause the minimum of disturbance to the fabric of the grains' surfaces. The samples were then shaken evenly onto sheets of clean paper. Aluminium stubs covered with double-sided adhesive tape were gently pressed into the sample therby obtaining a monolayer of grains on each stub. The mounted grains were then blown with a high pressure air jet, thus removing those grains which were not in proper contact with the stub. This procedure was necessary in order to prevent the sample from becoming charged and therefore difficult to view under the SEM. Subsequently the mounted samples were coated with gold, which allows optimal image resolution. A Cambridge Stereoscan 600 was used to view the grains equipped with an X-ray energy dispersion analyser (EDXRA).

APPENDIX 2

Trend Surface Analysis for the Glen Feshie Terraces

Upper Braided Reach

Fragment 21A: 1,000BP

$Z_i = 366.51 + 0.00164X - 0.0099Y$

Valley Slope = 0.01

Direction of slope = 12°E of N

%RSS accounted for by the best fit plane = 93.24

$$F = \frac{93.24/2}{(100-93.24)/42}$$

F = 289.65

Fragment 21 : 3,600BP

$Z_i = 362.67 + 0.0082X - 0.0883Y$

Valley Slope = 0.0121

Direction of slope = 14°W of N

%RSS accounted for by the best fit plane = 92.13

$$F = \frac{92.13/2}{(100-92.13)/58}$$

F = 339.49

Fragment 36 : 80BP

$Z_i = 365.0 + 0.0012X - 0.0133Y$

Valley Slope = 0.013

Direction of slope = 6°E of N

%RSS accounted for by the best fit plane = 88.4

F = $\underline{88.4/2}$

$$\frac{}{(100-88.4)/38}$$

F = 144.8

<u>Fragment</u> <u>28 :1,000BP</u>

$Z_i = 366.51 + 0.0016X - 0.0097Y$

Valley Slope = 0.01

Direction of slope = 10°W of N

%RSS accounted for by the best fit plane = 93.24

$$F = \frac{93.24/2}{(100-93.24)/42}$$

F = 289.65

<u>Fragment</u> <u>35 : 1,000BP</u>

$Z_i = 359.2 - 0.0064X - 0.0062Y$

Valley Slope = 0.0091

Direction of slope = 48.5°E of N

%RSS accounted for by the best fit plane = 89.2

$$F = \frac{89.2/2}{(100-89.2)/23}$$

F = 94.98

<u>Fragment</u> <u>17 : 1,000BP</u>

$Z_i = 364.38 - 0.0015X - 0.0091Y$

Valley Slope = 0.0093

Direction of slope = 20°E of N

%RSS accounted for by the best fit plane = 95.90

$$F = \frac{95.90/2}{(100-95.90)/69}$$

F = 807.0

<u>Fragment</u> <u>34 : 80BP</u>

$Z_i = 349.86 - 0.0039X - 0.0116Y$

Valley Slope = 0.0116

Direction of slope = 2°W of N

%RSS accounted for by the best fit plane = 94.6

$$F = \frac{94.6/2}{(100-94.6)/31}$$

F = 271.5

<u>Fragment</u> <u>31 : 1,000BP</u>

$Z_i = 361.69 + 0.0022X - 0.0101Y$

Valley Slope = 0.0104

Direction of slope = 70°W of N

%RSS accounted for by the best fit plane = 90.2

$$F = \frac{90.2/2}{(100-90.2)/29}$$

F = 133.45

<u>Fragment</u> <u>40 : 3,600BP</u>

$Z_i = 379.7 + 0.0074X - 0.015Y$

Valley Slope = 0.011

Direction of slope = 20°N of W

%RSS accounted for by the best fit plane = 96.99

$$F = \frac{96.99/2}{(100-96.99)/62}$$

F = 998.0

<u>Fragment</u> <u>39 :1,000BP</u>

$Z_i = 373.4 + 0.005X - 0.00107Y$

Valley Slope = 0.0120

Direction of slope = 21°W of N

%RSS accounted for by the best fit plane = 96.4

$$F = \frac{96.4/2}{(100-96.4)/42}$$

F = 562.0

<u>Fragment</u> <u>24 : 1,000BP</u>

$Z_i = 349.87 + 0.0022X - 0.00912Y$

Valley Slope = 0.0094

Direction of slope = 24°W of N

%RSS accounted for by the best fit plane = 86.72

$$F = \frac{86.72/2}{(100-86.72)/71}$$

F = 231.82

<u>Fragment</u> <u>32 : 80BP</u>

$Z_i = 344.91 + 0.004X - 0.0115Y$

Valley Slope = 0.0114

Direction of slope = 2°W of N

%RSS accounted for by the best fit plane = 92.1

$$F = \frac{92.1/2}{(100-92.1)/41}$$

F = 235.9

<u>Fragment</u> <u>5 : 1,000BP</u>

$Z_i = 346.86 + 0.0039X - 0.0116Y$

Valley Slope = 0.0116

Direction of slope = 42°W of N

%RSS accounted for by the best fit plane = 94.61

$$F = \frac{94.61/2}{(100-94.61)/31}$$

F = 272.07

<u>Fragment</u> <u>4 : 1,000BP</u>

$Z_i = 366.51 + 0.0016X - 0.0097Y$

Valley Slope = 0.0099

Direction of slope = 10°W of N

%RSS accounted for by the best fit plane = 93.24

$$F = \frac{93.24/2}{(100-93.24)/42}$$

F = 289.65

<u>Allt Garbhlach Reach</u>

<u>Fragment</u> <u>9 : 3,600BP</u>

$Z_i = 339.05 - 0.0076X - 0.0101Y$

Valley Slope = 0.0127

Direction of slope = 37°E of N

%RSS accounted for by the best fit plane = 90.35

$$F = \frac{90.35/2}{(100-90.35)/74}$$

F = 346.4

<u>Fragment</u> <u>10 : 1,000BP</u>

$Z_i = 334.92 - 0.0099X - 0.0069Y$

Valley Slope = 0.012

Direction of slope = 55°E of N

%RSS accounted for by the best fit plane = 92.9

$$F = \frac{92.9/2}{(100-92.9)/68}$$

F = 444.8

<u>Fragment</u> <u>12 : 3,600BP</u>

$Z_i = 341.57 - 0.0051X - 0.0141Y$

Valley Slope = 0.0150

Direction of slope = 20°W of N

%RSS accounted for by the best fit plane = 89.5

$$F = \frac{89.5/2}{(100-89.5)/27}$$

F = 115.07

<u>Fragment</u> <u>14 : 1,000BP</u>

$Z_i = 340.01 - 0.0038X - 0.0083Y$

Valley Slope = 0.009

Direction of slope = 26°E of N

%RSS accounted for by the best fit plane = 93.49

$$F = \frac{93.49/2}{(100-93.49)/20}$$

F = 143.6

Achleum Reach

Fragment 2 : 3,600BP

$Z_i = 314.73 + 0.0058X - 0.0161Y$

Valley Slope = 0.017

Direction of slope = 20°W of N

%RSS accounted for by the best fit plane = 93.02/2

$$F = \frac{93.02/2}{(100-93.02)/96}$$

F = 639.67

Fragment 1 : 1,000BP

$Z_i = 317.69 + 0.0012X - 0.0133Y$

Valley Slope = 0.013

Direction of slope = 11°W of N

%RSS accounted for by the best fit plane = 90.76

$$F = \frac{90.76/2}{(100-90.76)/41}$$

F = 201.36

Fragment 3 : 13,000BP

$Z_i = 342.71 + 0.0011X - 0.020Y$

Valley Slope = 0.0201

Direction of slope = 14°W of N

%RSS accounted for by the best fit plane = 96.9

$$F = \frac{96.9/2}{(100-96.9)/103}$$

F = 1610.12

Fragment 20 : 13,000BP

$Z_i = 330.62 - 0.0120X - 0.0136Y$

Valley Slope = 0.0182

Direction of slope = 41°W of N

%RSS accounted for by the best fit plane = 96.63

$$F = \frac{96.63/2}{(100-96.63)/63}$$

F = 903.2

Fragment 23 : 13,000BP

$Z_i = 334.65 + 0.0078X - 0.0198Y$

Valley Slope = 0.0198

Direction of slope = 3°W of N

%RSS accounted for by the best fit plane = 97.4

$$F = \frac{97.4/2}{(100-97.4)/32}$$

F = 599.4

Fragment 30 : 3,600BP

$Z_i = 283.37 - 0.0132X - 0.0041Y$

Valley Slope = 0.013

Direction of slope = 73°E of N

%RSS accounted for by the best fit plane = 92.3

$$F = \frac{92.3/2}{(100-92.3)/64}$$

F = 383.5

<u>Fragment</u> <u>31 : 1,000BP</u>

Z_i = 292.99 + 0.0047X - 0.0106Y

Valley Slope = 0.0118

Direction of slope = 24°W of N

%RSS accounted for by the best fit plane = 96.51

$$F = \frac{96.51/2}{(100-96.51)/36}$$

F = 497.7

<u>Fragment</u> <u>32 : 1,000BP</u>

Z_i = 287.81 - 0.0051X - 0.0108Y

Valley Slope = 0.012

Direction of slope = 27°E of N

%RSS accounted for by the best fit plane = 93.93

$$F = \frac{93.93/2}{(100-93.93)/20}$$

F = 154.7

<u>Fragment</u> <u>33 : 80BP</u>

Z_i = 292.99 + 0.0054X - 0.0107Y

Valley Slope = 0.0121

Direction of slope = 27°W of N

%RSS accounted for by the best fit plane = 96.78

$$F = \frac{96.78/2}{(100-96.78)/34}$$

F = 510.95

<u>Fragment</u> <u>22: 13,000BP</u>

$Z_i = 298.77 + 0.005X - 0.017Y$

Valley Slope = 0.0172

Direction of slope = 34°W of N

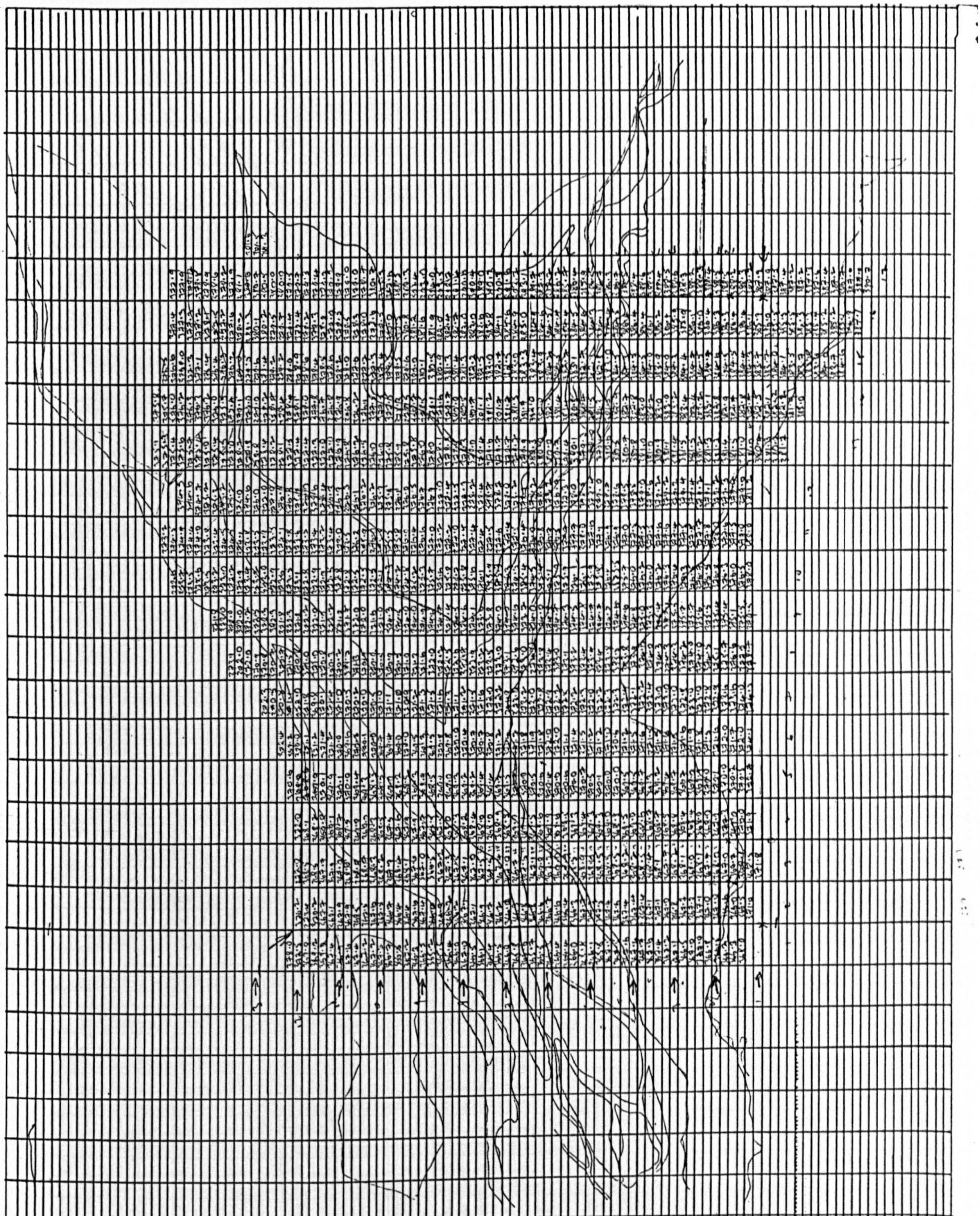
%RSS accounted for by the best fit plane = 85.76

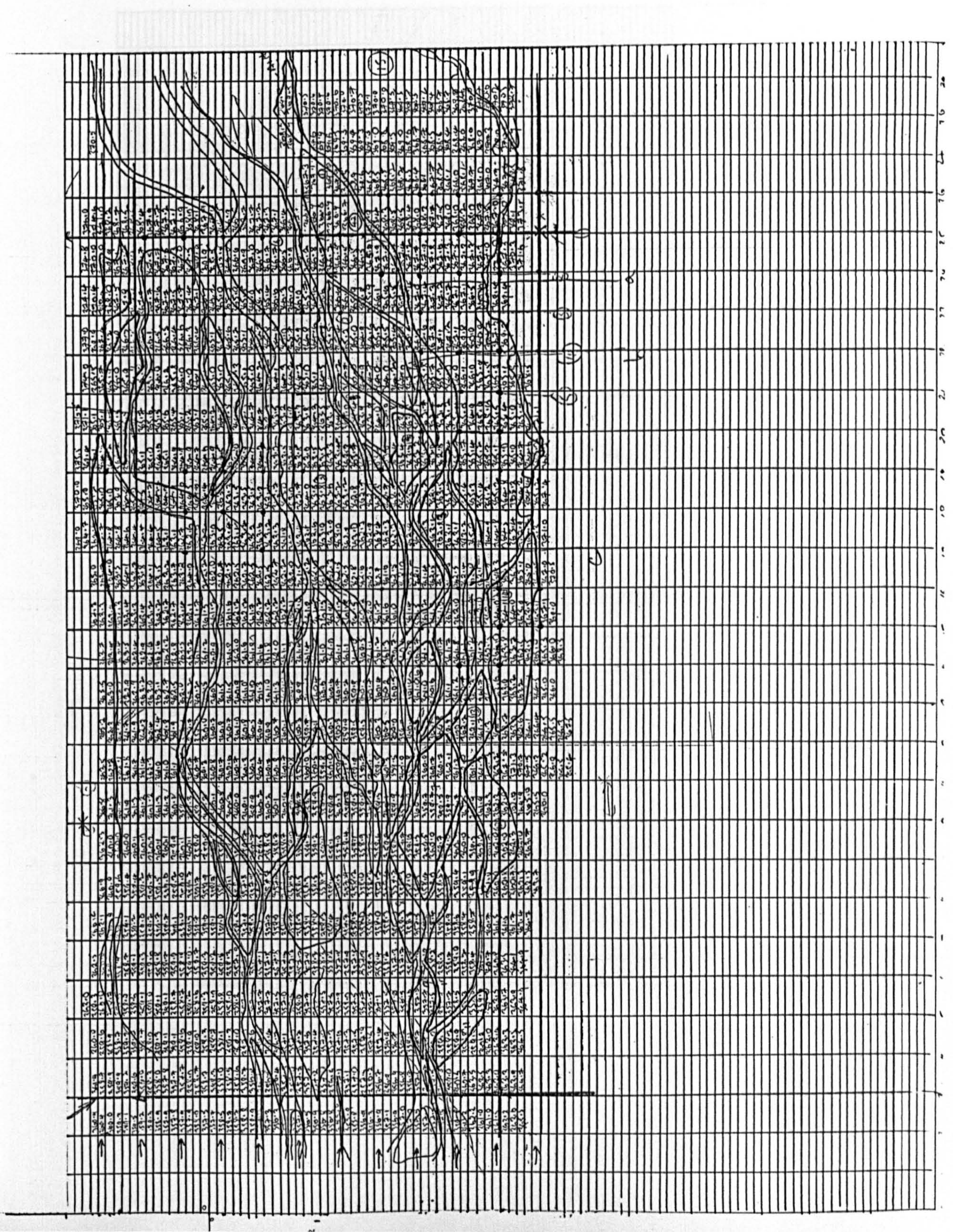
$$F = \frac{95.76/2}{(100-85.76)/43}$$

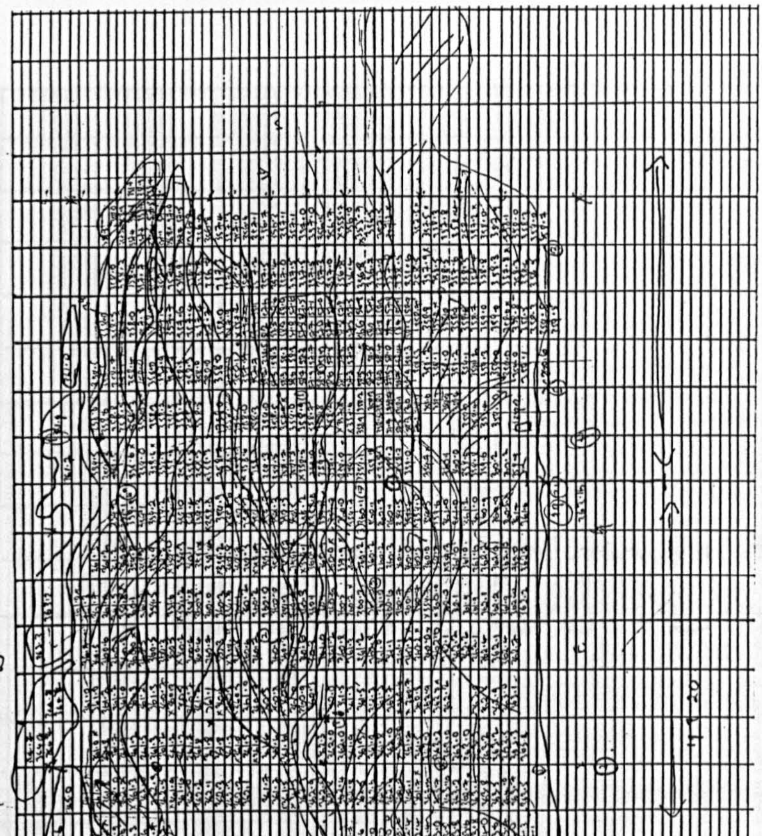
F = 129.48

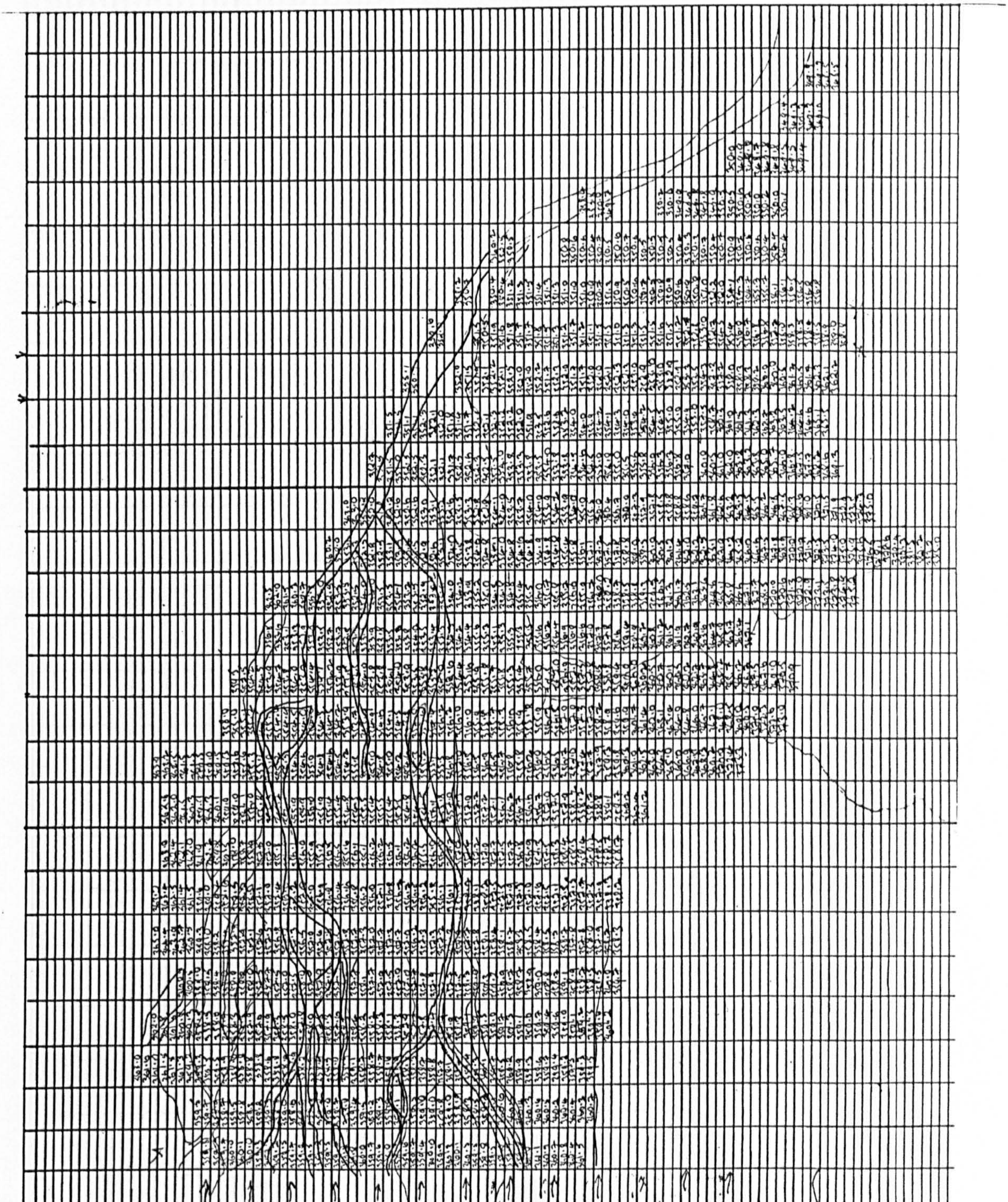
APPENDIX 3

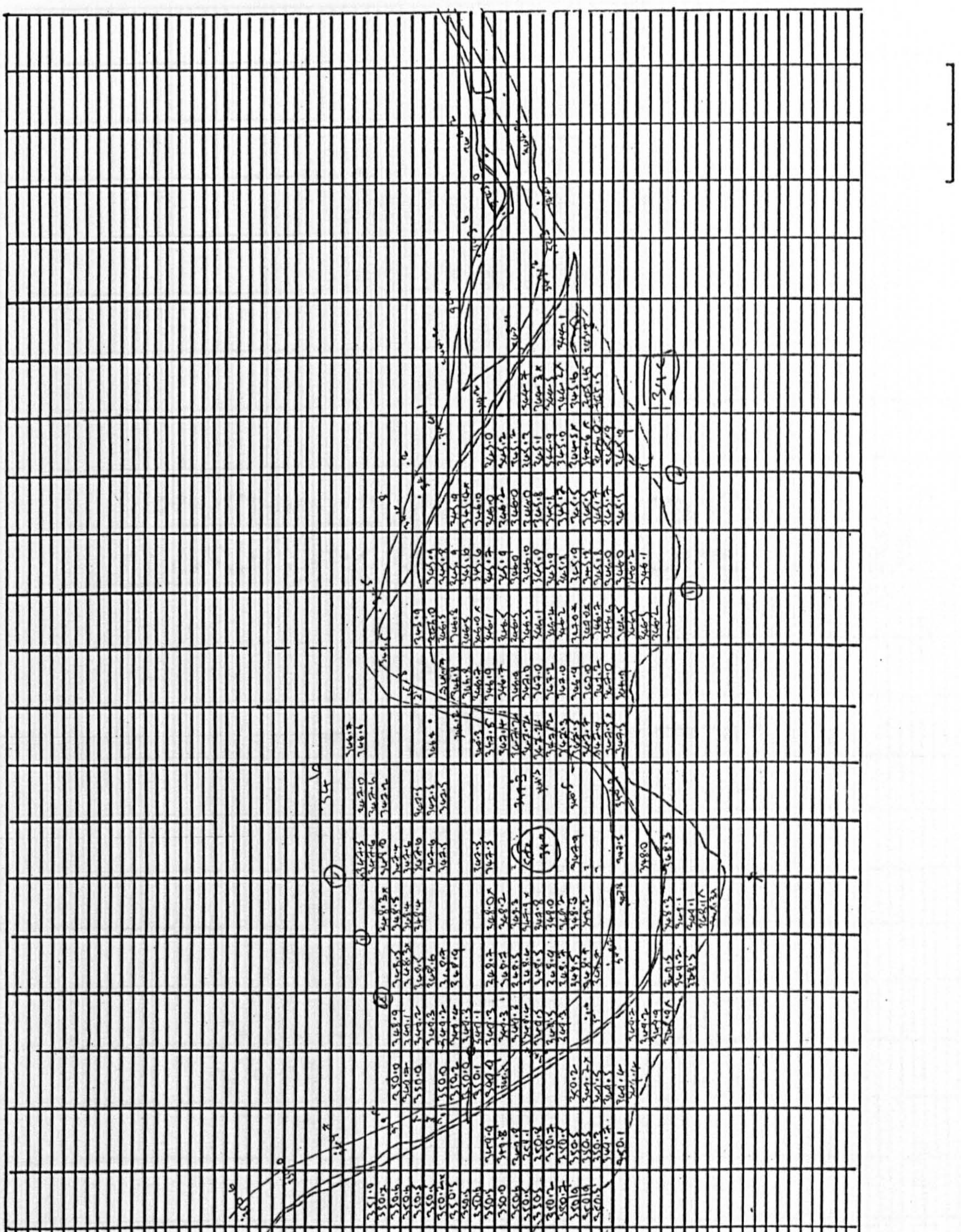
Altitude Matrices for the Glen Feshie Terraces







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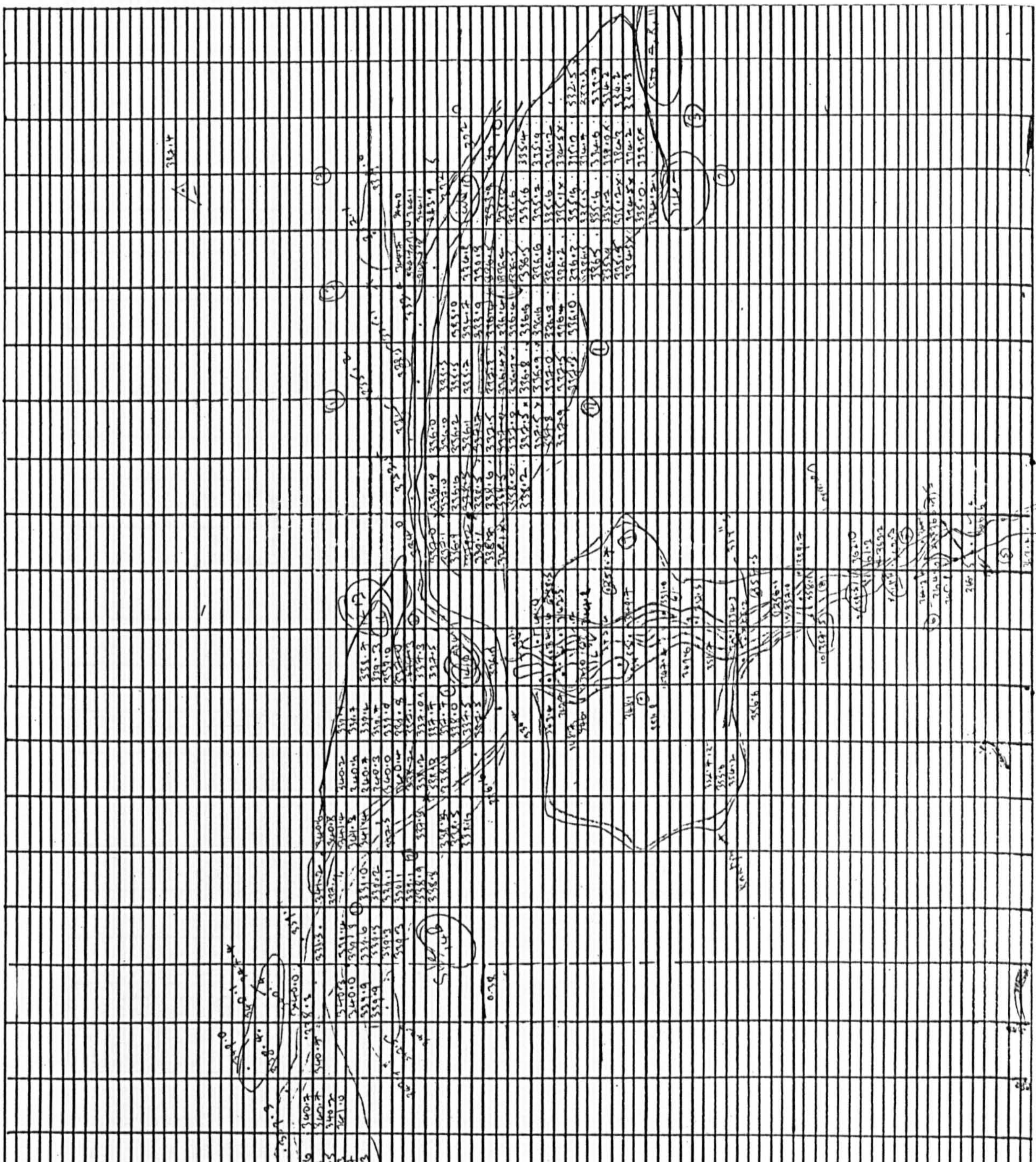

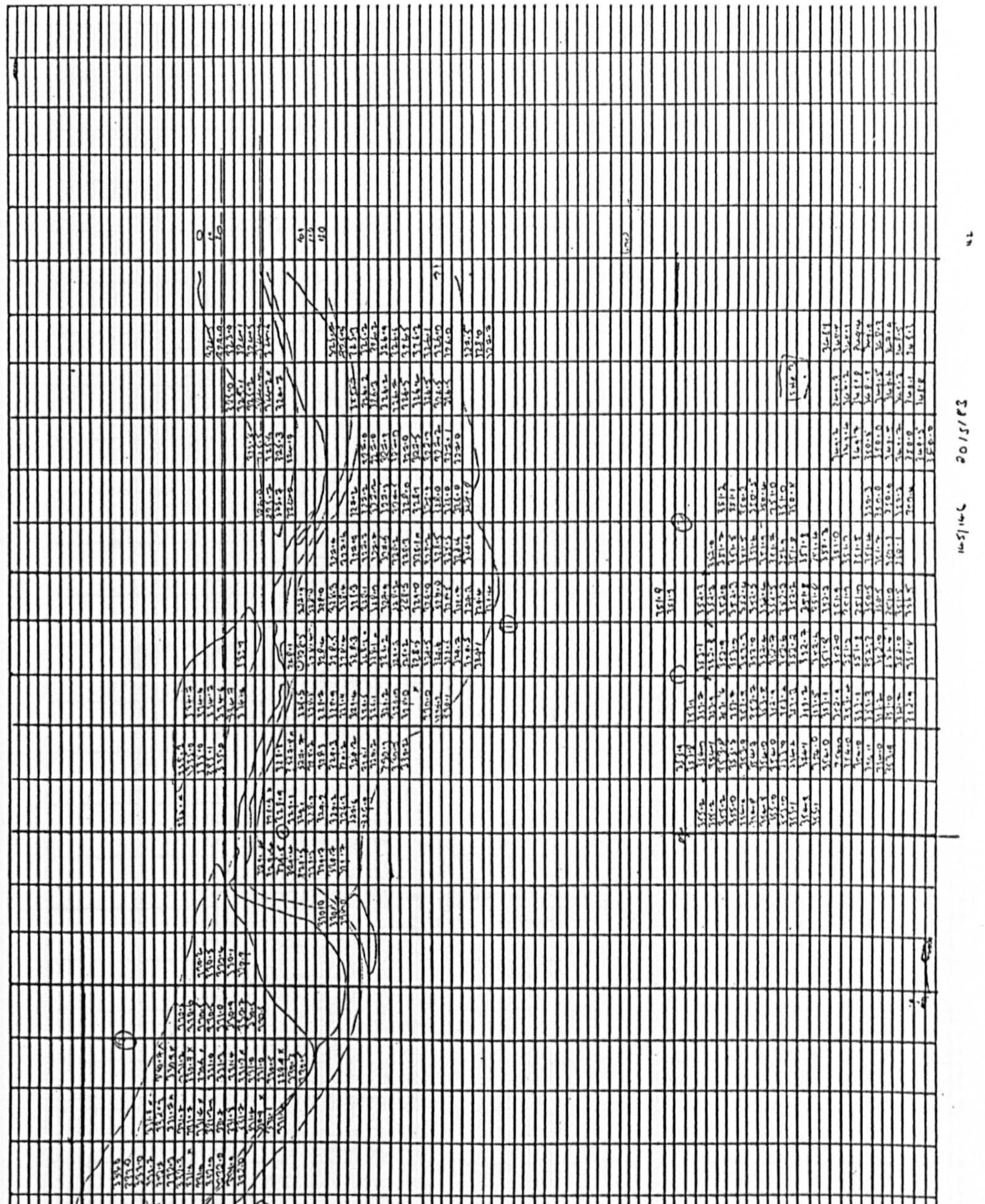

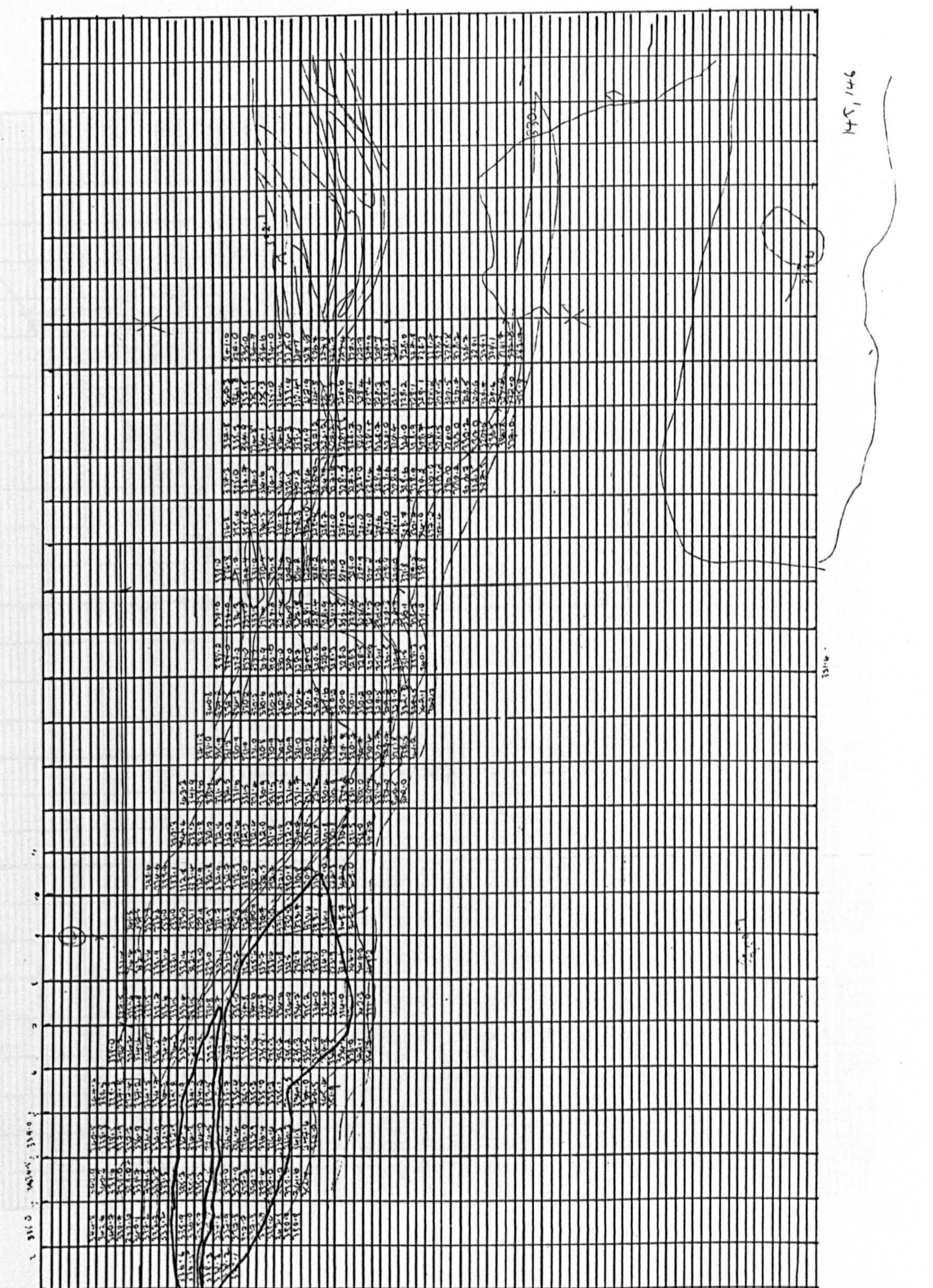

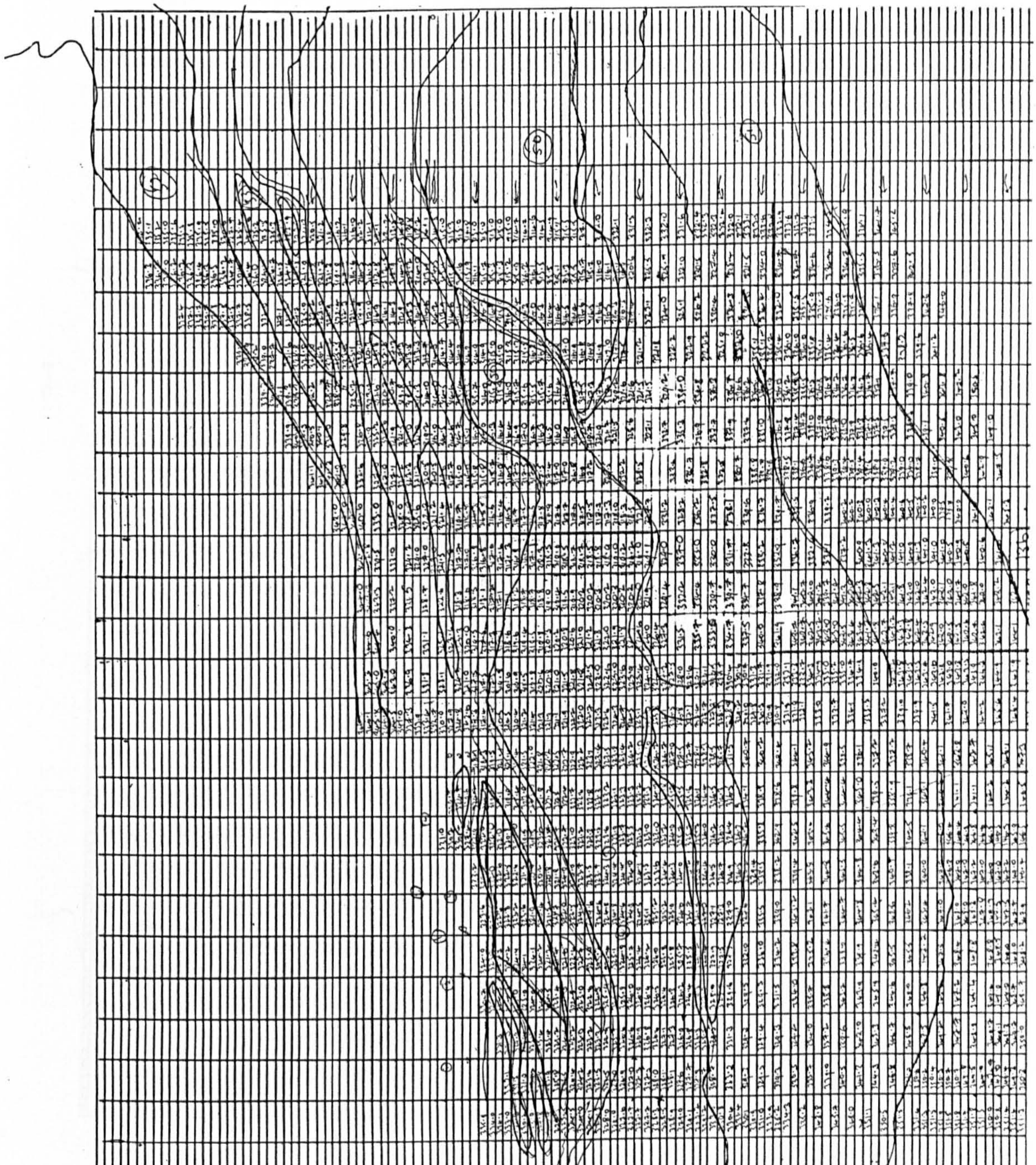

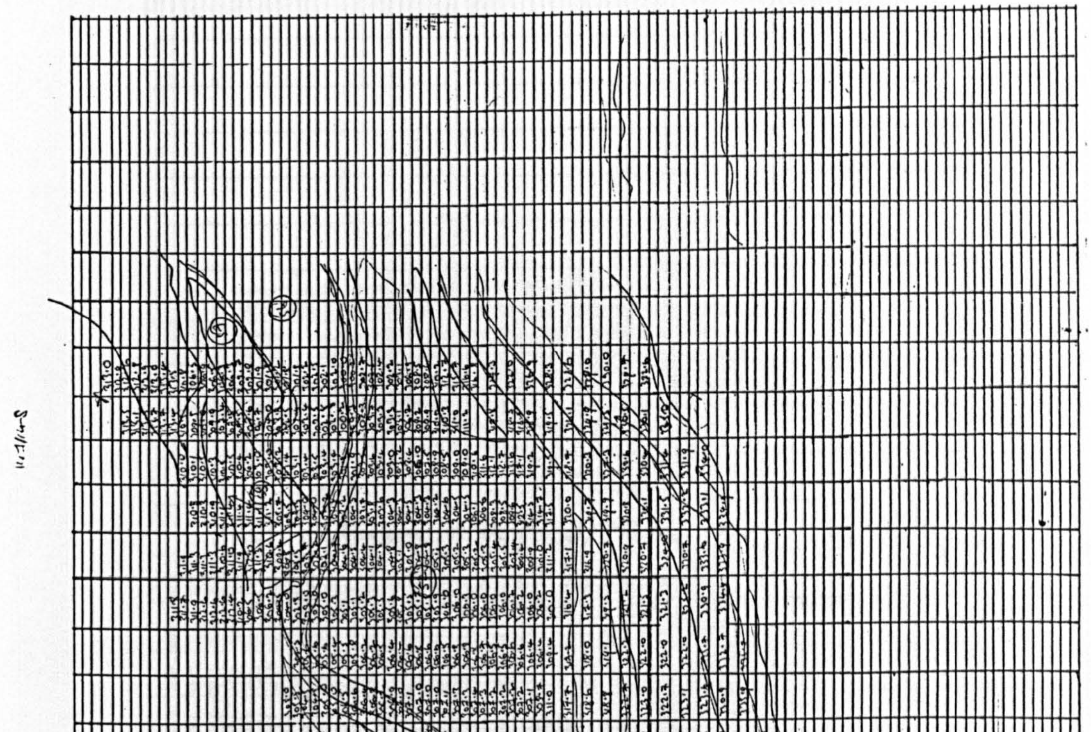

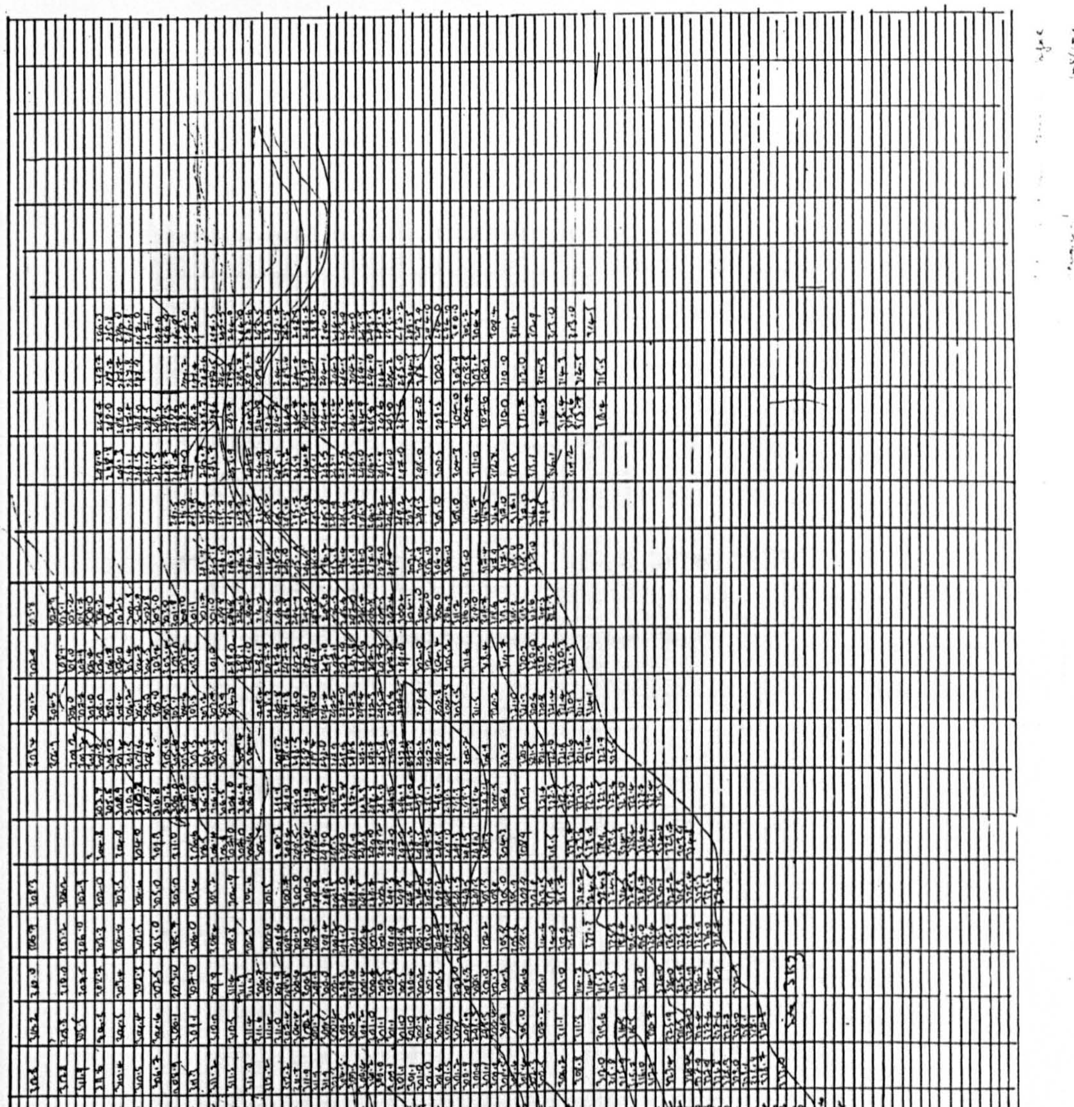

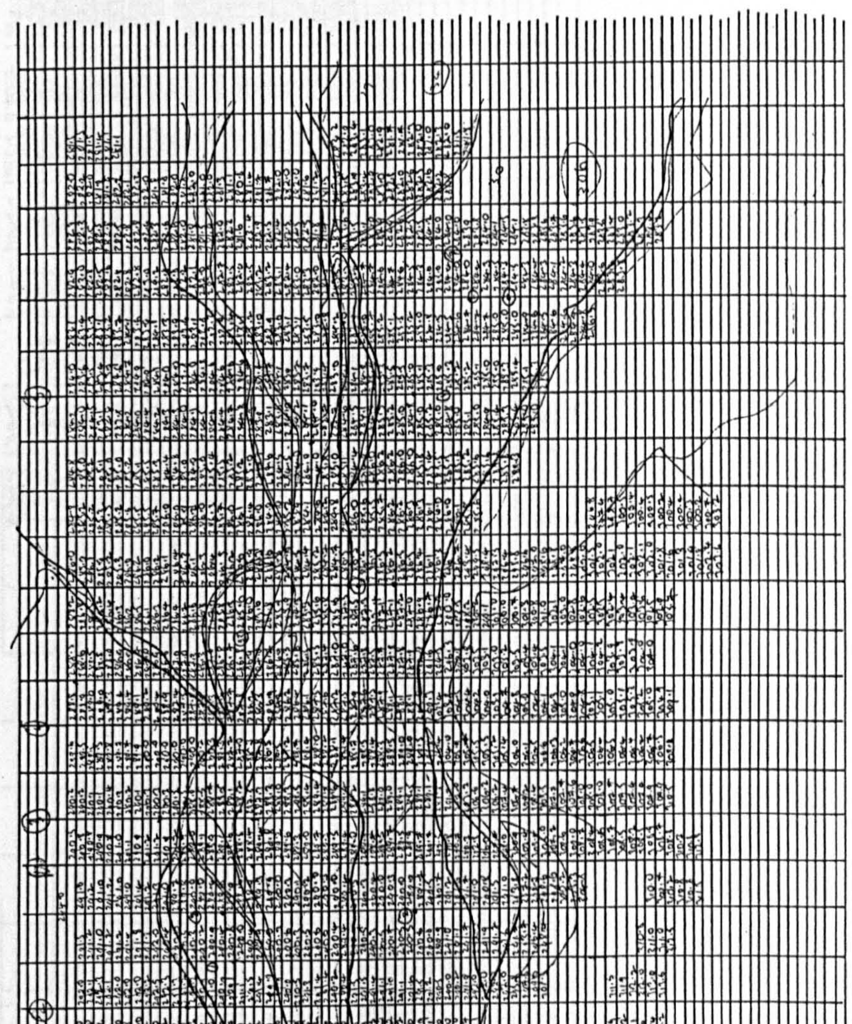

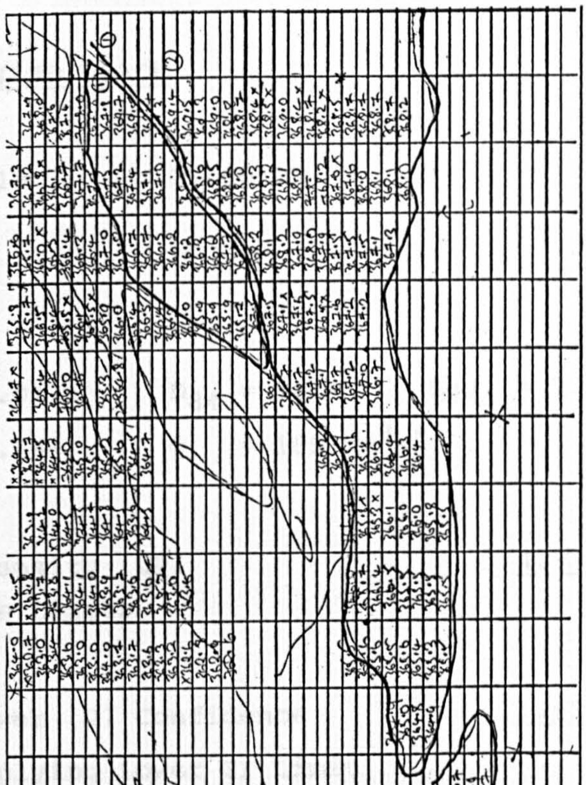

APPENDIX 4

Predicted Discharges for the Glen Feshie Terraces

UPPER BRAIDED REACH

FRAGMENT 19 3,600 BP

Network 1

Slope = 0.0121 D_{84} = 94mm Seg_d = 2.56 ΣP = 5.19

Ω = 2.045 Q = $169.0 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	4.97 - 5.40
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.63 \leq 1.77 \leq 1.91$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$2.18 \leq 2.33 \leq 2.48$
Combined Stream Power (Ω) Confidence Limits	$1.63 \leq 2.045 \leq 2.48$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$134.7 \leq 169.0 \leq 204.9$

FRAGMENT 19 3,600 BP

Network 2

Slope = 0.0121 D_{84} = 94mm Seg_d = 2.53 ΣP = 5.14

Ω = 1.98 Q = $163.5m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.92 \leq 5.14 \leq 5.35$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.44 \leq 1.58 \leq 1.72$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$2.11 \leq 2.26 \leq 2.41$
Composite Stream Power (Ω) Confidence Limits	$1.44 \leq 1.98 \leq 2.41$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$119.4 \leq 163.6 \leq 199.0$

FRAGMENT 21 1,000 BP

Slope = 0.010 D_{84} = 90mm Seg_d = 2.01 ΣP = 4.12

Ω = 0.92 Q = $91.7m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$3.90 \leq 4.12 \leq 4.33$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$0.69 \leq 0.81 \leq 0.91$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$0.94 \leq 1.07 \leq 1.20$
Composite Stream Power (Ω) Confidence Limits	$0.69 \leq 0.92 \leq 1.20$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$69 \leq 92 \leq 120.3$

FRAGMENT 17 1,000 BP

Slope $= 0.0093$	$D_{84} = 86mm$	$Seg_d = 2.04$	$\Sigma P = 4.22$
$\Omega = 0.967$	$Q = 104m^3s^{-1}$		

Total Sinuosity (ΣP) Confidence Limits	$\underline{4.00 \leq 4.22 \leq 4.43}$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$\underline{0.693 \leq 0.809 \leq 0.925}$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$\underline{1.01 \leq 1.141 \leq 1.27}$
Composite Stream Power (Ω) Confidence Limits	$\underline{0.693 \leq 0.967 \leq 1.269}$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$\underline{67.6 \leq 104 \leq 136.5}$

FRAGMENT 4 3,600 BP

Slope = 0.0093 D_{84} = 84mm Seg_d = 2.5 ΣP = 5.08

Ω = 1.77 Q = $190 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.87 \leq 5.08 \leq 5.29$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.39 \leq 1.53 \leq 1.67$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$1.87 \leq 2.03 \leq 2.18$
Composite Stream Power (Ω) Confidence Limits	$1.39 \leq 1.77 \leq 2.18$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$149.5 \leq 190 \leq 234.4$

FRAGMENT 36 80 BP

Slope = 0.013 D_{84} = 120mm Seg_d = 2.2 ΣP = 4.52

Ω = 1.505 Q = $115m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.31 \leq 4.52 \leq 4.73$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.12 \leq 1.28 \leq 1.44$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$1.58 \leq 1.75 \leq 1.92$
Composite Stream Power (Ω) Confidence Limits	$1.12 \leq 1.505 \leq 1.92$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$86.1 \leq 115.8 \leq 147.5$

ALLT GARBHLACH REACH

FRAGMENT 9 3600 BP

Network 1

Slope = 0.0127 D_{84} = 200mm Seg_d = 2.34 ΣP = 4.78

Ω = 2.52 Q = $198.3 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.56 \le 4.78 \le 4.99$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.903 \le 2.14 \le 2.38$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$2.35 \le 2.611 \le 2.863$
Composite Stream Power (Ω) Confidence Limits	$1.903 \le 2.52 \le 2.863$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$149.37 \le 198.3 \le 225.3$

FRAGMENT 9 3600 BP

<u>Network 2</u>

Slope = 0.0127 D_{84} = 200mm Seg_d = 2.41 ΣP = 4.91

Ω = 2.748 Q = $216.4 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.70 \leq 4.91 \leq 5.11$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$2.13 \leq 2.37 \leq 2.61$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$2.649 \leq 2.90 \leq 3.15$
Composite Stream Power (Ω) Confidence Limits	$2.13 \leq 2.748 \leq 3.15$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$167.7 \leq 216.3 \leq 248.8$

FRAGMENT 12 3600 BP

Slope = 0.015 D_{84} = 180mm Seg_d = 2.19 ΣP = 4.5

Ω = 1.903 Q = $173m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$\underline{4.29 \le 4.5 \le 4.71}$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$\underline{1.407 \le 1.63 \le 1.86}$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$\underline{1.99 \le 2.23 \le 2.46}$
Composite Stream Power (Ω) Confidence Limits	$\underline{1.407 \le 1.903 \le 2.46}$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$\underline{127.9 \le 173 \le 223.6}$

ACHLEUM REACH

FRAGMENT 2 3600 BP

Slope = 0.017 D_{84} = 126mm Seg_d = 2.9 ΣP = 5.83

Ω = 3.643 Q = $214.3m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$5.57 \leq 5.83 \leq 6.08$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$2.92 \leq 3.12 \leq 3.32$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$3.97 \leq 4.19 \leq 4.41$
Composite Stream Power (Ω) Confidence Limits	$2.92 \leq 3.64 \leq 4.41$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$172 \leq 214.3 \leq 259$

FRAGMENT 1   1000 BP

Slope = 0.0135 D_{84} = 112mm Seg_d = 2.39 ΣP = 4.87

Ω = 1.856 Q = $137.5 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.66 \leq 4.87 \leq 5.08$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.43 \leq 1.59 \leq 1.75$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$1.94 \leq 2.12 \leq 2.3$
Composite Stream Power (Ω) Confidence Limits	$1.43 \leq 1.856 \leq 2.3$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$105.9 \leq 137.5 \leq 170.3$

FRAGMENT 3 13000 BP

Network 1

Slope = 0.02 D_{84} = 135mm Seg_d = 4.27 ΣP = 8.38

Ω = 12.87 Q = 643.8m^3s^{-1}

Total Sinuosity (ΣP) Confidence Limits	$7.84 \leq 8.38 \leq 8.92$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$10.0 \leq 10.27 \leq 10.54$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$15.54 \leq 15.84 \leq 16.13$
Composite Stream Power (Ω) Confidence Limits	$10.0 \leq 12.87 \leq 16.13$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$500 \leq 643.8 \leq 806$

FRAGMENT 3 13000 BP

<u>Network 2</u>

Slope = 0.0198 D_{84} = 135mm Seg_d = 4.2 ΣP = 8.26

Ω = 12.25 Q = $619m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$\underline{7.74 \le 8.26 \le 8.77}$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$\underline{9.57 \le 9.84 \le 10.1}$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$\underline{14.67 \le 14.96 \le 15.25}$
Composite Stream Power (Ω) Confidence Limits	$\underline{9.57 \le 12.25 \le 15.25}$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$\underline{883.3 \le 619 \le 770}$

FRAGMENT 3 13000 BP

Network 3

Slope = 0.02 D_{84} = 140mm Seg_d = 4.3 ΣP = 8.45

Ω = 13.51 Q = $676m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$7.91 \leq 8.45 \leq 8.99$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$10.56 \leq 10.83 \leq 11.14$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$16.35 \leq 16.64 \leq 16.93$
Composite Stream Power (Ω) Confidence Limits	$10.56 \leq 13.51 \leq 16.93$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$528 \leq 676 \leq 846.5$

FRAGMENT 20 13000 BP

Slope = 0.0196 D_{84} = 136mm Seg_d = 4.28 ΣP = 8.41

Ω = 13.07 Q = $667.1 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$7.87 \leq 8.41 \leq 8.95$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$10.19 \leq 10.45 \leq 10.98$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$15.79 \leq 16.09 \leq 16.39$
Composite Stream Power (Ω) Confidence Limits	$10.19 \leq 13.07 \leq 16.39$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$520 \leq 667 \leq 836$

LOWER BRAIDED REACH

FRAGMENT 30 3600 BP

Network 1

Slope = 0.013 D_{84} = 113mm Seg_d = 2.75 ΣP = 5.54

Ω = 2.86 Q = $220 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$5.30 \leq 5.54 \leq 5.78$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$2.28 \leq 2.46 \leq 2.64$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$3.1 \leq 3.3 \leq 3.49$
Composite Stream Power (Ω) Confidence Limits	$2.28 \leq 2.75 \leq 3.49$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$175.3 \leq 220 \leq 269$

FRAGMENT 30 3600 BP

Network 2

Slope = 0.013 D_{84} = 113mm Seg_d = 2.9 ΣP = 5.83

Ω = 3.39 Q = $261 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$5.58 \leq 5.83 \leq 6.08$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$2.74 \leq 2.93 \leq 3.12$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$3.70 \leq 3.91 \leq 4.11$
Composite Stream Power (Ω) Confidence Limits	$2.74 \leq 3.39 \leq 4.11$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$210.7 \leq 261 \leq 316$

LOWER BRAIDED REACH

FRAGMENT 32 1000 BP

Slope = 0.012 D_{84} = 141mm Seg_d = 2.24 ΣP = 4.59

Ω = 1.76 Q = $147 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.38 \leq 4.59 \leq 4.80$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.44 \leq 1.49 \leq 1.54$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$1.97 \leq 2.038 \leq 2.1$
Composite Stream Power (Ω) Confidence Limits	$1.44 \leq 1.76 \leq 2.10$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$120 \leq 147 \leq 175$

FRAGMENT 22 13000 BP

Slope = 0.019 D_{84} = 140mm Seg_d = 4.34 ΣP = 8.52

Ω = 13.90 Q = $731.2 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$7.97 \leq 8.52 \leq 9.05$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$10.87 \leq 11.11 \leq 11.34$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$16.74 \leq 17.02 \leq 17.29$
Composite Stream Power (Ω) Confidence Limits	$10.87 \leq 13.90 \leq 17.29$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$572.1 \leq 731.2 \leq 910$

FRAGMENT 22 13000 BP

Slope = 0.019 D_{84} = 140mm Seg_d = 4.34 ΣP = 8.52

Ω = 13.90 Q = $731.2m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$7.97 \le 8.52 \le 9.05$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$10.87 \le 11.11 \le 11.34$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$16.74 \le 17.02 \le 17.29$
Composite Stream Power (Ω) Confidence Limits	$10.87 \le 13.90 \le 17.29$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$572.1 \le 731.2 \le 910$

FRAGMENT 33 80 BP

Slope = 0.0126 D_{84} = 153mm Seg_d = 2.29 ΣP = 4.61

Ω = 1.97 Q = $157m^3s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.48 \le 4.61 \le 4.90$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.638 \le 1.70 \le 1.765$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$2.22 \le 2.30 \le 2.373$
Composite Stream Power (Ω) Confidence Limits	$1.638 \le 1.97 \le 2.373$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$130 \le 157 \le 188.3$

FRAGMENT 31 1000 BP

Slope = 0.0118 D_{84} = 140mm Seg_d = 2.23 ΣP = 4.57

Ω = 1.71 Q = $145 m^3 s^{-1}$

Total Sinuosity (ΣP) Confidence Limits	$4.36 \le 4.57 \le 4.78$
Stream Power (Ω) Confidence Limits applying Lower ΣP Limit	$1.41 \le 1.47 \le 1.52$
Stream Power (Ω) Confidence Limits applying Upper ΣP Limit	$1.942 \le 2.00 \le 2.058$
Composite Stream Power (Ω) Confidence Limits	$1.41 \le 1.71 \le 2.058$
Derived Discharge ($Q_{2.3}$) Confidence Limits	$119.5 \le 145 \le 174.4$

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