RIVER CHANNEL PLANFORM CHANGES IN UPLAND SCOTLAND : WITH SPECIFIC REFERENCE TO CLIMATE FLUCTUATION AND LANDUSE CHANGES OVER THE LAST 250 YEARS (VOL. I)

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RIVER CHANNEL PLANFORM CHANGES IN UPLAND SCOTLAND, WITH SPECIFIC REFERENCE TO CLIMATIC FLUCTUATION AND LANDUSE CHANGES OVER THE LAST 250 YEARS.

BY

Lindsey Jo McEwen

Thesis presented for the Degree of Philosophiae Doctor,

University of St Andrews.

December, 1985

I, Lindsey Jo McEwen hereby certify that this thesis which is approximately 100,000 words in length has been written by me, that it is the record of work carried out by me, and that it has not been submitted in any previous application for a higher degree.

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I hereby certify that the candidate has fulfilled the conditions of the Resolution and Regulations appropriate to the degree of Doctor of Philosophy of the University of St Andrews and that she is qualified to submit this thesis in application for that degree.

date 10/12/85 signature of supervisor

I was admitted as a research student under Ordinance No. 12 on 1st September, 1981 and as a candidate for the degree of Ph.D. on 1st October, 1982; the higher study for which this is a record was carried out in the University of St Andrews between 1981 and 1984.

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The sources of all the diagrams extracted from published works are acknowledged on the page where they occur and also within the Table of Contents. Similiarly, all estate plans are acknowledged from their source, namely West Register House, Edinburgh.

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Abstract

Rates of river channel change in three contrasting Scottish upland environments have been studied within the context of climatic fluctuation and landuse changes over the last 250 years. The object of the research was to assess the spatial and temporal variation in channel types, the main controls on channel pattern and the dominant modes of channel adjustment. This was undertaken in a hierarchic framework with sites being investigated at three spatial scales.

At a macro-scale, the spatial and temporal variation in channel pattern was evaluated through a random sample of river channel segments for each study area, derived from the first and second editions of the 1:10,560 O.S. maps plus the 1:10,000 third edition. Each channel segment was classified within a map-based channel system typology, specifically constructed for upland Scotland. Measures of activity collected for each sample incorporated sinuosity, braiding and lateral shift indices. Flood histories were reconstructed for each study area on the basis of discharge records, long rainfall records and contemporary accounts, to assess if there was any evidence for climatic change, fluctuation or periodicities. Estimates of the recurrence interval of rainfall and runoff events of differing magnitude, frequency and duration were assessed. Data, mainly of a qualitative nature, were derived from contemporary sources and estate plans to evaluate whether any landuse changes could have changed the runoff regime and sediment mobility within each catchment.

At a meso-scale, 7 to 9 channel segments (already identified as "active" within the macro-scale study) were subject to a more detailed process-response analysis, using sequential aerial photographs. Finally at a micro-scale, the unit stream powers at these sites were studied in relation to specific runoff rates thereby relating channel process to channel form.

The strength of the controls on channel planform type varied in degree with the area studied. The glacial legacy, the positioning of local baselevels and sediment size were found to be dominant controls. In terms of channel dynamics, the position of the channel planform in relation to process thresholds and the existence of a quasi-equilibrium condition were both very important. In terms of process-response, the following general observations hold true. An extreme event of high RI (>100 years) will have a major disruptive impact if there is room for expansion of the channel system and providing thresholds for sediment transport are exceeded. If these thresholds are high, the fact that the channel has not recently been disrupted may also be important. The modes of expansion across the active area depend on the type of channel involved. Different study areas have different types of channel pattern present and thus a greater likelihood of certain types of planform adjustment. The role of more moderate events (10-50 years) varies principally with sediment size and channel slope. Small-scale modification may take place where stream powers associated with more moderate events exceed competence thresholds.

It was found that process rates were highly variable in both time and space and that present rates were not necessarily representative of the past 250 years. Even within this timespan, there have been periods of increased activity in response to increased discharges of moderate magnitude (eg. 1870s-1880s within the Dee study area) and random extreme magnitude floods (eg. between 1948-1956 in the Tweed study area). The impact of landuse change, especially in relation to sediment mobilisation (Dee and Spey study areas), and speed of runoff (Tweed study area) also appeared to be important.

Preface

(1) Within this thesis, there was the general problem that in the study of long data sets, by far the longer portion exists in imperial units. For purposes of standardisation however, here metric units have also been adopted. Thus, where imperial units are quoted in historical sources, the metric equivalent is also given.

(2) Those estate plans discussed within the text (Chapters 6 and 7) are referenced by their Register House plan number (RHP). Full details can then be obtained from the list of referenced estate plans at the end of the thesis (p 1126).

(3) All equations are in S.I. units except where indicated.

(4) A table of symbols used and their definitions occurs on Page lxiii

(5) For ease of reference a further copy of the following diagrams can be found within the envelope at the back of the thesis.

Figure 3.2.1.(i)	The relief of the Dee study area
Figure 4.6.(i)	Sampled river channel segments within the Dee study area
Figure 4.7.(i)	Sampled river channel segments within the Spey study area
Figure 4.7.(ii)	General location of catchments within the Spey

(6) Due to the bulky nature of the data sets used, the channel typology and rainfall data are available on magnetic tape with prior consultation with the author.

Symbols and their definitions

Parameter	Symbol	Units
Area of cross-section	A	m^2
Wetted perimeter	WP	m
Hydraulic radius	R	m
Width	W	m
Average depth	d	m
Maximum depth	d_{max}	m
Width: depth ratio	$W{:}d$	dimensionless
Average velocity	V	$m\ s^{-1}$
Discharge	Q	$m^3\ s^{-1}$
Bankfull discharge	Q_b	$m^3\ s^{-1}$
Sediment discharge	Q_c	$m^3\ s^{-1}$
Maximum flood on record	Q_{max}	$m^3\ s^{-1}$
Slope	S	dimensionless
Slope at bankfull	S_b	dimensionless
Size of particle for which x% of total particles are finer.	D_x	m
Manning's roughness coefficient	n	
Critical shear stress for initiating particle transport	τ_c	$N\ m^{-2}$
Darcy-Weisbach friction factor	F	dimensionless
Specific density of transporting fluid	ρ	$1000\ kg\ m^{-3}$
Specific weight of sediment	γ_s	$KN\ m^{-3}$

Parameter	Symbol	Units
Specific weight of fluid	γ_f	KN m^{-3}
Gravity	g	9.81 m s^{-2}
Mass	m	kg
Height	h	m
Gross stream power	Ω	W m^{-1}
Unit stream power	ω	W m^{-2}
Stream power index	Ω'	m^3 s^{-1}
Chezy resistance factor	C	dimensionless
Duboys' shear stress	τ	N m^{-2}
Shields competence at bankfull	D_s	m
Specific gravity of bed material	s	1000 kg m^{-3}
i th order statistic, the i th smallest in a sample	x(i)	
Location parameter in extreme value distributions	u	dimensionless
Scale parameter in extreme value distributions	α	dimensionless
Scale parameter in exponential distributions	$\varkappa$	dimensionless
Shape parameters of general extreme value distribution	k	dimensionless
Probable maximum precipitation	PMP	
Recurrence interval	RI	
Auto-correlation	AC	

Peak over threshold	POT
Aerial photograph interpretation	API
Flood Studies Report	FSR
Annual maxima	AM
Extreme value distribution 1	EV1
General extreme value distribution	GEV
Evapotranspiration	ET

CHAPTER 1

Introduction

"The rates at which floodplain processes operate are highly variable geographically, and .... even in Britain the surfaces of some floodplains have been entirely reworked by laterally migrating rivers within a hundred years or so. Other areas, by contrast, appear to have possessed stable rivers with at most only slow overbank sedimentation following floods for a matter of thousands of years. Both such environments have been considerably modified by human activity over the centuries, through canalization or channel stabilization, flood protection, and land drainage works." (Lewin, 1982 p21)

Scottish alluvial environments are known to have been highly variable in both space and time and thus it seems unlikely that fluvial processes reworking valley floors have operated at constant rates throughout the Holocene. Unfortunately, the length of the data base with which to analyse such change is rather limited. Runoff data are available extending from one to over forty years at selected gauging sites (Flood Studies Report, NERC 1975; Acreman, Ph. D. thesis in preparation), but from such brief records, it is difficult to assess the frequency of the runoff outliers in these series. There is also a great dearth of information about channel activity rates and channel response to discharge events of different magnitudes and recurrence intervals within upland Scotland, over a period of more than 40 years.

Occasionally, individual case studies of the geomorphic impact of a major flood have illuminated the current rates of river planform change. These have generally involved pre-flood reconstruction, peak discharge competence measurements and mapping of post-flood planform change (eg. Acreman, 1983; Werritty, 1984). It is however difficult to assess how typical, or atypical, these recorded changes are of the range of river channel types and modes of adjustment found in Scottish upland environments even within the last 250 years.

Thus, if we are to study channel pattern change from an applied viewpoint in order to assess how often we can expect different types of channel activity, we must study river regimes substantially earlier than the first gauging records. It is also clear that interpretation of the present rates of channel adjustment can only be considered as a legacy of the total range of a channel's behaviour. In this Hickin's recent comments are endorsed:

> "Our perspective on the fluvial system would benefit greatly from research focussed on channel changes during the last several hundred years rather than for shorter and longer timescales". (Hickin, 1983 p61)

There is therefore a great need to put present discharge events and their known geomorphic response into the context of the past record and to put present channel activity rates into the context of past rates.

Within this study, a series of questions are posed within the context of channel planform changes in upland Scotland over the last 250 years.

(1) What spatial range of channel planform types occur and what are the dominant controls?

(2) What rates of channel activity are evident and what are the dominant modes of adjustment?

(3) What is the planform response to runoff events of varying magnitude and frequency?

(4) Under what circumstances are rare floods geomorphically significant?

(5) Is there any evidence for change in the magnitude, frequency or duration of rainfall and runoff over the past 250 years?

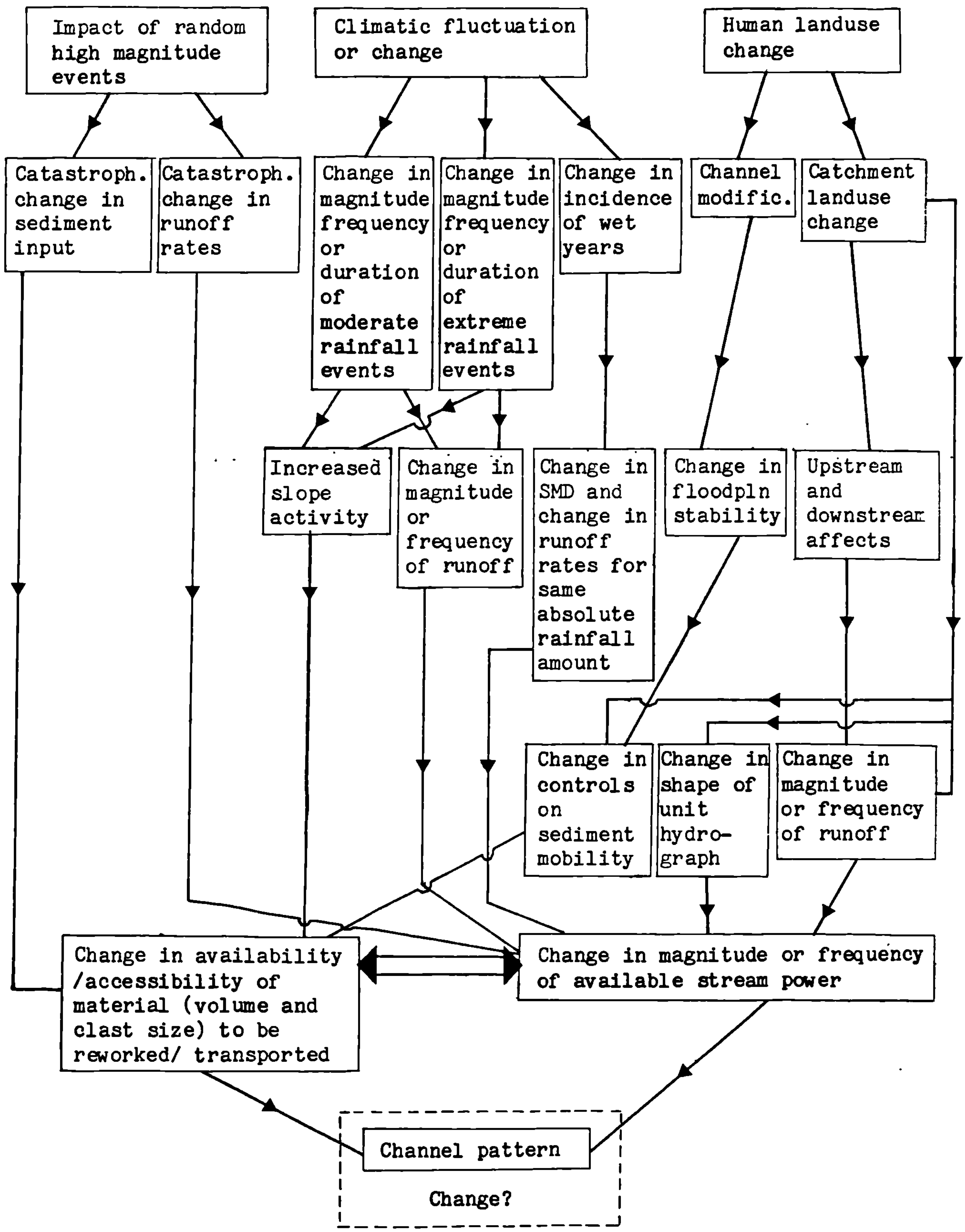
(6) Have there been any major landuse changes within the catchment which could have changed the runoff regime or sediment supply over the past 250 years?

It is only after the spatial variation in process rates has been initially assessed (Question 1) that more detailed questions can be answered. The temporal element is important (Question 2) in order to establish whether or not channel patterns have undergone major change over a 250 year period. If the magnitude and frequency of transport processes through time are clearly related to the magnitude and frequency of controlling events, it is necessary to ascertain what frequencies of controlling events are dominant in upland Scotland. It is important to evaluate what work rivers have the capacity to do across a range of flow conditions. Only after Question 3 is considered can the importance of inherent changes in the system be evaluated, as opposed to those arising in response to a change in system inputs. It must be ascertained from the past record whether present day channel patterns are tending towards equilibrium or disequilibrium with their environmental controls and how this is determined by the timescale of enquiry. It is important to appreciate what controls the past and present day geomorphic effects of extreme runoff and also what is the long term impact of extreme runoff as a landforming agent in this particular environment (Question 4).

If we are to study and assess the relative importance of landuse change and climatic fluctuations in understanding the controls on channel pattern, it is crucial to appreciate what changes in planform environmental variations could generate. A series of multiple working hypotheses (cf. Chamberlin, 1897) for explanation of channel planform change in upland Scotland are shown in Figure 1.(i). If associated planform changes have, or have not taken place, the importance of known climatic fluctuation (Question 5) and landuse change (Question 6) must be assessed. Then it can be evaluated whether climatic change or fluctuation has caused increased discharges of certain magnitudes. It also enables study of how robust the channel system is to fluctuation in

Figure 1.(i)

Multiple working hypotheses for explaining channel planform change in upland Scotland.



Abbreviations: modific. = modification
catastroph. = catastrophic
floodpln = floodplain

climatic inputs. Furthermore, landuse changes may have altered the speed of runoff rates, the shape of the unit hydrograph and both the rates of sediment mobilisation and transfer from slopes to channel over this 250 year period.

Although in an ideal world, these questions would be answered in relation to every reach of river channel in upland Scotland, this was obviously not practical. Clearly the area of study had to be constrained and focus was therefore placed on three contrasting upland areas in <u>eastern</u> Scotland. The west coast, in contrast, has smaller, more fragmented catchments, with a less complete data-base and thus was not selected for this study.

The three chosen study areas were as follows:

(1) The upper River Dee above Crathie, in Aberdeenshire.

(2) The River Spey between the Tromie and Avon confluences and the south to north flowing tributaries.

(3) The middle to lower River Tweed and the Leader, Whiteadder, Teviot and upper Bowmont Water tributaries.

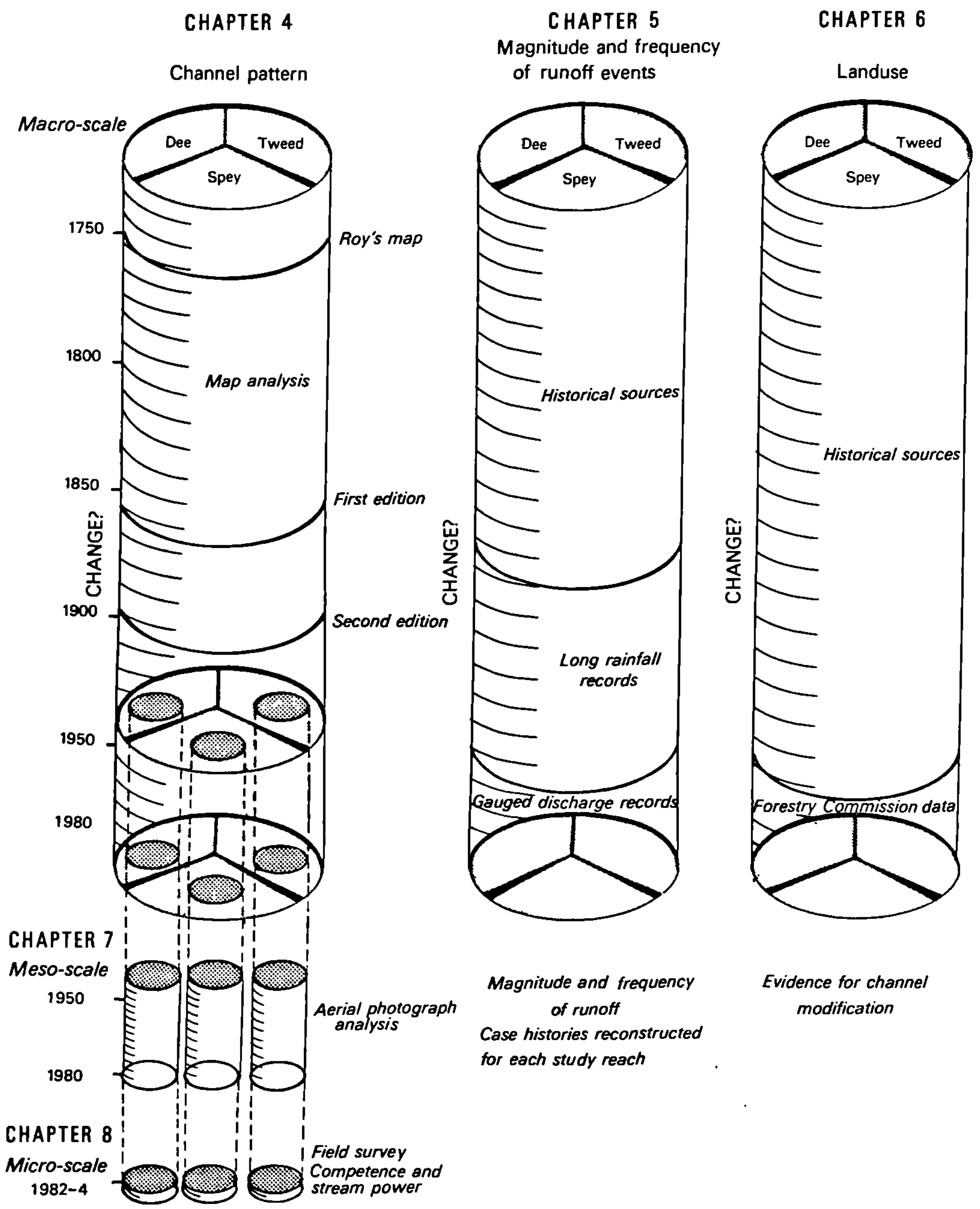
These three areas were chosen to highlight the contrasts both within the Cairngorms and adjacent mountain massifs but also to compare the differences between the Cairngorms and the Southern Uplands. The samples would therefore be expected to have certain similarities but also definite contrasts. The extrapolation of results beyond the study areas is clearly difficult to evaluate but the results are likely to be broadly correct for the Eastern Highlands and the eastern part of the Southern Uplands. West coast comparisons are likely to be unreliable, mainly because of climatic differences.

To answer satisfactorily the above geomorphological questions requires the integration of information from several fields of enquiry within geomorphology, climatology, hydrology, sedimentology and archival research. It must be remembered that, to a large extent, the data base is predetermined and inevitably there must be gaps. However, this approach holds the only key to an understanding of Scottish upland river systems in the three areas over a 250 year timespan.

The following research strategy was adopted, as indicated in Figure 1.(ii). Chapter 2 provides a literature review to place later chapters in context while Chapter 3 describes the three study areas, highlighting the principal contrasting environmental features. Chapter 4 involves the macro-scale analysis of map-based information to assess the range of form and rates of change within the framework of a channel typology. With rates of change established, Chapter 5 aims to reconstruct the magnitude and frequency of runoff events and discuss the relative importance of climatic fluctuation as opposed to random extreme flood events. The importance of the magnitude, frequency and duration of both rainfall and runoff events is evaluated. The existence of landuse change and the possible implications within the 3 study areas is

Figure 1.(ii)

The research strategy for the thesis



assessed in Chapter 6.

Having determined relationships on the macro-scale, a more detailed history of planform change and adjustment was carried out using aerial photograph analysis at selected high activity sites (Chapter 7). This was then assessed in relation to the magnitudes and frequencies of known flood records. Evaluation was made of the stream powers associated with sites of high activity to determine the range of flows that are geomorphologically important (Chapter 8). Finally, the results from these chapters are discussed in relation to the questions posed within this introduction and the ways forward for future research suggested (Chapters 9 and 10).

CHAPTER 2

Literature review

2.1 Introduction

The material of this chapter will be presented as an extension of the aims and questions outlined in Chapter 1 and thus provide a summary of the previous work from which this study arises. When reviewing the global and British literature on the spatial and temporal patterns of channel planform change, the discussion can be divided into four sections:

(1) The term "channel pattern" must be defined before any discussion can take place ie. what form are we studying?

(2) The basic types of channel pattern will be assessed together with the form properties of different channel planforms. This classification will be discussed in more detail in Chapter 4. The actual initiation of individual planforms will not be discussed in detail as it is not considered relevant to this study.

(3) Evidence for different types of change will be assessed and the known rates of change in contrasting fluvial environments within Britain will be discussed.

(4) The need for a unifying concept to determine the work done by a stream will be evaluated and the main variables in controlling and resisting this force are outlined.

It is necessary to put previous work in context primarily with research results from differing British situations, but also in terms of similarities and dissimilarities with other environments. In each section, specific information available in the Scottish context will also be analysed.

2.2 Definition of terms

It is first necessary to define the object of study. When a river's configuration is observed aerially, it reveals a planimetric variation of form, which represents the interface between the runoff regime and the sediment system and defines the particular channel pattern of the subsection of the river channel. Placed within the context of river channel dynamics, this pattern can be envisaged as representing a macro-scale mechanism of adjustment, which is carried out either in contrast or conjunction with meso-scale adjustments (eg. in cross-sectional geometry) and micro-scale changes (eg. in bed clast size composition). It is however extremely difficult to isolate the macro-scale planform controls since they cannot physically be considered independently of controls at lower levels within the hierarchy of resolution. Furthermore, the fluvial system is indeterminate at a range of scales from the channel planform down to the level of the individual cross-section (Maddock, 1970). The timescale of study is also highly important with different controls having a different status, dependent on the timescale considered (Schumm and Lichty, 1965).

To appreciate the importance of channel pattern within the fluvial system, it is first necessary to understand what planform represents in process terms. Pattern affects resistance to flow and in energy terms, the efficiency of the system. The existence of a particular pattern is to a large degree adjusted to two main factors: the amount, character and availability of sediment and the magnitude and frequency of the discharge regime. These factors will vary both spatially and temporally and the majority of previous studies may be criticised for giving little or no attention to upstream or downstream changes in these controls and the record of associated channel pattern responses.

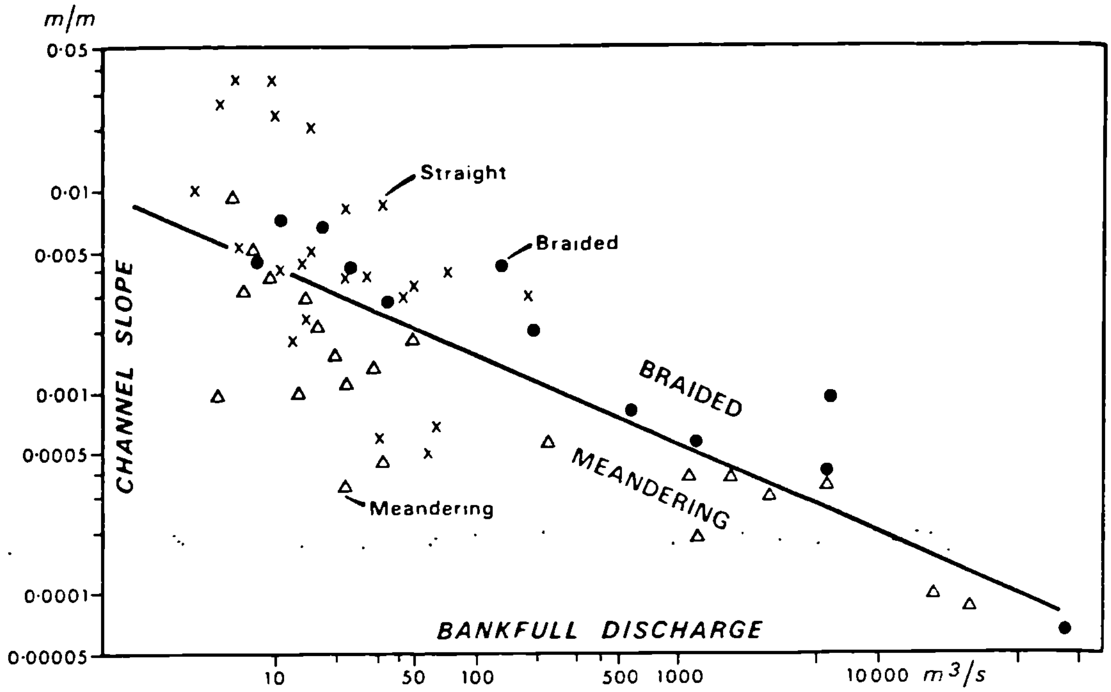
2.3 Channel classification

One major problem must be stressed at the outset of any attempt to classify pattern; river planform represents a continuum of response to a large interaction of control variables and thus represents a continuum of form from one extreme to another. There are no sharp distinctions for separation into fixed categories and thus any artificial attempt to do so will have its limitations. However, in terms of the processes involved, planform change is associated with crossing significant process thresholds (Schumm and Khan, 1972). Whether the two concepts of a form continuum and process thresholds are compatible is open to debate but there is a clear difference in the present interpretation of process and form.

Distinctions of form are possible in qualitative terms eg. between single or multiple channels or between straight and "wiggly" channels and such distinctions are useful. There is an important difference between a single and a split channel in process terms. To accommodate the same discharge, split channels tend to have greater boundary resistance, with higher W:d ratios. In contrast, there have also been several attempts to make such categorisation more detailed and quantitative.

A full discussion of the literature and problems of channel classification will be found in Chapter 4.3. There have been both inductive (eg. Schumm and Khan, 1972) and deductive (Chitale, 1970) attempts to relate pattern forms to certain controls, both in simulated flume experiments and field-based analysis, using a variety of techniques and measured parameters. For example, Leopold and Wolman (1957), using data from natural channels, plotted values of slope against bankfull discharge and distinguished a line which defined critical values to differentiate braided from meandering channels (Figure 2.3.(i)). For the purposes of this literature review, essentially four planform categories can be delimited, namely straight, meandering, braiding and anastomosing (Schumm, 1968). An example of such a genetically orientated channel classification is shown in Figure 2.3.(ii), with some indication of cause and effect relations eg. in terms of slope and relative stability (Schumm, 1981). Anastomosing channels are omitted for simplicity, as individual channels of an anastomosing planform can be meandering, straight or braided.

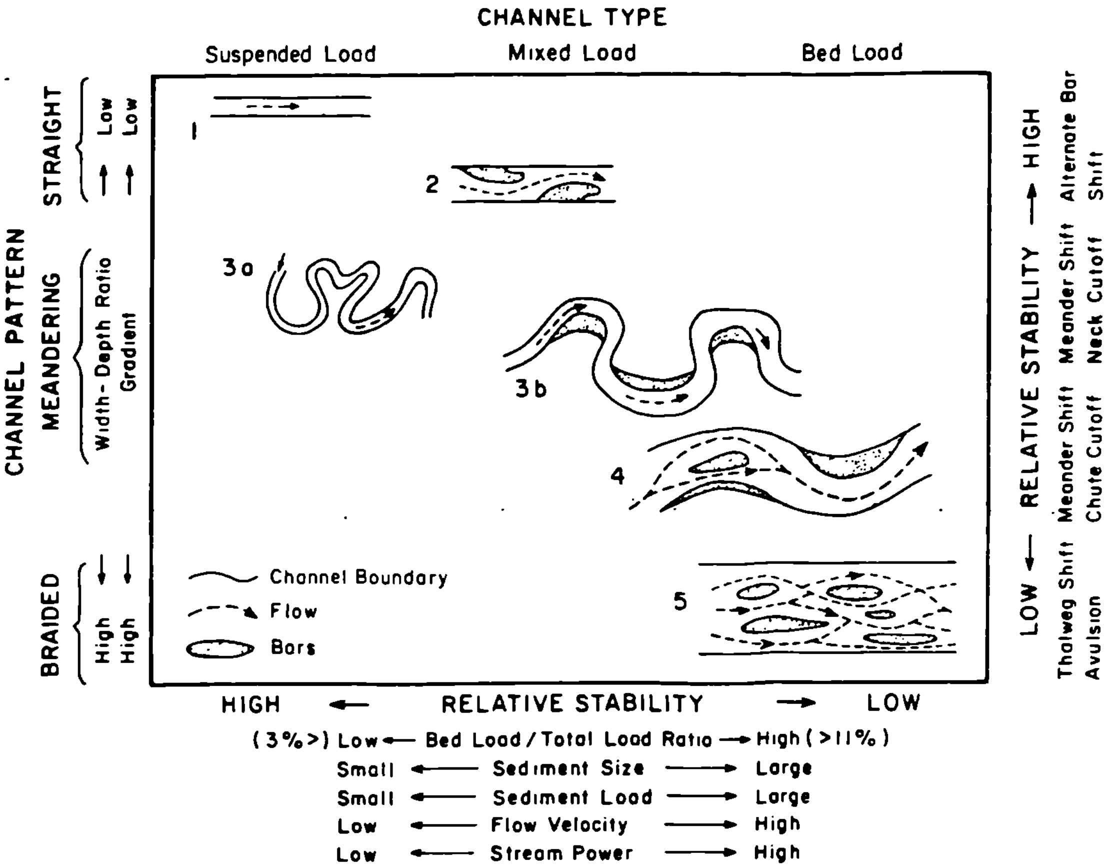
Figure 2.3.(i)



The relation of slope (S) to bankfull discharge (Q) of the form $S=0.13\ Q_b^{-0.44}$ was suggested by Leopold and Wolman (1957) to distinguish braided (above line) from meandering (below) patterns

(Source: Gregory and Walling, 1973 p252 Figure 5.5)

Figure 2.3.(ii)



Channel classification based on pattern and type of sediment load with associated variables and relative stability indicated

(After: Schumm, 1981)

(Source: Chorley et al., 1985 Figure 4 p24)

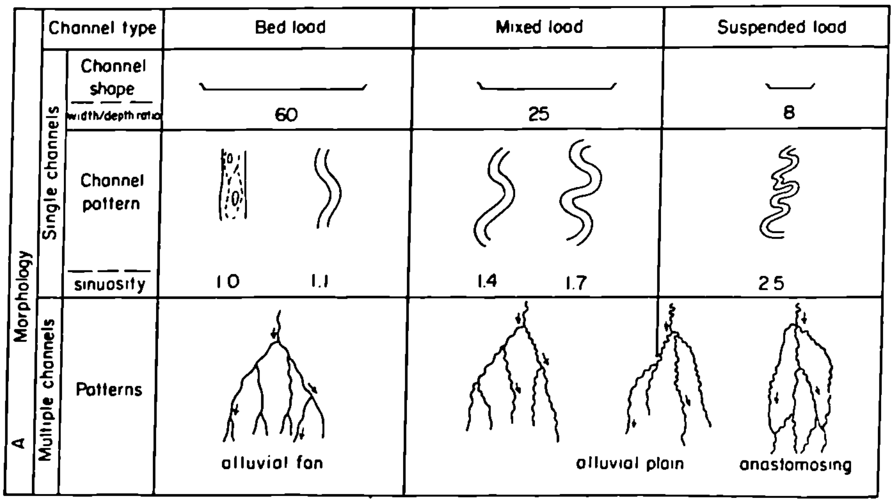
Different channel planforms, associated with different dominant sediment types, are typified by different cross-sectional forms defined by differences in W:d ratio. The W:d ratio is the function of the shear stress exerted against the shear strength of the channel boundary. The following general relationships were summarised by Simons and Senturk, (1977):

(i) channel width is directly proportional to both water and sediment discharge

(ii) W:d ratio is directly related to sediment discharge

In Schumm's (1968) classification of alluvial channels, bedload channels are characterised by high W:d ratios whereas mixed and suspended load channels have medium and low W:d ratios respectively (Figure 2.3.(iii)). However, although this distinction may apply to classical braided/ meandering channels, as described by Schumm (1977), the relative importance of different modes of sediment transport in determining W:d ratio is not so simplistic. For example, not all meandering channels are dominantly suspended or mixed load channels, such as the meandering planforms on the Calamus River described by Smith and Bridge (1985), which are dominantly bedload channels. Different modes of change are also associated with different planforms, and the processes of propagation will be discussed separately.

Figure 2.3.(iii)



Types of river channel

(After Schumm, 1968)

(Source: Schumm, 1977 p155 Figure 5.29)

Perfectly straight channels persisting over a long stream length, in excess of ten channel widths (Leopold et al., 1964), are unusual in natural channels since, as Gilbert (1914) observed, a free stream does not tolerate a straight channel. In nature, straight channels exhibit some tendency for the flow to follow a sinuous thalweg within the confines of their straight banks and Leopold and Wolman (1957) in their flume experiments confirm this (see also Ackers and Charlton, 1970; Schumm and Khan, 1972; Wolman and Brush, 1961). They also tentatively conclude that pools and riffles are a fundamental characteristic of all natural channels, including apparently straight channels. This would suggest that the processes forming straight channels are similar to those that will be discussed for meandering channels. It may be inferred that the straightness of a thalweg is a temporary disequilibrium state while the straightness of banks may be more constant in steady time.

There is a substantial literature concerned with the initial formation and the subsequent development of meander bends. Meanders occur when a channel is flowing on a surface that is too steep for the sediment load and water discharge carried out by the river. In providing greater length, it has been suggested that this pattern allows the more uniform dissipation of energy (Leopold and Wolman, 1960). The postulated mechanisms of meander initiation are numerous and problematic with no single theory explaining the formation of all meanders (Knighton, 1977; Friedkin, 1945). Keller (1972) produced a 5-stage model of meander development based on the energy distribution of pool and riffle sequences (Figure 2.3.(iv)), but again these sequences are

Figure 2.3.(iv)

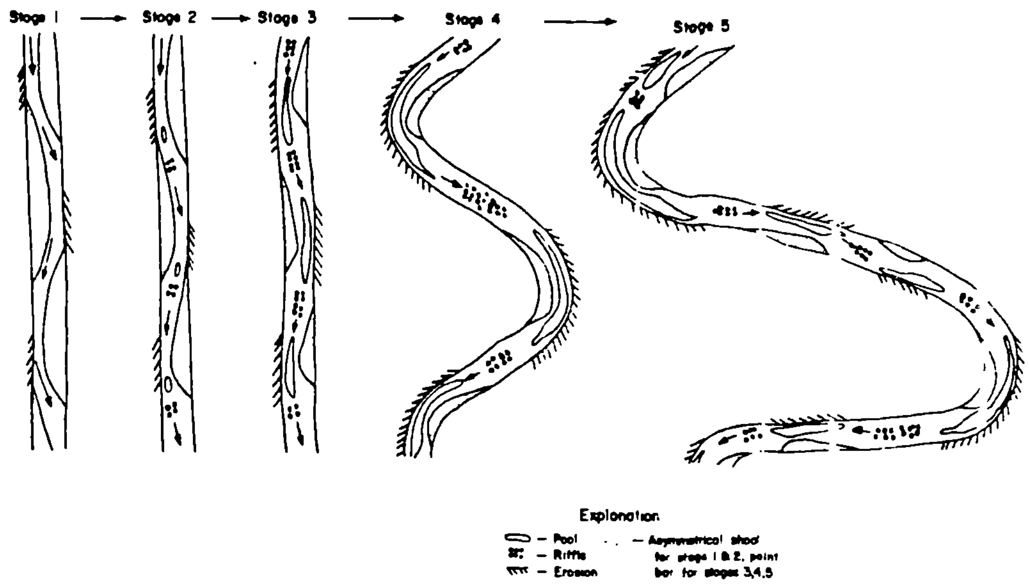


Illustration of the five-stage model of development for alluvial stream channels

Characteristics of the five-stage model of development for alluvial stream channels

Stage 1	No pools or riffles	Dominant bed forms are asymmetric shoals
Stage 2	Incipient pools and riffles at about 3 to 5 channel widths	Dominant bedforms are asymmetric shoals Pools and riffles are small
Stage 3	Well-developed pools and riffles with mean spacing 5 to 7 channel widths and mode 3 to 7 channel widths	Dominant bedforms are pools, riffles and asymmetric shoals (mostly point bars) Pools are about 1.5 times as long as riffles
Stage 4	Well-developed pools and riffles with mean spacing 5 to 7 channel channel widths and mode 5 to 7 channel widths	Dominant bed forms are pools, riffles and asymmetric shoals mostly point bars) Pools are generally greater than 1.5 times as long as riffles
Stage 5	Mixture of well developed pools and riffles with incipient pools and riffles mean spacing is generally 5 to 7 channel widths with a mode of 3 to 7 channel widths	Dominant bed forms are pools, riffles and asymmetric shoals Pools are generally much longer than riffles

(Source: Keller, 1972 p1534 Figure 4; Table 2) not universal for all meandering channels. An evolving pattern will only occur if discharge, wavelength and sinuosity are systematically increasing rather than remaining constant. Research has also focused on the controls of the size and shape of a meandering planform. Numerous measurements of the planimetric dimensions of meander patterns show relationships between the ratio of curvature to width (usually 2-3) and meander wavelength to width (usually 10-12) throughout a large range of stream sizes (Leopold and Wolman, 1957).

In contrast planimetrically, a braided channel shows division into several channels, which successively meet and divide. They are also characterised by a lack of development of distinct river bends, wandering of the twalweg at different rates and the development of pools and riffles (Bluck, 1976). Various explanations have been offered to explain the planform of braiding patterns including erodible banks, rapid and large variations in discharge, high regional slope, abundant load, local incompetence and a valley wide enough to braid (eg. Smith, 1970). Doeglas (1962) for example, suggested that large and sudden variations in runoff seem to form braided rivers. Henderson (1966) found that on steeper slopes with high stream power, banks are vulnerable to attack as the river dissipates its surplus energy through the formation of multiple channels.

There is an extensive literature on the processes associated with braiding initiation in a variety of contrasting environments and a variety of sediment sizes (eg. Leopold and Wolman, 1957; Ashmore, 1982). Smith (1970), studying the sandy River S. Platte in Colorado and Nebraska, concluded that a braided planform is initiated principally by the accretion and subsequent dissection of longitudinal bars and the dissection of transverse bars. Within gravel bed rivers, the dissection of longitudinal bars is more common (see Williams and Rust, 1969). Braiding patterns thus are frequently recognised as an instrument of aggradation. Accumulation of sediment may be stored both within the channel and in palaeofeatures within the floodplain.

Braiding is less frequently found as a means of degradation of the sediment store. For example, Fahnestock (1963) found two places in his White River study where there was a close association between degradation and braiding. Mackin (1948) also mentioned the existence of braiding in a stable or slowly degrading reach. From the literature, it seems that it cannot be inferred that a braiding reach is necessarily an accumulating sediment trap. In complete contrast, Wolman and Leopold (1957) argued that braiding could represent an equilibrium form. For example on the lower Spey, although approximately half the valley floor has been reworked over the last 100 years and large amounts of sediment have been flushed through the system, the overall level of surface sedimentation has not changed (Lewin and Weir, 1977). Obviously, it is hazardous to assume anything from study of only planform configuration without knowledge of the river system.

The fourth category, first identified by Schumm (1968), is that of anastomosing channels, which divide and reunite but these are small in size in relation to the size of the intervening islands. These will not be discussed here as there are no examples within this thesis.

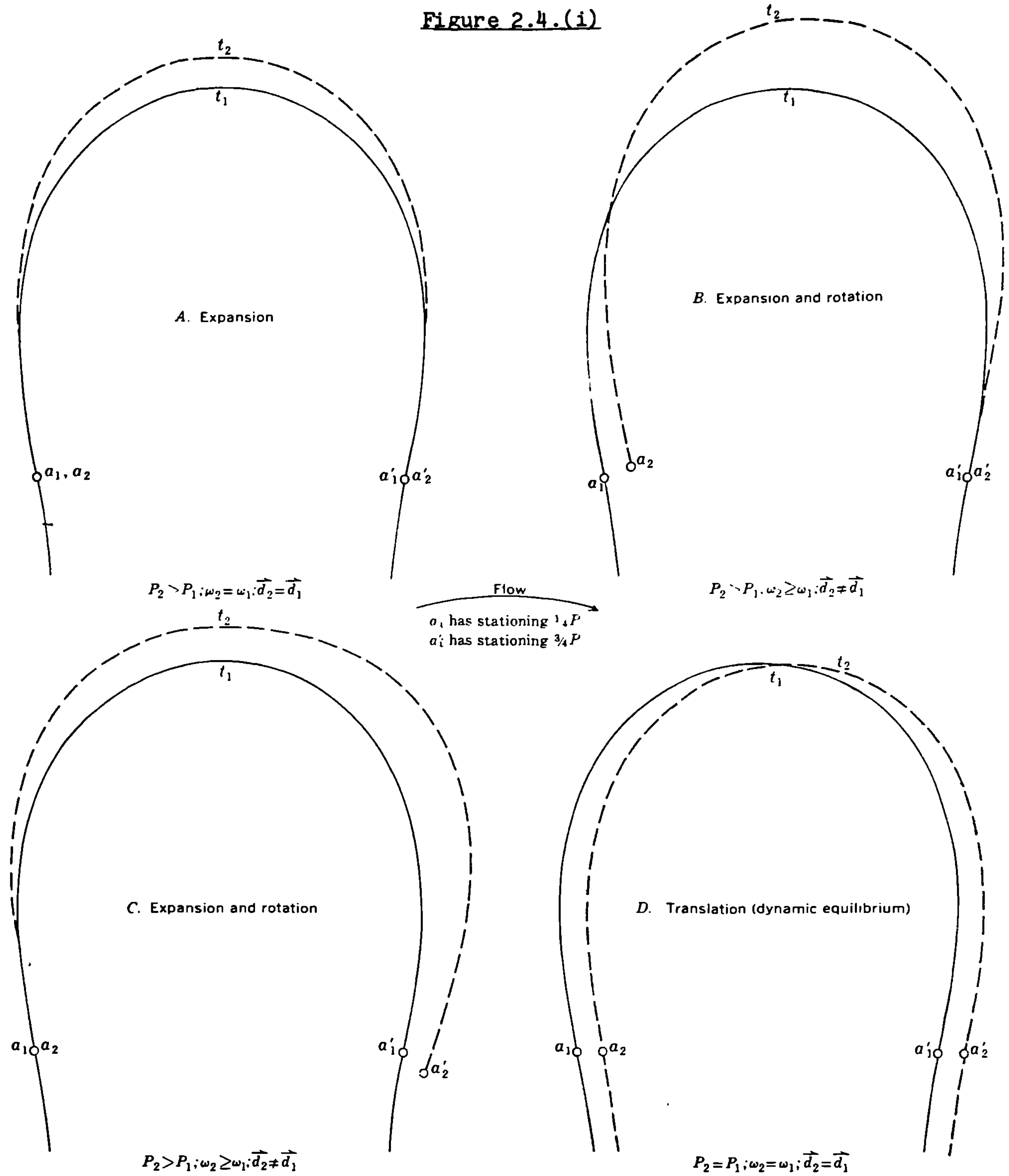
2.4 Rates of channel planform change

Channel planform change occurs when a new planform is maintained for a sufficient length of time to permit distinction between the previous and newly established pattern. Study of channel planform represents the study of equilibrium channel behaviour (Knox, 1977) while study of planform change focuses on the nature of any deviations from those equilibrium conditions. The scale of planform change can vary considerably, from localised bank erosion through avulsion to complete metamorphosis or transformation of planform morphology (Schumm, 1968; 1969; 1977).

Rates and types of such changes have been studied in a variety of fluvial environments both spatially and temporally by three complementary methods, which are dependent on the timescale of enquiry. Thus present day erosion may be studied by comparing sequences of planimetric maps or photographs for the same reach over a time-period determined by the length of study (often 1-10 years eg. Werritty and Ferguson, 1980; Harvey, 1977), or by erosion pins, which allow rates of bank erosion to be monitored (eg. Wolman, 1959). To study planform changes over a longer time-period of 10-150 years, available map (Moseley, 1975), aerial photograph (Hooke, 1977) and historic evidence can be analysed. Over a period of 100-500 years, floodplain vegetation successions may be analysed (eg. Hicken and Nanson, 1975). Other dating techniques using the Holocene stratigraphic record, such as radiocarbon dating and pollen analysis of soil horizons or organic deposits in slackwater deposits, allow the dating of individual or sequences of palaeodeposits well back into the Holocene (eg. Baker et al., 1983). The literature on spatial variation of changes throughout a basin is very limited although Hooke (1977) has attempted this for rivers in Devon.

Different types of channel pattern change have been documented for different types of channel pattern. Migration of meander bends is a characteristic process of alluvial rivers and one of the most conspicuous features taking place in the fluvial landscape. In terms of meander movement, once a meander pattern becomes established, hydraulic conditions in the bends and variations in the bank materials cause an enlargement of meander amplitude. It is now recognised that simple expansion into a sine-generated form (as proposed by Langbein and Leopold, 1966) is a gross oversimplification in most environments. Thus the complex process of channel movement involves the rotation, translation and expansion of meander loops and an increasing path length (Daniel, 1971; Figure 2.4.(i)) but the resultant regular pattern in these Indiana streams may be limited to more resistant boundary materials. Irregularities are more likely to occur due to differences in resistance of more easily eroded material; any one or combination of these modes of channel migration may occur depending on the spatially variable boundary conditions within the loop (Hooke, 1977; Hooke and Harvey, 1983; see Figure 2.4.(ii)). Thus, contrasting forms of planform development have been documented, including complex compound lobe formation (Brice, 1973), asymmetric lobes (Hooke and Harvey, 1983), neck, chute and multiloop cutoffs (Lewis and Lewin, 1983) and the rapid abandonment of lengths of cut bank and deep channel. Such complex loops are associated with rapid growth and a second inflection. Irregular patterns are unlikely to be in a quasi-equilibrium condition but if

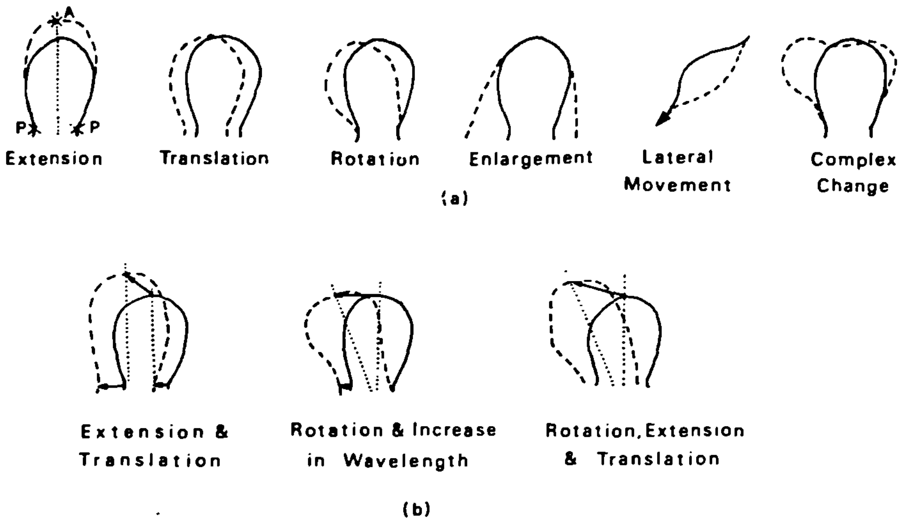
Figure 2.4.(i)



Effects of stationary and (or) translating end points on meander loop movement

(Source: Daniels, 1971 pA4 Figure 3)

Figure 2.4.(ii)



Types of meander movement: (a) primary elements (b) combinations

(Source: Hooke, 1977 p278 Figure 17:10)

other bends decline while others grow then the overall level of irregularity may remain more or less constant over time. It does seem difficult to appreciate how a regular pattern can exist spatially for long under natural variation in controls.

In terms of rates of meander bend migration, maximum values are recorded where the ratio of channel curvature to stream width approximates 3.0 (Hickin and Nanson, 1975). Table 2.4.(i).(a) shows the documented rates of change across a variety of environments. It is by no means exceptional to record bank recession rates of 10-15 m yr^{-1} in large rivers (Ferguson, 1977). Other bends on the same channel and over the same timespan may shift more slowly or not at all ie. rates are highly variable in space and time. A British example is Hooke's (1977) work on Devon rivers, where meander movement of up to 100 m in a 50 year period was found from map-based data.

Spatial variability in process rates within a meander planform causes the occurrence of relict features such as meander scrolls, meander cutoffs and oxbow lakes (Lewis and Lewin, 1983). What happens when actual cutoff processes occur? As meanders enlarge, they become increasingly sinuous with a reduction in gradient, the meander neck is reduced in size and a neck cutoff occurs. Aggradation will precipitate cutoffs and the subsequent formation of stagnant oxbow lakes (Smith and Stopp, 1978; NEDECO, 1959, Lewin, 1977). For example, the Beatton river planform, described by Hicken (1974), included many straightened reaches produced by neck cutoffs. Another type of cutoff has also been distinguished; Thorne and Lewin (1979) found chute cutoffs were the dominant form of planform adjustment within Welsh rivers.

Table 2.4.(i)

Documented rates of channel pattern change over a variety of fluvial environments

(a) Dominantly meandering planforms

River	Rate	Source
Rio Ucayali, (Amazon)	meander loop forms and cuts cuts off in 500 years.	Lathrap (1968)
Hernad, Hungary	meandering planform, 5-10 m yr^{-1}.	Laczay (1977)
Russian rivers	Meander bend migration 100 m yr^{-1}.	Kondratyev & Popov (1967)
Mississippi	Cutoff requires 1000 year period. Maximum rate was 14 m in 100 years. Natural relaxation time following cut off was 30-80 years.	Winkley (1970)
River Bollin	1872-1935 minor channel shifting. 1935-1975: 7 meander cutoffs during 38 years. Rates of erosion 2.1-5 m yr^{-1}.	Moseley (1975)

Table 2.4.(i) cont.

River	Rate	Source
Bollin-Dean	0.48 m in 18 months.	Knighton (1973)
Devon rivers	meander movement- up to 100 m in a 50 year period.	Hooke (1977)
River Cound	mean rate 1.71 m yr^{-1} over 1/1972- 9/1974.	Hughes (1977)
Welsh rivers 1000 km of channel eg. Teme and Severn	cutoffs: 1 per 5 years over 1880-1900; 1 per 2 years over 1950-1970.	Lewis and Lewin (1977)
N. Ireland	erosion of meander bends 0.03-0.06 m yr^{-1}.	Hill (1973)
Harthope Burn, Northumberld	1897-1976 Average shift 5 m Low sinuosity chute cutoffs.	Milne (1982)

Table 2.4.(i) cont.

(b) Dominantly braided planform

River	Rate	Source
Brahmaputra	Bank erosion 793 m yr^{-1} maximum. Bar movement in one flood was 1525 m.	Coleman (1969)
Yellow river	90-120 m yr^{-1} Rapid channel shift.	Chien Ning (1961)
River Feshie	Channel widened 136% from 1977 to 1981. Bank erosion locally exceeded 10 m yr^{-1}.	Werritty & Ferguson (1980; 1983)

Certain other features of the migration process in meanders should be noted. In uniform material, amplitude does not progressively increase, nor do meander loops cutoff during the downstream migration of bends although such steady state migration may occur in a flume (Friedkin, 1945). (Such a "steady state" would imply a balance between process and form.) If in contrast a meander pattern is progressively altering its average properties, then it is neither in a steady state nor in dynamic equilibrium but rather it is undergoing slow metamorphosis or evolution. Irregularly meandering planforms are less likely to maintain a steady state (Ferguson, 1977). However a pattern may maintain its aggregate degree of irregularity, despite changes in detail.

When reviewing the types of channel changes that can be expected to occur within a braided channel, several types of response recur. If erosion on one bank of a braided reach and deposition on the other are approximately equal, then a gradual shifting of the channel over the whole valley takes place without any appreciable change in channel size. Whole stabilised areas develop topographically higher than the active tracts as avalanche faces gradually build up, especially during more extreme discharges. Many old channel scars and dry channels are left behind, frequently plugged up with sediment (eg. Werritty and Ferguson's (1983) account of channel change on the River Feshie; see also Hitchcock, 1977; Church, 1983).

A process of rapid widening may occur when an increase in flow stress (such as a catastrophic flood) disrupts the system. The braided pattern may become accentuated at the periphery by the extra-channel reoccupation of old flood channels, both within the "active area" and, if stage permits, within the proximal floodplain. Competence allowing, this causes material to be deposited during the waning stages of these channel reoccuping events eg. Williams and Rust (1972) in the Donjek valley. In addition, widening of these flood channels may take place. Braided planform may also be increased by the excavation of new channels within the former active area, but high flows may too remove large quantities of sediment to produce flood channels, outwith the former active area. This can be accompanied by the development of new medial bars, as the channel anabranches. However, the upstream plugging up of formerly active anabranches either by sediment deposition or scouring of the channel bed will lead to a reduction in active area and thus may cause instability downstream (Hitchcock, 1977). In all instances, the morphology of palaeochannels reflects the hydrologic regime during the last occasion when the channel was active (which may of course be during a low recurrence interval extreme flood event).

In comparison with braided channels, rates of change associated with meanders gives an impression of apparent relative stability. Rates of change in high energy braided environments such as sandurs can be very rapid. Examples of rates of change in other environments are shown in Table 2.4.(i).(b). In Scotland, Ferguson and Werritty (1983) found the River Feshie in their study reach to have widened by 136% from 1977 to 1981, with local bank erosion exceeding 10 m yr^{-1}. However, such a rapid response is not necessarily universal; Leopold and Wolman (1957) suggest that braiding once established, may be maintained with only slow modifications. Obviously, all depends on what is classified as braiding; there is clearly a continuum of form under this broad category. Fahnestock (1963) found that on White River that the main channel may remain stable for several years, with much less shifting than individual anabranches. Stability of such features suggests that rivers with a braided planform may be as close to quasi-equilibrium as rivers possessing meandering or other planforms.

2.5 Equilibrium, thresholds and complex response

The two main categories of planform changes, which can be identified, are "autogenic" and "allogenic" changes (Lewin, 1977). Both are important in terms of interpreting pattern evolution but are also highly dependent on the timescale of study. In addition, there is the problem that these categories are not mutually exclusive and thus they may be more useful conceptually rather than in terms of application.

Over graded time, archaeological evidence supports the theory that most rivers are susceptible to constant autogenic or internal changes as a part of their morphological evolution and that the natural shift of channels with a reworking of the floodplain is a completely natural and expected process. Aggradation and erosion can take place without necessarily implying climatic or landuse change (Starkel, 1983). Thus, obvious palaeofeatures, such as terrace fragments, incised fossil fans or old meander cutoffs, can represent the long term adjustments internal to the workings of each individual basin. More recent features indicating channel migration, such as avulsion or recent swale and ridge sequences, represent shorter term internal adjustments. From such palaeochannels, it is possible to calculate the rate of past processes, as seen in Table 2.4.(i) (Lathrap, 1968; Alexander and Nunally, 1972). While a meander forms and cuts off on the Rio Ucayali over a period of 500 years (Lathrap, 1968), point bars have been deposited and destroyed in rapid succession over a much shorter time-period. Thus timespan is important because bank erosion and cutoffs may represent instability over a timeperiod of a few years but during 1,000 years, the process can be considered part of normal internal river behaviour and adjustment. This again poses the question of contrasting timescales, especially in terms of how stability is defined.

In contrast, allogenic change in channel pattern occurs in response to a systems change, which may be caused by an external shock to the system, resulting in a complex alteration in the controlling variables (Lewin, 1977). Changes may be instigated by random climatic fluctuations, such as an increase in extreme flood events, or by a change in mean climatic values. Sediment load and runoff rates may be altered through change in their availability by both man-induced and bioclimatic landuse changes.

The importance of change in channel pattern is therefore that it reflects the altered or changing trends of fluvial processes, in relation to parameters that are both internal and external. Just as the extent to which these variables will affect the system will vary, so too the extent to which the system will respond will vary. At the extreme case, metamorphosis need not be autogenic. How large such a change has to be to trigger a single change or a sequence of changes, will depend on the delicacy of the balance between controls in the environment in question (Brunsden and Thornes, 1979).

When the process-response system, reflected in river planform state, has achieved a balance between sediment to be transported and available discharge, a condition of equilibrium is attained. The application of the concept is however highly timescale dependent and will also tend to fluctuate around a mean condition. Its attributes are neatly summarised thus:

> "True equilibrium is a theoretical state towards which the system behaviour is tending with greater or lesser rapidity, by attempting to absorb the successive effects of a sequence of process inputs of differing magnitude and frequency." (Chorley et al., 1984 p11)

The concept of channel form tending towards a grade or equilibrium condition was first modified and extended by Mackin (1948) from Davisian theory. The grade condition was where:

> "Over a period of years, slope is delicately balanced to provide, with available discharge and prevailing channel characteristics just the velocity required for the transportation of all the load supplied from the drainage basin. The graded stream is a system in equilibrium; its

diagnostic characteristic is that any change in any of the controlling factors will cause a displacement of that equilibrium in a direction that will tend to absorb the effect of the change." (Mackin, 1948 p471)

This concept of adjustment between the parameters of process and form, has attracted considerable criticism (Knox, 1975, Hack, 1960), for example that this equilibrium condition was not supported by observation (Kesseli, 1941). However, Schumm and Lichty (1965) re-established the concept for dealing with landscape change at a intermediate or graded timescale. Leopold and Bull (1976) further defined grade to incorporate hydraulic adjustment over a longer period. In contrast, the concept of "quasi-equilibrium", with a more morphologic connotation, represented attainment of equilibrium over a much shorter timescale. This was associated with a balance between form and process within the channel, achieved over hours or days.

The time taken for a system to achieve a new equilibrium, following a change in input or internal operation, is also very important and the answer will obviously vary from system to system. Schumm (1973,1977,1979) has identified the importance of thresholds and complex responses intrinsic in some fluvial systems. As previously asserted, associated with the channel pattern form continuum, there exist abrupt thresholds between major process states, which occur when critical limits are exceeded (Schumm, 1973). Thus, stream gradients and valley floor altitudes do not change steadily and progressively through geologic time, instead relatively brief periods of instability and incision (ungraded) are separated by long periods of relative stability or grade (Schumm, 1975).

Two types of threshold have been identified within the fluvial system. These comprise intrinsic thresholds, associated with internal adjustments, which concern processes involved with sediment load and stream power eg. Schumm and Khan (1972). In contrast, extrinsic thresholds are triggered by an external stress and thus, for example, critical slopes may be exceeded and an extensive alteration in channel pattern may take place. The trigger could be a major flood or a clustering of major floods (Schumm, 1977). This may push the evolution of river channel planform in a new direction and result in channel metamorphosis (Schumm, 1969). Such a metamorphosis is described on the River Cimarron by Schumm and Lichty (1963). Thus, thresholds may be exceeded by alteration of the present process rates by external or internal controls. Therefore in some situations, adjustments from one stable state to another may be gradual. In contrast, other geomorphic systems operate with sudden changes from one stable state to another, for example through episodic erosion as process thresholds are exceeded.

Obviously, as stressed by Schumm and Lichty (1965), the time-period of observation is crucial to geomorphic equilibrium concepts (Figures 2.5.(i) and (ii)). Changes that may appear as smooth transitions over one period of observation, may appear as step functions when viewed in a different temporal perspective. However, it must be remembered that the difference between two rivers at a particular time may reflect a true change in the processes involved to gain quasi-equilibrium but this could also be accounted for quite naturally by temporal departures from

Figure 2.5.(i)

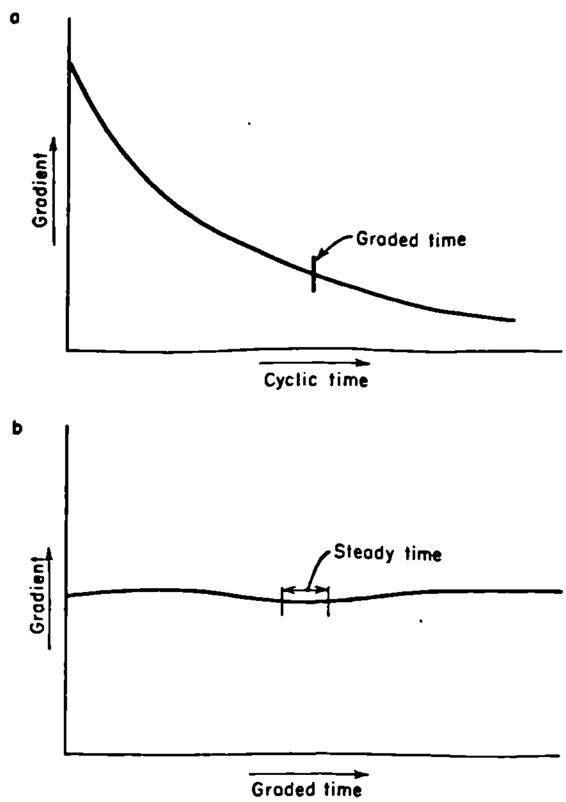


Diagram showing the concepts of cyclic, graded and steady time as reflected in changes of stream gradient through time.

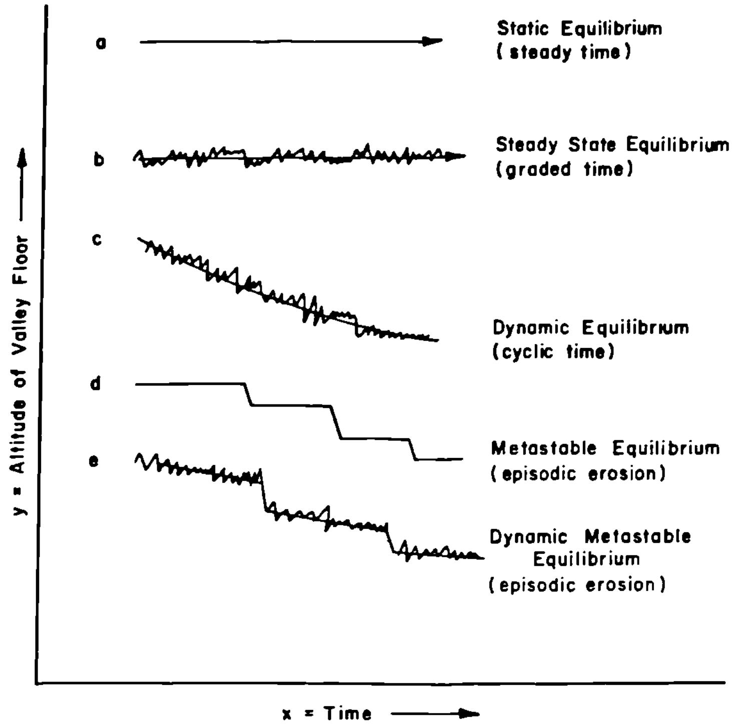
(a) Progressive reduction of channel gradient during cyclic time. During graded time, a fraction of cyclic time, the time

(b) Fluctuations of gradient above and below a mean during graded time. Gradient is constant during the brief span of steady time.

(After Schumm and Lichty, 1965)

(Source: Schumm, 1977 p11 Figure 1.4)

Figure 2.5.(ii)



Types of equilibria based on Chorley and Kennedy (1971). Each is shown with respect to changes of valley floor elevation with time.
(a) Static equilibrium- no change with time (steady time)
(b) Steady-state equilibrium- variations about a constant average condition (graded time)
(c) Dynamic equilibrium- variations about a changing average condition (cyclic time)
(d) Metastable equilibrium- static equilibrium separated by episodes of change as thresholds are exceeded.
(e) Dynamic metastable equilibrium- dynamic equilibrium with episodic change as thresholds are exceeded

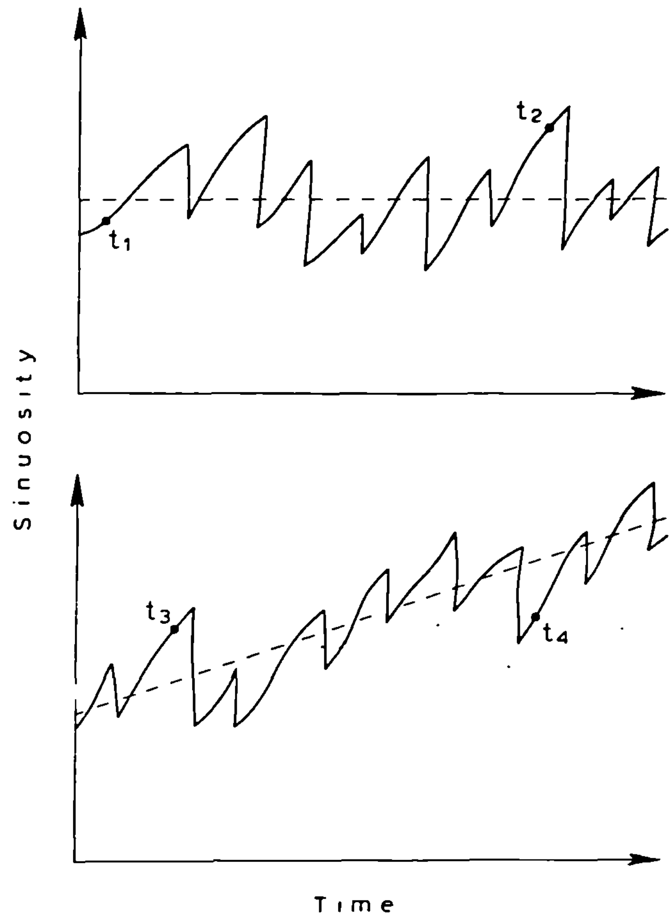
(Source: Schumm, 1977 p5 Figure 1-2)

a common equilibrium, as seen in Figure 2.5.(iii) (Ferguson, 1977). Simulated fluctuations of sinuosity with time reveal such problems of pattern interpretation in connection with equilibrium conditions (Gregory, 1977). For example, Fahnestock (1963) in his analysis of channel pattern change on the White River, Mt. Rainier, Washington over a yearly cycle found that a marked change from a meandering to a braided pattern took place with the onset of the high summer flows and the pattern returned to meanders with the lower autumn flows.

Associated with the concept of the exceeding of thresholds and the consequent disruption of the system, is the complex response of the channel as it searches for a new equilibrium. There is frequently no simple cause/ effect relationship, but rather a delayed transmission of information through the system. The concepts of relaxation time or recovery indicate the time taken to return to a quasi-equilibrium condition.

It is necessary now to assess the controls involved in regulating the stream power of individual systems. The problem of using channel pattern as an indicator of underlying process is that geomorphic theory is at present inadequate to explain or predict planimetric movement, because of the number of variables involved and the complexity of their interaction in the natural environment (Hooke, 1977). At present, no single theory can predict the processes of meander migration or braid formation or movement reliably. However some attempt will be made here to outline the main documented controls. A distinction must be made between the strength of controls which initiate the development of a particular planform towards an equilibrium state and those which might

Figure 2.5.(iii)



Equilibrium or change? The ambiguity of comparing an irregularly migrating meander pattern at two points in time

(Source: Ferguson, 1977 p96 Figure 15.1) upset that stability and initiate change.

2.6 Stream power

Bagnold's (1966) concept of stream power provides an important concept in explaining the spatial variation in activity rates, relating the work expended by a stream and the quantity of sediment transported. Controls on planform change can be subdivided into processes controlling the stream power input to the system (ie. controlling boundary shear stress) and the forces resisting such shear stresses. The temporal interaction of these controls determines the stream power available for carrying out sediment transport. Variables controlling inputs to the system are rainfall magnitude, frequency and duration and the frequency of hydrometeorological conditions that generate discharge events of different magnitudes. Within the actual channel, the main stress controlling variables are slope and depth indirectly reflecting cross-sectional form. In contrast, the most important resisting forces include the particle size distribution and geomechanical properties of the sediment and the confinement of the channel, in terms of its Glacial/ Holocene legacy and bedrock controls. The importance of vegetation cover must be considered both in terms of its hydrological significance and in its giving varying degrees of stability to floodplain, river banks and bar successions. Finally, a variety of anthropogenic controls on these variables must be assessed. The interaction of all these variables must be stressed.

In terms of channel hydraulics, total stream power (Ω) represents the liberation of potential energy (P_e=mgh) into kinetic energy (K_e=$^1/_2$ mv^2) as a river flows down its valley slope (Bagnold, 1966), as shown in Equation 2.1.

$$\Omega = \rho g Q S \ (W m^{-1}) \qquad (2.1)$$

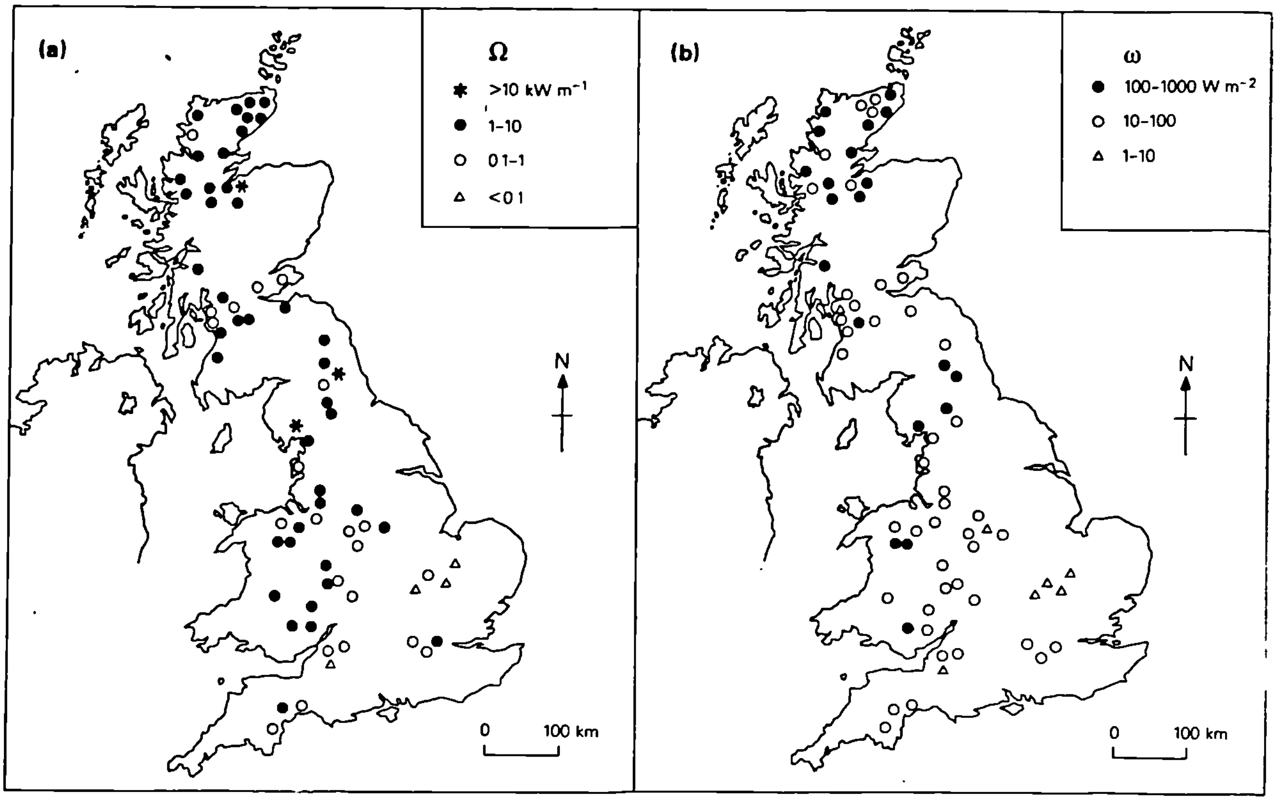
Specific stream power, ie. power per unit area of bed, can be defined either as the product of shear stress/ unit bed area and average velocity or gross stream power divided by channel width.. Bagnold (1977) argued that a stream can be conceptualised as a transporting machine whose efficiency in using the available power can be applied to the rate of sediment transport. As well as dissipation through the work of transporting sediment, stream power is also lost through frictional resistance and heat generation. Transport of bed material is directly related to stream power and concentration of fine material and inversely related to the fall diameter of the bed material (Simons and Senturk, 1977). Thus, if available stream power is less than the critical stream power necessary to transport average sediment load, aggradation will occur (Leopold and Bull, 1979). However, if the critical stream power is greater, degradation can be expected to occur. If both are equal, then a condition of grade or quasi-equilibrium (depending on the timescale of study) has been attained. Whether or not such a condition has been reached depends on the position of the stream at a particular threshold. Only a small variation in this ratio may be sufficient to push a stream out of an equilibrium condition eg. by the occurrence of a major flood event. Thus:

> "Excessive rates of power expenditure locally ... result in changes in channel form because of this [sediment] transport, until power expenditure is reduced and the channel stabilised at a minimum power threshold compatible with the threshold for the sediment, which is a geological constraint (Bull, 1979)." (Richards, 1982 p27)

The application of the concept has however been limited because of the problems of defining the critical stream power in precise quantitative terms. Schick and Lekach (1981) studied high bedload transport rates in relation to stream power of a major flood event in the watershed of the Wadi Mikeimin in SE Sinai mountains. This validated the proportionality of bedload transport rate to $^3/2$ power of unit stream power (in excess of that required for initial motion) for very high transport rates (Bagnold, 1977). However, stream power estimates at reconstructed high discharges may be unreliable.

According to Ferguson (1981), stream power is an important but neglected property. Maps of stream power at bankfull discharge in a variety of British fluvial environments exhibit a considerable range of between highland and lowland values (Figure 2.6.(i)). Ranges of stream power were also related to particular types of channel pattern. Ferguson found that underfit, straight or irregular rivers have bankfull unit stream powers from approximately 1-60 W m^{-2} while active free meanders ranged in power from 5-350 W m^{-2}. Rivers with high stream

Figure 2.6.(i)



Stream power at bankfull discharge in some British rivers (data from Nixon, 1959; Charlton et al., 1980; Riddell, 1980). The pattern of power per unit channel length (Ω) and per unit channel area (ω) is broadly similar.

(Source: Ferguson, 1981 p96 Figure 4.4)

power but apparently inactive were all incised or confined and thus, unable to use the power available.

2.7 Magnitude and frequency of channel forming discharge

In processes such as river discharge, the applied stress is provided by climatically controlled events. This poses a question as to the relationship between rainfall and runoff and whether extreme rainfall necessarily produces extreme floods (Finch, 1972). Similarly, the recurrence intervals of rainfall and peak runoff associated with the same event are not necessarily the same order of magnitude. It is important to establish the magnitude and frequency of the discharge history along a particular reach, to allow assessment of channel response to stream powers of different magnitude and frequency. As discussed, channel planform can be envisaged as attempting to attain an equilibrium state characterising this runoff regime and local sedimentary environment. Local changes in planform may occur when the competence of the discharge exceeds the erosional thresholds for bank and bed materials.

In terms of relating process to form, there has been considerable debate as to the contrasting role of extreme catastrophic events of high recurrence interval to the net effect of more frequent events of lesser magnitude (Wolman and Miller, 1960). The relative importance of such events in terms of fluvial processes and sediment delivery can be measured by the work done and more importantly, their geomorphic effectiveness in the landscape (Wolman and Gerson, 1978). It is

In terms of frequency, bankfull discharge is however not necessarily $Q_{1.58}$ (Williams, 1978) and many would argue against the concept of a single dominant discharge (Woodyer, 1968). In different regimes, frequency of both bankfull discharge and dominant discharge may differ (eg. Pickup and Rieger, 1979). Froelich, Klimek and Starkel (1972) found in the Dunajek valley that bedload transport takes place only during high discharges lasting normally several days per year. However, this channel manages to contain higher flows and thus undergoes little modification during such higher stage events.

In all these examples, this frequency of sediment transport was due to the low thresholds of sediment entrainment involved. However, the coarser the bed material, the larger the tractive force necessary before initiation of movement takes place. Where stresses generated by frequent events are incompetent to initiate transport of available materials, less frequent flood events of higher magnitude are obviously required. Correspondingly, there is an extensive literature on channel response to flood events of different magnitudes and recurrence intervals and only a few selected examples to illustrate the main characteristics can be discussed here (Table 2.7.(i)). A very good analysis of the long term impact of a major flood is provided by Schumm and Lichty (1963). In 1914, a major flood occurred and destroyed the floodplain of the River Cimarron of S.W. Kansas. Previous to that date, the river channel had been highly sinuous and relatively narrow (15 m wide), with associated low levels of sediment transport. However, post flood, the channel widened to 366 m, forming a wide straight bedload transport river. Obviously sediment transport rates were

Table 2.7.(i)

Magnitude and frequency of geomorphic activity within different fluvial environments

(a) Global examples

River	Rate	Source
Colorado River, Grand Canyon	RI between 1 and several years.	Wolman & Miller (1960)
Little River, (headwaters Shenandoah R. Appalachians)	Flood of 100 years RI. Lack of post flood recovery.	Hack & Goodlett (1960)
Bassin de Guil, French Alps	1000 year event initiated disequilibrium period of slope activity. More than total accumulation during the Holocene.	Tricart (1961)
Gila River, Arizona	8 major floods between 1905-1917 caused major widening (45-610 m). Recovery in 50 yrs.	Burkham (1972)

Table 2.7.(i) cont.

River	Rate	Source
Dunajec valley, Poland	Bedload transport only during high discharges commonly lasting several days per year.	Froehlich et al. (1972)
(b) British examples		
Langden Brooke, Forest of Bowland	Over period 1972-1975 planform development takes place by a sequence of major changes which occur in response to events with RI= 1 year. More frequent events are associated with minor changes.	Hitchcock (1977)
Grains Gill, Howgill Fells	Sediment removal and major channel changes about once per year.	Harvey (1977)

Table 2.7.(i) cont.

River	Rate	Source
River Feshie, Cairngorms	Flood with Q_{max} of 200 $m^3 s^{-1}$ - disruption of sinuous meandering channel pattern to unstable braiding form. Modified over 17 yr relaxation period to meandering thalweg.	Werritty & Ferguson (1981)
R. Cannich, Inverness-shire	Acute flooding with RI of 30-60 years caused Liatrie burn to change course over floodplain.	Wolf (1952)
Ardessie Burn, Wester Ross	Major planform disruption caused by flash flood, with Q_{max} of 60 $m^3 s^{-1}$. Severe localised bank erosion. Measured volume of flood deposits exceeded 1800 tonnes approx. equal to 14 years normal average annual non-dissolved sediment output.	Acreman (1983)

drastically altered during the systems change caused by this extreme climatic input, with intrinsic sinuosity thresholds exceeded.

According to Thorne and Lewin (1979), extreme events have associated shear stresses which appear out of balance with previous channel geometry. Process thresholds are crossed and phenomena such as rapid bank recession, abandonment of lengths of cut bank, cutoffs, areas of sedimentation marginal to the evolving channel and flood chutes may all occur. For example, loop abandonment on the River Severn at Maesmawr (Thorne and Lewin, 1979) has occurred through chute cutoffs, probably related to high magnitude floods. A rapid rise in the flood hydrograph is also known to initiate processes such as bank sapping. Therefore, bankfull discharges alone do not produce the full range of fluvial forms found in the field. However, we have to ask if it is possible to regard extreme events as part of the present process system, or to treat channel changes and the slow or rapid recovery as essentially allogenic and in terms of relaxation paths.

However, it is the persistence of features caused by such events that must be stressed, rather than their immediate landforming capacity at the time of the event. Floods may upset channel equilibrium but it is the rate of resultant form adjustment to a stable state that is important. This seems to be highly dependent upon the climatic environment, with flood events of similar magnitude possibly associated with varying rates of recovery. Both Gupta and Fox (1974) and Costa (1974) found that recovery of channels from flood induced disruption can be fast in comparison to the recurrence interval of the flood in hydrometeorological terms. For example, although Costa's Piedmont flood had a recurrence interval in excess of 100 years, after 5 months, channel width had nearly reached pre-flood proportions. The ability of the post-flood range of flows to reverse or modify flood effects is therefore crucial. If the flood deposits are coarse, the lag effect of high magnitude events will persist. In extreme contrast, Dury (1973) when studying the magnitude/ frequency relationships on the River Nene in the Fens, found that the estimated 1000 year event was associated with no widespread erosion or deposition. Even a catastrophic event of this magnitude was unable to leave its imprint on the landscape.

There are however a series of examples of catastrophic landforming flood events that have made a persisting impact within a variety of different climatic environments (Table 2.7.(i)). Stewart and LaMarche (1967), studying Coffee Creek in Northern California, concluded that floods as large as that of December, 1964 were infrequent because this destroyed floodplain forests uprooting many trees 200-300 years old and causing deep erosion of previously undisturbed deposits as old as 1700 years. Again, in 1954, Hurricane Alice generated catastrophic flooding (return period > 2000 years) along the Pecos River in S. Central Texas and at one particular site in an abandoned channel, a surface was buried that was 1300 years old (Patton and Baker, 1977). To give a British example, the Lynmouth flood in Exmoor in 1952 (studied by Dobbie and Wolf, 1953), gives a good instance of the disparity between rainfall and peak flow recurrence intervals (estimated 150 and 1000 years respectively) (Dobbie and Wolf, 1953 and FSR, NERC 1975). This resulted in channel deepening and the deposition of flood gravels that will survive the mean recurrence interval of the flood event (Anderson and Calver, 1977).

The spectrum of different discharges to which most natural channels are subject, suggests a more complex situation with channel morphology relating to a range of flows rather than a single discharge. Research by Neller (1980) on the Macquarie Rivulet, N.S. Wales, indicates that at least two sets of flow are responsible for major channel forms. Channels were found to be modified by flows with a 1-3 year recurrence interval, but over the last 150 years meander pattern has changed during infrequent high magnitude flows and the size of bends is adjusted to events with a recurrence interval of 25 years. Intermediate flow levels between these two sets probably prepare and reinforce such planform adjustments. Another example is provided by Rust (1972), who found on the proglacial River Donjek that the longitudinal gravel bars, the dominant bedform, only migrate during flood events due to the high proportion of coarse bed material and lower slopes. The actual recurrence interval of these floods was not known. At lower flows, sand accumulated in wedge shaped units lateral to the bars and thus again, at least two sets of flows are important due to the heterogeneous nature of the bed material. Knox (1975) also suggested that intrinsic threshold values are more likely to be exceeded after a sequence of floods as opposed to one event of very high recurrence interval. However, Hitchcock (1977) found on Langden Brook that although larger flood events are responsible for the major channel changes, for any given flood, the relative amount of change at different locations along the channel cannot easily be predicted from past flood events. This is a very important problem, especially in relation to river management.

"The fact is that after a century of relatively intense effort to understand the ways of rivers, we remain unable to predict accurate channel response to a simple change in flow regime at any timescale" (Hicken, 1983 p63)

In particular in the Scottish context, Khayat (1978) in his investigation of bankfull discharge both from records and local knowledge, showed that no single value of frequency was appropriate for all the Scottish rivers studied. One has also to be careful about the qualitative terminology used to describe the magnitude of event. Harvey et al. (1979), in NW England, describe that major events control the overall morphology and moderate events cause adjustment within these overall forms. However, their "major" event occurs 2-4 times per year and "moderate", 14-30 times per year. The terms "major" and "moderate" obviously depend on what timescale the study is concerned with. In certain environments, magnitude and frequency curves may simply shift up or down in magnitude (Kellerhals, 1972).

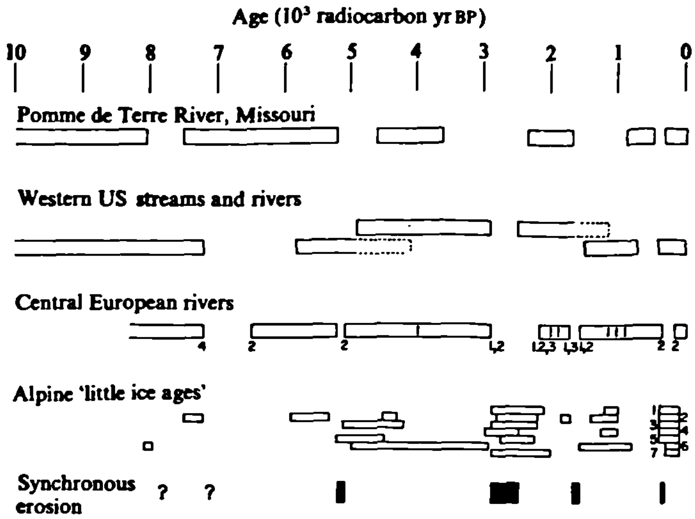
Incorporating the temporal dimension, the importance of the interspacing of major events, in relating channel form to magnitude and frequency of discharge events, has also been noted (Wolman and Gerson, 1978; Anderson and Calver, 1977). Werritty and Ferguson (1980) studied the 17 year post-flood recovery of the River Feshie from a fragmented and unstable braiding pattern to a meandering planform. Pattern was found to reflect in a cyclical manner the processes of flooding disruption and the progressive regaining of order by events at lower recurrence intervals, as the stream attains quasi-equilibrium (cf.

Starkel's (1983) cyclical patterns over a longer 1,500 time-period). It would seem from these examples that the effectiveness of the processes that control channel pattern, depends on the distribution or inter-arrival times of the events as well as their magnitude. It also poses questions as to what extent process rates are a direct response to immediate processes, accepting the possible transitory nature of the river system state.

Processes controlled by cyclic climatic phenomena may show cyclic or regularly periodic fluctuations. However what happens if this periodicity itself changes, causing variations in river and sediment discharges (Starkel, 1983)? Change in the frequency of landscape forming events is known to have implications in process terms. As meander length has been empirically related to the square root of dominant discharge, meander geometry is an indicator of the extent and occurrence of climatic change. For example, the meandering Mississippi shows considerable fluctuations in length through time and is a good example where present processes evident from present channel pattern cannot be extrapolated back to the past. Long term climatic change may produce extensive differences in mean pattern. Thus, the underfit meanders studied by Dury (1977) are typically 1/5 to 1/10 of the size of their meltwater eroded predecessors. Dury's palaeodischarges, based solely on the relationship between meander morphology and mean annual flood, must however be treated with caution.

The effect of climatic change is only clearly discernible when a trend in the palaeochannel width, or any other descriptive parameter, can be detected. In an extreme case, channel metamorphosis may take place, as documented by Kozarski (1983) in his study of the Warta valley in the vicinity of Poznan. Two major changes in planform (from braiding to large meanders to small meanders) took place over 7000 years due to system responses to climatic fluctuation and associated bioclimatic changes. Another example is provided by McDowell (1983) in her paper describing stream response to Holocene climatic change in Brush Creek, a small Wisconsin watershed. It is suggested that during the Holocene, climatic changes have triggered several episodes of channel adjustment, including erosion at a mid-Holocene precipitation minimum and floodplain construction paralleling a rapid increase in precipitation. However, the relationship between the direction of climatic change and the type of system response is highly complicated. Brakenbridge (1980) however suggested widespread fluvial erosion was synchronous with the timing of the beginning of Northern Hemisphere "Little Ice Age" phases, within Alpine catchments (Figure 2.7.(ii)). This was associated with increases in meriodional circulation and consequently increases in flooding of 1-2 year RI. Grove (1972) also found increased incidence of landslides and floods during the Little Ice Age in Western Norway.

Figure 2.7.(ii)



Holocene alluvial and glacial chronologies compared

(Source: Brakenbridge, 1980 p656 Figure 2)

Central European Rivers
1: Becker and Frenzel (1977)
2: Becker and Schirmer (1977)
3: Schirmer (1977)
4: Becker (1977)

Alpine "Little Ice Ages"
1: Alaska (Denton and Karlen, 1977)
2: Swedish Lapland (Denton and Karlen, 1977)
3: British Columbia and Western Washington (Duford and Osborn, 1977)
4: Norway (Griffey and Worsley, 1978)
5: Glaciations common to the eastern and western Alps (Patzelt, 1974)
6: Colorado Rocky Mountains (Benedict, 1973)
7: Mexican volcanoes (Heine, 1978)

2.8 Factors controlling boundary shear stress

In terms of factors controlling boundary shear stress, the channel slope is crucial as it represents a balance between the transport capacity of the stream and the size and quantity of the sediment load supplied (Simons and Senturk, 1977). Channel slope is inversely proportional to water discharge and directly proportional to both sediment discharge and median grain size (Schumm, 1971). Its significance however varies with the designated timespan of study (Schumm and Lichty, 1965). Therefore, valley slope is a dependent variable in cyclic time but becomes an independent variable in graded and steady time. In terms of channel morphology, slope is an indeterminate variable in cyclic time, dependent in graded time and independent in steady time. Thus while the long profile of a valley may exert a constant control, the slope of individual reaches may be modified within a steady timespan, through either aggradation or degradation.

In terms of basic hydraulic theory, slope is an important determinant of channel form since it not only directly controls the shear stress on the bed (Equation 2.2) but also determines the average velocity (Equation 2.3).

$$\tau = \gamma R S \qquad (2.2)$$

$$V = \frac{R^{2/3}\ S^{1/2}}{n} \qquad (2.3)$$

These when combined as a product give the specific stream power exerted at a given location within the river (Equation 2.4).

$$\omega = \tau V \qquad (2.4)$$

Various experiments have shown the importance of slope in differentiating between channel patterns. Leopold and Wolman (1957) found that the following relationship between slope and bankfull discharge, separated braided from meandering channels (Equation 2.5).

$$S = 0.013\ Q_b^{-0.44} \qquad (2.5)$$

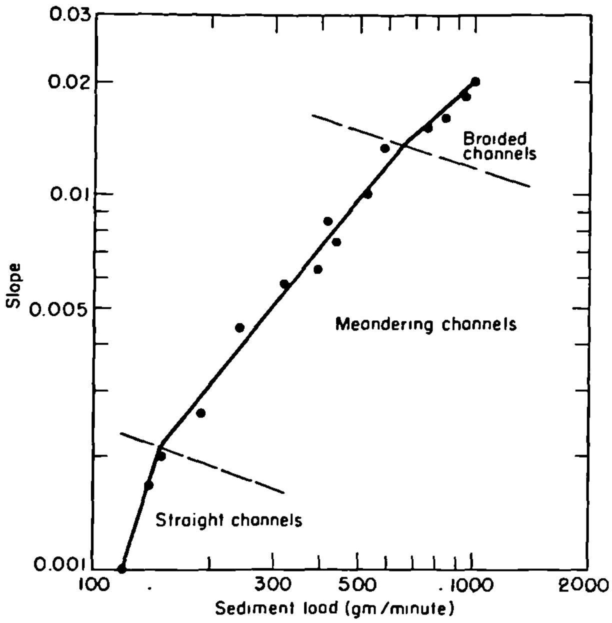
Braided channels were found to occur on higher slopes for a given discharge and also at a higher discharge for a given slope, than for corresponding meandering channels. It was difficult to find a clear relationship between slope and bankfull discharge for straight channels, but a large proportion lay above this line. However, Ackers and

Charlton (1970) identified another threshold between straight and meandering thalweg channels at very low slopes. This was further analysed by Schumm and Khan (1972), who linked slope with sediment discharge in a flume experiment. An increased rate of sediment transport was found to occur at the process thresholds where channel pattern changed (Figure 2.8.(i)). Braiding is expected to occur where there is a high stream power, associated with steeper slopes. The process of braiding dissipates excess stream power by dividing the flow over a number of channels (Henderson, 1966).

2.9 Factors controlling boundary resistance

One important resisting force within the channel is sediment size and its availability both in bed and banks ie. the character of the boundary material. Braids appear to develop in coarser deposits than those supporting meanders. However, the absolute size of the available sediment may be less important than the manner in which it moves through the system, since channel pattern stability depends on the balance or lack of balance between sediment load and its transportability. For example, Ackers (1964) found in flume studies that although meander wavelength is strongly dependent on the dominant discharge, there is a strong secondary dependence on grain size and mode of sediment transport. The actual pattern of sediment transport may also be very episodic and spatially discontinuous, with large quantities of material stored along the channel. In the extreme case, "sedimentation zones" may occur along a river (as observed by Church, 1983 on the Bella Coola River, British Columbia). In these areas, sediment entrained as bedload

Figure 2.8.(i)



Relation between sediment load and flume slope showing increased rate of sediment transport at thresholds of pattern change

(After Schumm and Khan, 1973)

(Source: Schumm, 1977 p130 Figure 5.15)

from river banks and tributary sources is stored and these zones are characterised by lateral instability and locally reduced gradient. These sediment stores are interspaced by more stable reaches.

Sediment available for transport may be stored in the floodplain and periodically, or continually, be eroded from the channel banks. According to Burkham (1972), sediment may be accreted in alluvial floodplains in five general ways:

(1) by the development of islands in the stream channel and their subsequent attachment to one bank by channel abandonment
(2) by direct deposit on the floodplain
(3) by deposit in the stream channel along the banks
(4) by the formation of natural levees
(5) by the deposition on alluvial fans at the mouths of tributary streams.

The quantities of sediment readily available to rivers in Britain may vary considerably from negligible, in the case of bed-controlled reaches, to substantial, in the case of the major sediment sources in the active gullies eg. Grains Gill (Harvey, 1977).

This sediment availability for transport is very important. Decline in upstream sediment reserves can have major implications on downstream channel stability. Church (1983) found that a decline in the transport of unusual sediment volumes, associated with the erosion of Neoglacial moraines of Alpine glaciers (18th-19th C), caused a stabilising tendency within the wandering channel planform to be propagated downstream. Sediment availability has also been noted to be an important control in the geomorphic response to major flood events within Scotland. The vast flushing out of fluvioglacial sediment and associated planform change during a flash flood on a mountain torrent within the Cairngorms (McEwen, 1981), can be compared with flood impact in areas of more restricted sediment supply. For example, Werritty and Metcalfe (1985) studied geomorphic response to an extreme flash flood on the Caldwell Burn within the Southern Uplands. Scouring of the floodplain occurred and more restricted modes of planform adjustment.

Thus, the rate at which channel migration takes place depends not only on the forces exerted on the channel banks, but also on bank resistance. For example, Daniels (1971) suggests that the amount of increase in meander wavelength is proportional to the discharge pulse and inversely proportional to the % silt/clay in the boundary material.

There have been several studies of the mechanisms by which sediment from channel-side banks is incorporated into the channel sediment system. These demonstrate that it is necessary to study both the modes of erosion and the associated causes (Table 2.9.(i)). Such processes fall into two main groups, namely fluvial entrainment and subaqueous weakening and weathering (Thorne, 1982). Thorne and Lewin (1979) made observations on bank processes and bed material movement at a meander bend on the river Severn. In general, mechanisms of bank failure were dependent on bank size, geometry, structure and composition and were dominated by fluvial undercutting and mechanical failure of the cantilevers in the upper bank. Tracer experiments showed that bank

Table 2.9.(i)

Documented modes of bank erosion within the British Isles

River	Mode	Source
British rivers (gravel bed)	(i) abrasion and scour ie. removal of soil from the surface of the bank. (ii) bank collapse ie. collapse or sliding of part of the bank.	Charlton (1982)
Afon Crewi, Penegoes	slumping involving combination of gravity fall and hydraulic entrainment.	Blacknell (1981)
Severn	(i) cantilever failures (ii) fluvial undercutting	Thorne and Lewin (1979)
Cound, Schropshire	4 possible mechanisms: (i) vertical retreat (ii) basal scouring and undermining (iii) collapse of overhanging blocks (iv) rotational slipping	Hughes (1977)

Table 2.9.(i) cont.

River	Mode	Source
N. Ireland	(i) slipping (ii) lateral corrasion (iii) hydraulic action on lower stream banks	Hill (1973)

retreat rates were fluvially controlled, though the actual failure mechanisms were not. Thus processes of fluvial undercutting, flows of a magnitude to carry this out and cantilever failure were responsible for planform development in the short term. In contrast, Blacknell (1981), when monitoring an upland catchment in Wales, found that bank erosion was apparently greatest when freeze-thaw processes operated on a thoroughly moistened river bank and Hill (1973) found that frost action was very important on till-banked rivers in N. Ireland (cf. Wolman and Leopold, 1957 in the east coast of the USA). Slumping involving a combination of gravity fall and hydraulic entrainment was found to be the most important process in this environment. Thus particle movement from the bank toe only occurs if river flows permit.

In composite banks, modes of failure are more complex and Thornes (1982) identified three main types, namely rotational slip failure, plane slip failure and cantilever failure. Hydrometeorological conditions are also very important and incorporating more information into her model, Hooke (1979) suggests that river bank erosion is a function of rainfall characteristics, discharge rise, antecedent river conditions, soil moisture conditions and temperature. In her studies of river banks in Devon, she identified two main methods of bank erosion, namely corrasion and slumping. These appeared to be associated with the influence of river flow levels and especially antecedent precipitation conditions. Thus, the duration of the storm may influence the degree of wetting of the bank material. Twidale (1964) again stressed the importance of bank moisture conditions and demonstrated the significance of "falling stage" phenomena, along with associated rapid changes in perimeter pressures. The efflux of ground water was also found to be important in bank slumping. Finally, it should be emphasised that there can be considerable spatial variation in both types of process and process rates, on neighbouring reaches of the same river (eg. Hill, 1973).

A second major factor affecting boundary resistance is channel confinement. The literature on the influence of channel confinement and planform change is very limited. Nevertheless, floodplain width can exert an important constraint on room to migrate and also on availability of sediment. For example, meander planform may have very different controls, depending on whether it is "free" to migrate or whether it is restrained by rock walls (Lane, 1957) or else confined by glacial drift (Knighton, 1977). As Lewin and Brindle (1977) note, contemporary channels frequently pass to and fro across alluvial valley floors and impinge with some frequency against valley sides of very restricted erodibility, so that major distortions of the meander pattern can result. Degree of confinement can vary from minor bank irregularities to distortion of meander loops and Lewin and Brindle (1977) make the division into three categories, dependent on this degree of confinement. These ranged from first degree confinement where the channel regularly comes into contact with confining agents, such as valley sides, to third degree confinement where meander geometry is unable to develop due to large-scale natural or artificial restrictions.

The third factor affecting boundary resistance is the vegetative cover of floodplain banks and bars. The importance of protective riparian vegetation as a control has been documented, in terms of its enhancement of bank stability. Alteration from a braided to a non-braided character is sometimes associated with a change from dense vegetation along the banks to sparse or no vegetation. Width of the river may also be subject to constant readjustment if the banks are not well stabilised by vegetation.

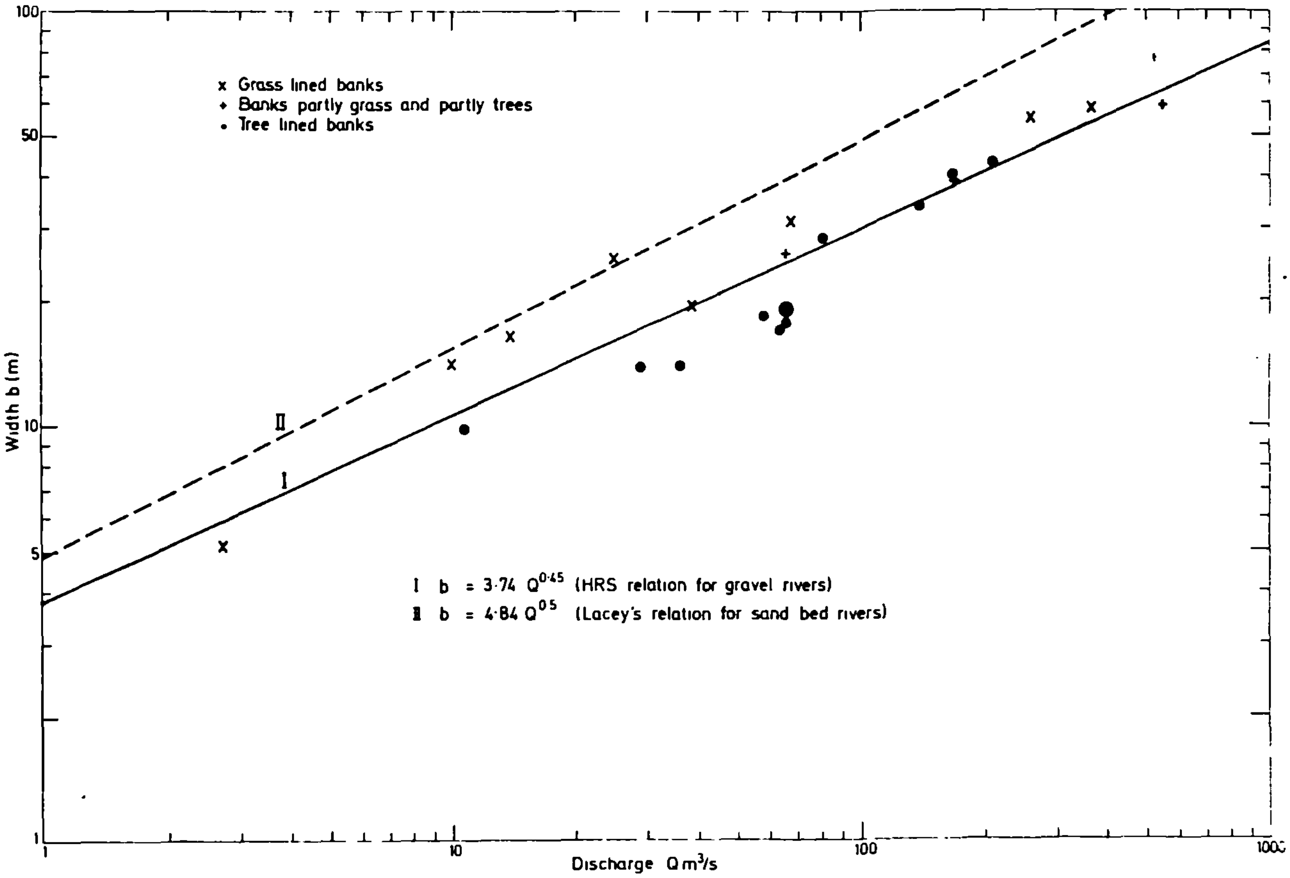
Charlton et al. (1978), in their study of the hydraulic geometry of some gravel bed rivers in Britain, found a significant relationship between channel width, bankfull discharge and type of vegetation on the banks (Figure 2.9.(i)). Light vegetation (generally grass) allowed width to be 30% greater than the average width of the 23 rivers studied. Heavy vegetation (shrubs and trees) resulted in a width about 30% smaller. However, the important factor was whether vegetation existed during the formative period (when it did have an effect) or whether it was introduced later, when it may have no affect.

Burkham (1972), in contrast, found that trees on the floodplain had a mixed influence on channel widening on the Gila River, Arizona. Although the trees restricted the flow of water onto the floodplain during the major floods and concentrated the flows within the stream channel, there were other features that could increase erosion. The concentrated flow increased stresses along the stream channel banks and thus may have caused erosion, but this had only a minor effect on channel widening. However, during major floods, floating debris hung on

Figure 2.9.(i)

Relation between width and discharge at bankfull flow

(Source: Charlton et al., 1978 Figure 35)



the fairly rigid cotton trees and the forces applied to the trees caused them to be uprooted, carrying large chunks of alluvium with them and leaving the easily erodible material exposed. Klimek (1974) found that vegetation cover of trees and shrubs was important in controlling erosion on the Wisloka banks in the Polish Carpathians. The root system bound the loose alluvial deposits and this prevented them from liquifying or slumping. However, when the water level fell below root level due to increased vertical erosion, the bank stability decreased causing bank slumping and collapse. Blacknell (1981) suggested however that although bank vegetation may influence rates of bank erosion, increased rates of erosion probably reflect the properties of hydrodynamic flow around the bend rather than any changes in bank properties.

Vegetation may also be important in the restabilisation of the flood plain after a major flood event (Rust, 1972) and can control the the rate of depositional processes from suspension onto bars and floodplain (Froehlich et al, 1972). The role of vegetation influencing planform change is thus complex and highly varied. It depends in detail on the type and nature of the vegetation both in the banks and the adjacent floodplain. The stability of the underlying bank material (with or without vegetative cover) is very important and no simple generalisation is possible.

2.10 Anthropogenic controls affecting channel morphology

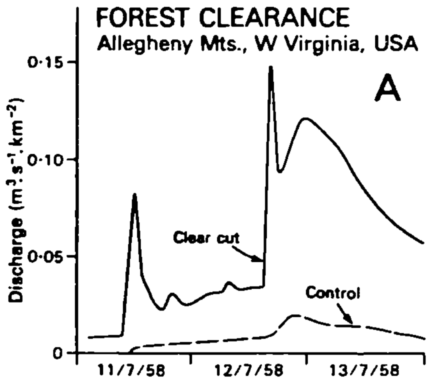
Anthropogenic controls can affect the fluvial system both at the basin level and through direct channel manipulation, and this can affect boundary shear stresses or resistance, depending on extent and management. The impact of changes in sediment input, through altered erodibility of soils, and changes in rates of runoff have been studied both in controlled experiments and retrodictive studies. Catchment controls include deforestation and afforestation (the presence or absence of associated ditching, the affect on peat hydrology), farming methods (eg. in relation to snow melt), burning of moorland or urbanisation. The possible impact of such change on the runoff regime and on sediment yields must be assessed. Changes which affect the actual channel include the construction of dams, weirs and bridges and the impact of flood protection schemes; all at a variety of scales. The rapidity with which man-induced change can effect the system must also be stressed.

2.10.1 Implications of deforestation/ afforestation

There is an extensive literature on both the hydrological and sediment yield implications of such drastic land-use changes as afforestation and deforestation. Specific controls related to forestry practices include species type and the nature of the harvesting procedures (eg. Hoover, 1944). The key issues in terms of this study are the effects on the speed of runoff, the shape of the unit hydrograph and sediment supply. At the catchment level, vegetation can influence infiltration rates and thereby surface runoff, due to its influence on rates of interception, degree of surface compaction and root storage. Removal of a protective covering of vegetation may also make highly erodible material more available and the impact of logging and road building in such catchments can also increase rates of sedimentation (Bethlahmy, (1974). Unfortunately, many studies have been carried out in different climatic regions and under more extreme geomorphic conditions (Rothacher, 1970; Hewlett, 1967) and it is difficult to extrapolate results to upland Britain. Documented effects on flood peaks therefore vary considerably (Table 2.10.1.(i)).

Forests have long been considered to have a substantial effect in flood peak reduction (Rutter, 1958; Pereira, 1973), although there is debate over the scale of this effect (Hewlett and Helvey, 1970). With deforestation on the Coweeta watersheds, quick flow was found to increase 10% but time to peak showed no significant difference. Gregory and Walling (1981) cite the example of Harrold (1971) who compared the responses of clearcut and undisturbed catchments to a heavy summer storm in the Allegheny Mts, West Virginia (Figure 2.10.1.(i)). Flood magnitude was found to increase about 6 times. However, the forest influence becomes less important as rainfall increases and as the major mechanisms of interception and greater diversion to soil moisture storage become relatively less effective, when their capacities are exceeded by the volume of storm rainfall.

Figure 2.10.1.(i)



Man's impact on catchment response

(After Harrold, 1971)

(Source: Gregory and Walling, 1981 p67 Figure 5.5A)

Table 2.10.1.(i)

Additional documented changes in the flood hydrograph associated with rural landuse change

Location	Effect	Source
West Virginia	Response on clear cut catchment to heavy summer storm increased in magnitude 6 times.	Harrold (1971)
Kenya	After 76.2 mm of rain in 30 mins: In 13 ha clearings, peak flow was 6.5 $m^3 s^{-1}$. Peak rate from forest was less than 0.17 $m^3 s^{-1}$.	Pereira (1976)
Upper Severn	Suggested that afforestation and drainage aggravated problem of flooding, causing increased amount and speed of runoff after major storm events.	Howe, Slaymaker and Harding (1967)

Several studies have been carried out in Britain, notably at the Plynlimon research catchments in mid Wales. There the rapidity and magnitude of runoff response to a unit depth of precipitation was compared between a catchment with a cover entirely of upland pasture and another with coniferous forest cover (Sitka spruce/ Norway spruce) by Newson (1976). During very long prolonged rainstorms, the interception effect was found to be minimal. Deforestation of the floodplain may also lessen floodplain stability (Brice, 1973), causing irregularities in planform.

Change in vegetation also affects slope stability with implications in terms of sediment availability and sediment yield (Wolman, 1967). It has been shown that periods of deforestation and the development of agriculture in the United States have been paralleled by increased runoff and increased erosion of slopes and sedimentation in the river valleys (Knox, 1977; Sopper and Lull, 1967). In the humid environments of New Zealand, destruction of natural forest vegetation on the steep slopes and the subsequent reseeding with grasses, have precipitated large-scale erosion problems. Stable narrow sinuous rivers have changed to wide straight channels due to the coarse sediment input, and both aggradation and an increase in flood peaks were observed (Grant, 1950). Disruption of the soil surface can be more important than the actual tree-removal (Hewlett and Helvey,1970, Reinhart, 1964). Increases in sediment mobility are also confirmed within British studies. For example, the wooded Tanllwyth sub-catchment at Plynlimon produced 4 times more bedload than the pastureland of the Cyff basin (Clarke and McCulloch, 1979).

Other forms of landuse change may also have a impact. Klimek and Trafas (1972), in a basin study of the Polish Carpathians, suggested that a change from corn to potato growing in the 19th century led to soil erosion, with sediment yield increased 100 times. This caused subsequent bar development and change of channel pattern from a meandering to braided planform. Similarly, Mycielska Dowgiallo (1977) found that over the past 200 years, there had been a change from a meandering to braiding planform in the Vistula valley. Although increased sediment supply due to soil erosion and deforestation were the main causes, the destruction of mill-waters and small retention pools developed during industrialisation in the 18th century was also important (Dembinska, 1972).

2.10.2 Land drainage

Land drainage may be instigated to lower the water table or to remove surface water (Walling, 1981). Thus, the implication of land drainage on the varying magnitude and frequency of discharge events must also be considered. However, the impact is highly variable depending on a variety of factors:

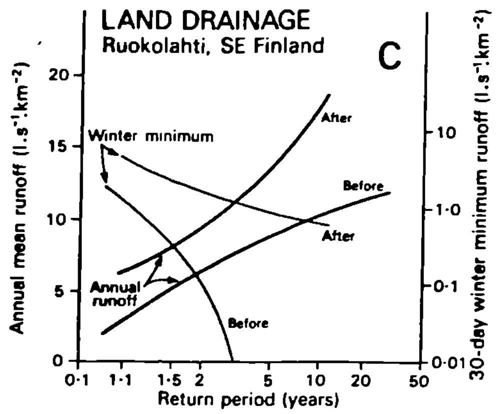
> "The hydrological effects of field under drainage depends upon the type of under drainage adopted and soil characteristics, particularly interflow properties. If satisfactory outflow is provided for the drains, there will be a lowering of the average level of the water table in the

soil." (Green, 1979a p10)

Very important are the antecedent conditions of the soil before a rainfall event. Green (1979a) suggests that as the magnitude and frequency of surface runoff is reduced, the number of major discharge events would also be reduced. As the volume of subsurface runoff is increased, the heights of the peaks are flattened in comparison with those created by surface runoff. However as the amount of under drainage in a small catchment is increased, the heights of the peaks tend again to rise slightly but the recession is more rapid than before. Thus, the effect would appear to be ambiguous; it seems that either an increase or reduction in flood peaks can take place depending on other factors such as rainfall intensity, soil moisture storage and catchment size. More work is required to understand the relative contribution of field under drainage and rainfall over varying thresholds. It seems certain that the possible alteration is large. Work carried out in Finland (Mustonen and Seuna, 1971) on the effects of peat drainage on forestry found that drainage caused lowering of the water table, reduced evapotranspiration and an increase in annual mean runoff of over 40% (Gregory and Walling, 1981; see Figure 2.10.2.(i)).

Drainage associated with the Forestry Commission double throw ploughs can cause rapid runoff of rain falling into the furrow (Binns, 1979). Because up to 20% of the land surface may consist of ditches, 20% of any storm will fall directly into these ditches and be evacuated from the site within a very short time. The Institute of Hydrology has studied the hydrological impact of ploughing at Coalburn, Northumberland (Robinson, 1980), the main effect being registered in the period

Figure 2.10.2.(i)



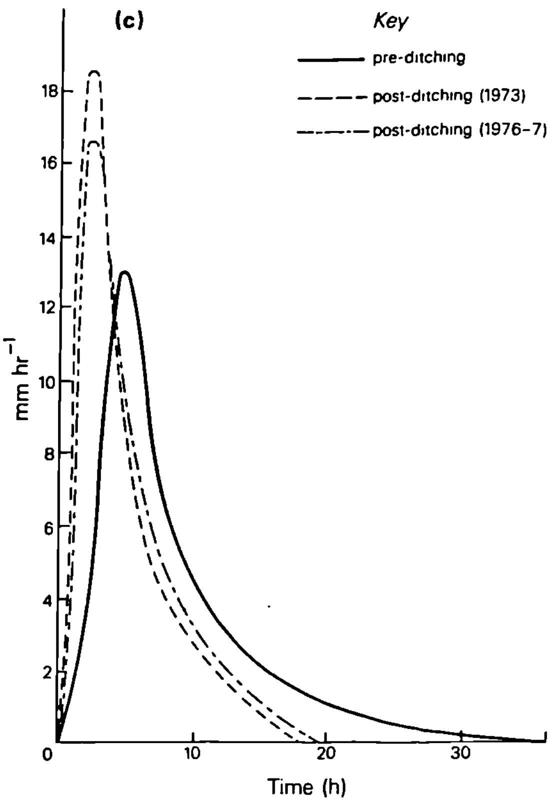
The impact of land drainage on catchment response

(After Mustonen and Seuna, 1971)

(Source: Gregory and Walling, 1981 p67 Figure 5.5C) immediately after ploughing. A unit hydrograph study demonstrated that peak discharges increased by 40% and the time to peak was halved. Figure 2.10.2.(ii) shows the pre- and post-ditching hydrographs for a forestry plantation at Coal Burn, Northern Pennines. Newson (1981b) and Howe, Slaymaker and Harding (1967), within the Severn catchment, suggest that forestry and agricultural drainage have been an important factor in changing the patterns of peak discharges, though the triggering mechanism was an increased incidence of intense storm events. Lewin (1981) however is sceptical about the supposed increase in flood frequency claimed due to forest ditching in the Welsh Borders. He suggests that the period of greatest Forestry Commission activity has been over the last 20 years during which time the trend of large floods has in fact decreased. The time-base of such a study is therefore very important.

An important factor seems to be the percentage of the catchment ditched and thus an increase in the effective drainage density. Acreman (1985) studied changes in the flood hydrograph due to afforestation on the upper Ettrick, during two main periods of ploughing. The initial ploughing of the lower basin caused increased time to peak and decreased peak flow, whereas upstream this was reversed. As Werritty and Acreman (1985) point out, the shapes of the hydrographs reflected the differences in arrival times of runoff from different sub-areas of the catchment. However, it is difficult to extrapolate these results to differing percentages of landuse change; position within the catchment is clearly crucial.

Figure 2.10.2.(ii)



Pre- and post-ditching hydrographs for a forestry plantation at Coalburn, Northern Pennines

(Source: Newson, 1982 p75 Figure 3.10)

The impact of ditching on sediment mobility is also important, with open drainage ditch catchments giving higher sediment yields than unditched catchments (Newson, 1980b). Early studies within the Severn and Wye valleys found that sediment concentrations in storm runoff after ditching were two orders of magnitude greater than before ditching (Painter et al., 1974). As trees increase in leaf-surface area, more water is evaporated, the ridges become degraded and gradually a less flashy hydrograph returns. The degree to which this recovery takes place requires more study. Subsequent study within the Tanllwyth catchment of the upper Severn indicated that afforestation ditching increased bedload yields from under 10 t km^{-2} yr^{-1} to over 30 t km^{-2} yr^{-1}. Bedload was increased more than five times over grassland yields (Calder and Newson, 1979; Newson, 1981b). The material within which the ditches are cut is also very important. Unconsolidated colluvium underlying peat was found to be a major sediment source in upland Wales (Newson, 1980b).

Other landuse changes can also have an important impact. For example, Edwards and Rowntree (1980) found from a study of lake sedimentation in Braeroddach Loch, Aberdeenshire that there is evidence for increased sedimentation over the past 400 years, related to modern agricultural practices, especially the excavation of drainage ditches.

2.10.3 Flow regulation

Flow regulation, though of lesser importance within the studied areas, must be briefly discussed. Alteration of the natural balance of water storage within a catchment, through damming for reservoirs or for power, may cause modification of the frequency distribution of downstream discharges. It may also act as a sediment trap for sediment moving downstream, causing depletion effects. Obviously, the surface area of the reservoir as a percentage of the catchment area is highly important; even more so may be the percentage of this area that is upstream of the reservoir.

In the study catchments in this thesis, none are significantly affected by the impoundment of large surface areas of water and so discussion here is restricted. However, smaller scale damming has taken place, both present day and historically. Within the context of peak flows, such storage and absorption may dampen the hydrograph downstream. The resulting effects on the frequency of bankfull discharges are important in predicting any associated change in channel form. Moore (1969) found that if the reservoir is not full then inflow may be contained within the reservoir so that peak discharges downstream may be reduced by 98%. Even if the reservoir is near capacity level, the routing of a flood peak through it, can lead to a peak discharge downstream reduced by more than 50%.

The extent to which flow reductions from regulating affects of upstream dams can be related to changes in channel width was studied by Williams (1977), in terms of the shrinking channels of the river Platte in Nebraska. Decreases in channel width were related to discharge decreases, principally due to the effects of major upstream damming. In Britain, Petts (1977) has studied channel response to flow regulation on the River Derwent in Derbyshire, and found that while the magnitude of more frequent events will be markedly reduced after damming, the regulation of the rarer events is less effective.

In terms of sedimentation rates, depleted sediment load may lead to scour depending on the hydraulic conditions beyond the spill way. Also if sediment sources are unequally distributed around the catchment, the trapping of upland sediment sources, which are working their way through the system, may lead to considerably reduced sediment supply downstream.

2.10.4 Channel modifications

As well as catchment alteration, local and upstream modifications to the channel through channelisation and river training may be important in attempting to force a new artificial equilibrium on the channel. Channelisation has been used to facilitate navigation, improve stream alignment, control bank erosion (Keller, 1975, 1976; Hicken, 1977), improve flood control due to increased channel capacity and increase general channel stability. Other localised modifications may have less dramatic effects on water and sediment discharge, with consequences which depend on the way in which abstraction of water and sediment relates to peak flow rates. Directly induced change thus occurs where a reach has been artificially modified to improve its efficiency.

Modification may be carried out in several ways, namely widening, deepening, straightening (eg. through artificial meander cutoffs, Laczay, 1977), clearing and snagging, diking and bank stabilisation (Keller, 1975, 1976; Cooke and Doornkamp, 1974). However, the controls of the hydraulic geometry of the natural mobile boundary must be understood for success in channelisation (Charlton, 1982). The effect of creating a larger channel can lead to increased water velocity so that the frequency of peak discharges may be increased downstream of the modified reach, due to a reduction in channel capacity. If for example, meander loops are too accentuated, the resistance to flow becomes over large and thus flood levels are more likely to over flow the banks. According to Henderson (1966), in such river alteration it is normal to cutoff loops that are too deep, while the basic meandering planform is preserved.

Localised stabilisation frequently has undesirable effects outside the reaches directly involved. For example, the steepening caused by artificial meander cutoffs may gradually progress upstream causing increased flooding, aggradation downstream and incision upstream, with headwater gullying (Bray and Kellerhals, 1979; Emerson, 1971). These effects may occur long after the initial channelisation. Channelised reaches may undergo general readjustment with increased flooding downstream while deepening may also alter the baselevel of tributary channels. After artificial cutoffs, there will be increased meandering and bank erosion (Laczay, 1977) and this must be prevented by proper bank protection and training works. The aim of river training is thus to induce river erosion and deposition in only desired locations. The success of such engineering works is however debatable:

> "Regardless of the degree of channel stability, man's local activities may produce major changes in river characteristics both locally and throughout an entire reach. All too frequently, the net result of a river improvement is a greater departure from equilibrium than that which originally prevailed. Good engineering design must invariably seek to enhance the natural tendency of the stream toward posed conditions. To do so, an understanding of the direction and magnitude of change in channel characteristics, caused by the actions of man and nature is required." (Simons and Senturk, 1977 p40)

A Scottish example of effective channel planform diversion as a means of flood alleviation, with associated containment and stabilisation of unconsolidated channel side deposits, is provided by Young (1980) around Loch Moy, Inverness-shire. A variey of measures were adopted between 1805 and 1870; for example, meanders on tributary burns were artifically cutoff, during realignment and canalisation of the mainstream (Funlach Burn). Loch Moy was lowered 1.4 m to reclaim and drain neighbouring land and barriers were built to prevent streams readopting their former alignment during subsequent floods. This artificial impact must be compared with change over a longer graded timescale:

> "It is possible that in the course of one century [19th], man-made changes to the details of the local physical landscape were as great as those produced accumulatively by the naturally occurring processes of the whole post-glacial period." (Young, 1980 p166)

Other forms of channel alteration may take place due to industry, eg. the construction of mill lades, but there is little literature specifically on the geomorphic impact of these features (Beckinsale, 1972). However, Hooke and Kain (1982), when discussing the former importance of water abstraction for mill-lades in Devon, suggest it is possible that the decline in mill working and setting up of lades had increased discharge in the natural channels. Renewed meandering may also take place in formerly artificially straightened reaches (Hooke, 1977).

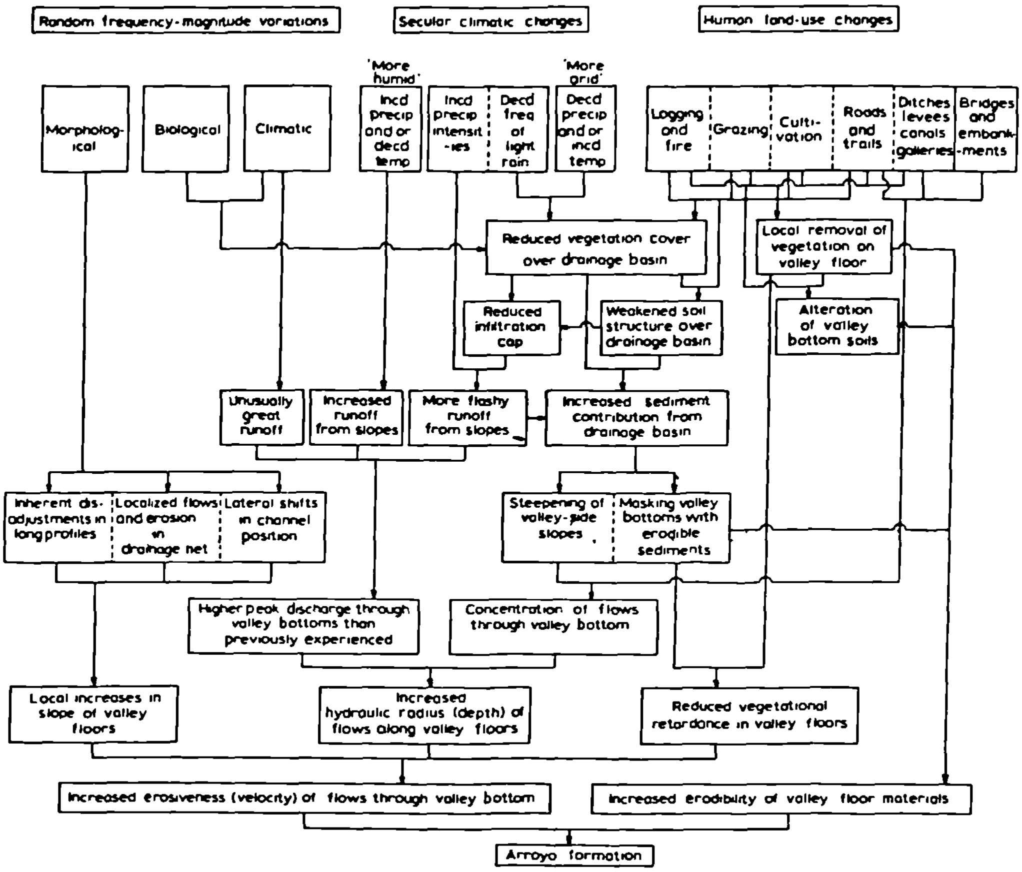
2.11 The landuse change versus climatic change debate

Having discussed the possible effects of these changes separately, there has been considerable debate as to their relative importance to the river system. Are the changes on controls from either landuse or climatic change sufficient to cause a change in channel pattern, or is the random magnitude/ frequency variation of extreme rainfall events more important? Unfortunately this has to be assessed through interpretation of the equifinality of channel form. The relative influences on controls are usually intermixed and it is difficult to differentiate between cause and effect.

Perhaps the most detailed case study of such a problem is that on the formation of arroyos, with many ephemeral channels in the SW United States undergoing rapid widening over the past 200 years (Cooke and Reeves, 1976; Graf, 1979). There are a large number of hypotheses concerning the possible effects of change in controls, associated with both secular climatic and landuse change (Figure 2.11.(i)). For example, Bryan (1928) and Antevs (1952) proposed that aridity led to incision while Leopold (1951) suggested that there was an increase in heavy rains around 1880 in New Mexico. In terms of the landuse changes, it has been proposed that overgrazing leading to reduced vegetative cover and the introduction of ditches, caused increased runoff. The many alternative hypotheses thus include a complex combination of climatic and landuse effects. The third major hypothesis is however the importance of random magnitude and frequency variations eg. the occurrence of unusual rainfall events (Thornthwaite et al., 1942)). It is clearly difficult to find evidence to support totally any hypothesis, as in most cases change will be due to an interaction of causative factors and it may be a question of identifying the dominant one.

Other studies addressing this problem have tended to be over much longer term adjustments than will be considered in this thesis eg. Knox (1972, 1977) in Wisconsin or Starkel (1983) and Falkowski (1972, 1975) in Poland. However, Knox found that the conversion of natural land cover to agricultural land in the 1830s caused a three to five fold

Figure 2.11.(i)



A model of arroyo formation

(Source: Cooke and Reeves, 1976 p16 Figure 1.2) increase in the magnitude of those floods that occur more frequently than once in five years in the tributary watersheds. This was accompanied by erosion and sedimentation, leading to major change in the physical characteristics of stream channels. Despite the identification of three episodes of variation in the intensity of the landuse impact, channel evolution was more complex, with only relatively small climatic variations occurring since the 1830s. Nevertheless, the conversion of the watershed from a natural to agricultural cover normally causes surface runoff and sediment yield to become more sensitive to these small climatic variations ie. landuse augments rather than causes change. In fact, Knox (1977) attempts to demonstrate that valley alluviation cycles occur when climate and vegetation are out of balance. Unfortunately, the problem of equifinality must be confronted in that the same form may be attributed to different processes.

Studies in Poland have also concentrated on the relative importance of climate and landuse on channel pattern controls within the Carpathian valleys, particularly over longer timespans (eg. 15,000 years, Starkel, 1981, 1983; Kozarski and Rothnicki, 1977) as well as more recent changes. Although precise dating varies in different valleys, a cycle of braiding-meandering-braiding has been identified in the Postglacial development of these river systems. The initial braiding period, associated with aggradation, occurred up to the pre-Boreal. However, with the onset of the Boreal until 4200 BP, a meandering planform developed with an erosional tendency, causing a progressive lowering of the valley floor. This was due to a reduction in discharge and sediment supply. From 4200 BP, with more continuous forest cover, a more stable meandering planform evolved, associated with floodplain stability (eg.

Falkowski, 1972, 1975 in the Wisloka valley). As climate indirectly affects channel pattern through type of natural vegetative cover, there will be a lag effect between climatic change and vegetation change, and thus cause and effect are again difficult to differentiate.

This meandering planform only dramatically altered during the 17th century to be replaced again by aggradation and a tendency to braid. However when this sequence of changes from 1700-1900 were studied, it was found that phases of intensive activity were followed by periods of relative stability (see Szumanski, 1972). Various changes had periodically disturbed the meander system, namely the introduction of potatoes, phases of higher flood frequency and artificial meander cutoffs. Cyclic tendencies may thus occur in channel pattern controls, with phases of alluviation or degradation as thresholds for change are exceeded. The shifts in these controls may be potentially attributed to climatic, bio-climatic or man-induced landuse changes. In reality, an interaction of such changes in controls may cause the greatest planform disruption.

CHAPTER 3

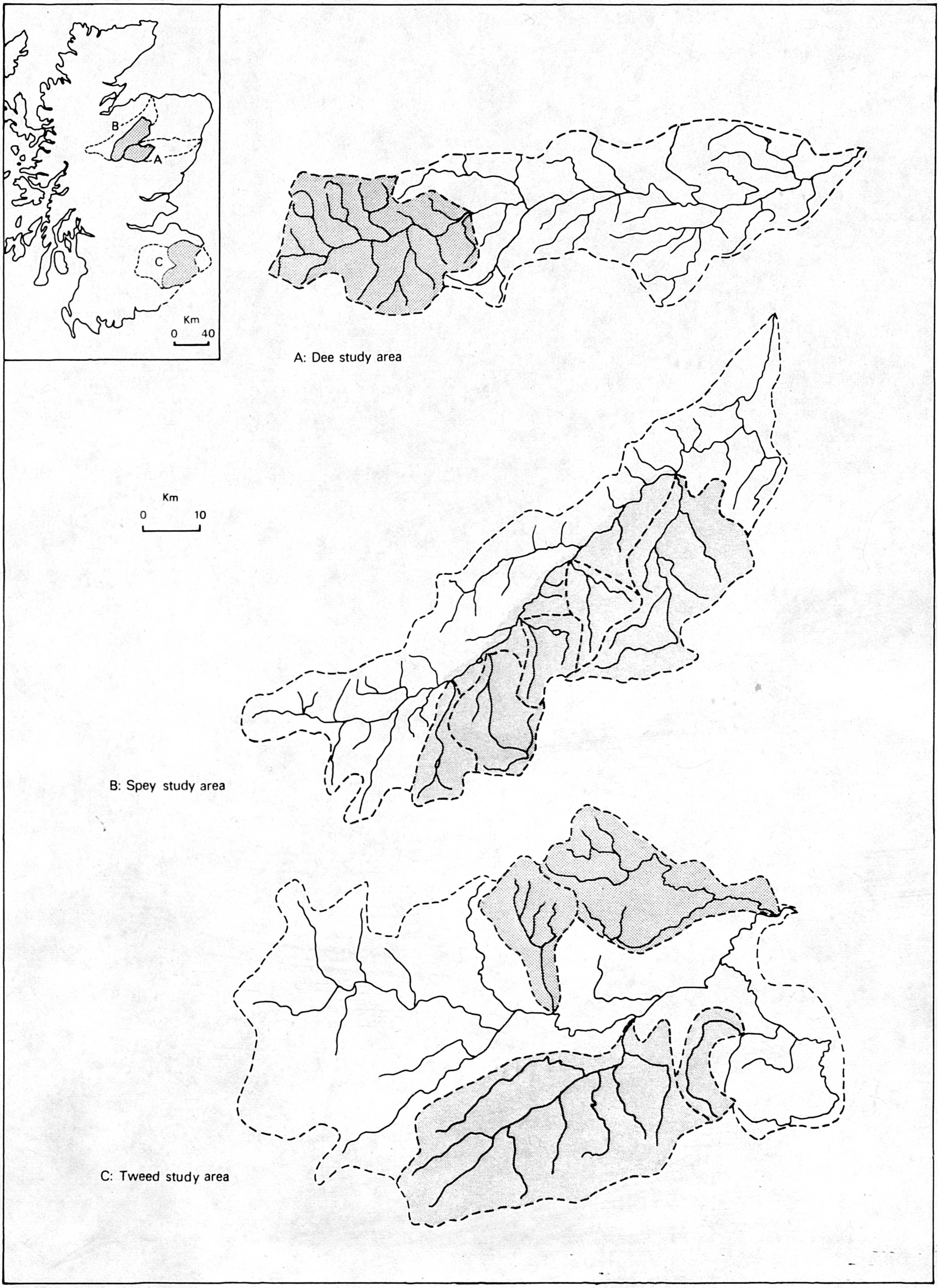
The physical environments of the three study areas.

3.1 Introduction

Prior to evaluating rates of channel change, the macro-scale controls on the fluvial system must be identified in terms of the environmental setting, at both local and basin levels. Three major study areas were chosen, principally due to contrasts in terms of their glacial legacy (sediment availability or accessibility), catchment size, valley slope, vegetative cover, rainfall and runoff hydrology but also, very importantly, availability of information. The locations of each of these areas, in relation to Scotland, are shown in Figure 3.1.(i). The first study area was the upper River Dee in Aberdeenshire above Crathie and its tributaries, located in Figure 3.1.(i).A. The second incorporated the middle River Spey in Moray/ Inverness-shire between the confluences of the rivers Tromie and Avon, and the south to north flowing tributaries, namely the Tromie, Feshie, Druie, Nethy and Avon, located in Figure 3.1.(i).B. The third was the lower Tweed and its Scottish tributaries, the Teviot, Leader, Whiteadder/Blackadder and Bowmont Water (a tributary of the Till), located in Figure 3.1.(i).C. It can be seen that the choice of these three study areas facilitated comparison between rates of fluvial processes within the Cairngorm massif and the Southern Uplands, and also between catchments with different controls within the Cairngorms. The environmental background for each area was divided into six sub-sections, namely catchment

The general situation of the three study areas

Figure 3.1.(i)



characteristics, geology, geomorphology, climatology, hydrology, and landuse. The standardisation of such background information was recommended by Starkel and Thornes (1981).

Standardised basin characteristics, collated by the Flood Studies Report (FSR) (NERC, 1975), are tabulated for drainage basin areas above gauging stations within or proximal to the study areas for comparative purposes. Precise definitions of the parameters given are as follows. The parameter STRMFREQ gives a measure of the stream network above a gauging station, which will have implications in terms of the timing of storm runoff. S1085 gives a measure of mainstream slope and provided a useful inter-catchment comparison. Having determined the length of the mainstream, the S1085 slope was calculated, using the lengths and elevations of the 10% and 85% points (FSR, NERC 1975). LAKE, the fraction of the catchment draining through lakes or reservoirs, was derived from the O.S. 1:250,000 maps.

Additional catchment characteristics were then collected using a map data base (O.S. 1:63360 series) and the parameters extracted were drainage basin area (km^2), amplitude of basin relief (m) and average basin relief (m). Basin relief is the difference in metres between the height at the outlet of the basin and the highest summit in the basin. The length of the drainage basin was also calculated (km) and thus an indication of the average slope or relief ratio of the catchment (m km^{-1}), as recommended by Starkel and Thornes (1981). Solid and drift geology for each area were derived from 1:63360 O.S. Geological Survey maps and memoirs. Geology is important in terms of catchment slope, channel confinement, erodibility of rocks and the size of the clasts transported. Local outcrops may give particular characteristics to subareas of the catchment.

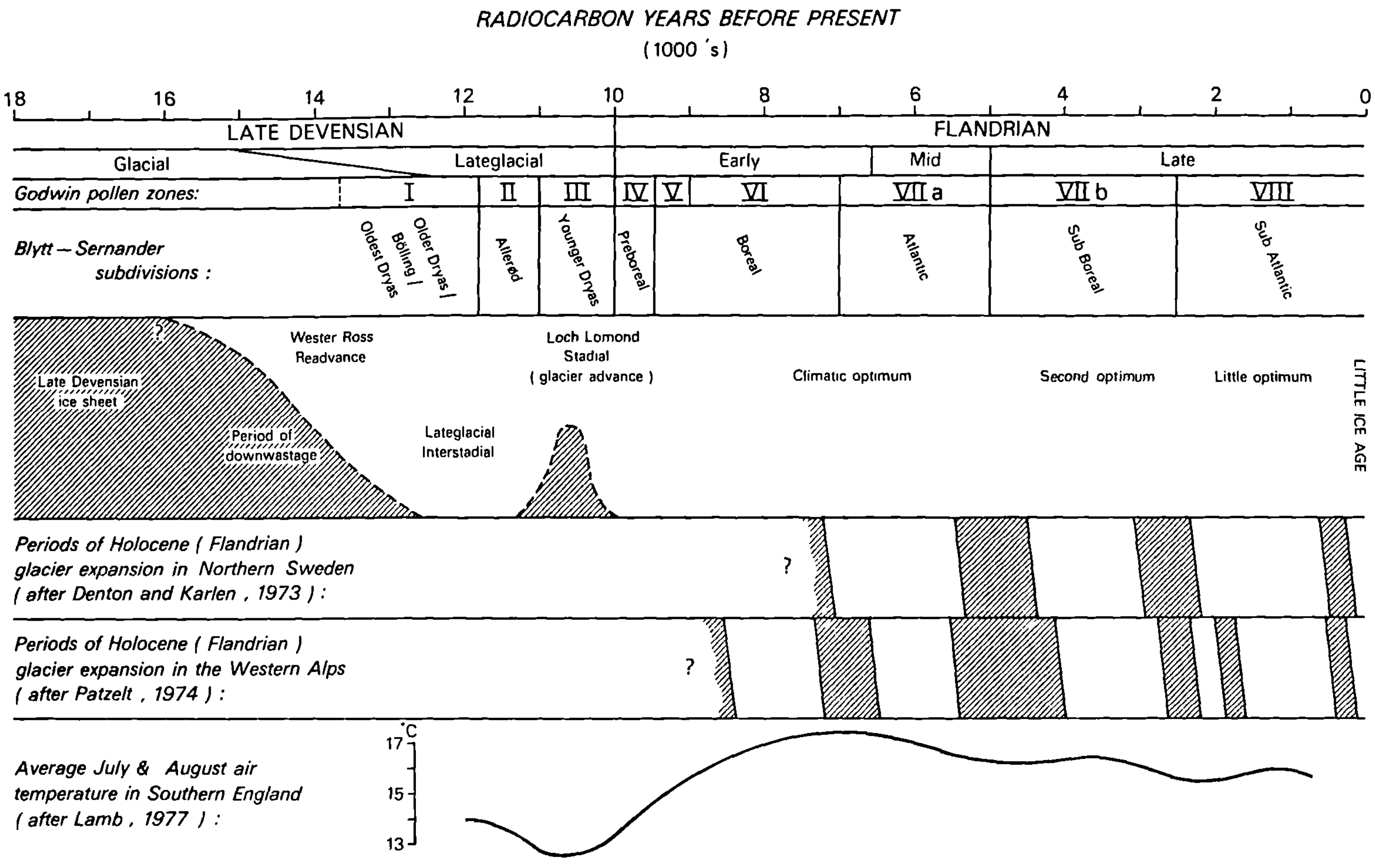
The geomorphology will be discussed in relation to the Lateglacial legacy and Holocene modifications of erosional and depositional landforms. This is important in assessing sediment availability, accessibility and channel confinement, by either the original or reworked deposits. Unfortunately, the depths of depositional materials were rarely obtainable from the literature. Periglacial activity must also be mentioned, but present-day periglacial processes were noted to be relatively insignificant (cf. much higher rates during the Loch Lomond Stadial), with the slope and channel systems operating largely as independent units in terms of sediment transport. This was confirmed by Ballantyne (1981) on the west coast on An Teallach. However, the west to east coast climatic gradient and contrasting rock types in Wester Ross, in comparison with the study areas, must be noted. Locally, debris flow activity and colluvial fans may be important, but only in certain areas. Occasionally, there may be extensive outwash deposits or localised, entrenched kame deposits, eg. the Allt Mor on the northern slopes of Cairngorm, and then direct sediment inputs are possible.

In terms of climatic conditions, it was first necessary to understand Holocene climatic changes, which apply more generally to Scotland. When the glaciers began to decay around 10,300 BP, there were dispersing meltwaters of high competence, and this date marks the commence of the Postglacial or Holocene in Scotland. Following the downwasting of the last major Scottish icesheet, there was a period of climatic warming, followed by a period of approximately 1000 years cooling associated with the Loch Lomond readvance (Figure 3.1.(ii)). The Loch Lomond readvance was important in terms of its effect on runoff and sediment regimes. Runoff rates would have been higher but still relatively modest, in comparison with those associated with the downwasting of the Late Devensian icesheet. Floodplains were unvegetated and sediment transport rates much higher, partly in response to increased periglacial activity. The complex response to the different conditions of the Loch Lomond Stadial (ie. in terms of paraglacial processes) would be expected to take several thousand years to be completed, with reworking of mobilised sediment (cf. Church and Ryder, 1972).

The following sequence of environmental change (Table 3.1.(i)) has been established for the whole of Scotland by climato-vegetation studies through pollen analysis (Gray and Lowe, 1977; Pennington, 1974). Average rainfall and temperature have clearly changed through the Holocene. Blytt and Sernander's Scandinavian chronology does not always correspond with evidence from the British Isles, but the basic structure of the chronology is still recognised as valid (Pennington, 1974). At present, the climate of Scotland is thus within the Sub-Atlantic period. The climatology sub-section will outline the basic climatic parameters for each area, as rainfall regimes can vary markedly on a regional basis. Climate and climatic fluctuations over the past 400 years will be discussed in Chapter 5 and this section only aims to place that detail in context. All three study areas are to the east of the highest mean annual precipitation (ie. in rain-shadow locations) and thus would be expected to have lower annual rainfall totals than the west of Scotland (Figure 3.1.(i)).

Environmental change, 18000 BP to present

Figure 3.1.(ii)



Source: Ballantyne (1981) p8 Figure 2-1

Table 3.1.(i)

Blytt and Sernander's division of the Postglacial

Date of boundary	Blytt and Sernander period	Climate
10,000 BP	Pre-Boreal	sub-arctic
10-7,000 BP	Boreal	dry and warmer
7-5,000 BP	Atlantic	warm and wet (climatic optimum) oceanic
4-3,000 BP	Sub-Boreal	warm and dry (climatic optimum) continental
2500 BP to present	Sub-Atlantic	wet and cool (continuing) oceanic

(Source: Pennington, 1974 p45 Table 1)

Hydrologic regimes will be discussed in relation to the gauged discharge data base. A measure of soil moisture deficit (SMD) was calculated by the FSR (NERC, 1975) from:

> "A water balance between daily rainfall and a Penman estimate of actual transpiration assuming a notional catchment, under 50% short rooted vegetation, 30% long rooted vegetation and 20% riparian areas and estimates the root constant for these three zones." (FSR, NERC 1975, p307)

"RSMD" is the five year recurrence interval, 1 day rainfall excess. The FSR (NERC, 1975) also provided a measure of winter rainfall acceptance potential to give an indication of the infiltration capacity of catchment soils.

> "Winter rain acceptance is broadly infiltration potential and the reverse of runoff potential. It is favoured by combinations of high permeability, low ground water level or gently sloping terrain. Rapid movements of water likely to cause severe flooding during and after heavy rain, tend to occur near or over the soil surface; movements through deeper soil, or permeable geological deposits, tend to affect streamflow later or not at all." (FSR, NERC 1975 p303)

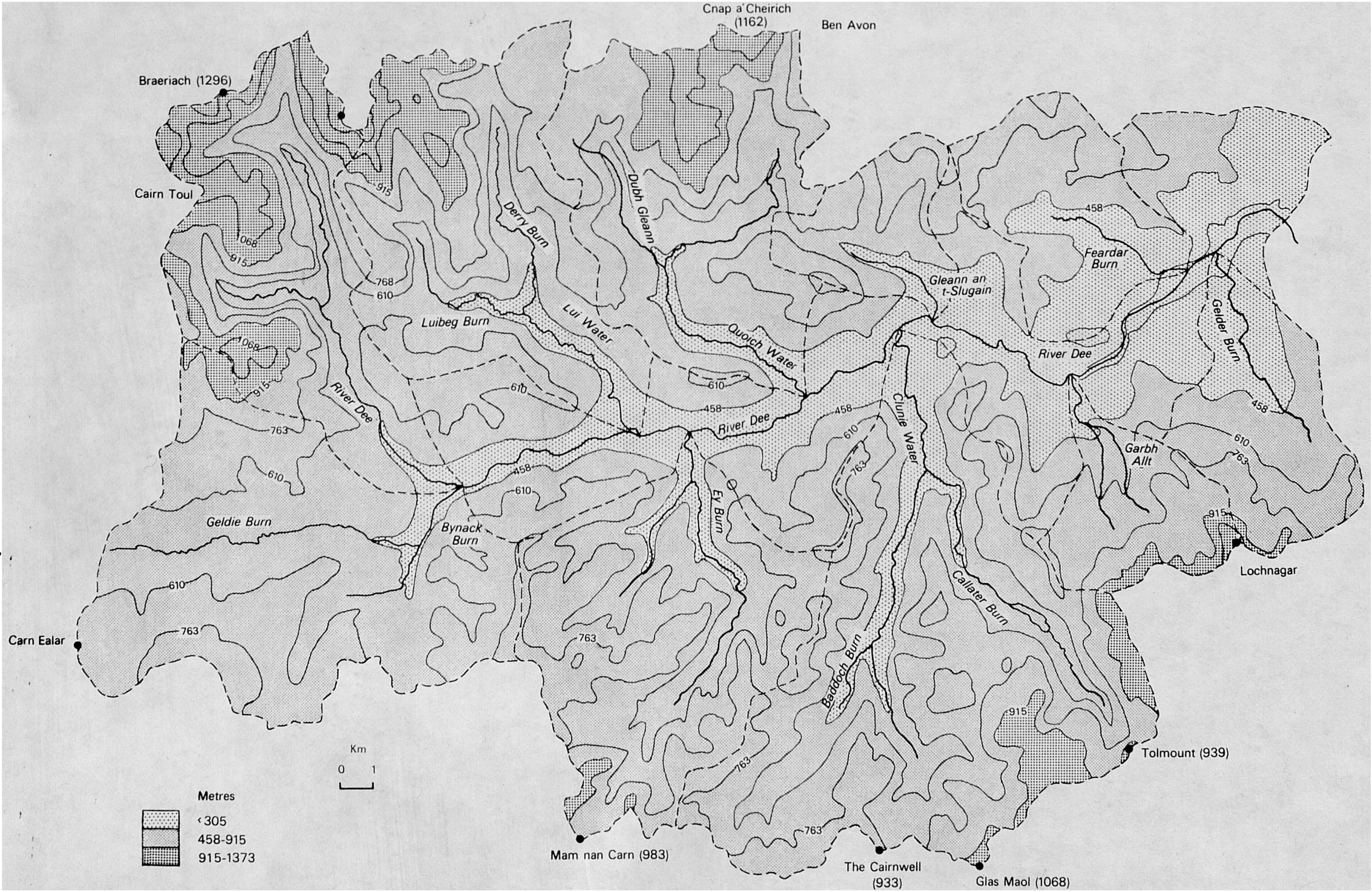
Landuse change has been both bioclimatic and man-induced in all three study areas, but to differing extents and over different time-periods. Landuse changes, discussed briefly here, are detailed in Chapter 6; this section is only intended to present a longer timescale overview and specific details, relevant to the 150-250 years of study, will be elaborated on later in the text.

3.2.1 Dee study area: Catchment characteristics

The upper River Dee drains a semi-circular shaped catchment comprising the steep south-facing slopes of the Cairngorm massif (including Ben MacDhui 1310 m and Cairntoul 1294 m) and also the north-facing slopes of the Grampians (eg. Glas Maol 1068 m and Lochnagar 1154 m). The Gleann an t-Slugain, Quoich, Lui tributaries and the mainstream upper Dee all drain approximately north to south, while the Ey, Clunie and Gelder flow south to north (Figure 3.2.1.(i)). These tributaries and their sub-basins have their sources in contrasting upland massifs. A summary of the associated catchment properties are presented in Table 3.2.1.(i). In terms of basin relief ratio, the smaller catchments have highest values while the Clunie has by far the lowest with 47.6 m km^{-1}. The profile of the river bed on the mainstream Dee (Roberts, 1919) showed the considerable increase in slope above Ballater (Figure 3.2.1.(ii)) and longitudinal profiles for individual catchments, digitised from 1:63360 O.S. maps, are shown in Figure 3.2.1.(iii). From the <u>FSR</u> (NERC, 1975), the catchment above Pollhollick (see Section 3.2.5) has a STMFREQ of of 2.26, a S1085 of 5.94 and a

The relief of the Dee study area

Figure 3.2.1.(i)



Contours in metres

Figure 3.2.1.(iii)

LONGITUDINAL PROFILES WITHIN THE DEE STUDY AREA

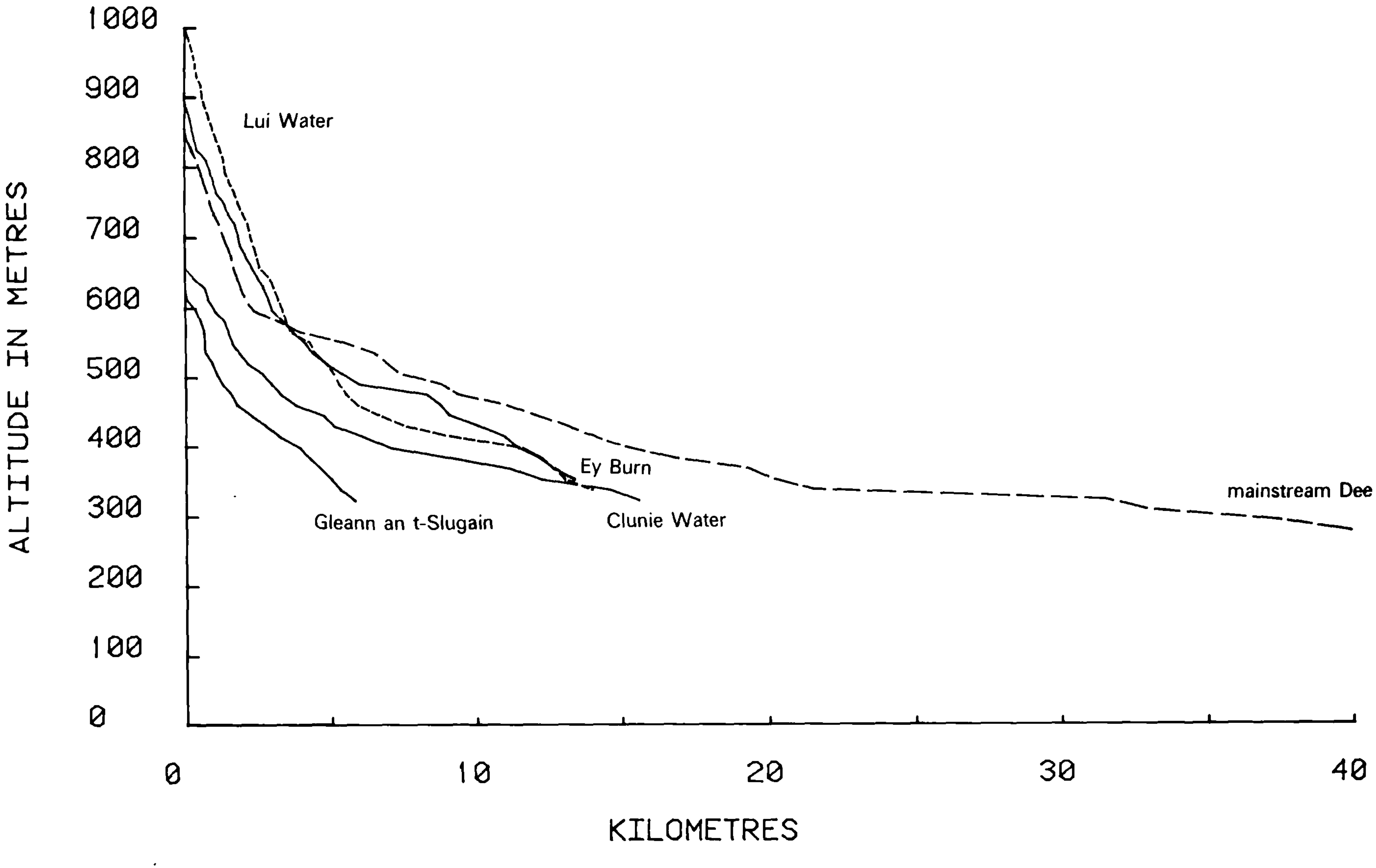
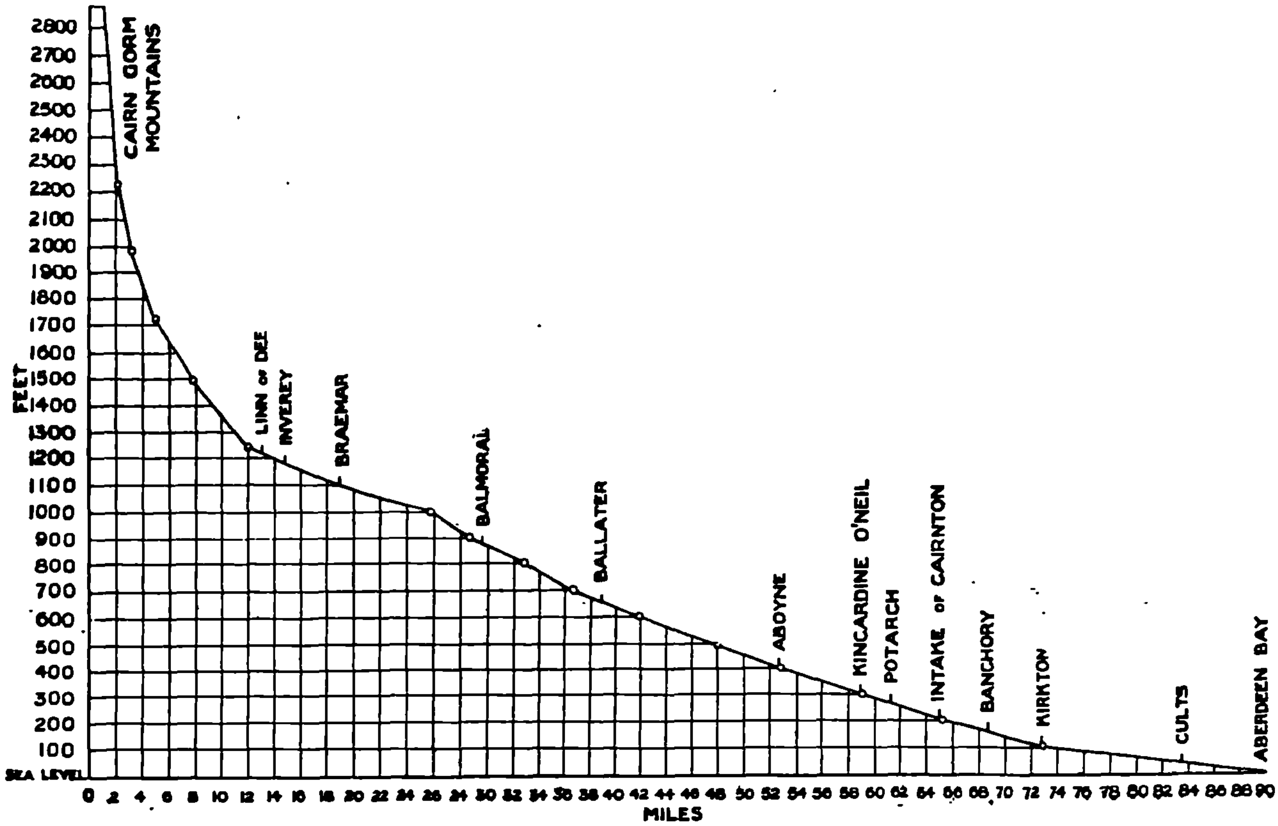


Figure 3.2.1.(ii)



Profile of the riverbed of the Dee from source to Aberdeen

(Source: Roberts, 1919 p4 Figure 1)

Table 3.2.1.(i)

Dee study area: Catchment parameters

Catchment	Basin Area (m^2)	Max. Ht. (m)	Min. Ht. (m)	Basin Relief (m)	Basin Length (km)	Relief Ratio ($m\ km^{-1}$)
(a) River Dee	573.2	1295	271	1024	73.5	13.9
(b) Clunie Water	105.4	1068	320	748	15.7	47.6
Callater Burn	34.0	1062	366	696	9.9	70.3
Baddoch Burn	23.0	974	411	563	7.3	77.1
Cairnwell Burn	22.7	1068	411	657	5.0	113.3
(c) Geldie Burn	79.8	1090	404	686	10.8	63.5
(d) Ey Burn	61.6	1044	335	709	11.8	60.1
(e) Lui Water	58.3	1197	358	839	13.0	64.5
Luibeg Burn	23.9	1155	424	731	6.8	107.5
Derry Burn	26.2	1155	424	731	8.2	89.1
(f) Quoich Water	58.1	1196	450	746	10.4	71.7
Dubh Gleann	15.9	1196	326	870	6.5	133.8

Table 3.2.1.(i) cont.

Catchment	Basin Area (m^2)	Max. Ht. (m)	Min. Ht. (m)	Basin Relief (m)	Basin Length (km)	Relief Ratio ($m\ km^{-1}$)
(g) Feardar Burn	27.6	900	290	610	5.7	107.0
Feindallocher	20.0	1088	308	780	6.2	125.8
Gleann an-t S.	16.7	863	331	532	5.0	106.4
River Gelder	29.2	1155	287	868	8.7	99.8

value for LAKE of 0.029. The only major water body is Loch Callater (Figure 3.2.1.(i)), which occupies only a small % of the total catchment area.

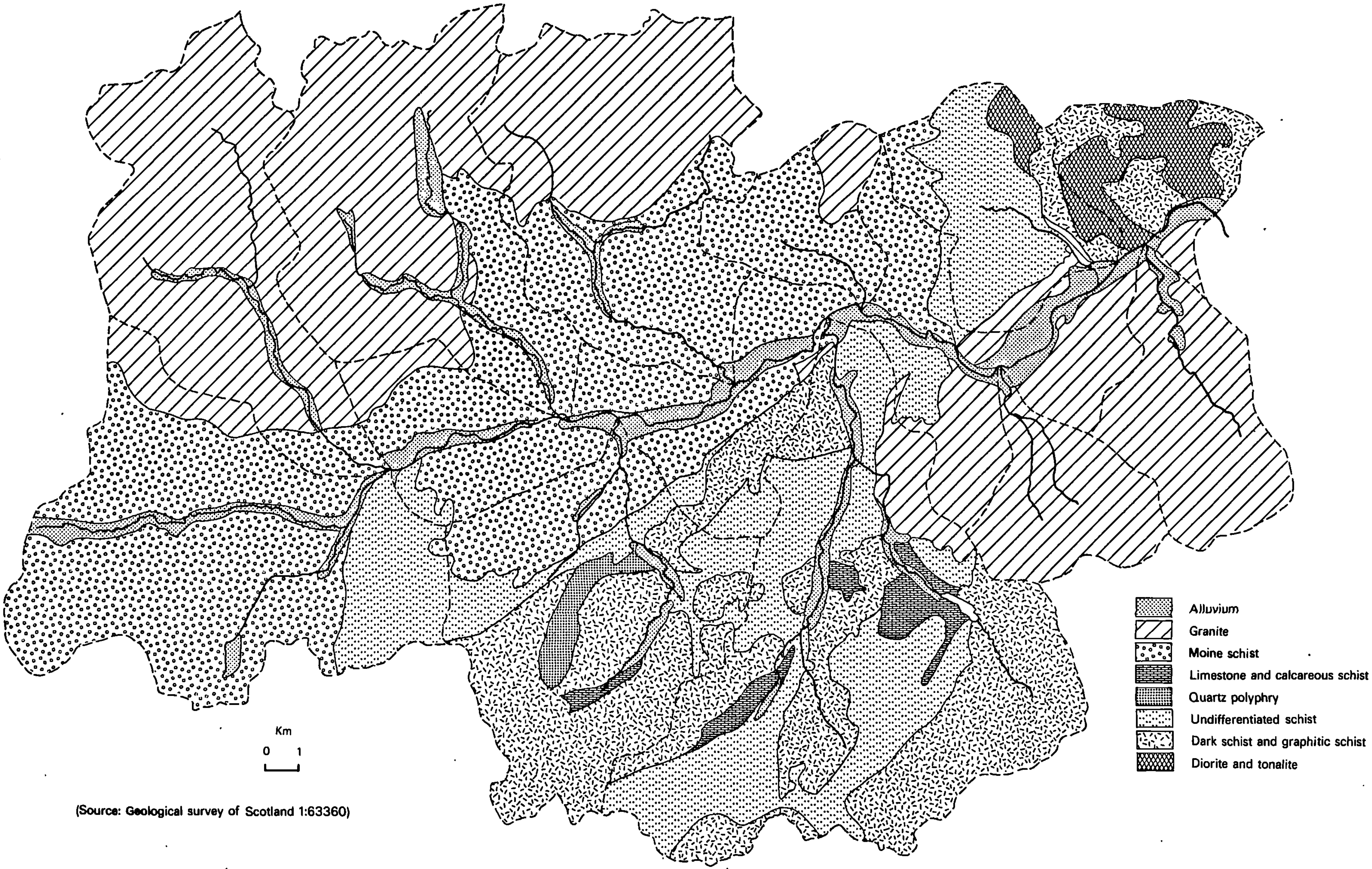
3.2.2 Dee study area: Solid and drift geology

Even within a relatively small area, the individual catchments display a variety of lithological conditions. The solid geology of the River Dee catchment above Crathie is shown in Figure 3.2.2.(i). The solid geology of the Quoich catchment is predominately Moine gneiss but, in the upper reaches of the Quoich and the Dubh Gleann, there are large granite intrusions, which form the highest peaks. The Lui and Gleann an t-Slugain have a similar pattern with some quartzite/ quartzite schist.

The simplest geology was found in the smaller catchments eg. Gelder, upper Dee and Geusachan, which are entirely underlain by granite, and the Gleann an t-Slugain and Geldie Burn, which are principally Moine gneiss. In contrast, the Clunie and the Ey have a much more diverse geology of metamorphic rocks, both have Moine schist near their confluences and varying amounts of graphitic and quartz schist in their upper catchments. The geology of each catchment can be studied in detail in Figure 3.2.2.(i). In terms of the drift geology, the distribution of glacial, outwash and alluvial material, as indicated on the 1:63360 Geological survey, is shown in Figure 3.2.2.(ii). The catchments are typified by localised widening into a U-shaped valley or alluvial basin, where there is an abundance of available trapped material to be reworked eg. upper Clunie and Ey catchments.

The solid geology within the Dee study area

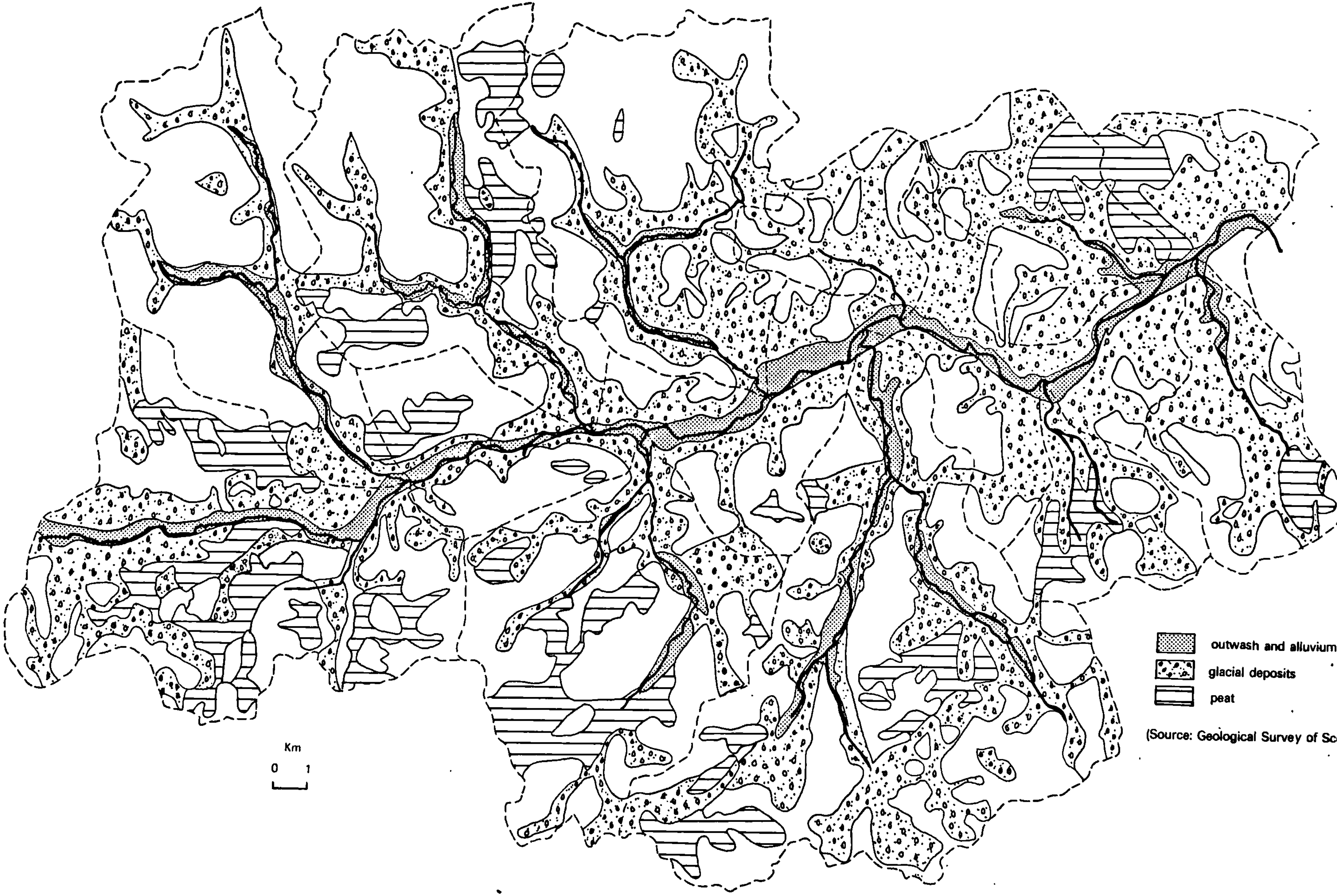
Figure 3.2.2.(i)



(Source: Geological survey of Scotland 1:63360)

Drift geology within the Dee study area

Figure 3.2.2.(ii)



(Source: Geological Survey of Scotland 1:63360)

3.2.3 Dee study area: Geomorphology

The glacial history and subsequent legacy of this area are very important, both in terms of erosional and depositional landforms. The Dee study area has been subject to several severe glaciations; particularly important was the decay of the last major icesheet which, according to Sissons and Walker (1974), reached its maximum extent 17000-18000 radiocarbon years ago. The Cairngorm massif has been subject to a great deal of controversy concerning its Lateglacial history, principally in relation to the importance of the Cairngorm equivalent to the Loch Lomond pollen zone III readvance (Figure 3.1.(ii)) but also concerning the nature and importance of the more recent "Little Ice Age" (see Section 5.3). A series of papers reflect chronologically the changing ideas of the main rivalling viewpoints of Sissons (1967, 1974, 1976), Sissons and Grant (1972) and Sugden (1970, 1980). However, only the geomorphic legacy of this glacial history is relevant to this study.

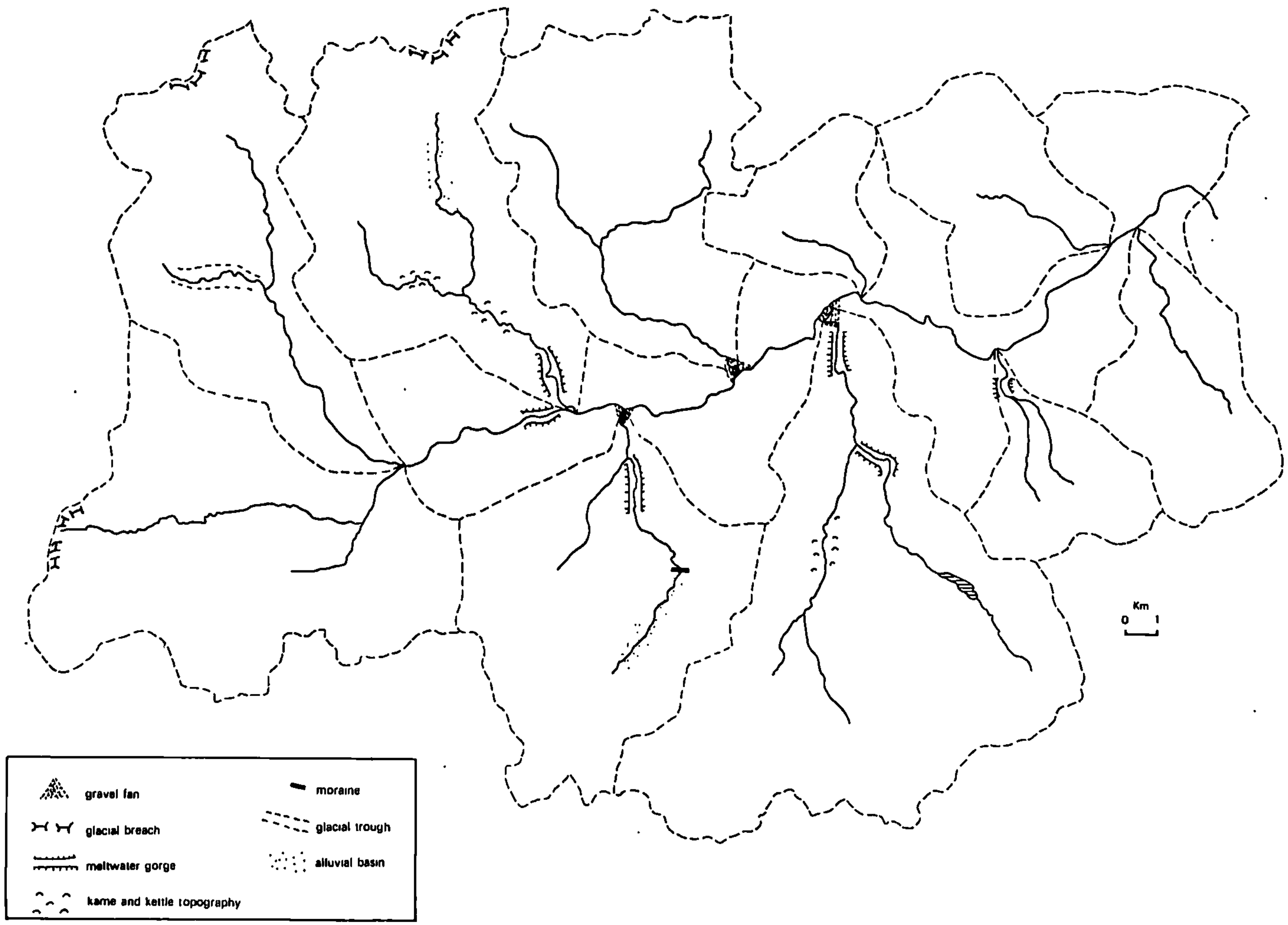
The glacial modification of the landscape is most easily presented working systematically downstream and some important examples are annotated schematically on Figure 3.2.3.(i). The upper parts of the catchments are dominated by erosional landforms. Steep sided glacial troughs occur as in Glen Geusachan; such troughs being excavated along lines of pre-glacial drainage that were suitably orientated to take the north-easterly discharge of the Grampian ice sheet (Whittow, 1977). Where preglacial valleys stood at right angles to the dominant direction of ice movement, as in Glen Lui, Glen Quoich and Gleann an t-Slugain, such valleys kept much of their original fluvial formation. The breaching of pre-glacial watersheds has also occurred in some places eg. upper Feshie/ upper Geldie burn (Linton, 1949) and upper Glen Derry/ upper Glen Avon (Figure 3.2.3.(i)). Steep-sided corries, sometimes with moraines, are found in the granite of the Cairngorm massif (eg. Lochan Uaine and Coire nan Clach) and south of Braemar, on Lochnagar.

Frequently in the upper reaches, there are steep sided "U-shaped" alluvial basins with gently sloping valley floors eg. in Glen Geusachan, Glen Derry and Glen Ey, which form sediment stores for a substantial amount of outwash and alluvial material. A series of alluvial basins separated by relatively steep rock controlled reaches may have significant effects on the baselevels of stream activity. Morainic deposits can also be identified and act as barriers to downstream sediment transfer eg. in Glen Ey where a large moraine subdivides a U-shaped valley.

Another locally important slope activity, in terms of sediment transfer to rivers in upland areas, is debris flow activity eg. Lairig Ghru and Glen Geusachan (Sugden and Ward, 1980; Baird and Lewis, 1956; Innes, 1982). These inputs are however periodic, only occurring occasionally during rare major storm events. Furthermore such debris flows only rarely input sediment directly into river channels eg. in the Lairig Ghru, and it is much more frequently a two stage process. Alluvial fans from smaller, steep tributaries, eg. on the middle Lui and upper Clunie, also contribute sediment to the channel through basal undercutting and in localised flood events where the fan is active, with the flushing out of sediment from these small basins.

Selected geomorphological features within the Dee study area

Figure 3.2.3.(1)



Frequently, in the middle reaches of most of the tributaries, there are undulating fluvioglacial deposits, where the valley widens and is reduced in slope. Such kames, for example those in Glen Lui, Glen Luibeg and Glen Clunie are being reworked in varying degrees by the present river system. Terrace features also provide sediment stores to be reworked, but there are fewer sequences of terraces than in Speyside. The depth of fluvioglacial material varies, but along localised reaches the rivers have cut down to bedrock eg. on the middle Clunie.

Further down valley near the confluences with the mainstream Dee, meltwater gorges exist at a variety of scales, channelling the flow through a narrow rock-cut section. For example, there is a large example at Linn of Dee (with a channel 4 feet [1.2 m] wide and 300 yards [275 m] long,(Mackie, 1911)) and several others exist at a smaller scale eg. Linn of Quoich and the Clunie Water at Braemar. These deep gorges must have been the product of one or more glacial phases and the scour features indicate that they must have been eroded by a considerable stream power. Clearly the meltwaters derived from the Loch Lomond readvance glaciers were not on their own sufficient to cause such scouring. These are entrenched meanders in the absolute sense of the term. The lower reaches, characterised by a considerable widening of the valley floor, comprise areas of outwash fluvioglacial gravels and low level gravel fans. The tributaries then flow into the lower slope mainstream Dee.

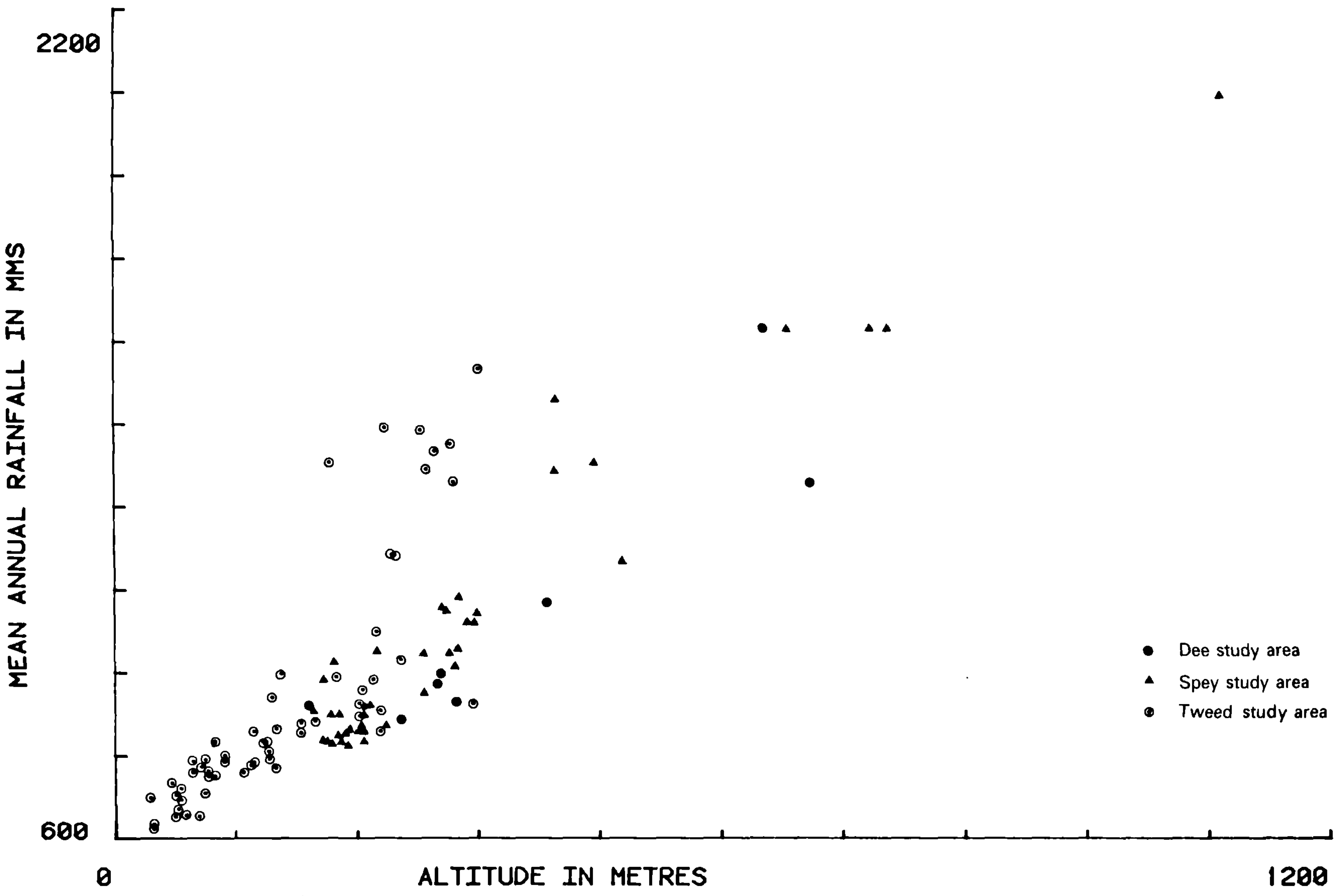
3.2.4 Dee study area: Climatic conditions

In terms of present day rainfall conditions, derived from the standardised Meteorological Office tabulation "Rainmaster", the mean annual rainfall for the standard period (1941-1970) varies from 845.0 to 1600.0 mm (see Table 3.2.4.(i)) for locations and altitudes). This is highly dependent on altitude and aspect, as tends to be the case in mountainous regions. The FSR (NERC, 1975) gave the annual average rainfall (1916-1950) for the catchment above Pollhollick as 1164 mm. The relationship between magnitude of average annual total and altitude can be seen in Figure 3.2.4.(i), and this is compared with the other two study areas. There is a large increase in rainfall with altitude from this gauged record, especially on the north-facing slopes of the Clunie catchment. However, vast areas of the upper catchment have no raingauges (Figure 3.2.4.(ii)), and the average rainfall may be underestimated; the unpredictability of the rainfall regime is very important, especially in relation to periodicity of summer convective storms (Chapter 5.4.7).

3.2.5 Dee study area: Hydrology

The entire Dee catchment has 4 gauging stations with varying lengths of record (Table 3.2.5.(i)), maintained by the North East River Purification Board and located in Figure 3.2.5.(i). The Dee catchment above Crathie was not gauged at any point until very recently (the Mar Lodge record commenced 1982). However, there is a station at Pollhollick, above Ballater, with a catchment area of 690 km^2 (see

Figure 3.2.4.(i) Mean annual rainfall (1941–1970) in relation to altitude in the three study areas



The location of rainfall gauge sites within the Dee study area

Figure 3.2.4.(ii)

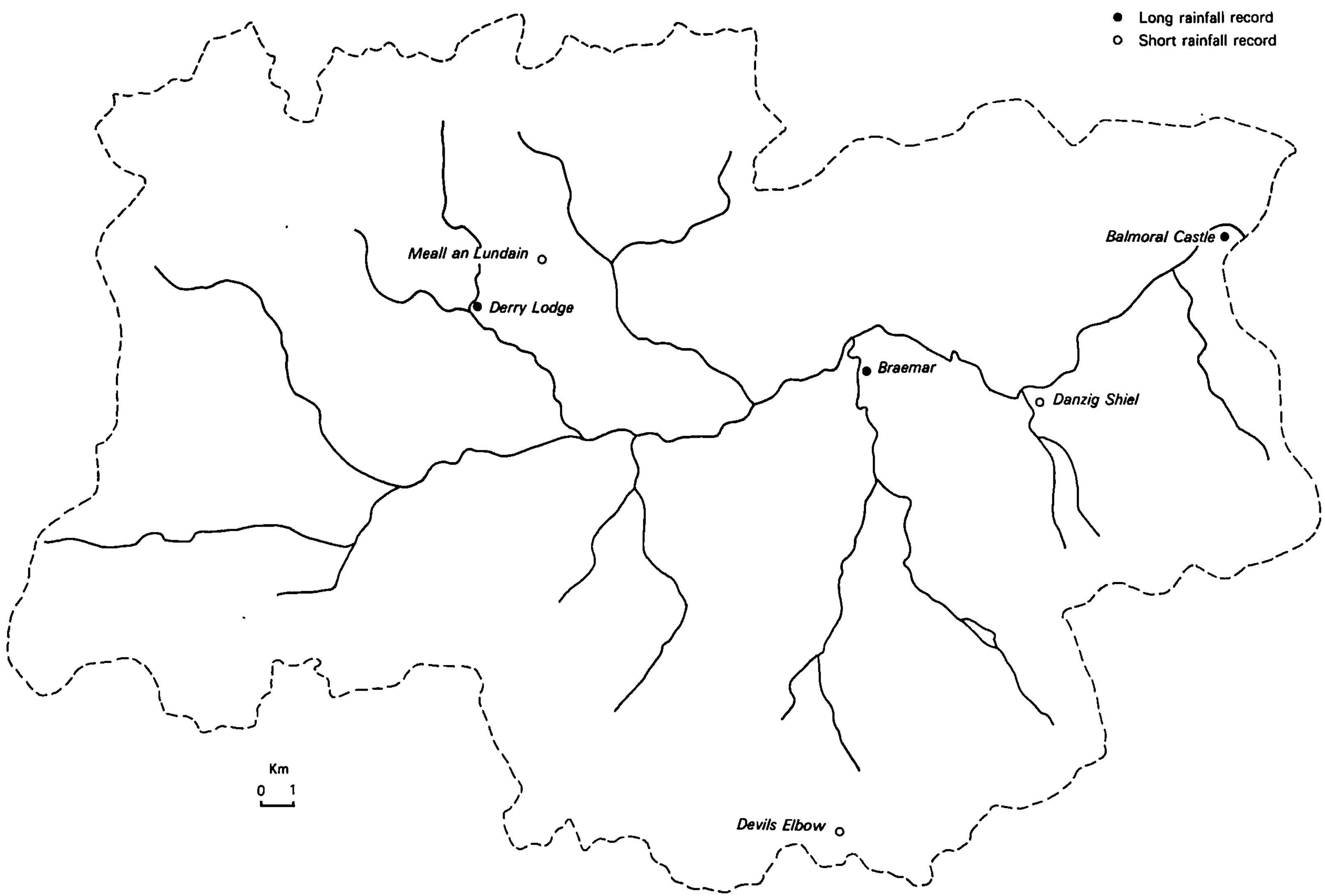


Table 3.2.4.(i)

Mean annual rainfall totals for stations within upper Deeside (1941-1970)

Station	Grid Reference	Altitude (m)	Rainfall (mm)
Ballater	3380 7965	193	866.0
Balmoral	3260 7946	283	845.0
Danzig Shiel	3202 7904	320	910.0
Braemar	3152 7914	339	879.0
Derry Lodge	3036 7932	427	1072.0
Devils Elbow	3139 7780	640	1600.0
Meall an Lundain	3054 7957	686	1300.0

Table 3.2.5.(i)

Discharge gauging stations on the River Dee

Station	Grid Reference	Area (m^2)	Length of Record
Mar Lodge	NO 098895	289	commenced 1982
Pollhollick	NO 343965	690	18/6/1975-31/12/1982
Woodend	NO 632960	1370	1/10/1929-16/2/1983
Parkbridge	NO 798983	1844	10/10/1972-22/1/1982

Table 3.2.5.(ii)

Summary statistics for discharge gauging stations on the River Dee

	Mean flow (m^3 s^{-1})	Mean annual flood (m^3 s^{-1})	Maximum instantaneous flow (m^3 s^{-1})
Mar Lodge	****	****	230.0
Pollhollick	21.7	291.7	690.0
Woodend	35.7	462.1	1370.0
Park bridge	44.4	593.6	1844.0

**** data not available

Figure 3.2.5.(i)). This gives some indication of flow frequency for the basin as a whole over the period 18/6/1975 to 31/12/1982. Fortunately, no major rivers join with the Dee between Crathie and Pollhollick as the Rivers Garry and Muick flow into the River Dee after the gauging site. Further downstream, there are gauges at Woodend and Parkend, with drainage areas of 1370 and 1844 km^2 respectively. The Woodend record is exceptionally useful in a Scottish context, since it extends from 1/10/1929 to 16/2/1983 while Park Bridge is comparable in length with Pollhollick.

The catchment area above Pollhollick has a soil moisture deficit of 5.3 mm and RSMD of 55.7 mm (FSR, NERC 1975). The winter rain acceptance potential for rain falling on the valley floor is categorised as high (category 2), but all of the catchment above that narrow band was classified as very low ie. much of winter rainfall would be expected to runoff at or near the surface (Figure 3.2.5.(ii)). In terms of catchment runoff response, large areas of the upper catchments eg. Ey and Clunie have blanket peat (see Figure 3.2.2.(ii); Durno, 1959, 1967).

3.2.6 Dee study area: Landuse

Present vegetation in upper Deeside can be subdivided into several sub-categories, namely woodland, heathland and agricultural land. In terms of forestry, there is the contrast between the old Caledonian *Pinus sylvestris* woodland, (excellent examples being the forests of Ballochbuie and Mar), and newer planting by private landowners. The

Figure 3.2.5.(i)

The location of discharge gauging stations within the River Dee catchment

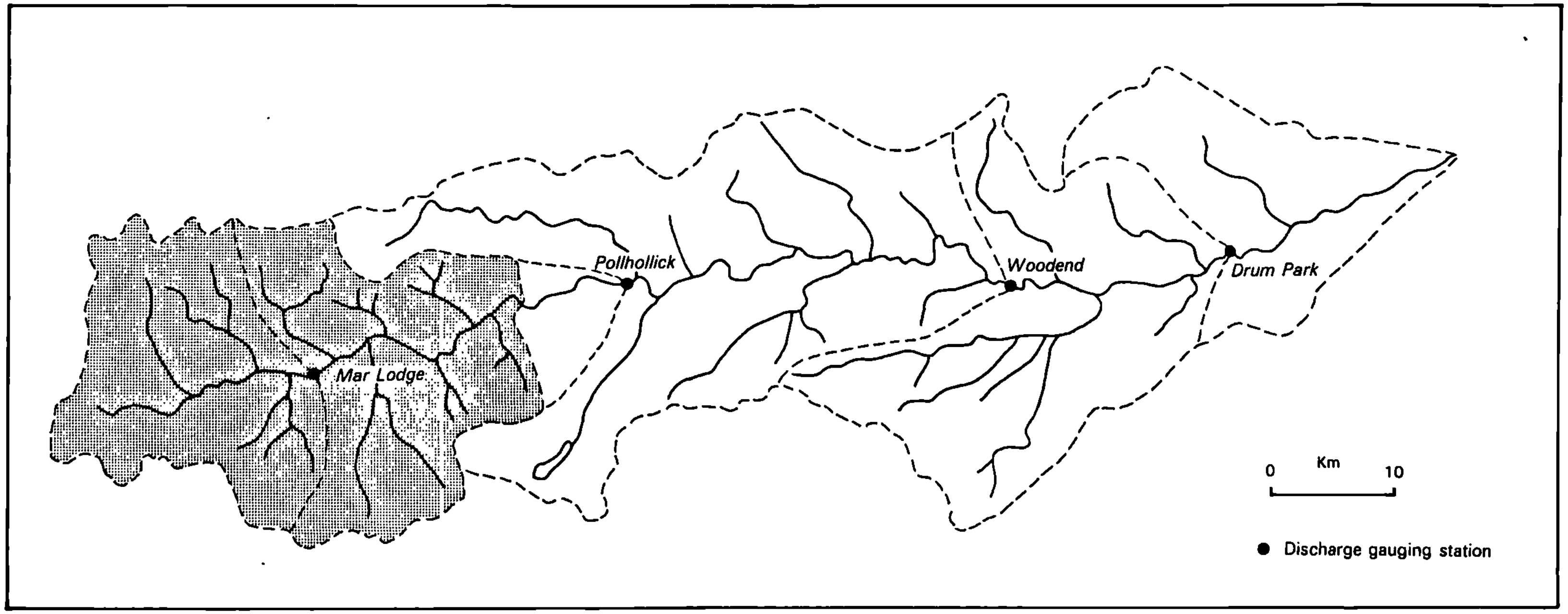
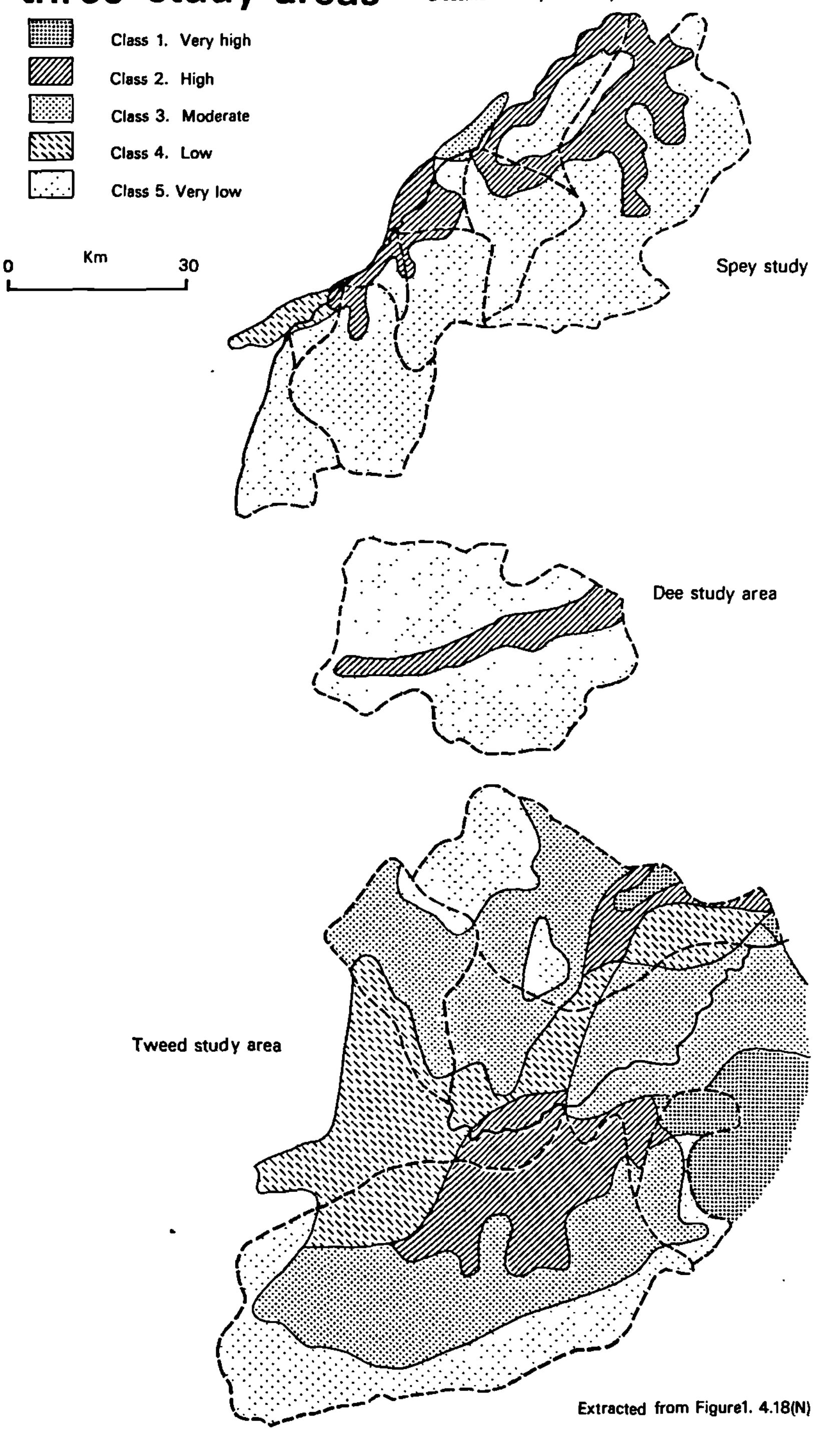


Figure 3.2.5.(ii)

Winter rain acceptance potential within the three study areas

Classification by Mackney, Smith and Thomasson



Extracted from Figure1. 4.18(N) in FSR (1975) extent of the Caledonian forests has been considerably reduced over the past three centuries due to the varying importance of the timber industry. In terms of agricultural landuse, there is some sheep grazing in the valleys, but considerable emphasis and income is derived from deer and grouse, and the estates are managed with this in mind.

3.3.1 Spey study area: Catchment characteristics

The catchment parameters of the Spey tributaries, which drain the steep north-facing corries of the Cairngorm massif, are tabulated in Table 3.3.1.(i). The relief maps for each catchment are shown in Figures 3.3.1.(i).A-F. The tributary basins in the Spey study area are larger in size than in Deeside, ranging from the Avon catchment (401.5 km^2) to the Tromie (126.5 km^2). As can be seen from Figures 3.3.1.(i).A-E, there is a gradual rise in the height of the upper catchment from west (Tromie) to east (Avon). The Avon has its source near Loch Avon and the north-facing slopes of some of the highest mountains within the Cairngorm massif, eg. Ben MacDhui (1310 m) and Ben Avon (1172 m) (see Figure 3.3.1.(i).E). The neighbouring Nethy catchment drains a steep-sided semi-circular basin, up to 1246 m high.

Progressing further upstream on the Spey, the next catchment is drained by the Druie which, with its tributary the Allt Mor, drains the north-facing corries of Cairngorm (1265 m) (Figure 3.3.1.(i) C). The Am Beanaidh has its source in Loch Einich and neighbouring corries. In contrast, the Feshie and Tromie catchments (Figures 3.3.1.(i).A-B) drain the western side of the Cairngorm massif. The Tromie catchment is long,

Table 3.3.1.(i)

Spey study area: Catchment parameters

Catchment	Basin area (m^2)	Max. Ht (m)	Min. Ht (m)	Basin relief (m)	Basin length (km)	Relief ratio ($m\ km^{-1}$)
(a) Spey	2514.0	226	142	84	60.5	1.4
(b) Tromie	126.5	952	226	726	21.8	33.3
(c) Feshie	234.2	1265	229	1036	23.4	44.2
Eidart	32.0	1265	519	746	10.2	73.1
Allt Chromaig	44.0	858	281	577	11.9	48.5
(c) Druie	126.5	1265	207	1058	16.5	64.1
Allt Mor	16.4	1246	328	918	7.0	131.1
Am Beanaidh	59.4	1265	229	1036	15.3	43.0
Luineag	63.8	320	229	91	4.2	67.7
(d) Nethy	123.8	1265	195	1070	19.6	54.6
Dorback	36.2	821	317	504	10.6	47.5
(e) Avon	401.5	1310	142	1168	54.0	21.6
Livet	136.0	688	258	430	16.2	41.7
Conglass	49.0	792	271	521	15.7	33.2

Figure 3.3.1.(i).A

The relief of the River Tromie catchment

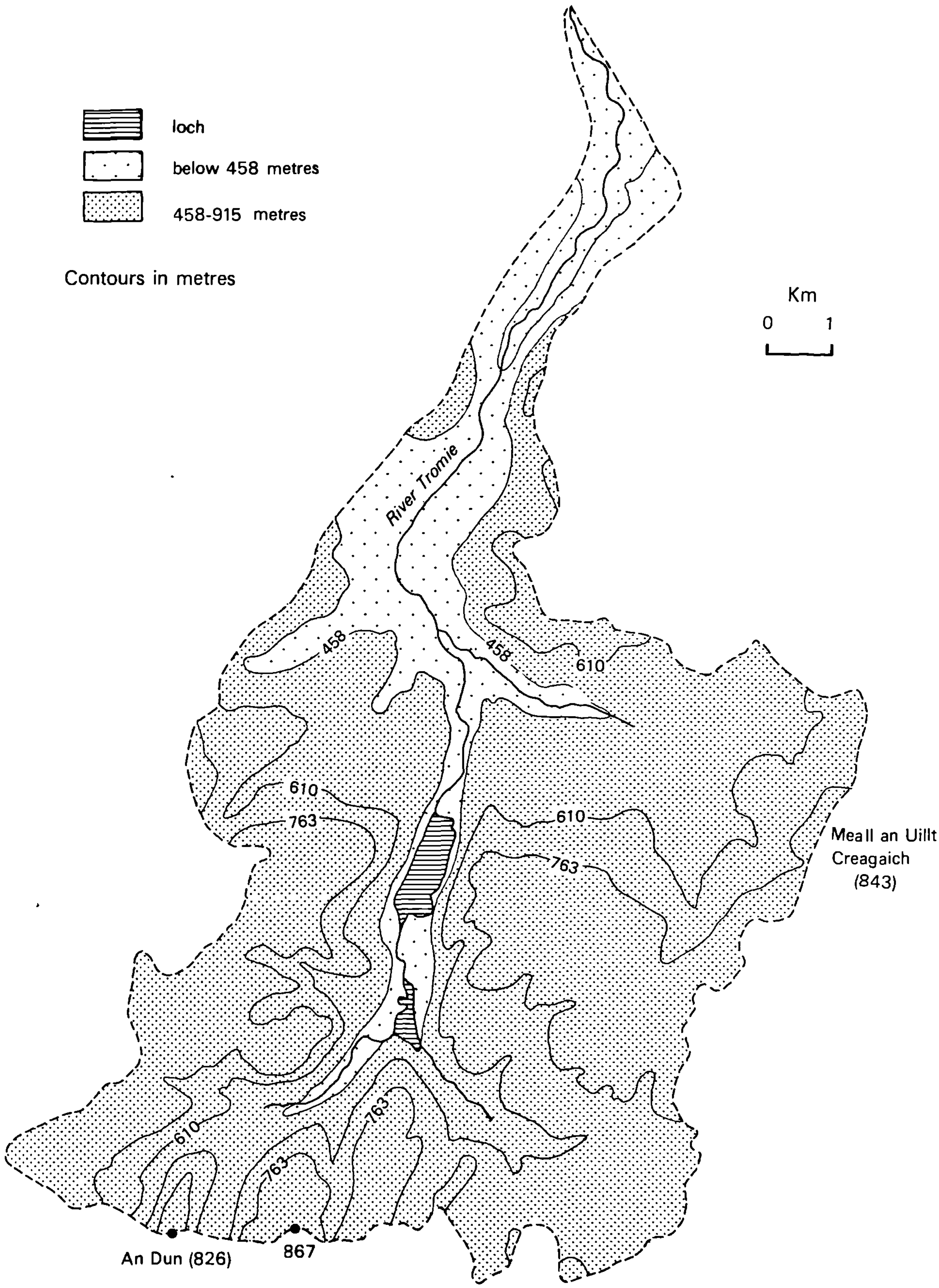


Figure 3.3.1.(i).B

The relief of the River Feshie catchment

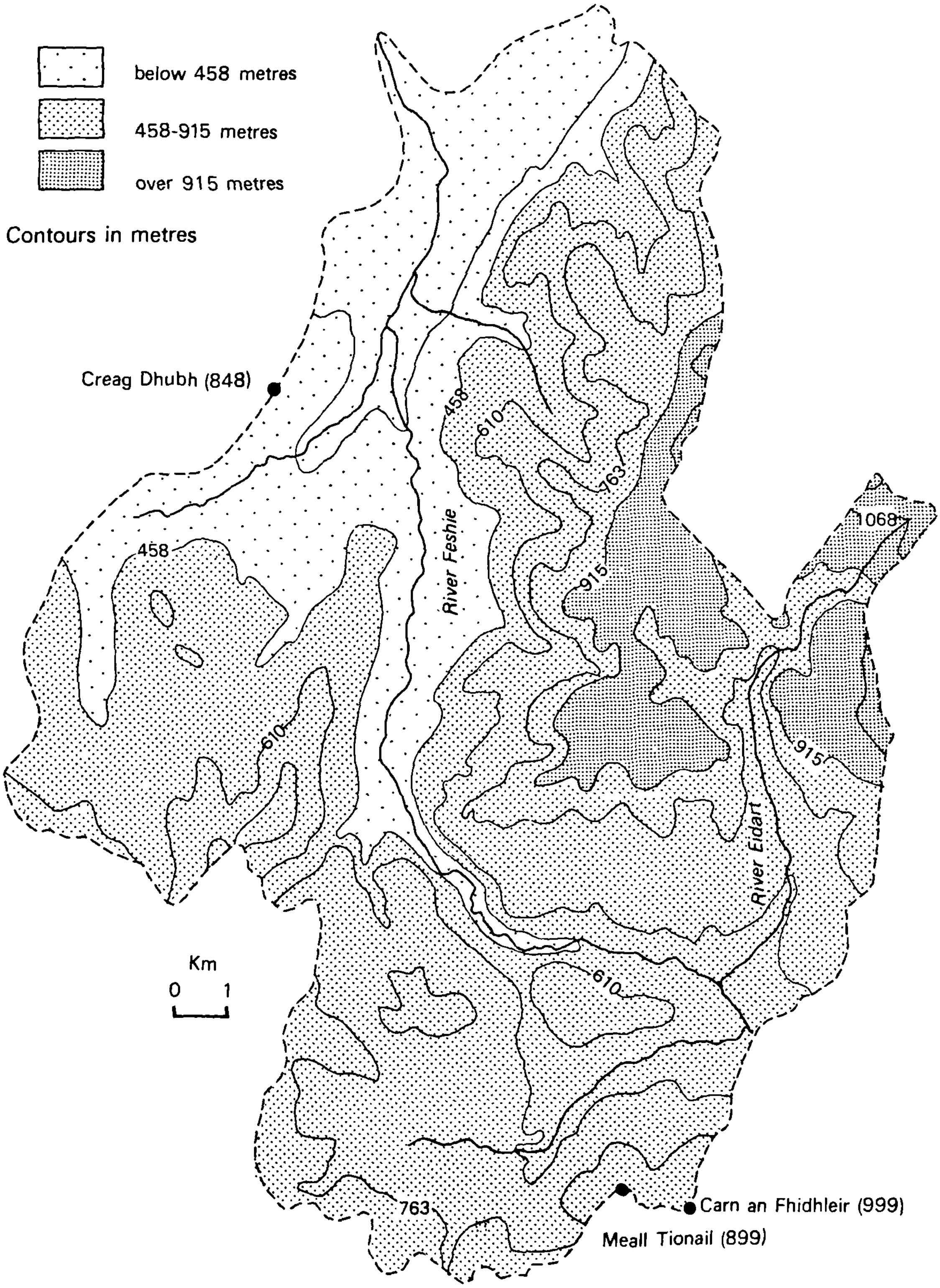


Figure 3.3.1.(i).C

The relief of the River Druie catchment

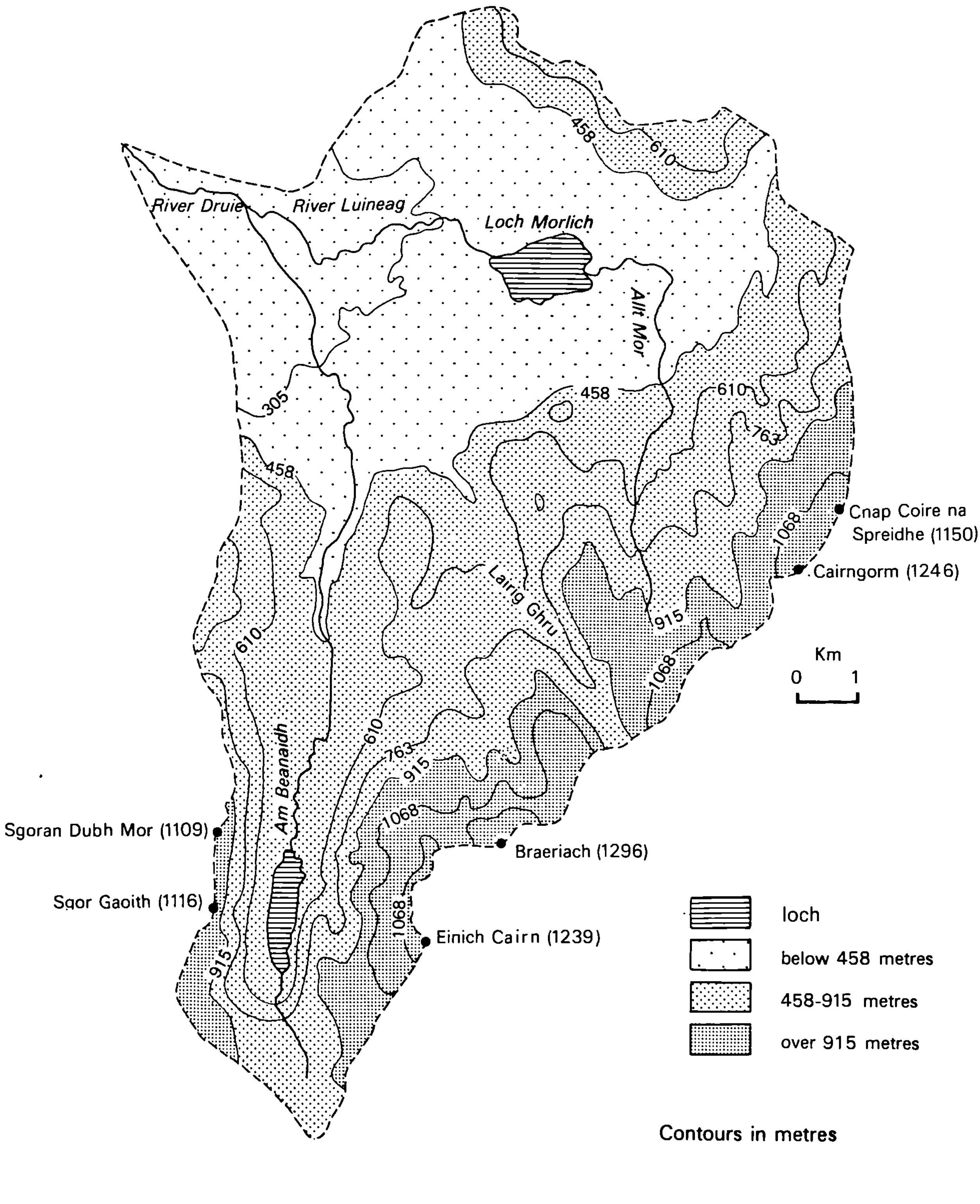


Figure 3.3.1.(i).D

The relief of the River Nethy catchment

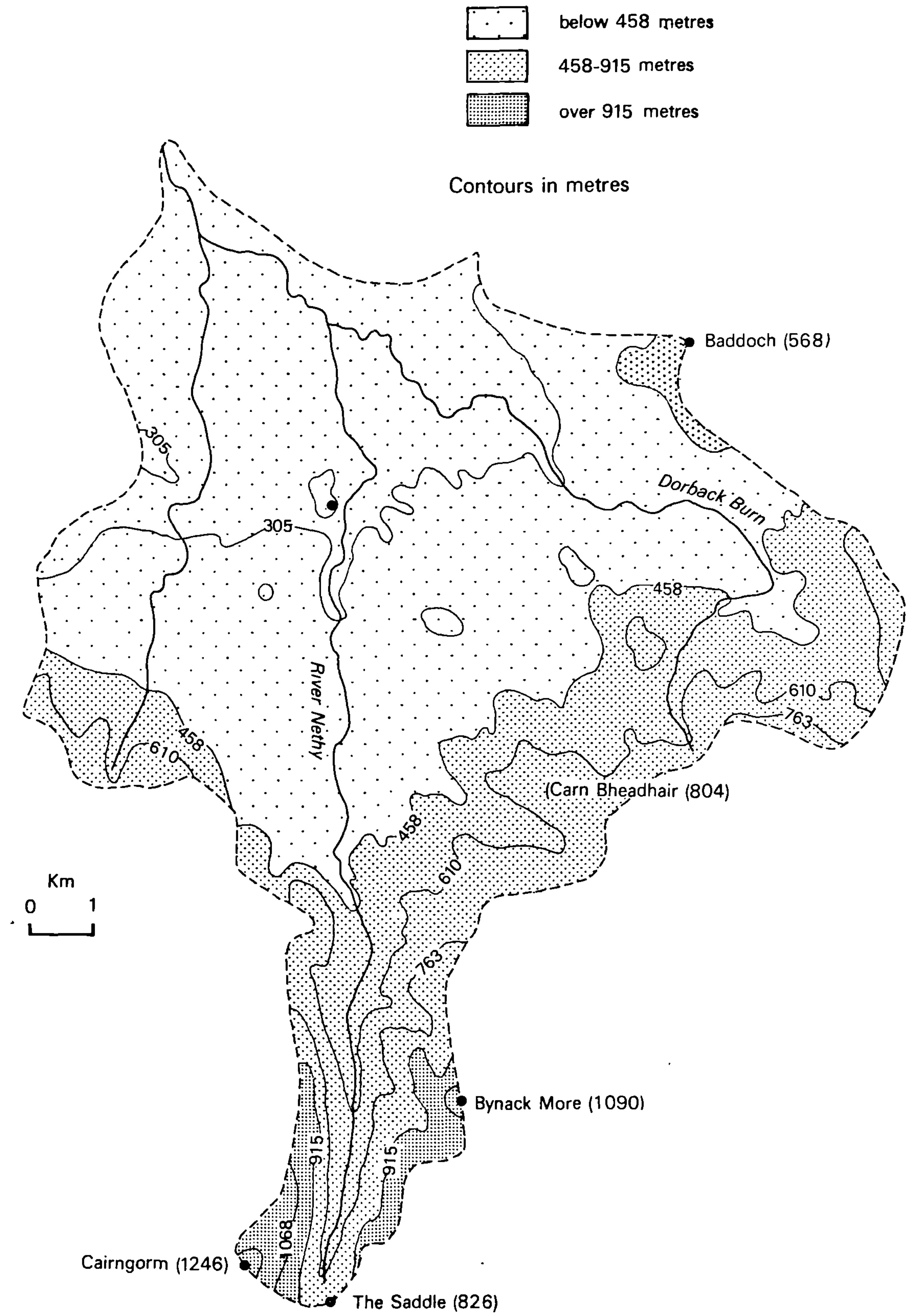


Figure 3.3.1.(i).E

The relief of the Avon catchment

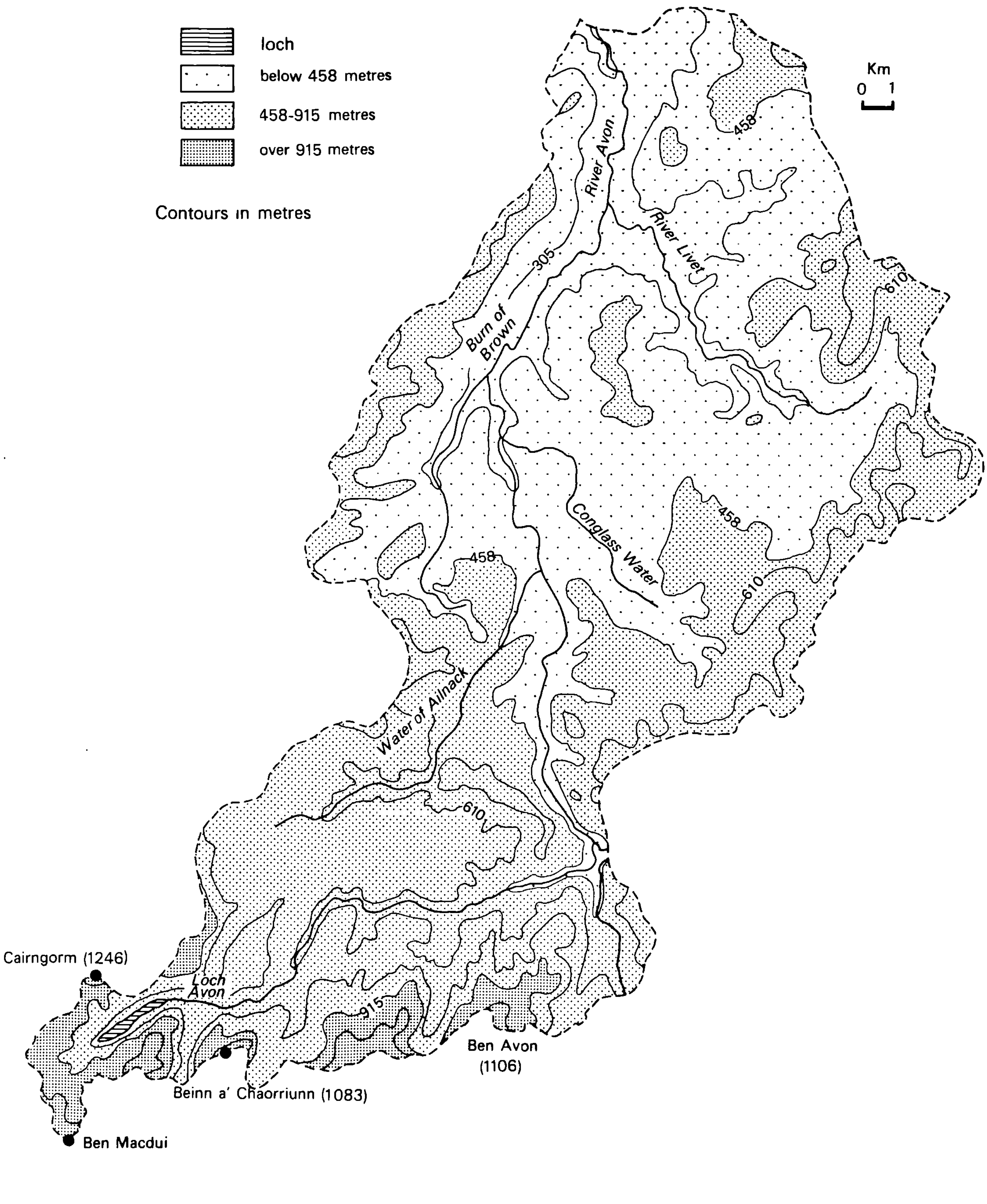
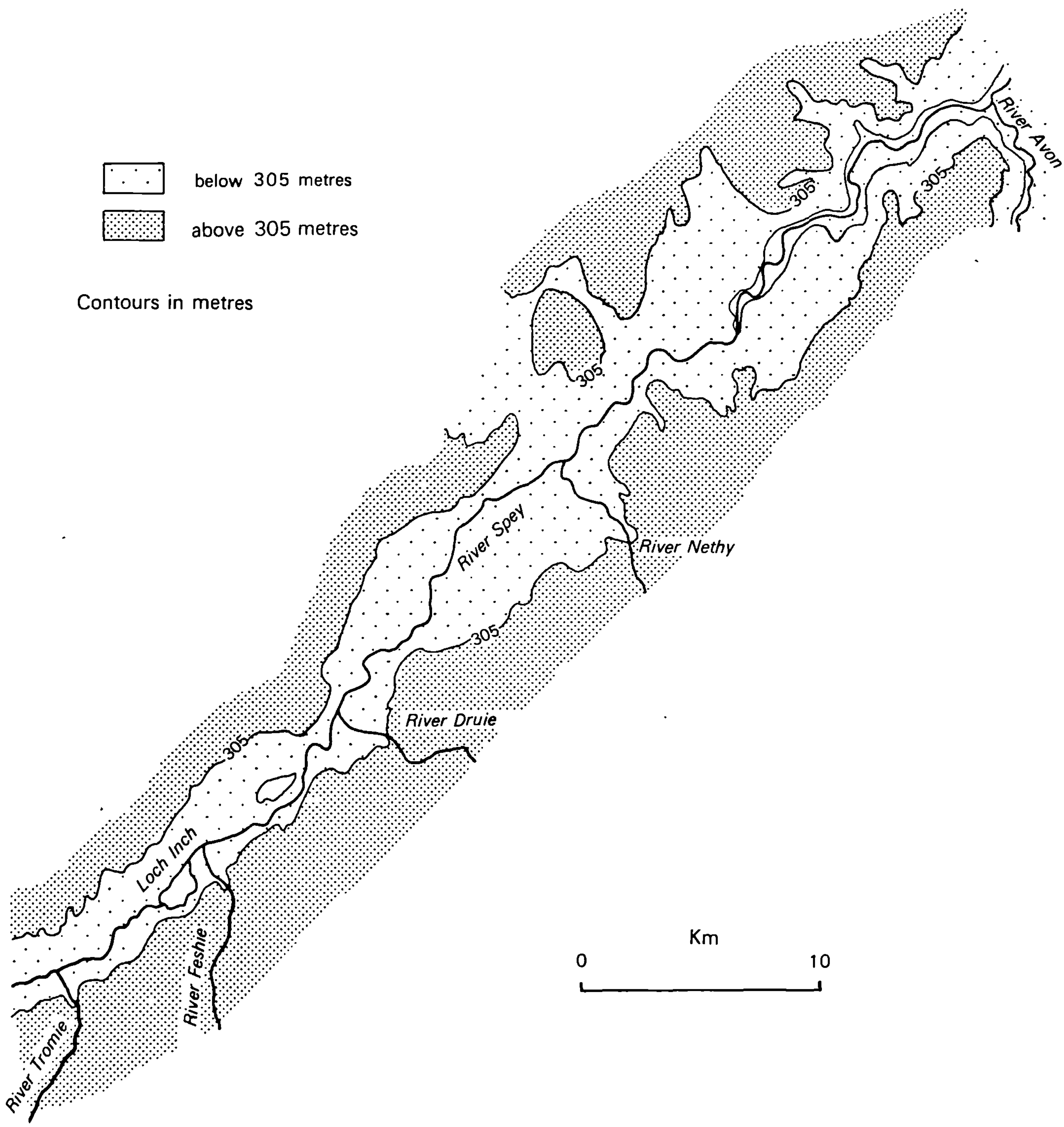


Figure 3.3.1.(i).F

The relief of the mainstream Spey



narrow and steep sided in its middle to lower reaches and thus has a limited width of floodplain. Unlike the other catchments, there are no major tributaries joining below 380 metres. The river has its source in very steep-sided alluvial basins, which are partially lake filled eg. Loch an t-Seilich (Figure 3.3.1.(i).A). The gauged catchments have the following basin characteristics, shown in Table 3.3.(i) (Source: FSR, NERC 1975). For the relevant gauge locations see both Figure 3.3.5.(i) and Section 3.3.5.

It is important to note the difference between the mainstream Spey and its gauged tributaries, with both S1085 and STRMFREQ considerably higher in the Tromie and Avon catchments. The Feshie, too, has an average gradient of 10 m km^{-1} along much of its length (Young, 1976). The digitised long profiles of the catchments are shown in Figure 3.3.1.(ii), displaying the step-like nature of the profiles and also the comparative steepness of the Druie and Nethy.

3.3.2 Spey study area: Solid and drift geology

The solid geology is much simpler than in Deeside, with larger areas of each catchment categorised by similar rock type. It is simplest to discuss the geology of each tributary catchment separately; for appropriate detail see Figure 3.3.2.(i). The mainstream Spey, within the study area, has cut deeply into the Moinian schists. This is partially because its course is closely adapted to the regional strike and follows the SW to NE Caledonian grain for much of its length (Whittow, 1977). The Tromie, too, is mainly underlain by a uniform

Table 3.3.1.(ii)

Basin characteristics above discharge gauging stations within the Spey catchment (Source: FSR, (NERC, 1975)

Station	AREA (km^2)	MSL (m)	S1085 ($m\ km^{-1}$)	STRMFREQ	LAKE
Grantown	1750.0	95.1	2.17	0.86	0.077
Boat o' Garten	1270.0	80.2	2.25	1.02	0.106
Kinrara	1010.0	64.4	2.36	1.18	0.066
Ruthven Bridge	534.0	47.2	3.07	1.24	0.022
Invertruim	401.0	37.3	3.58	1.19	0.014
Tromie bridge	130.0	22.7	9.89	2.61	0.415
Delnashaugh	544.0	59.1	10.55	1.44	0.000

Figure 3.3.1.(ii)

LONGITUDINAL PROFILES WITHIN THE SPEY STUDY AREA

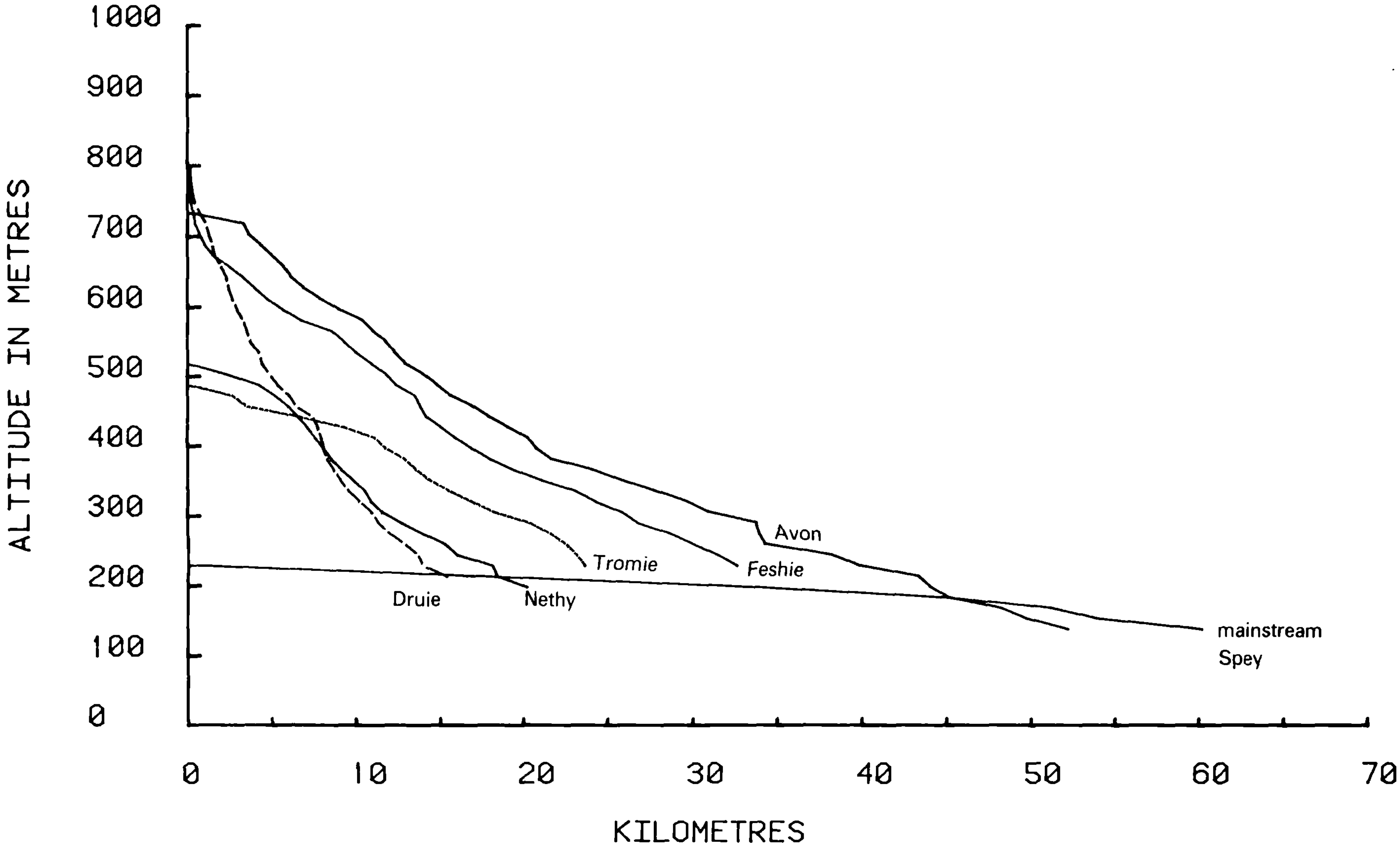
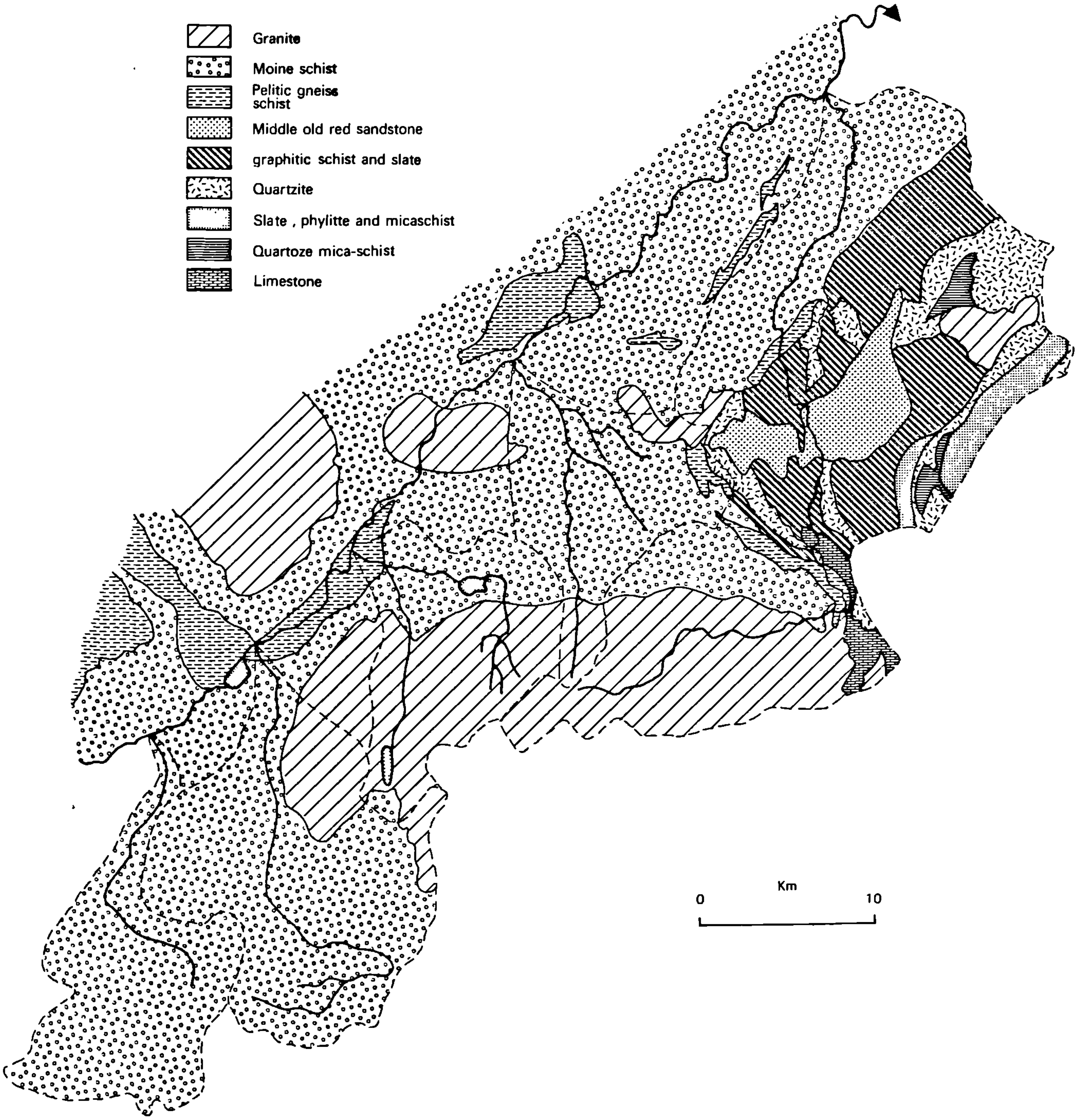


Figure 3.3.2.(i)

The solid geology of the Spey study area



bedrock comprised of Moine schist. The middle reaches of the Nethy, Feshie and Druie catchment have principally the same solid metamorphic geology. However, in contrast, the Feshie catchment is bounded to the east by the major granitic batholith that provides the solid geology for the upper Druie, Nethy and Avon and the highest peaks in the Cairngorm massif that form their upper catchments.

Large percentages of the middle to lower catchments are underlain by extensive and deep deposits of fluvioglacial material, in much more extensive quantities and depths than in upper Deeside (Figure 3.3.2.(ii).A-C). Sissons (1976) reported that sections high on the sides of the Cairngorms revealed 60 metres, sections by the Spey exposed a similar thickness and as much as 75 metres of fluvioglacial and morainic deposits were found in Glen Tromie. There is however a general thinning with altitude to around 455 m (Sugden, 1970) but there is generally no shortage of available sediment along the valley floors.

3.3.3 Spey study area: Geomorphology

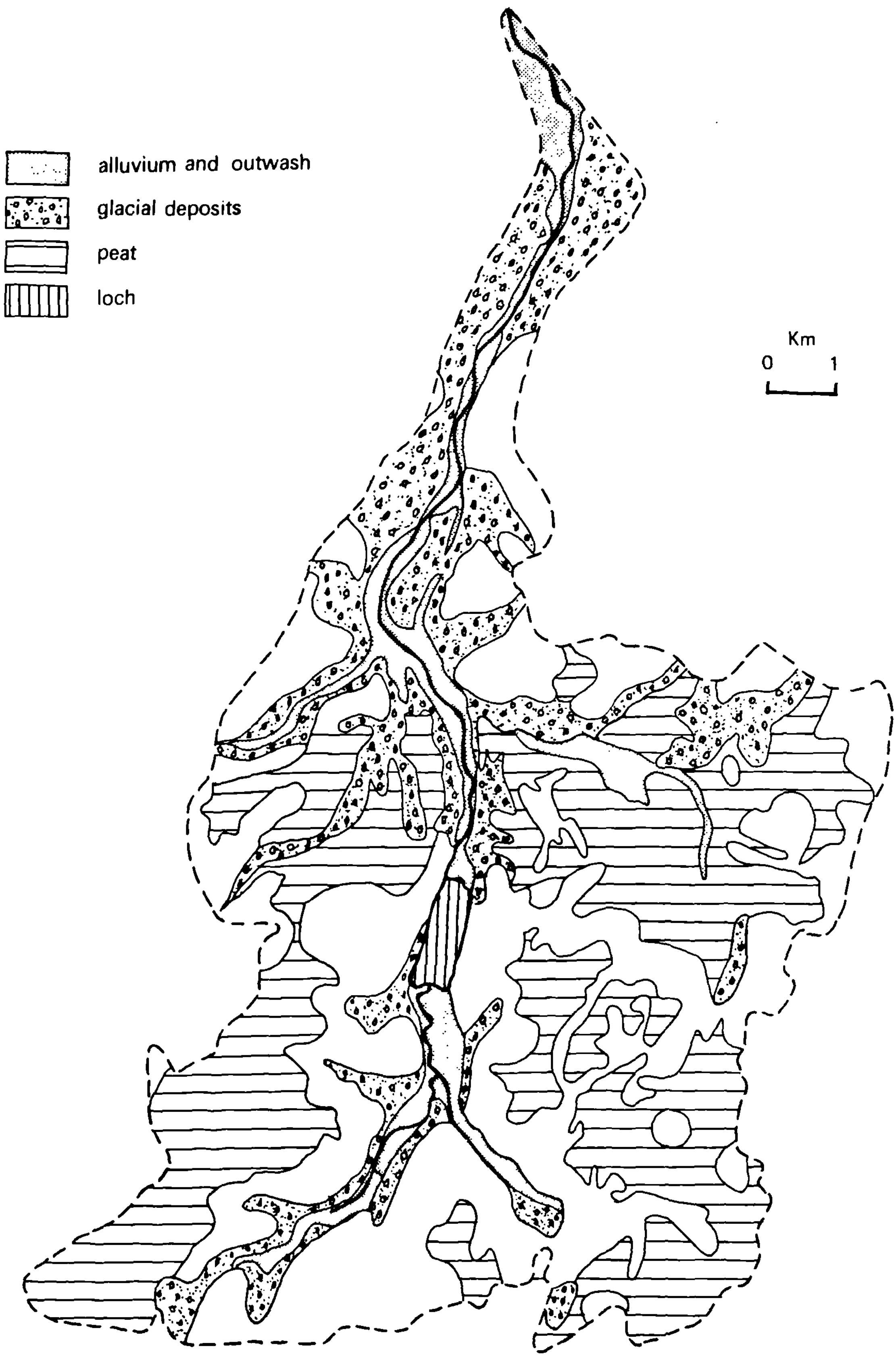
In terms of chronology, the glacial history of the Spey study area is similar to that outlined for upper Deeside. During the Devensian, the main Spey glacier had several distinct branches with contributing valley glaciers eg. the Glenmore glacier (Young, 1974). Again, this has left a series of erosional and depositional features that effect different glens to varying extents. Within Glenmore, Abernethy and Glen Feshie, the impact of ice wastage has been discussed in detail by Young (1974; 1975a; 1975b) and in the upper reaches, there are abundant features associated with glacial erosion, such as glacial troughs, breaches and corries. Thus, the upper Feshie flows to the east but, when about to join up with the upper Geldie as it did preglacially, the Feshie abruptly turns away and descends down through a large glacial breach cut through the former watershed (Linton, 1949). By this means the Feshie captured the former head waters of the upper Dee (Figure 3.3.3.(i)). Another example of a glacial diversion of drainage occurs on the Water of Caiplich on the Avon (Linton, 1954).

Glacial troughs with steep sided rock walls occur in Glen Einich, Loch Avon and the upper Feshie basin. There are also numerous meltwater channels, for example those carrying meltwaters from the downwasting Glenmore glacier were cut in a series approximately parallel to the ice margins. Within the Nethy catchment, there is one of the best examples of a cliff-sided glacial meltwater channel in Scotland, the Eag Mor. Glacial erosion has also carved out wide U-shaped valleys of varying sizes in the middle reaches of valleys eg. the Avon and the Feshie. Important features of glacial erosion are shown in Figure 3.3.3.(i). In terms of recent slope modification, slow periglacial activity has been documented on the slopes (King, 1971; Sugden, 1971), but this is a negligible contributor to channel sediment supply.

Massive fluvioglacial accumulations of considerable depth, associated with icesheet decay, are found in all catchments (Figure 3.3.2.(ii).A-C). Only a few examples can be highlighted here eg. middle to lower Glen Feshie (Figure 3.3.2.(ii) B). These deposits were laid down by meltwater streams of much greater competence than present-day rivers, as evidenced by the large clast sizes (over 1 m

Figure 3.3.2.(ii).A

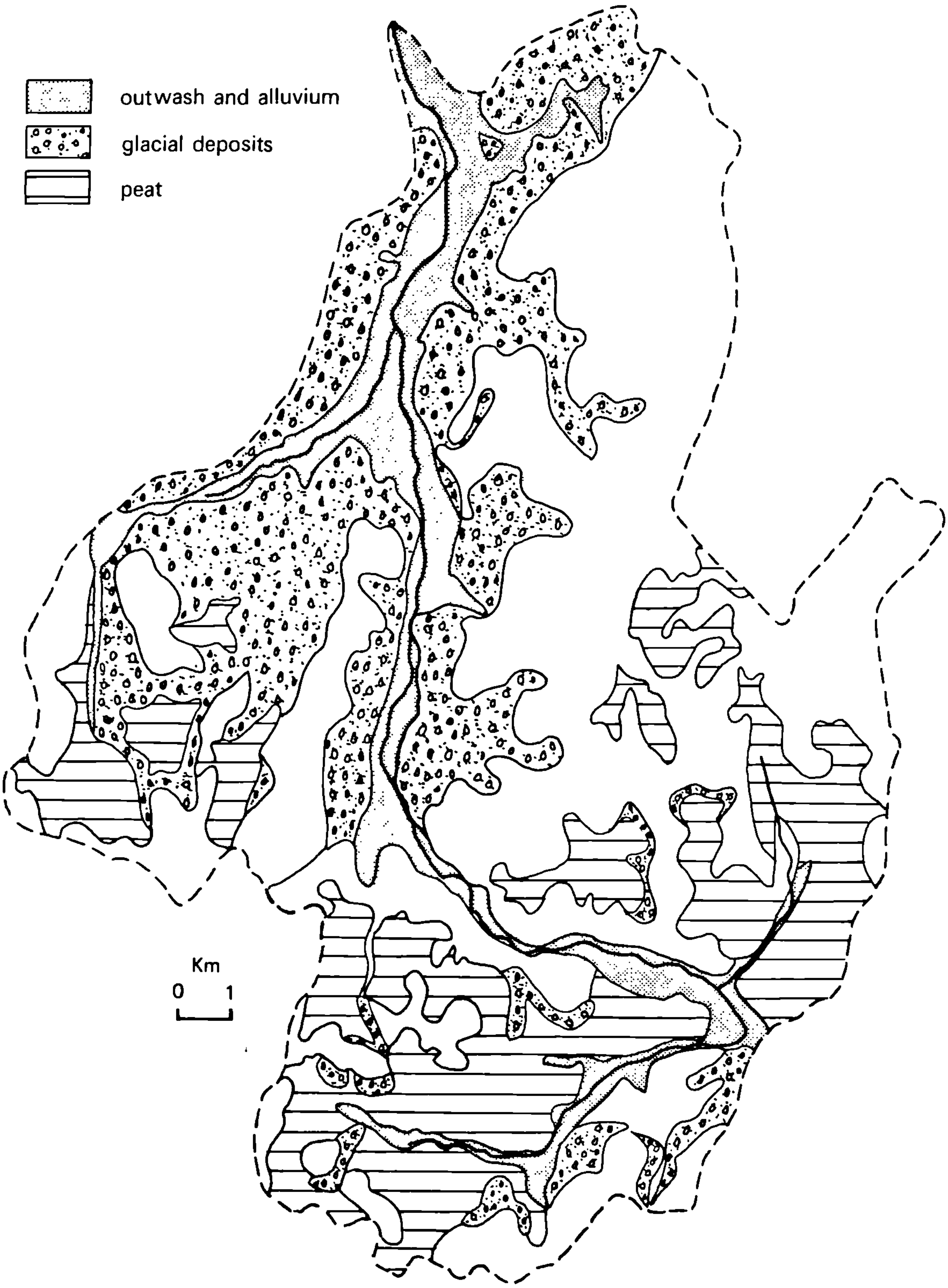
The drift geology of the Tromie catchment



(Source: Geological survey of Scotland 1:63360)

Figure 3.3.2.(ii).B

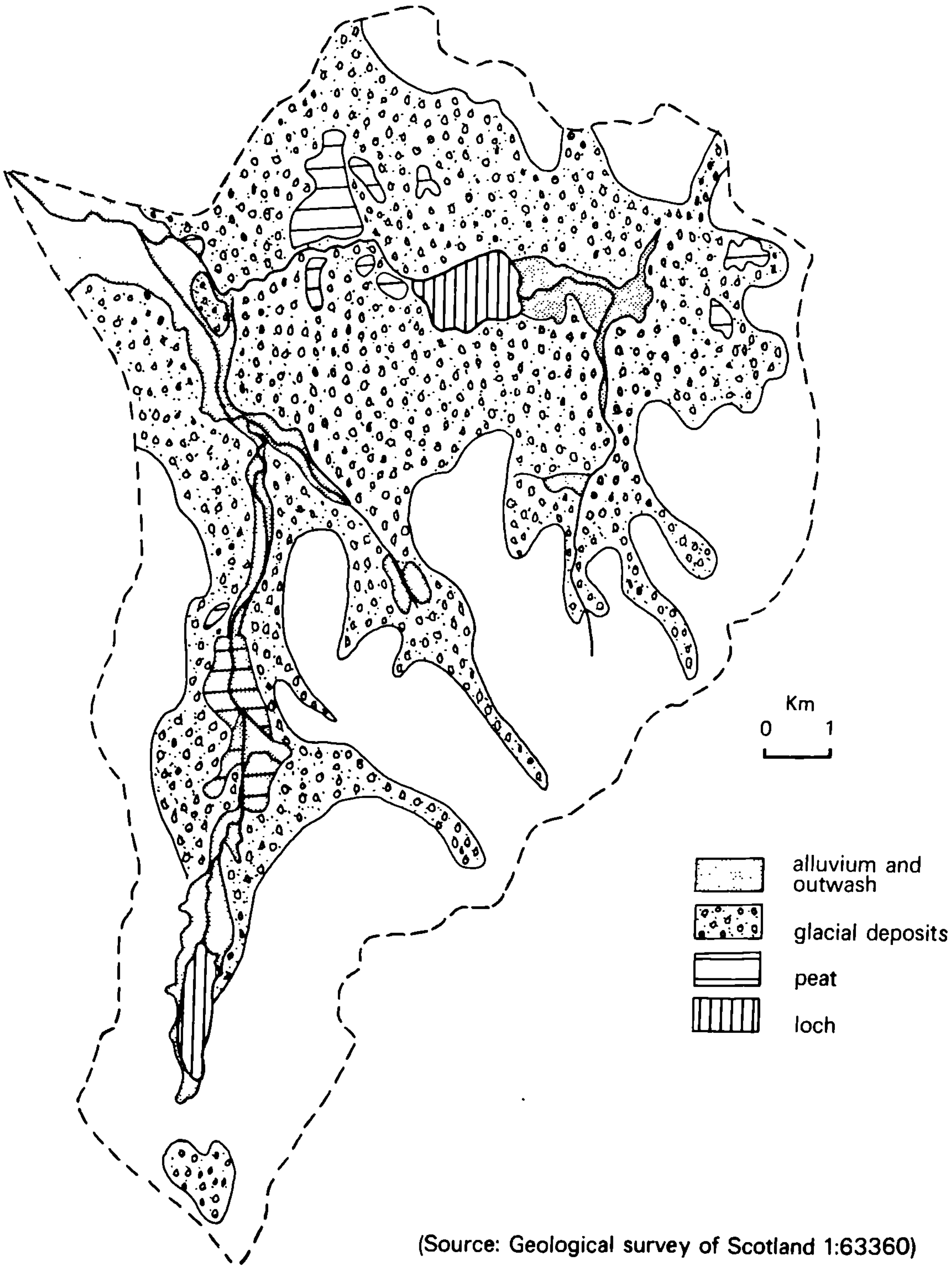
The drift geology of the Feshie catchment



(Source: Geological survey of Scotland 1:63360)

Figure 3.3.2.(ii).C

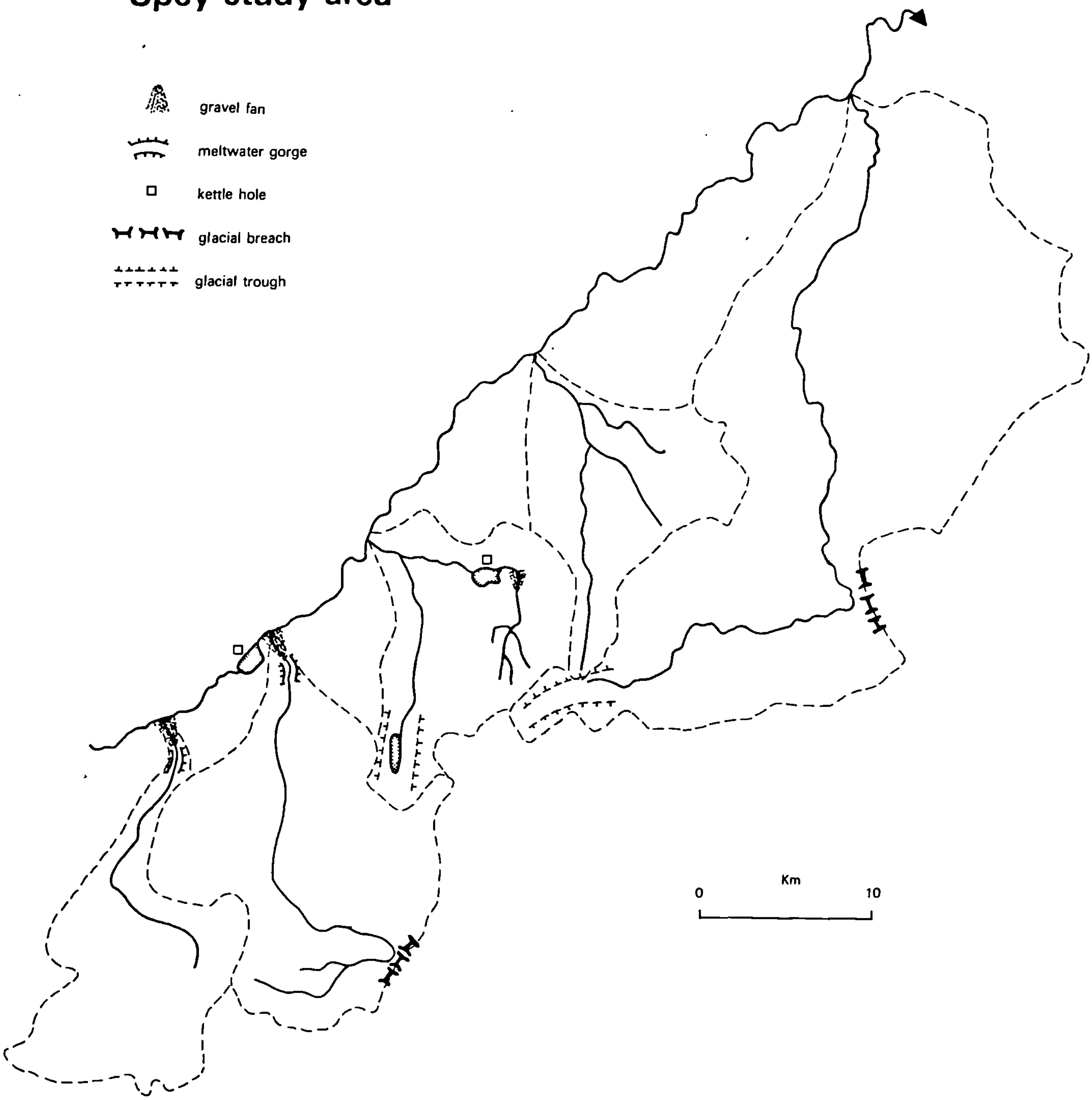
The drift geology of the Druie catchment



(Source: Geological survey of Scotland 1:63360)

Figure 3.3.3.(i)

Selected geomorphological features within the Spey study area



intermediate axis), embedded in some deposits. Frequently such deposits have been incised to form complex terrace sequences, which date from the Lateglacial to the present day. Such a terrace sequence in Glen Feshie was analysed by Young (1975a; 1976), who speculatively identified two stages of terrace development; one associated with the meltout of the Devensian icesheet and a second, associated with the meltout of the zone III valley glacier. An almost continuous suite of terraces, never more than 5 m above modern stream level and representing "the most recent adjustments of the contemporary river Feshie to erratic conditions of hydrology and bedload" (Young, 1975a p100), is implied to be of Holocene age. In that particular example, the fluvioglacial material has been eroded down to bedrock in only two places. In Strath Nethy and Glenmore, there are also extensive kame terraces and eskers.

Other important features associated with deglaciation, which have a lasting impact on the fluvial system, include Loch Morlich, a large kettle hole derived from the Glenmore icesheet (Figure 3.3.3.(i)). In the Nethy basin, in the lower lying areas beyond the Eag Mor, fluvioglacial landforms dominate the topography with such features as kames, eskers and kettle holes. Such features are typical of the middle reaches of all of the catchments to greater (Glen Feshie) or lesser (Glen Tromie) extents. Fluvioglacial sediments incorporate both granite and schist in varying proportions, with contrasts in both particle size and shape reflecting their diverse origins. Overall particle sizes range from boulders to fine sand. Rock controlled meltwater gorges exist above the confluences of both the Feshie and the Tromie. Large, low gradient, gravel fans spread out below these gorges, associated with the reduction in meltwater competence (see Figure 3.3.3.(i)); remnants of the fossil fans of much greater extent that existed during deglaciation. For example, the depositional history of the Tromie fan is detailed below:

> "At the mouth of the Tromie, a 500 m wide by 1250 m long outwash spread lies 10-15 m above present day river level. Kettle holes pit the surface of the outwash fan, which is also scored by the casts of braided rivers that once ran across it, leaving 2-3 m deep depressions. The Tromie skirts the eastern edge of the outwash fan and has cut a 10-12 m gorge at Tromie bridge. Only remnants of the fan lie east of the Tromie and form a series of terraces that are separated by small risers. Downstream of the Tromie gorge, a second, clearly convex fan has been spread out into Strathspey on both sides of the Tromie. Like the massive fan at the mouth of the Feshie (Young, 1976), the lower level Tromie fan is also an abandoned feature that has only been slightly eroded by the later rivers" (Young, 1978 p89).

Within the glacially deepened and broadened Spey valley, with its extensive reserves of sands and gravels, there is a broad sequence of fluvioglacial terraces and dead ice kame and kettle topography eg. Loch Insh and Loch Alvie (see Young, 1975c). Above Loch Insh, the gradient of the Spey is very low, and is explained by the former existence of a 13 km ribbon lake from near Alvie upstream. However, this was infilled by sediments and now, in a considerably reduced form, only Loch Insh remains (Whittow, 1977). This constitutes a very important local baselevel.

3.3.4 Spey study area: Climatic conditions

The Spey study area has numerous raingauges at a range of altitudes. The average annual rainfall totals (1940-1971), from a selected number of these gauges, are shown in Table 3.3.4.(i). Locations of the raingauges are shown in Figure 3.3.4.(i). Steven and Carlisle (1959) give the average annual rainfall for Rothiemurchus and Glen Feshie forests as 1143 and 1143-1270 mm respectively. Thus over 1000 mm per annum must be expected in the tributary catchments. Figure 3.2.4.(i) also shows the impact of altitude on the annual rainfall in this area, with a raingauge fortuitously placed on the summit of Cairngorm (1090 m) recording an average annual rainfall total of 2050 mm. The pronounced, steep, almost linear relationship between average annual rainfall and altitude can be seen.

3.3.5 Spey study area: Hydrology

The Spey is gauged by the N.E. River Purification Board at several locations along its main channel and at single locations along certain of the study tributaries (Figure 3.3.5.(i)). Within the study area, the mainstream Spey has 5 discharge gauging stations, with 2 further downstream (Aberlour and Boat o' Brig). The Tromie, Nethy and Avon are all gauged in their lower reaches. The Nethy station is not equipped with a cableway, so that flows above approximately 5 $m^3 s^{-1}$ are obtained by extrapolation. The Feshie was gauged at Feshiebridge by D.A.F.S. from 27/7/1951 to 28/9/1970 and jointly, St Andrews and Stirling universities have maintained a gauge upstream since 1978. The only

Figure 3.3.4.(i)

The location of studied long-length raingauge sites within the Spey study area

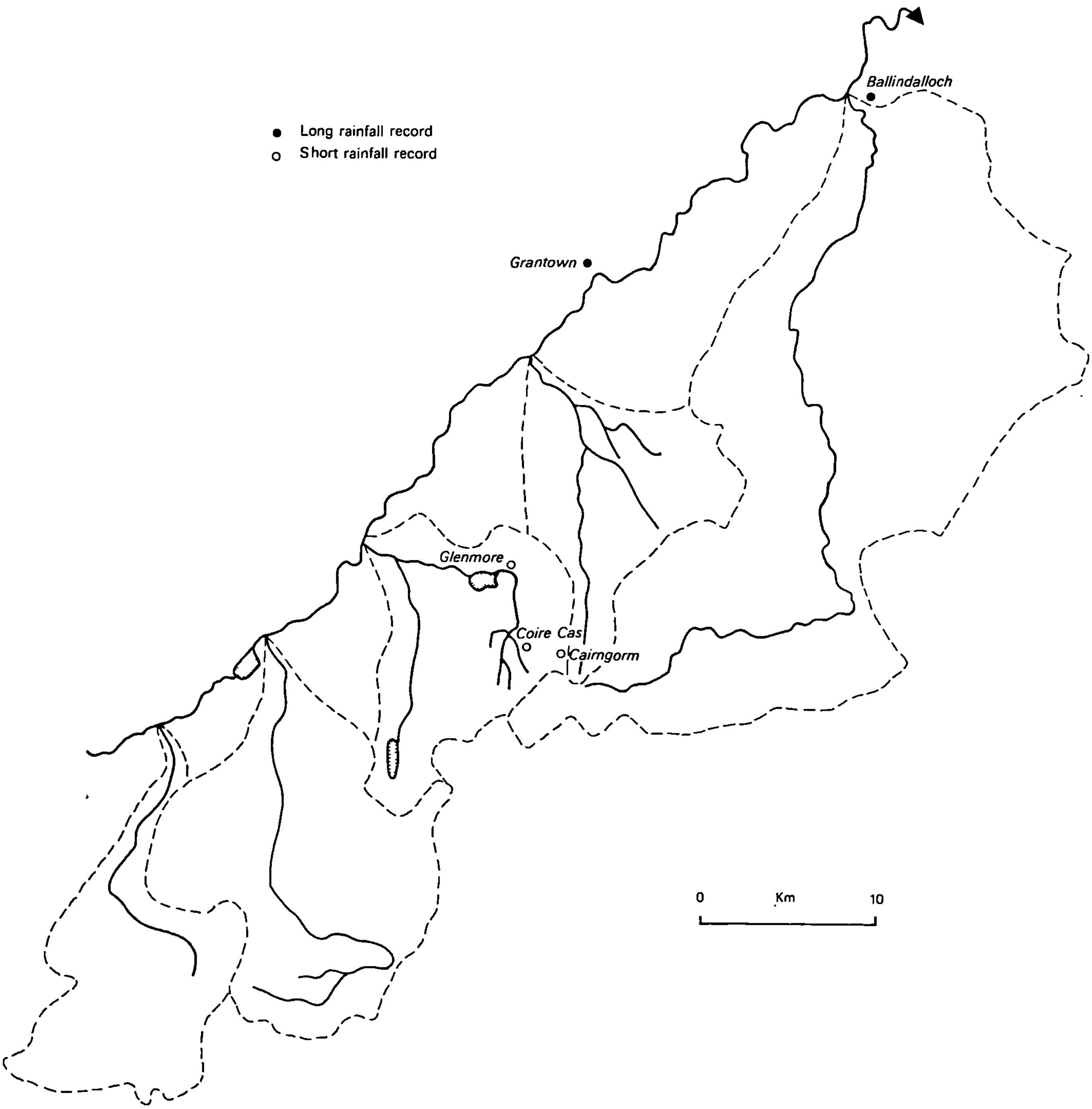


Table 3.3.4.(i)

Average annual rainfall totals for stations within the Spey study area (1941-1970)

Station	Grid Reference	Altitude (m)	Rainfall (mm)
Cairngorm	3005 8049	1090	2050
Coire Cas	2995 8054	762	1600
Glenmore	2986 8095	341	1077
Kingussie	2755 8007	224	830
Grantown	3028 8274	219	793
Ballindalloch	3185 8369	192	883
[Fochabers	3350 8595	32	755]

Figure 3.3.5.(i)

The location of discharge gauging stations within the Spey study area

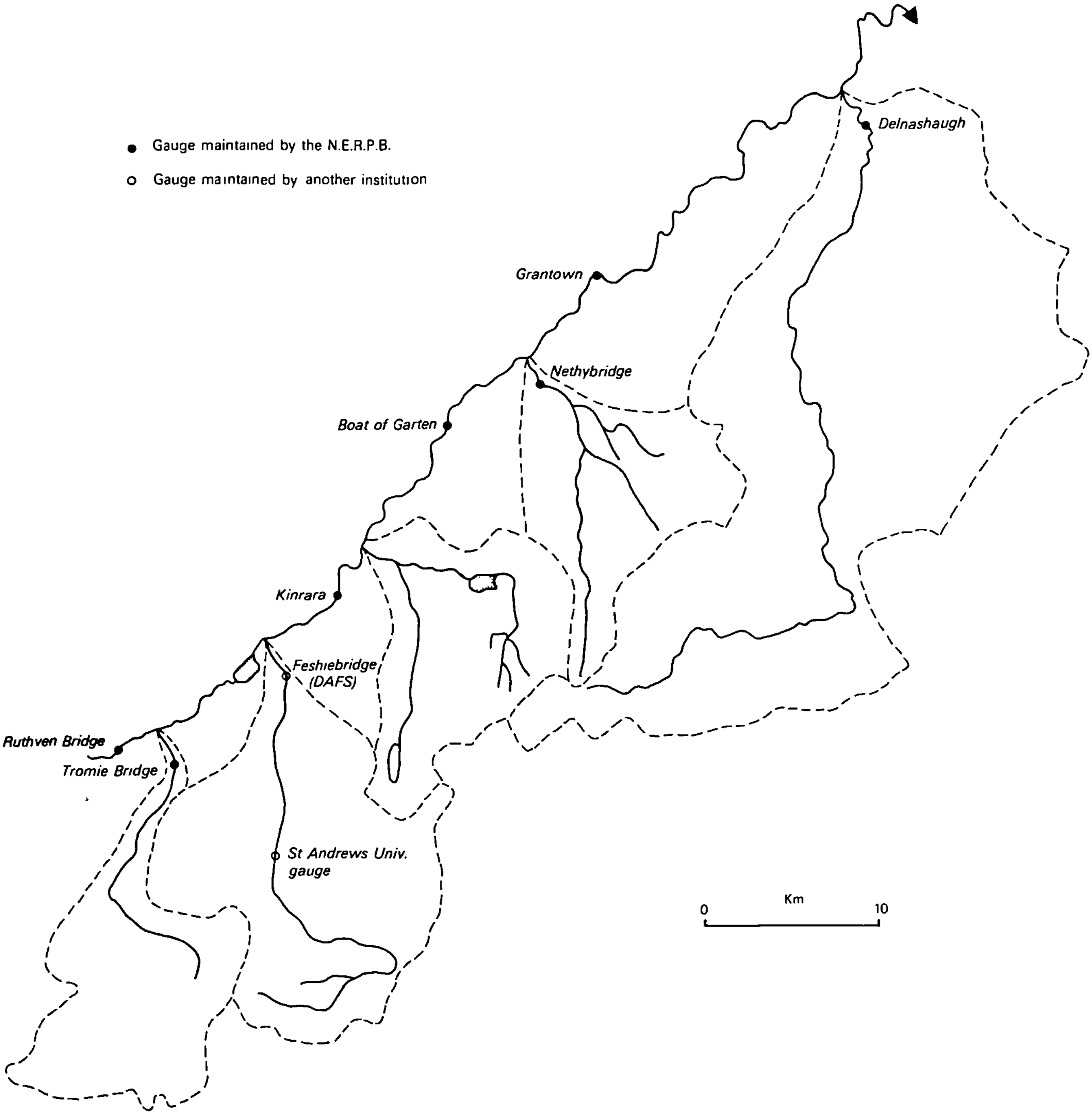


Table 3.3.5.(i)

Discharge gauging stations on the River Spey

River	Station	Grid Reference	Area (km^2)	Dates of record
Spey	Boat o' Brig	NJ 318518	2861	10/8/1952-31/12/1982
Spey	Aberlour	NJ 278439	2640	1/9/1938-31/12/1938
Spey	Grantown	NJ 034268	1750	29/11/1951-31/12/1982
Spey	Boat of Garten	NH 946191	1270	29/8/1951-31/12/1982
Spey	Kinrara	NN 881082	1010	7/8/1951-31/12/1982
Spey	Ruthven bridge	NN 759996	534	6/8/1951-10/3/1974
Spey	Invertruim	NN 387962	401	16/9/1952-31/12/1982
Tromie	Tromie Bridge	NN 789995	130	8/9/1959-30/9/1970
Nethy	Nethy Bridge	NN 999207	101	6/6/1975-2/11/1982
Avon	Delnashaugh	NJ 186352	543	3/8/1951-31/12/1982

Table 3.3.5.(ii)

Summary statistics for discharge gauging stations on the River Spey

Station	Mean flow ($m^3\ s^{-1}$)	Mean annual flood ($m^3\ s^{-1}$)	Maximum instantaneous flow ($m^3\ s^{-1}$)
Aberlour	****	430.9	1241.8
Boat o' Brig	63.2	468.6	929.8
Grantown	35.1	249.5	487.5
Boat of Garten	27.4	176.9	410.8
Kinrara	20.5	141.1	387.8
Ruthven Bridge	****	107.7	223.3
Invertruim	5.6	106.2	256.9
Delnashaugh	14.4	232.3	521.3
Tromie Bridge	2.4	73.4	155.1

**** unavailable data

tributary, which has no discharge data, is the Druie, as the river bed was found to be too active in its lower reaches to calibrate a rating curve. The details of each station and the available length of record are shown in Table 3.3.5.(i). The FSR (NERC, 1975) records soil moisture deficits of 5.0 to 5.8 mm and an RSMD of 40-60 mm. Figure 3.2.5.(ii) shows winter rain acceptance potential with high acceptance levels (categories 2 and 3) in the Spey valley, but lower values (categories 4 and 5) on the higher ground of the tributary catchments. The importance of this in terms of winter flooding will therefore be assessed later.

3.3.6 Spey study area; Landuse

Bioclimatic change is reflected in known tree-line fluctuation (Pears 1967, 1975) with pine growing at altitudes of between 750 and 850 m in the Cairngorms between 6,000 and 4,000 BP. More recently, landuse within the Spey study area has undergone a similar sort of deforestation as in upper Deeside. The natural Caledonian forests, though considerably reduced, are still more widespread than in most other areas of Scotland (Steven and Carlisle, 1959) and there has been Forestry Commission planting in several areas eg. Inchriach forest in Glen Feshie. There is also some agriculture, mainly pastoral, on the lower valley slopes and floodplains/ river terraces. The upper valley slopes are frequently heather covered, with grouse and deer hunting again the major estate interests. In certain areas, there are extensive blanket peat bogs eg. in the upper Nethy catchment, by the Faesheallach Burn, and the Allt Mor catchment (Figure 3.3.2.(ii).A-C). Landuse change over the past 200-300 years is studied in much more detail in Chapter 6.3.

3.4.1 Tweed study area: Catchment characteristics

In comparison to the Deeside and Speyside catchments, those of the Tweed study area are considerably larger. The entire Tweed basin has been likened to a horseshoe, slightly tilted to the north-east, and draining an area of approximately 5000 km^2. The study catchments represent the four main Scottish tributaries entering the Tweed furthest downstream (see Figure 3.1.(i)). The rivers Leader and Whiteadder, draining the Lammermuirs with highest hills of over 500 m (eg. Meikle Says Law, 535 m), flow south to join the mainstream Tweed. In contrast, the Teviot and Bowmont Water (a tributary of the Till) flow predominantly northwards and have very different higher source areas within the Cheviot Hills. Within this massif, the highest peak is the Cheviot (815 m), and tributaries of the Teviot (such as the Jed and Kale) and the Till (eg. Bowmont Water) have deeply dissected this upland area (see Figures 3.4.1.(i).A-E).

The summary basin characteristics for the Tweed study area are collated in Table 3.4.1.(i). It is notable that the relief ratio in these basins is considerably lower than in either of the other two study areas. From the FSR (NERC, 1975), the following additional catchment characteristics can be collated (Table 3.4.1.(ii)). Slopes equivalent to the Speyside tributaries are only maintained in the upper catchments eg. the Teviot above Hawick. The longitudinal profiles for the mainstream in each catchment are shown in Figure 3.4.1.(ii). These show

Figure 3.4.1.(i).A

The relief of the Leader catchment

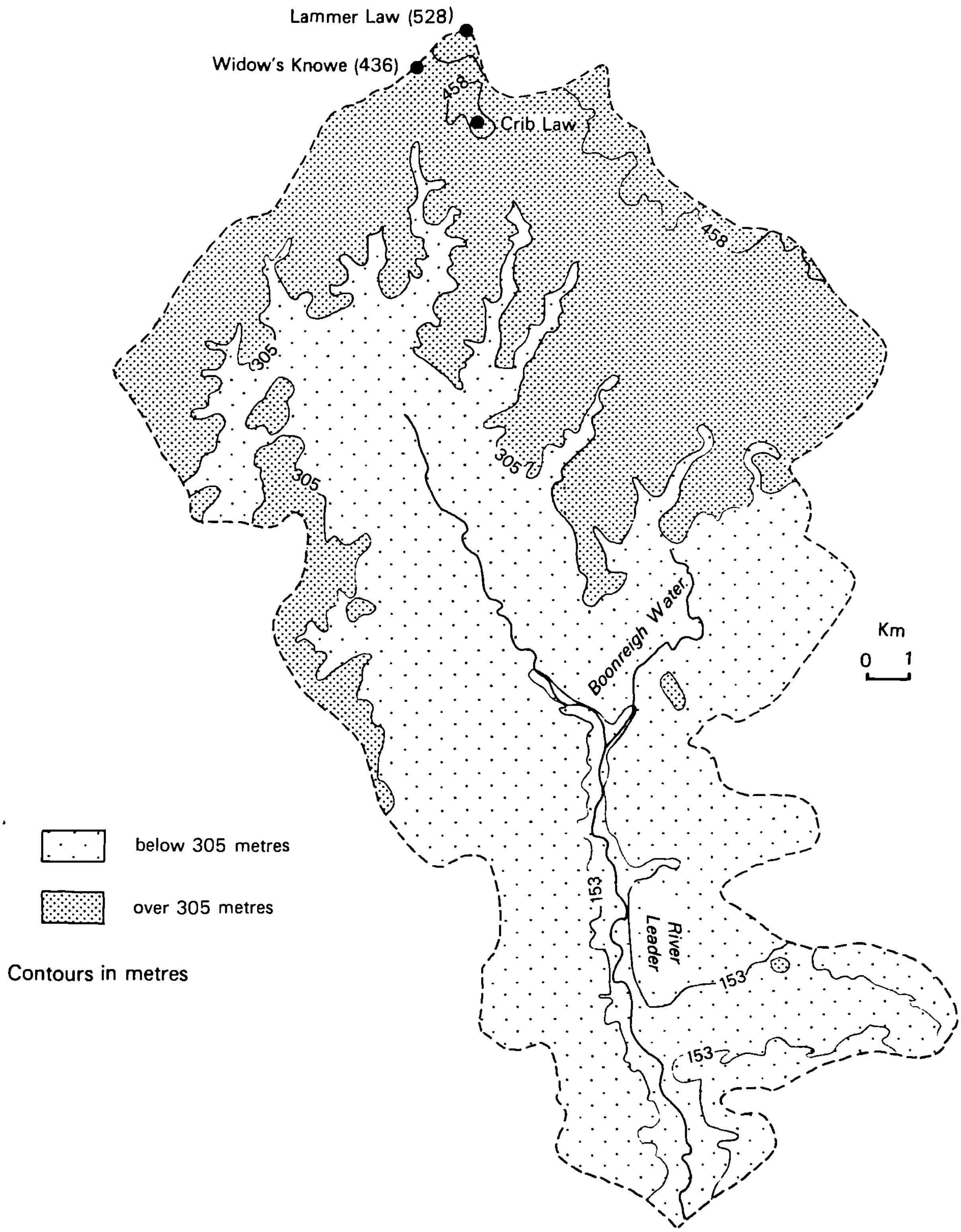
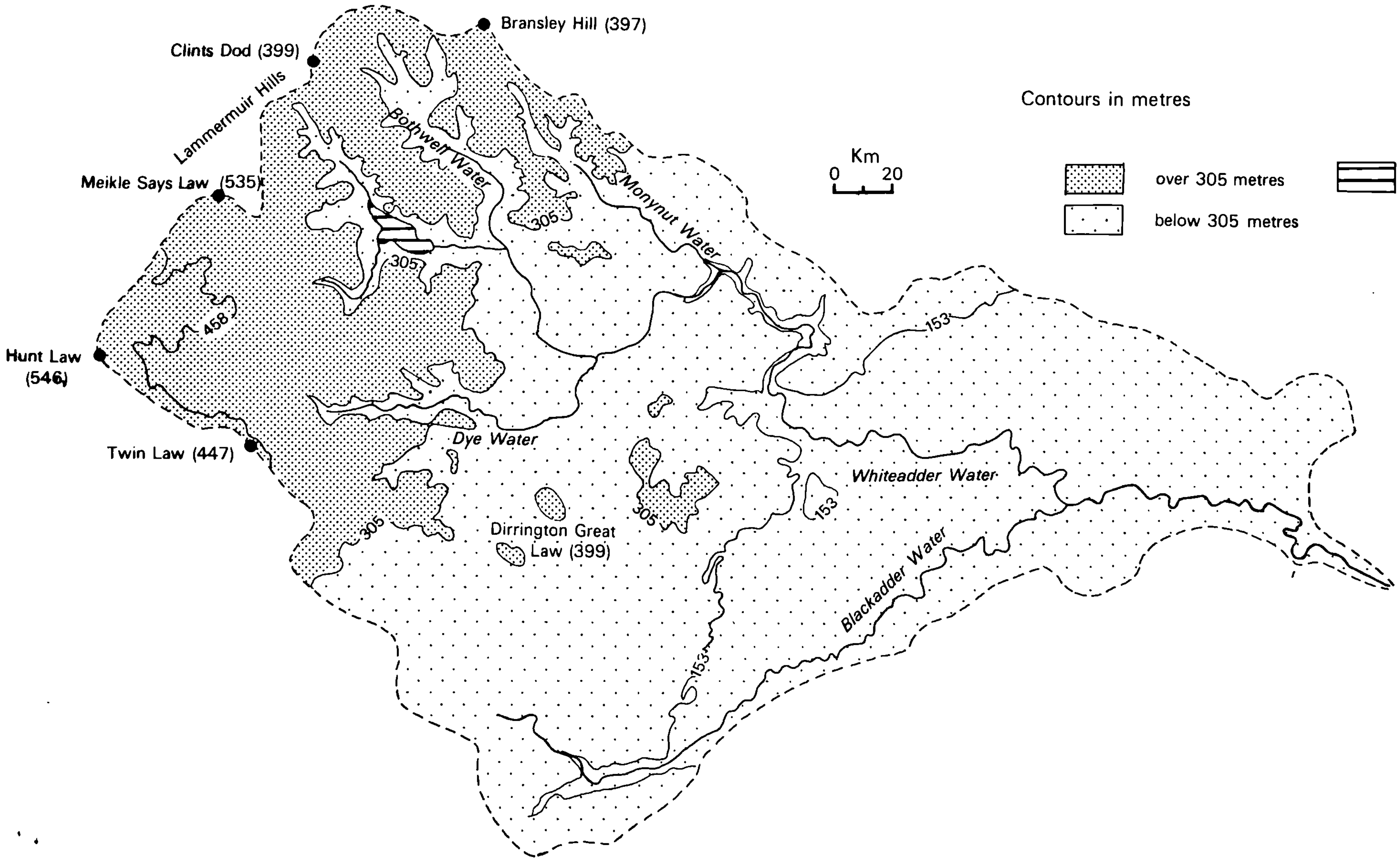


Figure 3.4.1.(i).B The relief of the Whiteadder catchment



The relief of the Teviot catchment

Figure 3.4.1.(i).C

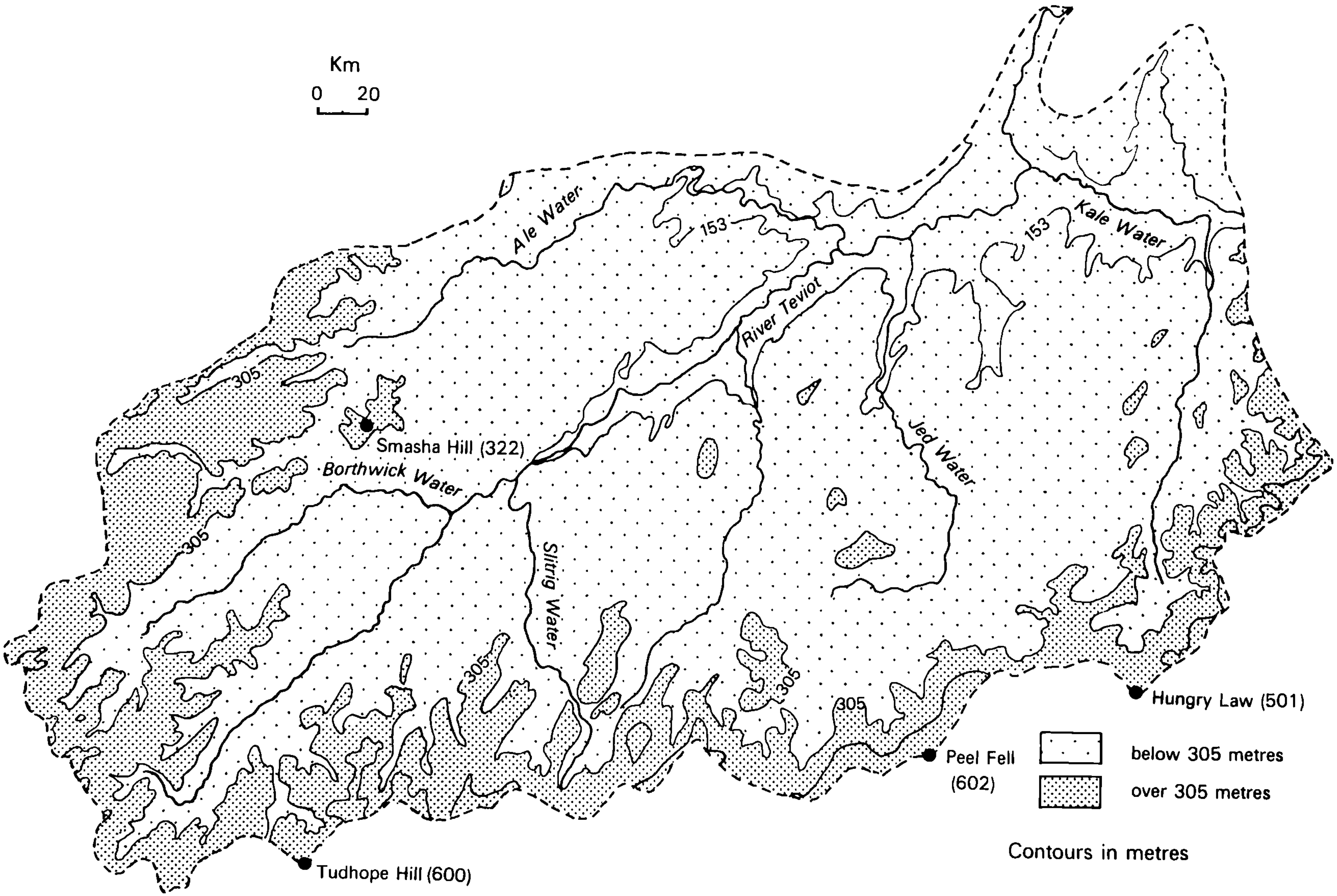
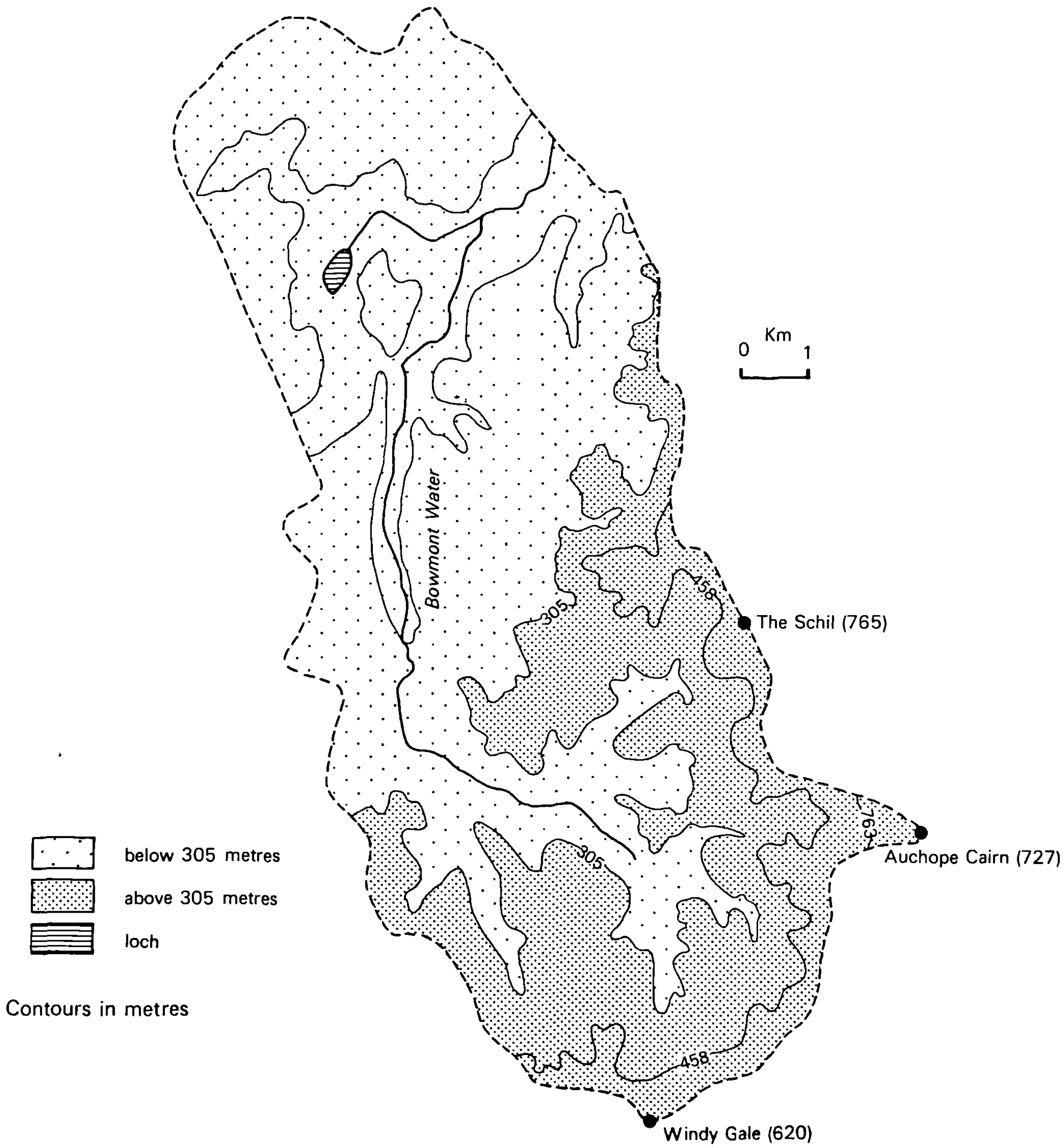


Figure 3.4.1.(i).D

The relief of the Bowmont catchment



The relief of the mainstream Tweed

Figure 3.4.1.(i).E

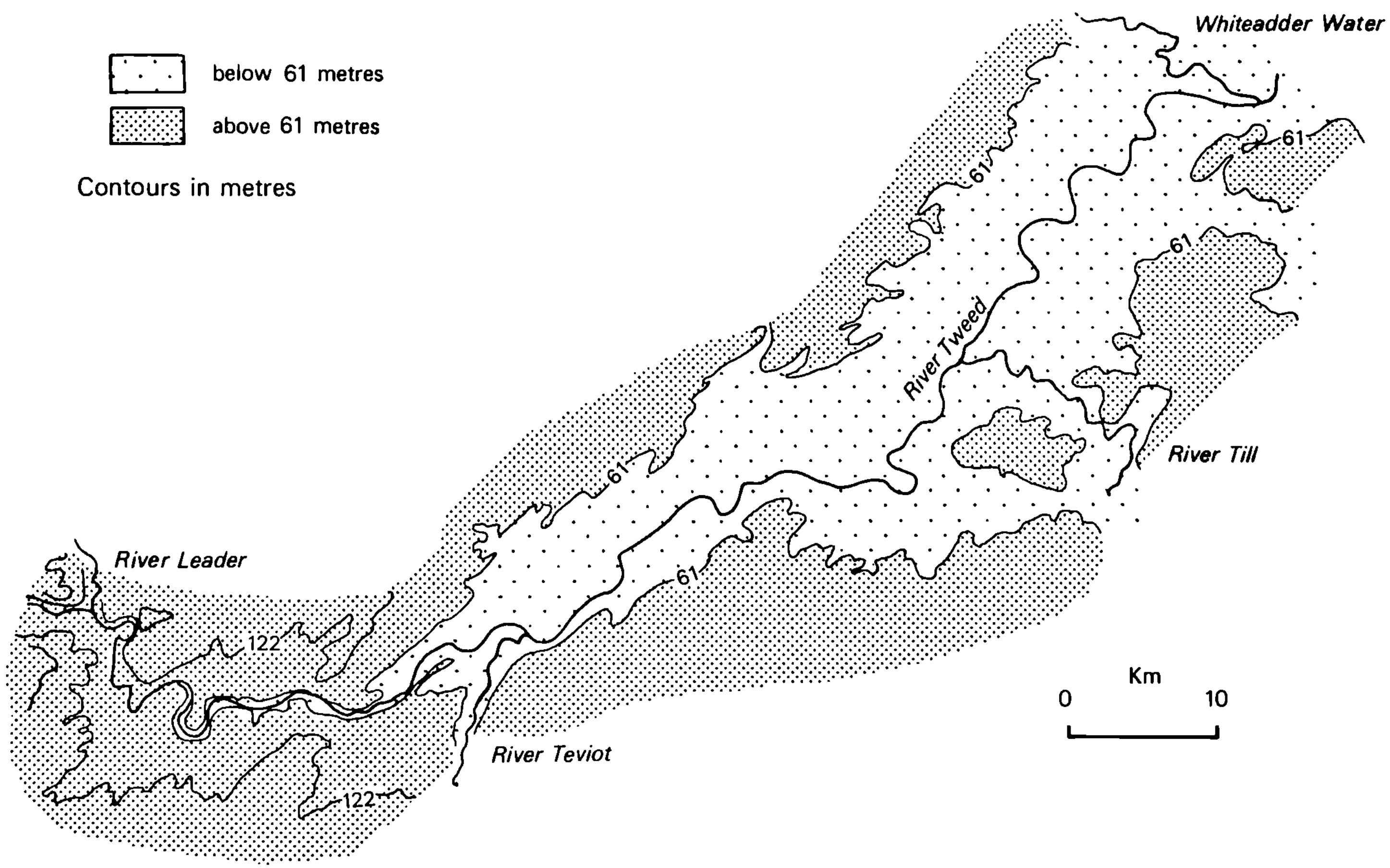


Table 3.4.1.(i)

Tweed study area: Catchment parameters

Catchment	Basin Area (m^2)	Max. Ht. (m)	Min. Ht. (m)	Basin Relief (m)	Basin Length (km)	Relief Ratio (m km^{-1})
(a) Whiteadder	377.3	534	35	499	40.8	12.2
Blackadder	142.6	399	61	338	24.3	13.9
(b) Leader	274.8	528	85	443	27.0	16.4
Boonreigh	55.0	447	143	304	11.0	27.6
(c) Teviot	1109.1	609	32	577	52.5	11.1
Kale Water	141.6	562	52	510	21.8	23.4
Ale Water	170.0	383	58	325	29.5	11.0
Borthwick	91.0	476	99	377	18.5	20.4
Slitrig	62.5	609	107	502	13.8	36.4
Jedburgh	145.4	549	61	488	23.8	20.5
(d) Bowmont Water	100.0	738	76	662	14.5	45.7

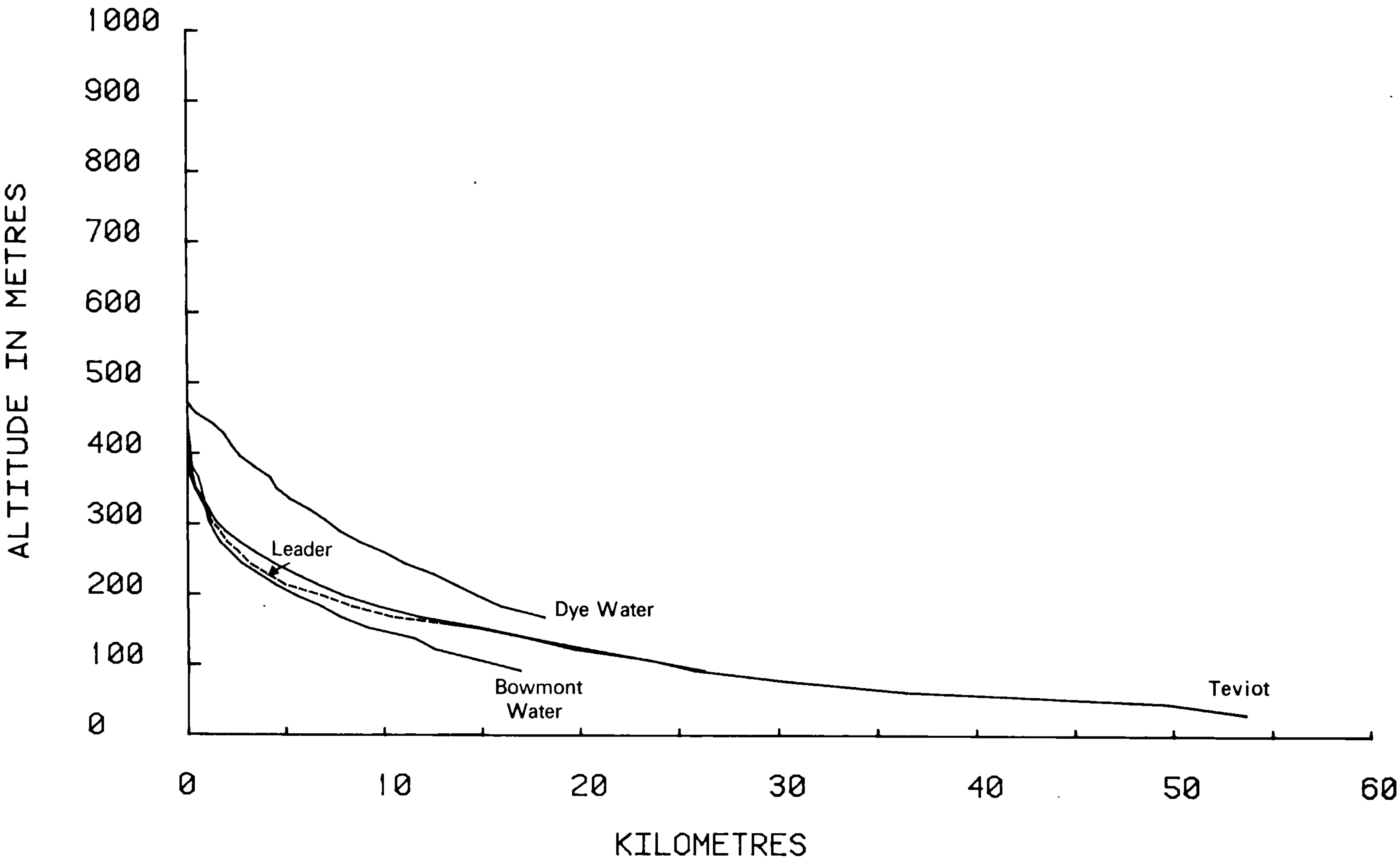
Table 3.4.1.(ii)

Tweed basin characteristics from the FSR (NERC, 1975)

Station	Area	MSL	S1085	STRMFREQ	LAKE
Earlston	239.0	29.3	9.47	1.13	0.000
Hutton Castle	503.0	48.9	5.39	1.14	0.003
Hungry Snout	45.6	13.5	15.05	1.49	0.857
Mouth Bridge	159.0	34.0	5.02	0.81	0.000
Ormiston Mill	1110.0	48.5	3.81	1.44	0.002
Hawick	323.0	22.4	10.17	2.11	0.010
Ancrum	174.0	46.0	5.30	1.40	0.009
Jed	139.0	66.1	3.39	1.45	0.000
Etal	648.0	11.2	14.2	5.23	0.000
Kirknewton	199.0	----	4.2	2.09	0.063
Lyneford	373.0	35.9	4.68	1.89	0.089
Peebles	694.0	42.6	3.95	1.71	0.056
Boleside	1500.0	75.2	2.43	1.82	0.226
Dryburgh	2080.0	89.8	2.24	1.60	0.081
Sprouston	3330.0	113.0	1.61	1.50	0.065
Norham	4390.0	140.3	1.62	1.41	0.046

(Kirknewton data source: Batley, 1983)

Figure 3.4.1.(ii) LONGITUDINAL PROFILES WITHIN THE TWEED STUDY AREA



a more uniform long profile, in comparison with the Dee and Spey study areas.

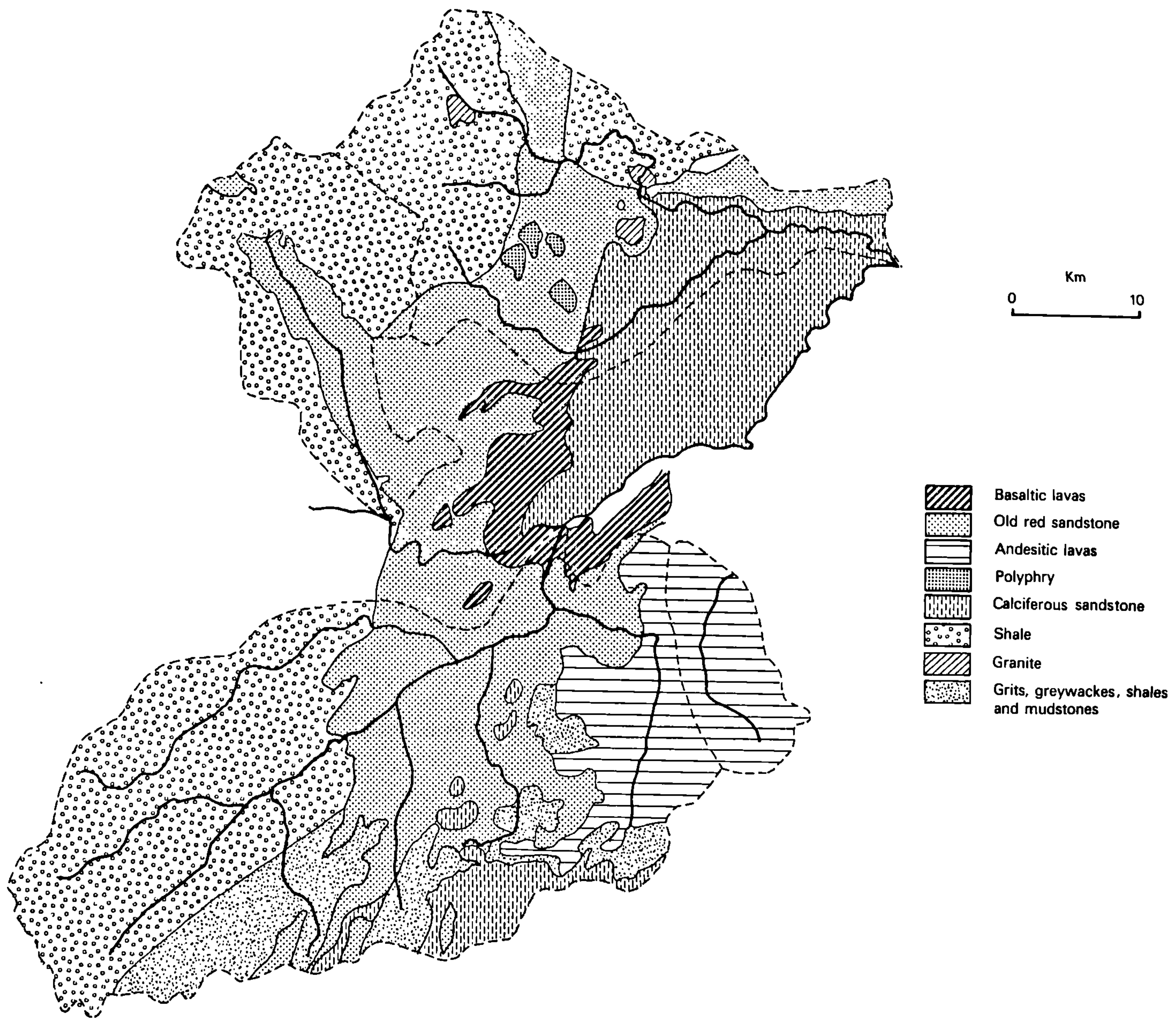
3.4.2 Tweed study area; Solid and drift geology

The solid geology of each catchment is shown in Figure 3.4.2.(i) and it is simplest again to discuss each catchment separately. The mainstream Tweed below Kelso is scarcely in contact with the underlying calciferous sandstone, which is buried beneath a mantle of glacial drift (cf. the outcrops of bedrock above Kelso). The Whiteadder/ Blackadder catchment has greywackes and sandstone at, or near, the surface within most of its catchment, giving rise to the convex slope profiles typical of the Lammermuirs and the Southern Uplands more generally. The lithology has important implications in terms of clast size:

> "Weathering of these rocks rarely results in the production of boulders [cf. Speyside] but instead produces mainly small flattish stones and lesser debris that contribute further to the rounded appearance [of the clasts]". (Sissons, 1976 p8)

There are also localised igneous intrusions eg. within the Whiteadder catchment (see Figure 3.4.2.(i)). The Leader valley is an ancient trough from which the tongues of upper old red sandstone strata have not yet been eroded. Conglomerates predominate, composed chiefly of the rounded greywacke pebbles described above. The main geological

Figure 3.4.2.(i) **The solid geology of the Tweed study area**



feature to the south is the massif of granite and andesitic lava flows of the Cheviots. Within the upper Teviot catchment, the Cheviot massif is predominately andesitic lavas but, in its middle reaches, it has cut a deep trench into Silurian rocks. Teviotdale also has prominent conical hills eg. Minto Hill, carved from a variety of igneous rocks formed during the Carboniferous. The main impression of the Teviot catchment's solid geology is its upstream/ downstream contrast; bedrock comprises greywackes, mudstones and shales from just below Hawick upstream, while downstream of that point, there are carboniferous sandstones and limestones (Figure 3.4.2.(i)). The upper Bowmont in contrast has an igneous bedrock, with andesitic and basaltic lavas. Downstream the pre-glacial channel, now containing Yetholm loch, may however have developed in response to a fault in the sandstone ridge.

In terms of drift geology, it has been estimated that in some parts of the Tweed basin, the solid rock surface is masked with 90-120 m of glacial debris (White, 1973). Typically within the upper catchments, bedrock is at, or near, the surface eg. Teviot and Whiteadder. Unfortunately a map is not available in terms of complete coverage, but the following comments are based on sound data. In the middle reaches, fluvioglacial deposits border and are incised by the present river system, whilst the surrounding glacial tills are tough and compact, containing pebbles mainly of local origin. The middle to lower Teviot and Whiteadder are characterised by drumlins.

3.4.3 Tweed study area: Geomorphology

During the last major glaciation of southern Scotland, the study area was covered by part of the Southern Uplands icesheet. The subsequent Loch Lomond readvance was only localised in scale, with glacier ice accumulation on the Cheviot (Clapperton, 1970; 1971). Passage of the debris-laden Devensian ice caused grooves and striae on the bedrock, which indicate that the ice sheet spread from the west or south-west towards the north-east. For example, the upper Teviot catchment above the Rule Water has indication of a down valley movement of ice. The ice preferentially eroded the softer Palaeozoic shales and the Devonian and Carboniferous strata, leaving harder greywacke beds standing out as positive features. A strong north-east to south-west lineation, due to the strike of these greywacke beds and the differential glacial ice erosion, is evident from the present day topography. Again the controversies over whether the Cheviot Hills possessed a separate ice cap during the Devensian will not be entered into here, but the legacy in terms of landforms is very important.

Glacial erosion is evident at a variety of scales, from macroscale river diversion to microscale glacial striae. In terms of river diversion through meltwater modification, the Bowmont water has an abandoned channel, which is now occupied by the marshy hollow of The Stank. As can be seen in Figure 3.4.1.(i).D, its previous course has been blocked by drift (Sissons, 1976). Another example is provided by the River Jed which, instead of continuing to flow north-east from Mossburn Ford, cuts a steep-sided gorge to the Tweed at Nisbet. Bowmont Water also has evidence of meltwater channels aligned in a north-westerly direction, paralleling the present drainage pattern (Clapperton, 1970). The lack of meltwater channels to the west, for example in the Teviot, is explained by the fact that most drainage flowed down pre-existing valleys where it removed large amounts of the till infill.

The effect of the lowland deposition of glacial materials, varying in thickness up to a depth of 100 m, was a substantial modification of the pre-glacial drainage systems. During the subsequent Lateglacial interstadial and the redevelopment of river systems, some rivers were unable to follow their previous courses (Clapperton, 1971). Remnants of wasting ice formed barriers behind which temporary lakes developed (Sissons, 1967). The eventual failure of such ice dams released large quantities of water into the valleys, causing an abrupt lowering of baselevels, and thus incising and partially eroding the outwash sand and gravel deposits into terraces eg. the River Teviot. A fluted topography of till with drumlins and drumlinoid forms occurs in the lower to middle Teviot, with increasing bedrock at or near the surface as one progresses upstream. In the valley of the Bowmont Water, fluvioglacial sands and gravels formed kames and eskers, which can be traced into the meltwater modified valley of the Yetholm. The Tweed below Kelso only infrequently comes into contact with solid rock, as there are extensive drift deposits, both in terms of depth and width, and periodic incision has led to several prominent river terraces, as described by Rhind (1969).

3.4.4 <u>Tweed study area: Climatic conditions</u>

Average annual rainfall at selected gauges in the Tweed study area, cver the period 1941-1970, are shown in Table 3.4.4.(i). The locations of selected rainfall gauges are shown in Table 3.4.4.(i). Average annual totals over a similar period in the Dee study area are approximately 100 mm higher. However, where stations exist at the same altitude in the three study areas, the Tweed tributaries have much greater annual rainfall totals (Figure 3.2.4.(i)). This may be due to a lower rainshadow effect.

3.4.5 <u>Tweed study area: Hydrology</u>

Within the Tweed study area, the mainstream Tweed has 4 discharge gauging stations with at least one on each of the studied tributaries, all being maintained by the Tweed River Purification Board (see Figure 3.4.5.(i) and Table 3.4.5.(i)). Soil moisture deficits, as computed in the <u>FSR</u> (NERC, 1975), range from 4.7 mm (Norham) to 11.0 mm (Hungry Snout). RSMD varied from 38.2 to 53.9 mm at the same two stations, reflecting the disparity between the upper catchments and the mainstream Tweed. The varying winter acceptance potential for the Tweed study area is shown in Figure 3.2.5.(ii). Values of winter acceptance potential were generally higher than in Deeside, though there is clearly a decrease with increasing altitude.

Table 3.4.4.(i)

Average annual rainfall totals within the Tweed study area

(1941-1970)

Station	Grid Reference	Height (m)	Rainfall (mm)
Kingside Loch	3344 6133	360	1520
Braehead	3406 6007	264	1406
Manderston House	3809 6547	108	772
Old Graden	3798 6299	152	740
Marchmont House	3743 6484	152	779
Duns Castle	3775 6538	137	791
Cowdenknowes	3575 6372	91	732
Hawick S.W.	3512 6156	96	792
Jedneuk	3656 6234	82	655
Floors Castle	3707 6345	59	698

Figure 3.4.4.(i) The location of studied long length raingauge sites within the Tweed study area

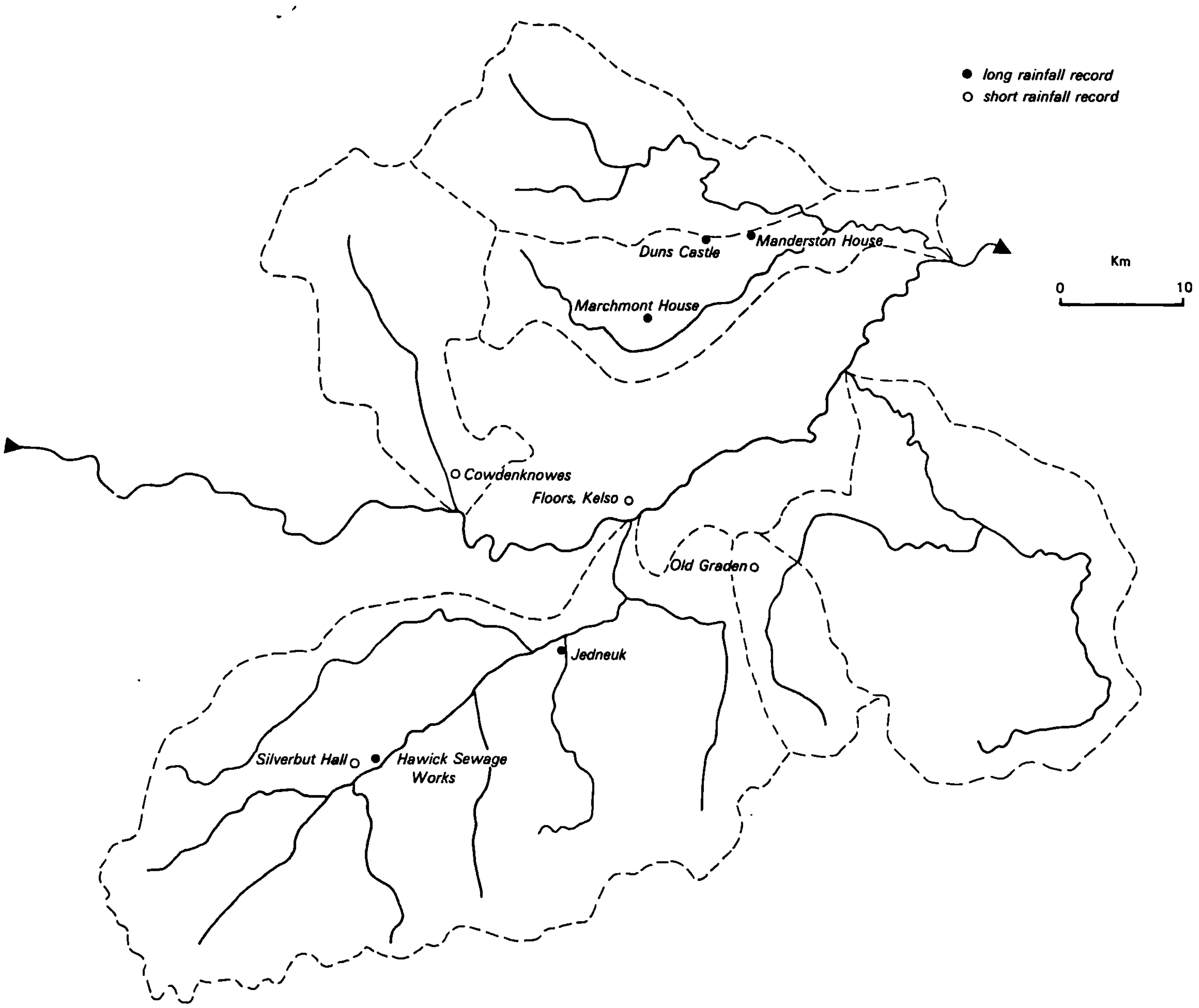


Table 3.4.5.(i)

Discharge gauging stations within the Tweed study area

River	Station	Grid Reference	Area (km^2)	Record
Tweed	Boleside	NT 498334	1500.0	11/7/1961-31/7/1983
Tweed	Dryburgh	NT 588320	2080.0	25/2/1949-7/1/1983
Tweed	Sprouston	NT 752354	3330.0	1/6/1969-31/7/1983
Tweed	Norham	NT 898477	4390.0	17/6/1959-31/7/1983
Leader	Earlston	NT 565388	239.0	1/10/1966-31/7/1983
Teviot	Hawick	NT 522159	323.0	18/9/1963-31/5/1983
Teviot	Ormiston Mill	NT 702280	1110.0	1/10/1960-30/6/1983
Jed	Jedburgh	NT 655214	139.0	21/7/1971-31/2/1983
Ale	Ancrum	NT 634244	174.0	1/10/1972-30/6/1983
Blackadder	Mouthbridge	NT 826530	159.0	1/11/1973-31/7/1983
Whiteadder	Hutton Castle	NT 881550	503.0	1/9/1963-31/12/1982
Whiteadder	Hungry Snout	NT 663633	45.6	30/12/1957-16/6/1968
Till	Etal	NT 927396	648.0	7/12/1955-30/9/1969
Glen	Kirknewton	** NT 913306	199.0	1/9/1961-28/2/1983

** approximate location

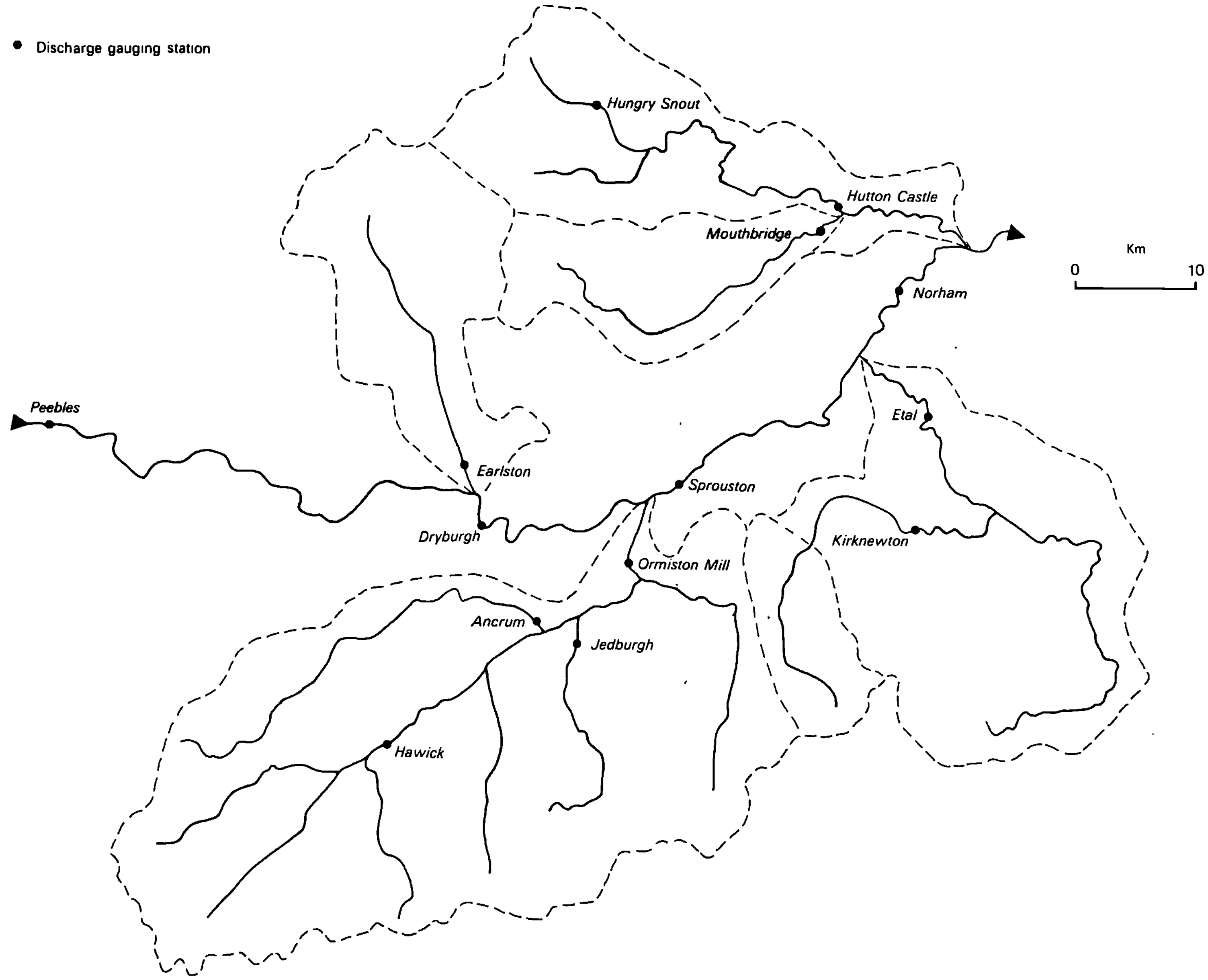
Table 3.4.5.(ii)

Summary statistics for discharge gauging stations within the Tweed study area

Station	Mean flow (m^3 s^{-1}) (1983)	Mean annual flood (m^3 s^{-1})	Maximum instantaneous flow (m^3 s^{-1})
Boleside	31.4	408.5	1156.0
Dryburgh	****	482.0	1174.1
Sprouston	60.9	698.6	1649.5
Norham	76.8	746.7	1556.8
Earlston	3.5	60.8	117.6
Hawick	8.4	193.3	310.4
Ormiston Mill	20.0	298.0	549.2
Jedburgh	2.6	57.5	90.5
Ancrum	2.4	42.0	80.6
Mouthbridge	1.8	44.5	92.8
Hutton Castle	7.5	114.4	265.9
Hungry Snout	****	19.4	63.1
Etal	****	87.9	299.6
Kirknewton	****	****	43.5

**** not available

The location of discharge gauging stations within the Tweed study area **Figure 3.4.5.(i)**



3.4.6 Tweed study area: Landuse

In contrast to the Dee and Spey study areas, the Tweed basins have a principally agricultural landscape. Above the woodlands of the valleysides, improved agricultural land extends up to 350 m (Whittow, 1977). There is some remnant woodland, which reveals nothing of its former extent, and depletion is largely caused by man. Formerly, there was an extensive oak forest in the valleys, with pine and birches on the higher ground. In the twentieth century, this has been augmented by new planting, both by the Forestry Commission and by private landowners. In contrast to the impression of Ledger (1981), who stated that the Tweed was still essentially unaffected by man's activities, the rivers have been considerably altered by weirs and mill lades, associated with both the 18th and 19th century textile industry (Stevens, 1951) and agricultural activities.

3.5 Summary

The main contrasts between the three study areas are as follows. In terms of geomorphic setting, the contributing areas and average heights of the catchments vary in scale from the Dee (smaller catchments, higher average heights) to the Tweed (larger catchments, lower average heights). Relief ratios are also considerably higher within the Dee catchment cf. the Tweed. The severity of glacial erosion has been most extreme within the Dee and Spey study areas and this has implications for both sediment availability and accessibility, the positioning of local baselevels, variation in valley slope and the potential for widening and reworking floodplain sediments.

In terms of annual rainfall, the Tweed study area has higher totals for a given station height cf. Spey and Dee. However, average rainfall amounts on the upper slopes of the Cairngorm massif will be higher than those on the upper slopes of the Southern Uplands, due to the differences in altitude. Contrasts within the hydrological regime will be detailed in Chapter 5. The Tweed region has higher winter rainfall acceptance potential than the other two study areas.

CHAPTER 4

A macro-scale study of rates of channel planform change based on map analysis

4.1 Introduction

Focusing on the very different study areas in upper Deeside, Speyside and the Tweed valley (Chapter 3), the following analysis was carried out to find which variables are important in determining the spatial range of channel pattern type at a moment in time (Chapter 1: Question 1), but also the associated rates of channel activity and the variables that influence the likelihood of change at a particular site (Chapter 1: Question 2). This enables one to assess how stability of channel planform varies with position in the drainage basin. In this chapter, the available data base will be discussed and a suitable channel typology developed. Then, the types of indices of channel change will be identified, bearing in mind the limitations of the data base. Finally, the results of this analysis will be presented in sequence for each study region.

4.2 The available map data-base

In this study of rates of channel activity, an attempt was made to extend planform records of present day river channels back into the mid-19th century and earlier. Within the major part of this timescale however, there are no aerial photographs and estate plans tend to be sparse and localised. The most complete data sources available were the first (1858-1873) and second (1897-1903) editions of the 1:10,560 O.S. maps and the metric 1:10,000 (or third edition 1:10,560) O.S. maps, surveyed or extracted from aerial photograph analysis (1957-1982). These provide quite detailed information about the nature of individual reaches at these different instances in time.

To extend the record back even further, the planform as depicted on William Roy's Military Survey of Scotland (1747-1755), at a scale of 1:46,200 [National Library Copy], was compared with that of the first edition. The quality and reliability of this source is commented upon by Arrowsmith (1809) in Skelton (1967):

> "The courses of all the Rivers and numerous streams were followed to the source and measured; all the Roads and the many lakes of salt water and Fresh were surveyed." (Skelton, 1967 p4).

Roy is known to have surveyed up the river valleys and recorded the larger channels accurately using an 18th century version of a theodolite (Moir, 1973; O'Donoghue, 1977). Therefore, the detail depicted on these maps can be expected to have a reasonable level of reliability. In this study, Roy's map is thus used to make the qualitative distinction between straight, meandering and braided channels and thereby establish a planform datum for 1750. Checks with contemporary literature (eg. local estate records) and old maps/ plans, where possible, confirmed the reliability of Roy's military maps as a source. There is however a difference in draughtmanship between the fair and protracted copy (Moir, 1973; Whittington, pers. comm.) and thus certain areas of the Tweed basin are less accurately covered.

Returning to the first and second edition O.S. maps, it should be stressed that problems occur in determining how much one can rely or interpret from such a data source. Taken to extremes, maps are only an engraver's representation of a cartographer's representation of a surveyor's representation of reality and thus in terms of quantitative analysis must be treated with some caution. It was necessary therefore to understand the minimum thresholds of accuracy. In Harley's (1975) descriptive manual on the Ordinance Survey, it states:

> "In surveying for the basic 1:10,560 Regular Series sheets, rivers and streams are shown to scale by double lines if 5m or more wide and, by a single line, if less than 5m." (Harley, 1975 p87)

One has to bear in mind however that the O.S. would be cautious to admit their inaccuracies and such limitations in the resolution of the data must be carefully considered in subsequent analysis. The general reliability of map-based data in delimiting stream networks was illustrated by Ovenden and Gregory (1980) and by Carr (1962) within coastal geomorphology. In general, field checking of map information showed a high degree of reliability, within all three study areas. Sample channel widths were compared with the metric edition O.S. map and they were found generally to be accurate to 2-4 metres, although variation in width along reaches was not always indicated. Of course, there was a time-lag within this comparison and subsequent erosion may have taken place. Selected palaeochannels on the first and second edition O.S. maps were also field checked to confirm their accuracy. Similarly, much older features on the ground could be related to channel planforms in 1750 from Roy's map.

It must also be remembered that map data provides only a set of "stills". There may be no indication of what has occurred between times of survey and this has implications for any conclusions that may be taken from the final results. It may be mere chance that the sequence of available maps picks up a sequence of change; on the other hand, the reverse may occur in that map-dates may capture the river at similar equilibrium conditions despite some degree of inter-mapdate change. In unknown "black box" terms, "black" gaps can be made "grey" from other sources such as rainfall evidence, newspaper information and contemporary literature details, as seen in Chapter 5. Another unknown is that of the stage at the time of surveying. Planform can be highly stage dependent; for example, old anabranches or flood channels are frequently reused or reactivated at higher stages in multibar channels while in other reaches, low-stage bars within the normal channel limits may be drowned out at higher flows. It is necessary however to be positive and map-based data does provide a valuable source of information which, when tapped, can provide insight into late Holocene rates and types of channel pattern change over the past 200 years and which cannot be obtained from any alternative source.

4.3 Construction of a channel typology

Classification is a crucial stage in scientific development (Johnston, 1976) and to provide a sound basis for comparison of channel pattern and activity rates at the macro-, meso- and micro-level, a detailed channel typology had to be constructed.

Many attempts have been made in the past to classify channel planform. The simplest classic typology was Leopold and Wolman's (1957) straight/ meandering/ braided distinction, determined empirically by the relation of slope (S) to bankfull discharge (Q_b), as stated in the following equation:

$$S = 0.13\ Q_b^{-0.44}$$

This may have proved satisfactory in sub-humid environments of the western USA, which were characterised by large drainage basins and fine-grained sediments. Its more practical limitations however are neatly summarised by the following comment:

"The well known division into straight, meandering and braided (Leopold and Wolman, 1957) is only satisfactory for reaches shorter than a few channel widths. Over longer stretches of valley floor, the traditionally descriptive terms are neither mutually exclusive or exhaustive". (Ferguson, 1981 p106)

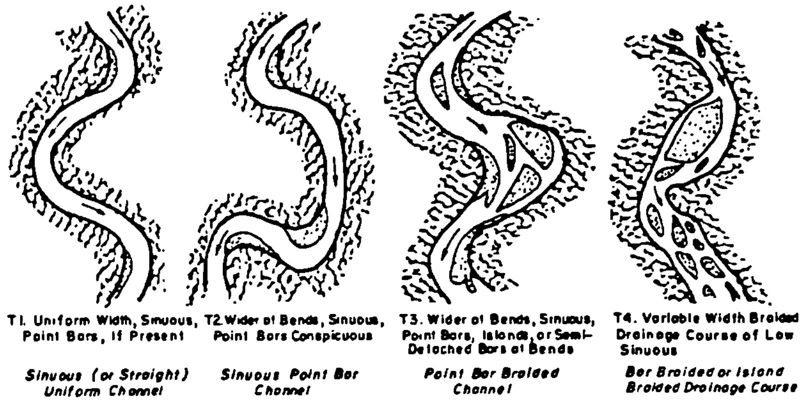
It was evident from preliminary map study, random aerial photograph study and previous field observation that within a more general Scottish upland context (ie. outwith the confines of the study areas selected), there existed a much wider range of channel pattern types than could be satisfied by the three-fold Leopold and Wolman typology.

There have been several other attempts to classify planform, for example that developed by Schumm (1968), which first identified anastomosing channels as a separate category. Although this is more elaborate in terms of a genetic orientation, it is still too simplistic and restrictive for upland Scotland (see Chapter 2.3; Figures 2.3.(ii) and (iii)). In contrast, Chitale (1970; 1973) suggested a more logical distinction was between single and multiple channels, recognising that braided and meandering channel patterns are not mutually exclusive. The categories were further subdivided into regular and irregular meanders and simple and compound channels, but the basic meandering/ braiding distinction remained.

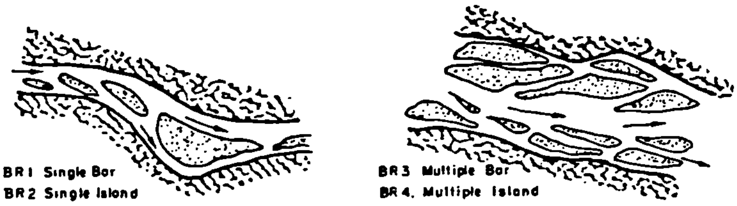
For a typology to have any value it must bring out the similarities and distinctions in channel regime within a variety of fluvial environments. The danger with an over-elaborate scheme is ultimately to study only the unique so that each channel reach produces a entry in the typology and occupies a unique niche. Thus, concentrating only on differences may conceal similarities while alternatively, the use of too few classes may obscure important differences. Cuthbertson et al. (1967) suggested there were subclasses within the major types of straight, meandering and braided channels eg. in terms of variability of unvegetated channel width, types of sinuosities and types of vegetal patterns (Figure 4.3.(i)). Here we are getting closer to what the Scottish upland environment demands.

A more detailed example of a channel system typology was provided by Kellerhals, Church and Bray (1976) for Canadian rivers (Figure 4.3.(ii)). This was a highly detailed typology designed for high resolution field based study. It included a classification of reach characteristics, such as bank vegetation, as well as subdivision of channel features into three different headings, namely channel pattern, islands and channel bars/ major bedforms. This seemed to provide the most suitable format for the aims of the present study and thus, the typology developed below for this study in upland Scotland is derived from this Canadian-based system.

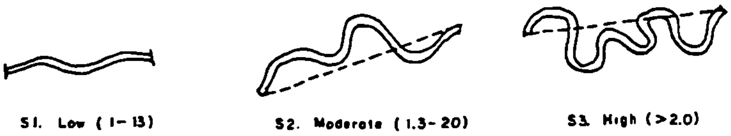
Figure 4.3.(i)



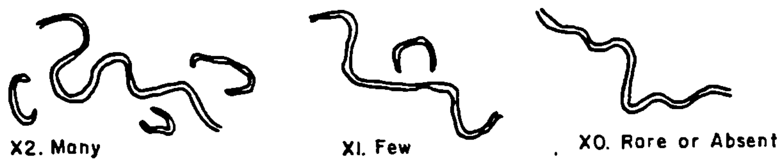
(a) Variability of Unvegetated Channel Width: Channel Pattern at Normal Discharge



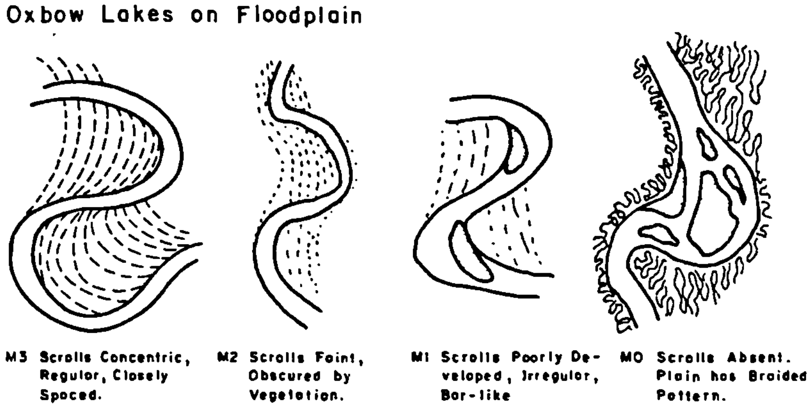
(b) Braiding Patterns



(c) Types of Sinuosities

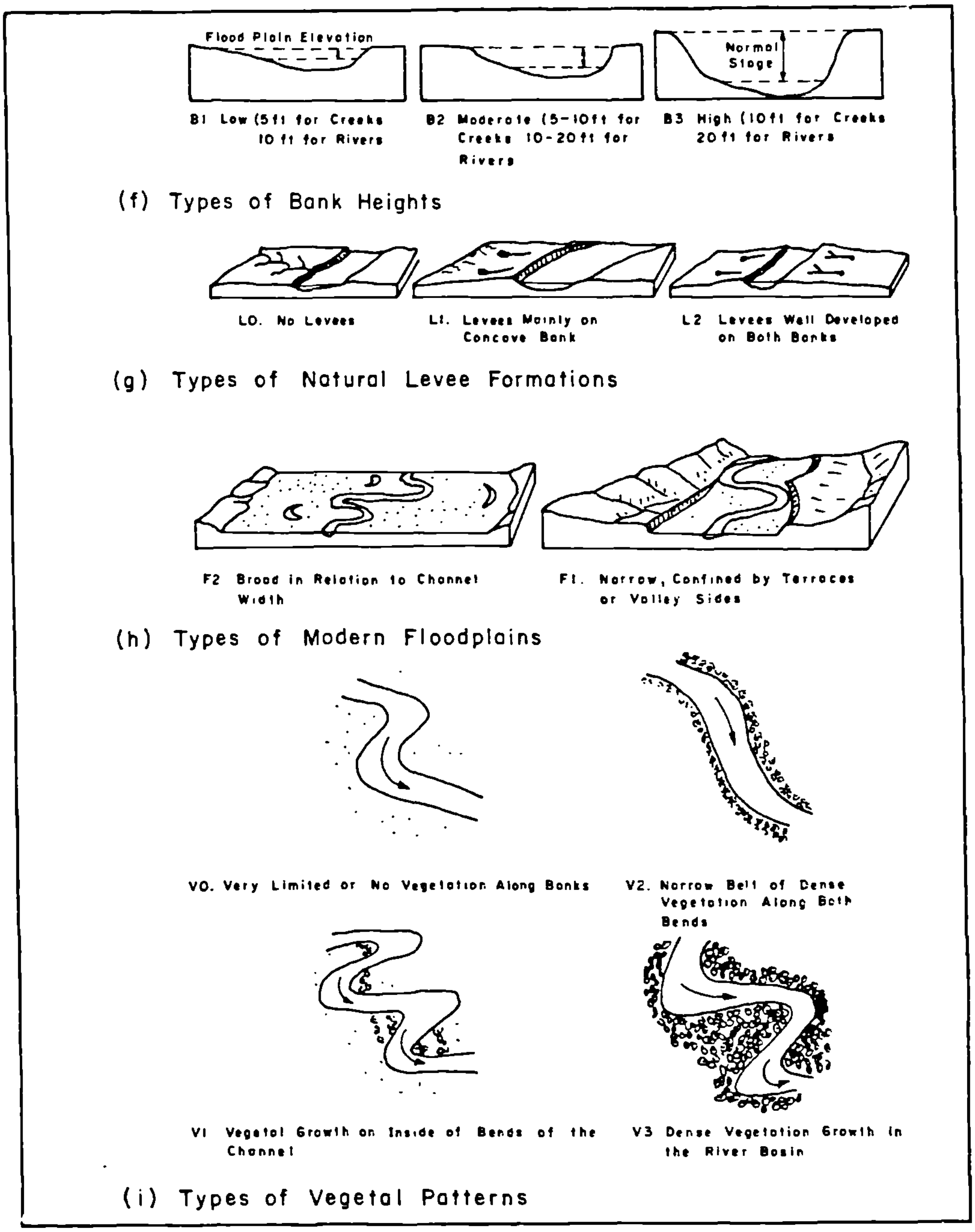


(d) Oxbow Lakes on Floodplain



(e) Types of Meander Scroll Formations

Figure 4.3.(i) cont.

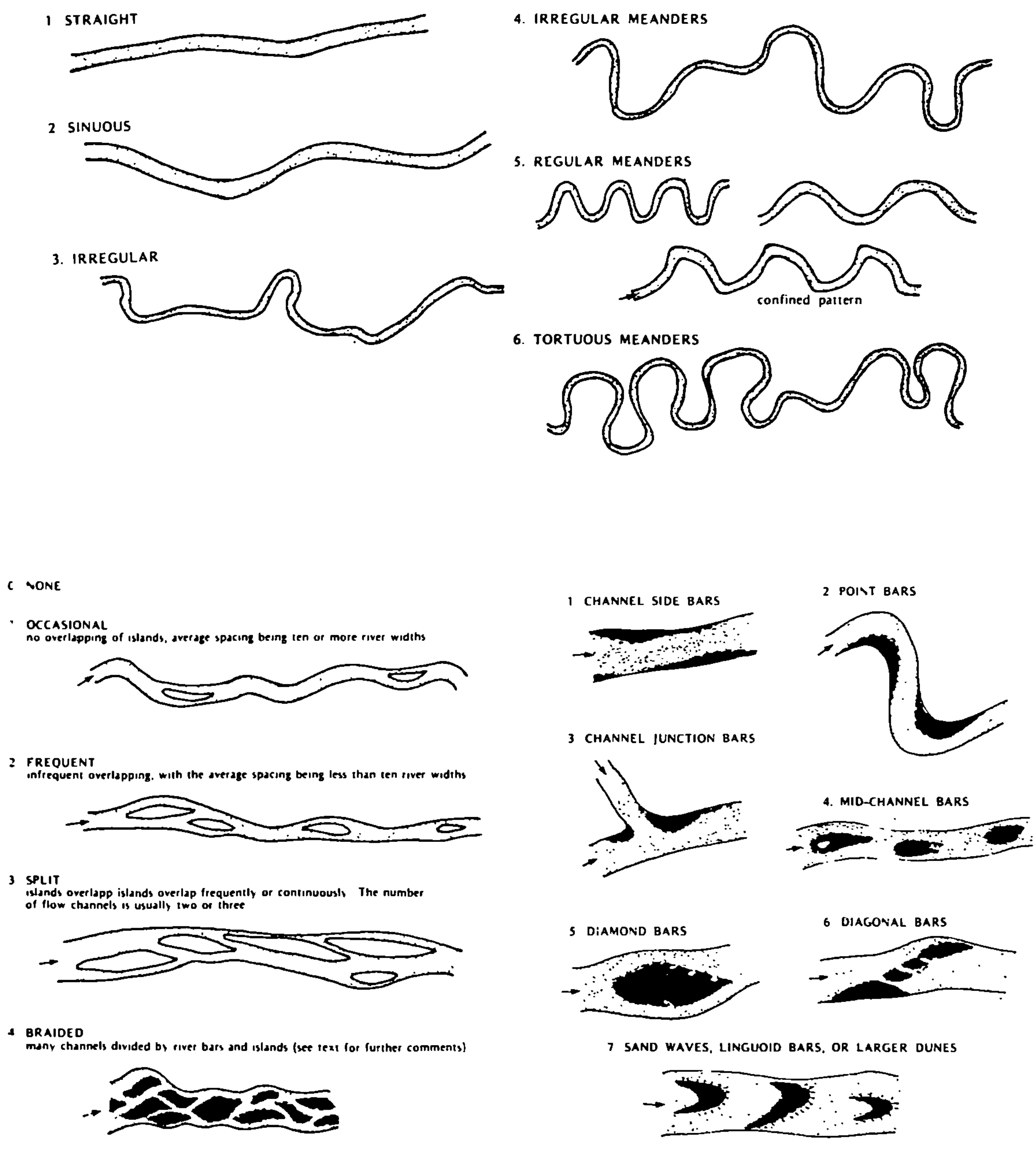


Classification of river channels

(After Cuthbertson et al., 1967)

(Source: Simons and Senturk, 1976 pp41-43 Figure 2.11)

Figure 4.3.(ii)



Codification of river channel patterns

(Source: Kellerhals et al., 1976 p823 Figures 2-4)

It was then necessary to decide what specific typology was required for a map-based study. In the previously mentioned preliminary survey of Scottish planform and river regime, a variety of important features and various possible controls were noted. The required typology has to be developed in such a way that it is suitable for accessing information from the desired scale of enquiry: map evidence, aerial photograph analysis, and field checking. These provide information on a macro-, meso- and micro-level respectively, with greater limitations as one is forced higher up the scale. In terms of the typology, this study initially focuses on information derived on the macro-scale.

At the outset, certain points should be stressed. Firstly, at the map or aerial photograph scale, some variables are inevitably qualitative rather than quantitative. Whereas in a field-based survey, attempts could be made to measure specific parameters (eg. sediment availability and slope), at a map-based resolution of information, any attempt to quantify such variables would be entirely speculative. Qualitative distinctions, made by objective assessment,can be better indications of reality. A second point to be noted is that the size of reach to be classified is especially important, as sometimes units are easily but artificially delineated. Obviously, the smaller the reach, the more homogeneous the controls on processes are likely to be and the easier it will fit into a classification, but this may not be representative of that channel system at a larger scale. For this reason, for example, some indication of sediment availability and accessibility must be included upstream of the reach under study, to see if there are local constraints on downstream supply.

Thirdly, apart from the obvious problem of the stage at the time of photography or map-surveying, merely studying aerial photographs or maps gives little indication of the local temporal variability of discharge. If this information was to be required, selection of sites would be very restricted in a Scottish upland context, due to the limited number of reaches with locally available discharge or rainfall data. Fourthly, a channel typology examines channel configuration as an instant in time. The temporal element is however crucial in any classification as two rivers may possess a similar planform at a particular moment in time, but may have differing rates of activity and/or be at varying degrees of equilibrium with the prevailing environmental conditions. It is therefore obvious, for example, that in some channels vegetation succession on bars and banks should be qualitatively incorporated into the typology. Also, in some cases, palaeochannels should be recognised in terms of their location, frequency and characteristics.

Finally, it must be realised that in a sense, a typology is provisional. The end result of placing a river reach within a series of classifications will be an elaboration on the features which originally placed <u>that</u> reach in <u>that</u> category. All classifications are arbitrary and are in need of constant reassessment and ultimate replacement. In the end, it may be necessary to return to the original classification. Perhaps in the future, allocation in one class will be deemed unsatisfactory with additional knowledge and experience of the channel system. It may also be the case that the original channel typology will need to be modified to bring out what have since been found to be the most important features differentiating distinctive reaches. The present typology can thus be defended as an initial framework but flexibility of structure is essential.

For the map-based typology applied in this study (detailed in Table 4.3.(i)), the first set of independent variables assessed were general features of the catchment and its regime. Contributing drainage basin area (in km^2) above the start of each channel segment under study was digitised from the O.S. 1:63,360 map, using a Tektronix micro-computer program. Average height (m) for each study reach was taken from the O.S. 1:10,000 map as were all the characteristics described below.

In terms of the actual site-specific characteristics, several parameters were considered. To gain some indication of sediment supply, availability or, more importantly, accessibility of sediment* was assessed qualitatively on a scale of 0-3 (Table 4.3.(i) and Figure 4.3.(ii)). This ranged from no sediment within or by the channel to extensive reserves of sediment, both within and proximal to the channel. Such supply was evaluated both locally in situ and upstream of the reach. Interpretation of this availability must be treated with care, especially as to the timescale of occurrence. If a map was surveyed immediately after a flood event, large amounts of sediment may be indicated, which are not necessarily typical of a longer period (eg. 10 years).

* Sediment may be available within or by the channel but only accessed under certain stage conditions or when thresholds are exceeded for slope failure.

Table 4.3.(i)

A map-based channel system typology (illustrations in Figure 4.3.(iii))

(A) General features of catchment and regime

Geomorphological setting

(a) contributing area above reach (km^2)

(b) average height of sample reach (m)

(B) Indications of sediment supply both upstream and local to site

Existence of bare sediment/ gravels/ boulders

(a) upstream of sample reach (UPSEDMT)

(0) no evidence of upstream sediment sources

(1) localised evidence of unvegetated sediment stored within channel

(2) large amounts of unvegetated sediment, stored both within the channel but also within the neighbouring floodplain

(3) extensive reserves of unvegetated sediment, both within the channel and the floodplain

(b) local to sample reach (LOCSEDMT)

(0) no evidence of upstream sediment sources

(1) localised evidence of unvegetated sediment stored within channel

(2) large amounts of unvegetated sediment, stored both within the channel but also within the neighbouring floodplain

(3) extensive reserves of unvegetated sediment, both within the channel and floodplain

Table 4.3.(i) cont.

(C) Indication of the temporal variability of discharge

(a) "Liability to flooding" noted

(b) boulders on floodplain near channel

(c) flood boulders indicated, marking maximum instantaneous flood stage

(D) Vegetation cover

(categorised by hydrological significance and degree of stability provided)

Classification for (a) floodplain, (b) bank and (c) bar vegetation

(1) trees (deciduous/ coniferous/ mixed)

(2) coppice (trees and shrub/ trees and heath/ trees, shrub and heath)

(3) trees with rough-grassland (trees, shrub and grassland/ trees, heath and grassland/ trees and agriculture)

(4) shrub (shrub and heath/ heath)

(5) agricultural (pastoral/ arable/ mixed)

(6) rough grass-land

(7) marsh

(8) unvegetated *

* assumption: no indication = unvegetated

Table 4.3.(i) cont.

(E) Indication of possible active area

(a) Ratio of maximum width of flood plain versus width of present channel at that point (MAXWID)

(F) Channel description

(see Figure 4.3.(iii))

(a) Channel pattern	(b) Islands
(CHANPATT)	(ISLANDS)
(1) straight	(1) none
(2) straight & sinuous	(2) occasional
(3) sinuous	(3) frequent
(4) wandering	(4) split
(5) irregular	(5) braided
(6) irregular meanders	(6) braided & anabranching
(7) regular meanders	(7) anabranching
(8) tortuous meanders	(8) reticulate

Table 4.3.(i) cont.

(G) Evidence of past lateral activity (ACTIVITY)

(a) Evidence within the grid square

(1) no evidence

(2) unvegetated sediment by channel

(3) progressions and cutoffs indicated by isolated areas of water or oxbow lakes

(4) irregular channel activity

(a) anabranching

(b) deltaic distributary fan

(c) gravel fan

(5) avulsion

(6) evidence of palaeochannels/ old anabranches/ flood channels

(7) marsh and unvegetated sediment

(8) evidence of old anabranches and unvegetated channel side sediment

(9) evidence of older palaeochannels through differential vegetation eg. former channel marked by trees.

Figure 4.3.(iii)

Illustration of the characteristics of classifications within the channel typology

(for detailed key, see Table 4.2.(i))

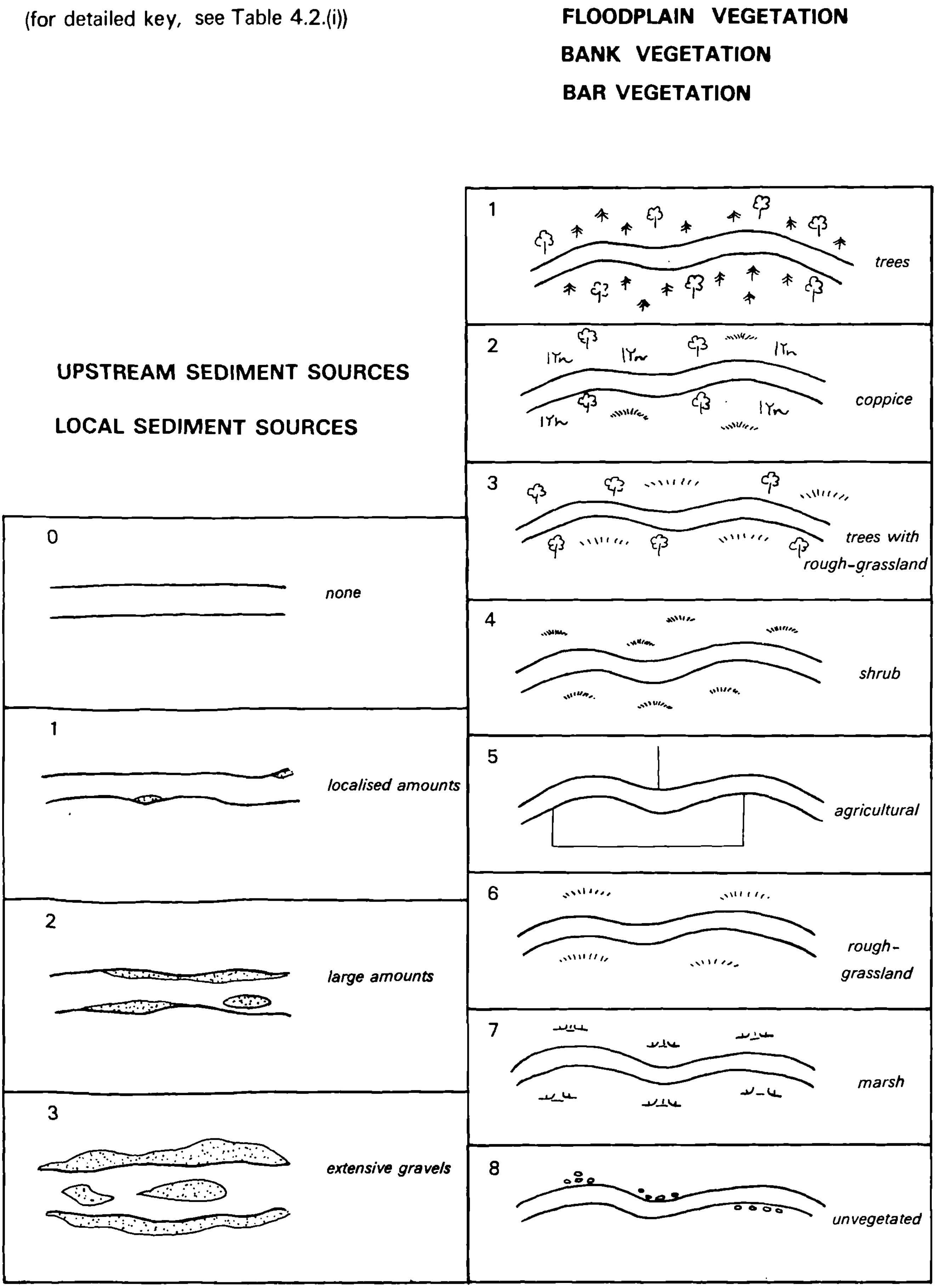
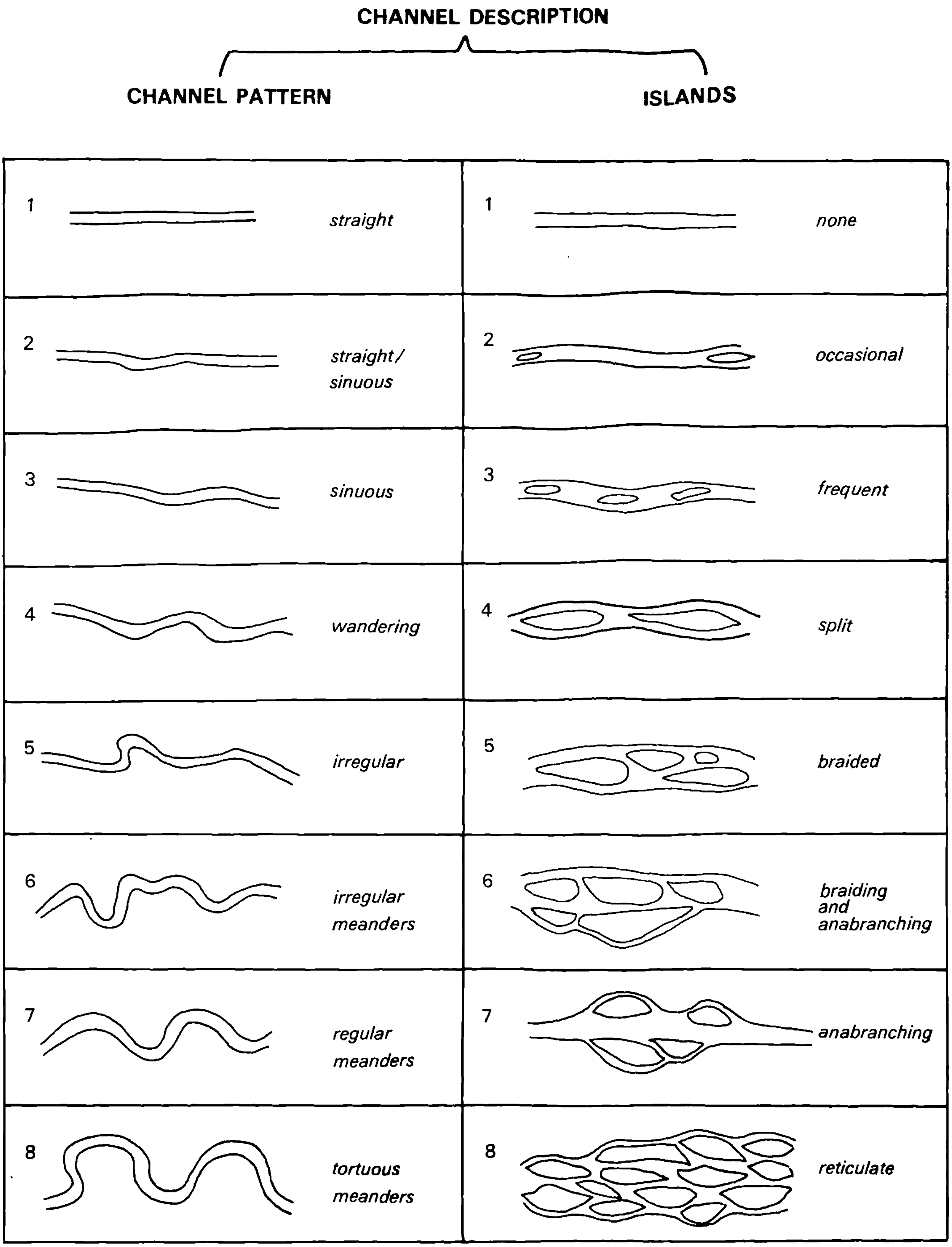


Figure 4.3.(iii) cont.



Maps tend not to give any indication of the temporal variability of discharge but when this did occur, it at least indicated an area which at some time was subject to frequent inundation. Any indication of flood frequency eg. existence of flood gates, flood stones, or "liability to flooding" were recorded on a presence/absence basis. Presence at one map-date and absence at another may imply change in frequency of inundation (or man's perception of such frequency). However, this also depends on the reliability of this information as displayed on successive editions. Although presence does give positive indication of flooding, absence does not mean that no inundation occurs.

Vegetation cover was another important site parameter, and the dominant vegetation of floodplain, banks and bars was studied. Hydrological significance and stability were the main considerations as suggested by the literature on bank erosion (see Chapter 2.8). These were subdivided into categories ranging from dense homogeneous tree cover at one extreme (higher infiltration rates, more difficult to erode) to unvegetated gravels at the other, as indicated on the 1:10,000 O.S. map. This allowed the identification of seres, especially on bars, which were thus becoming stabilised and more permanent landforms.

Next, in order to gain some indication of confinement, it was necessary to include an indication of the potential active area. It is however difficult to assess from map evidence whether a channel is free to rework its entire floodplain without obstruction, or whether terrace fragments or fluvioglacial meltwater deposits prevent or slow down the rate of possible channel migration. For the purposes of this study, it was considered that some parameter was required even if it only gave an estimation of the possible active area of the channel at some unknown time in the past. The parameter chosen was the ratio of the maximum width of the floodplain versus the width of the present channel at that point (MAXWID).

Having criticised other channel typologies, it was necessary to find a suitable form of channel description which could cope with the fact that individual channels of a braiding planform may have a tendency to meander, especially at lower flows. In contrast, another channel which may be dominantly meandering in planform, may have a varying frequency of bars. Channel description was therefore subdivided into actual mainstream channel pattern and the frequency/existence of islands or bars. Channel pattern (CHANPATT) ranged from a principally straight channel, notoriously rare in nature, through the continuous range of channel planform types in artificial steps, to tortuous meanders (Figure 4.3.(iii)). Classification of islands (ISLANDS) ranged on a scale from complete absence to a dense reticulate pattern at the other extreme (Figure 4.3.(iii)). The identification of bar-forms, as applied in the Kellerhals et al. (1976) typology, proved difficult from a map-base and thus channel description was confined to using only two criteria.

Finally, some indication of lateral activity was included to reconstruct a history of the reach and to see if any obvious significant adjustment had taken place in the past, which could be related to the planform at the time of Roy's map (1750s) or earlier. This ranged from "no evidence", which would obviously occur in more stable channels, to the existence of a variety of features indicating channel migration or shift. These features included palaeochannels (marked by differential vegetation), obvious flood channels, marshy areas proximal to the present channel, isolated areas of still water, progressions, avulsions and cutoffs, and even the existence of bare gravels by the channel side. These were all noted as indicating some movement at some perhaps indeterminable time in the past.

Of course, these independent variables rest at different levels of measurement eg. vegetation and existence of sediment by the channel must be nominal, while drainage basin area and height are ratio. This inflicts strong limitations on any subsequent statistical analysis. With these points in mind, the typology shown in Table 4.3.(i) was constructed. Consistency and objectivity in allocating sites to classes were checked by repeated analyses after a suitable period of time.

4.4 Sampling procedure and measures of channel activity

The sampling strategy was a simple random sample of 75 to 85 National Grid squares (1 km^2), located along the river channels. The map-base for data collection for the purpose of this study was the river channel depicted as a double blue line on the 1:25,000 Ordnance Survey maps. The proportion of total channel length actually sampled was by necessity variable, dependent on the size of the drainage basin and the density of possible change in controls. Sampling was thus slightly denser for upper Deeside than the Spey and Tweed study areas.

The great merit of a simple random sample is that it should give every possible sample of that size from that population the same chance of selection (Stuart, 1968) ie. unbiased estimates of the true population. Thus, the range of channel planform categories and typology characteristics can be assessed. This part of the present study has two contrasting aims. It is necessary to be able to infer overall aggregate behaviour of the total population on the basis of the random sample. It is also important to have analysis of aberrant sites, which for good geomorphic reasons behave rather differently. The other possibility would have been systematic random sampling but with no knowledge of the expected patterns, random sampling was more appropriate. The bias-free nature of the sample was ensured by the generation of random numbers to select the grid squares. For each of the randomly selected channel segments, three different variables were measured, their definitions being given below. Acronyms are used for simplicity in the subsequent text and these are defined for reference in Table 4.4.(i) (copy in back pocket).

(1) The first measurement taken was an assessment of lateral shift (LATSHIFT). The sampled river channel was compared sequentially through a series of three digitised map comparisons, using a Tektronix micro-computer program. Standardisation to the same scale was done by finding fixed points of reference that occurred on each of the three map-dates, a difficult and time-consuming task. Map-derived points (consistent at all three map-dates) in upland areas were very infrequent and potentially unreliable if not selected with care. Even in lower lying areas, the landscape in human terms has altered considerably over

Table 4.4.(i)

Acronyms used in the channel typology and activity analysis

MAXWID	Ratio of channel width to maximum valley width
UPSEDMT	Availability of upstream sediment
LOCSEDMT	Availability of local sediment
CHANPATT	Channel pattern
ISLANDS	Frequency of islands in channel
ACTIVITY	Evidence of past activity
BI	Braiding index
BRAID	Total change in braiding index
AVBRAID	Average annual change in braiding index
ABBRAID	Absolute average annual change in braiding index
SIN	Sinuosity index
CHASIN	Total change in sinuosity
AVSIN	Average annual change in sinuosity
ABAVSIN	Absolute average annual change in sinuosity
LSMAX	Total maximum lateral shift
LSAV	Total average lateral shift
AVLSMAX	Maximum annual lateral shift
AVLSAV	Average annual lateral shift
1	First edition
2	Second edition
3	Third edition

Eg. AVSIN1-2 is defined as average annual change in sinuosity between first and second edition.

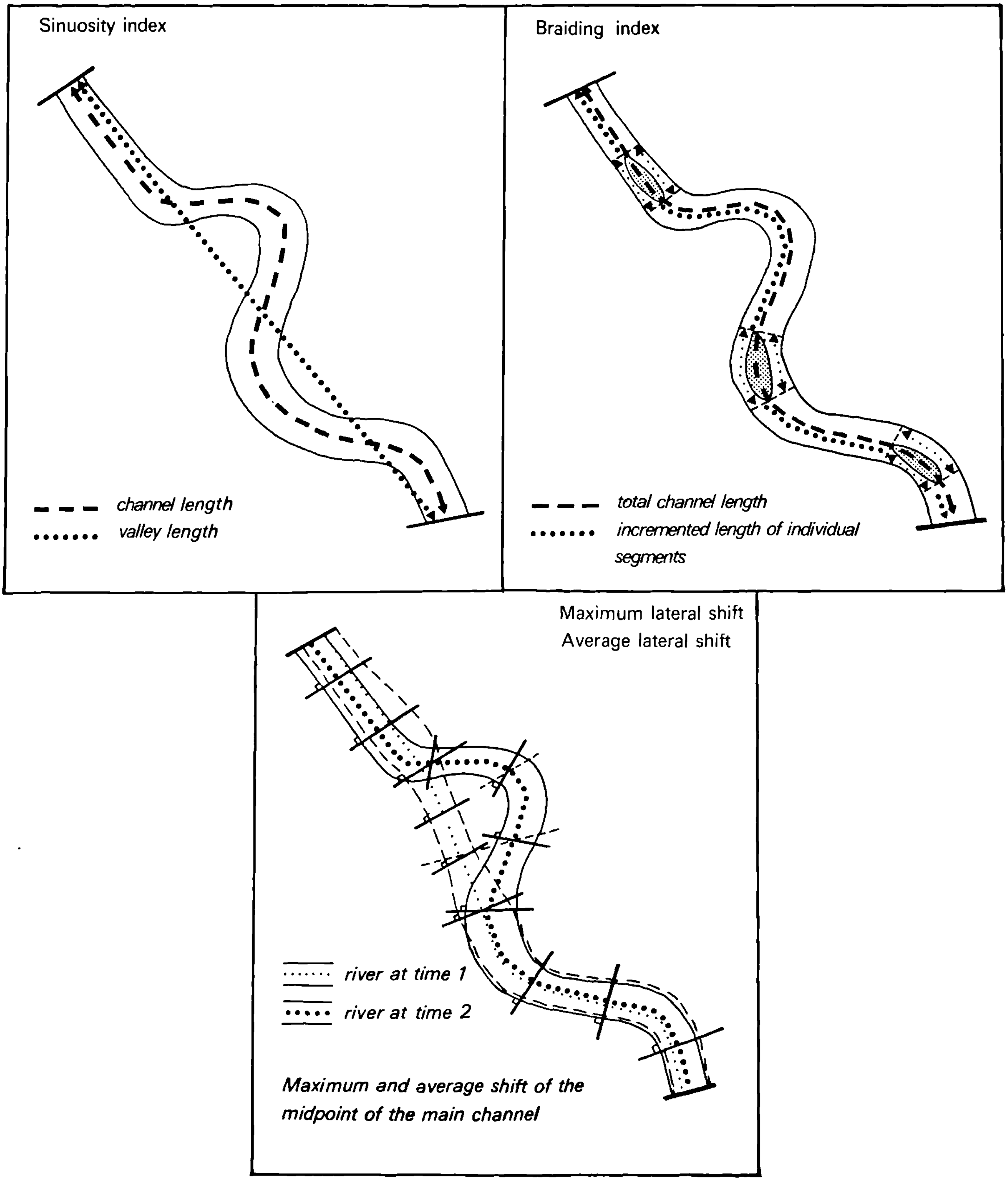
the last 200 years. A series of cross-sections was drawn along the sampled reach at approximately 100 metre intervals, as one would do if the reach was ground surveyed. The shift of the mid-point of the main channel was recorded in metres at each cross-section, between both the first and second edition maps and the second and metric edition maps, to gain an overall maximum and mean value of shift, as indicated in Figure 4.4.(i).

The assumption here is that the main channel is the widest one and that the thalweg runs down the middle of the channel. Maps are only a 2-D representation of a 3-D world and the mid-point is the most reliably represented point of the river channel; in contrast width and position of banks depend on map convention. This measurement shows how frequently and to what extent each channel segment was reworking its floodplain. Obviously, the importance of channel shift varies with the width of the river. Lewin et al. (1977) used width as a criterion of change on meandering Welsh rivers (see also Lewin, 1977) but unfortunately this was deemed unreliable in an upland Scottish context, especially as field survey indicated width of channel can vary considerably along small sections of an upland reach. The above method is justified in that it is physically and directly related to the amount of floodplain reworking.

(2) The second measurement was change in sinuosity of the middle channel point over time (SIN). Sinuosity was defined as the ratio of downstream valley length to actual channel length (Schumm, 1963) and thus both downstream length and reach length were digitised three times and averaged to gain as representative a value as possible (Figure

Figure 4.4.(i)

Definition of the indices of channel pattern change



4.4.(i)). This was because the definition of sinuosity was sometimes difficult due to the problems of determining valley length, as discussed by Gregory and Walling (1973). Interpretation of valley length may be variable in irregular catchments. Of course, there will also be operator error but this can be considered small in terms of values of sinuosity change. Either downstream mid-valley length or length of downstream active area were used within this study, depending on the location of the channel segment within its valley (Figure 4.4.(i)).

(3) The third measurement was a measure for overall braiding (BI), adapted from Milne (1982) and Brice (1960,1964). Downstream distance was taken as a ratio of the cumulative length of all the channel segments, . to give a measure of the overall subdivision of the reach (Figure 4.4.(i)). This measurement accounts for activity taking place between the cross-sections of index (1) and gives an indication of the variability and persistence of channel division down a reach. It has limitations however in that one reach might have high but very localised incidence of islands, and another reach could have occasional bars throughout the reach; both would have a similar BI value despite being distinctly different.

In all digitised measurements however carefully carried out, there must be an element of operator error. Efforts can only be made to minimise this "noise" with selective rejection of samples where error was deemed above a reliable threshold.

Channel pattern change in this study can be envisaged as occurring on three dimensions, namely degree of lateral shift (LATSHIFT), change in sinuosity (SIN) and change in braiding index (BI). Only when these measures were applied to a Scottish upland context could their individual sensitivity to the most frequent types of form adjustment be assessed. It must also be remembered in subsequent analysis that change may not be gradual. While for the sake of comparison of the two intermap periods, change between channels has to be subdivided by the number of years, it may all have taken place in a few hours. This depends on the energy relationships in the particular environment of study and this will be discussed later in the meso- and micro-scale analysis of Chapters 7 and 8.

4.5 Statistical analysis of the collected data

With the data collected, the following statistical analyses were carried out and the results are presented below. In all statistical tests, the conventional usage will be that "significant" represents a probability equal or in excess of the 0.05 level and "highly significant" equals or exceeds the 0.01 level. It was first necessary to ascertain what range of channel types and controlling variables were apparent ie. the spatial range of form as indicated by SIN and BI at each map-date. This was studied both in the main study areas and also at an individual sub-basin level, where there were obvious intra-study area disparities. Then it could be ascertained how frequently certain channel characteristics and patterns occur at each of the three instants of time. Simple frequency analyses were carried out and activity values were displayed on maps of the drainage basins to assess their spatial distributions. In terms of the statistical parameters chosen, because the distributions were frequently highly skewed and obviously non-normal, the median and inter-quartile range were used in preference to the mean and standard deviation (ie. no underlying distribution was assumed). The BI, for example, had an almost exponential distribution. The date in brackets within the tables is the average date for each edition.

Next, an attempt was made to see if any significant temporal change had occurred in channel pattern over the period of study. If change had occurred, it was necessary to assess what types of change were most and least prevalent in each sample, whether certain categories of channel pattern were more susceptible to change and whether such change involved an increase, decrease or maintenance of the active area dimensions. Maps were constructed to annotate the absolute change in activity and to indicate the spatial extent of change. Then, the sinuosity and braiding indices were plotted against year of map surveying in a series of diagrams, both at the regional and subregional levels. To simplify the information portrayed, sites which had undergone no change were omitted from these diagrams.

Again due to the skewness and kurtosis of the distributions, these values were subjected to the Wilcoxon Matched-Pairs Signed-Rank Test (Siegel, 1956) to assess whether there was any significant statistical difference between first/ second, second/third and first/third edition data. This non-parametric test was chosen because not only was the presence/absence of differences between paired observations analysed, but the magnitude of that difference was also used in the analysis. The power efficiency of this test whilst usually lower than the corresponding T-test, in other respects compares very favourably with it (Robson, 1973). In this Wilcoxon two-tailed test, the observed Z statistic is significant if it exceeds 1.96. With this data set however, it is necessary to combine the need to test the overall statistical significance of the sample as well as to preserve the geomorphic individuality of each contributing sample reach. It is thus possible for there to be no statistically significant change within the total sample, but still have important geomorphic change occurring at particular sites.

The next step was to see which independent, channel characteristic variables, from the map-based typology, were important in controlling rates of channel activity. Change in channel planform can either be in a positive or negative direction and thus in the analysis, both this range and the absolute change independent of direction were studied. The absolute value had to be considered because there was the danger that changes in one direction would cancel out changes in the other. Here a series of simple linear regressions were carried out using a statistical computer package (SPSS; see Nie et al., 1975). Each of the three measures of activity (dependent variables) were correlated with the following quantitative independent variables: contributing area, average height, and maximum width of floodplain ratio (MAXWID). They were also correlated with the actual size of the initial activity index to which subsequent change had taken place to see, for example, if highly sinuous channels were more likely to change their sinuosity than straight/ sinuous channels. The resulting correlation coefficient indicated the degree to which the variation in the dependent variable (activity) was related to the variation in reach/basin characteristics. This also allowed a method of comparing the strength of relationship between different pairs of independent variables and activity rates. Even when no major correlation was found, this allowed qualitative assessment of the distribution of points on a scattergram.

A problem occurred with subsequent analysis that the independent variables are often highly correlated with each other. This would have made the b estimates in multiple regression unstable due to the problem of multicollinearity. For example, floodplain vegetation and height might be expected to have a significant positive correlation in many cases. Almost every watershed characteristic is in some way correlated with drainage basin area (Anderson, 1957).

To assess qualitatively the amount of variation in the nominal reach classifications in steady time, data from different map-dates were cross-tabulated in a matrix to find out if the same independent variables at different map-dates were similar or very different ie. have these variables shifted through time. Deviation from the dominant axis gave a descriptive measure of category shift. As a result of the unusual nature of the data-set (ie. the inevitability of having many expected cell frequencies less than 5), Chi squared and associated formal tests of statistical association could not be computed. For these qualitative independent variables, eg. floodplain vegetation, existence of local sediment and channel description, different forms of statistical analysis were necessary to assess any association with activity. A breakdown program (Nie et al., 1975) analysed the mean and standard deviation for each activity variable, subdivided eg. by each hydrologically significant category of floodplain vegetation. This also allowed one-way analysis of variance to be computed with an F statistic produced, to see if there was a statistically significant difference between categories (Silk, 1981). One-way analysis of variance had the advantage that equal totals within each category (which could not be guaranteed in a random sample) were not required.

4.6 Dee study area: Results of the map analysis

The results of the analysis for the Dee study area are now presented. A reference map showing the location of the 75 sampled grid-squares and a general location map are found in Figures 4.6.(i) and 4.6.(ii), respectively.

4.6.1 Basic range of activity indices

4.6.1.1 Spatial variations in sinuosity index (SIN)

The basic summary statistics for the sinuosity index at each of the three map-dates are collated in Table 4.6.1.(i). The frequency distribution for the sample of sinuosity measurements at each map-date is shown in a series of histograms (Figure 4.6.1.(i)). The skew of each distribution was highly positive (1.3-1.6) and thus normality could not be assumed in subsequent analysis.

The spatial distribution of sinuosity values within the Dee above Crathie study area are shown in Figures 4.6.1.(ii) to (iv). Not only were certain reaches considerably more sinuous than others, but within all three map editions certain patterns repeated themselves, implying a spatial recurrence of the underlying controls. High sinuosities seemed to occur most typically in upland alluvial basins, tributary confluence sites and isolated large-scale meander features on the mainstream Dee (see Figure 4.6.1.(ii)). There may however be considerable variation in sediment size, slope and channel type between each of these three situations, arising in response to the glacial legacy (Chapter 3). It

Sampled river channel segments within the Dee study area

Figure 4.6.(i)

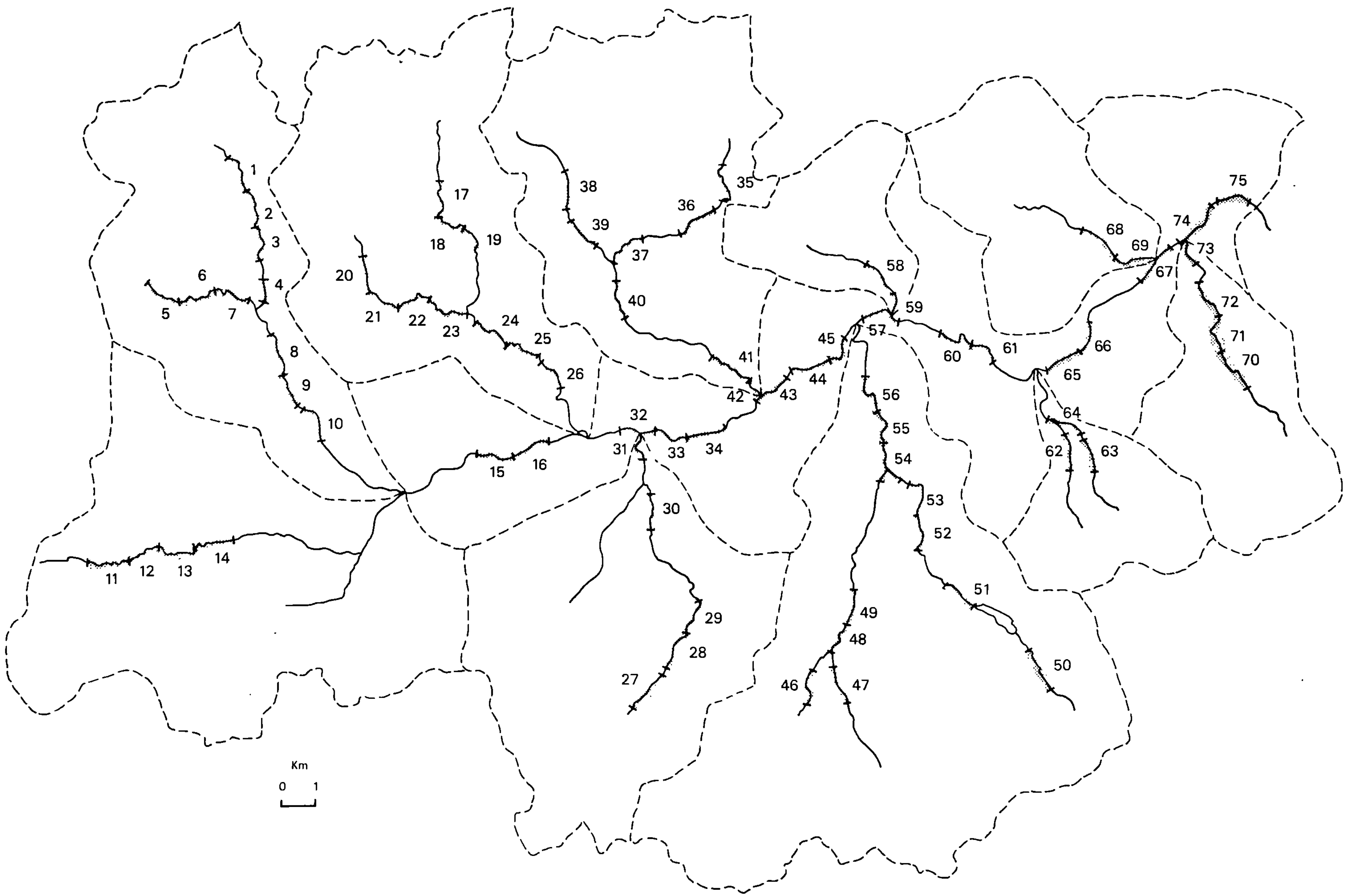


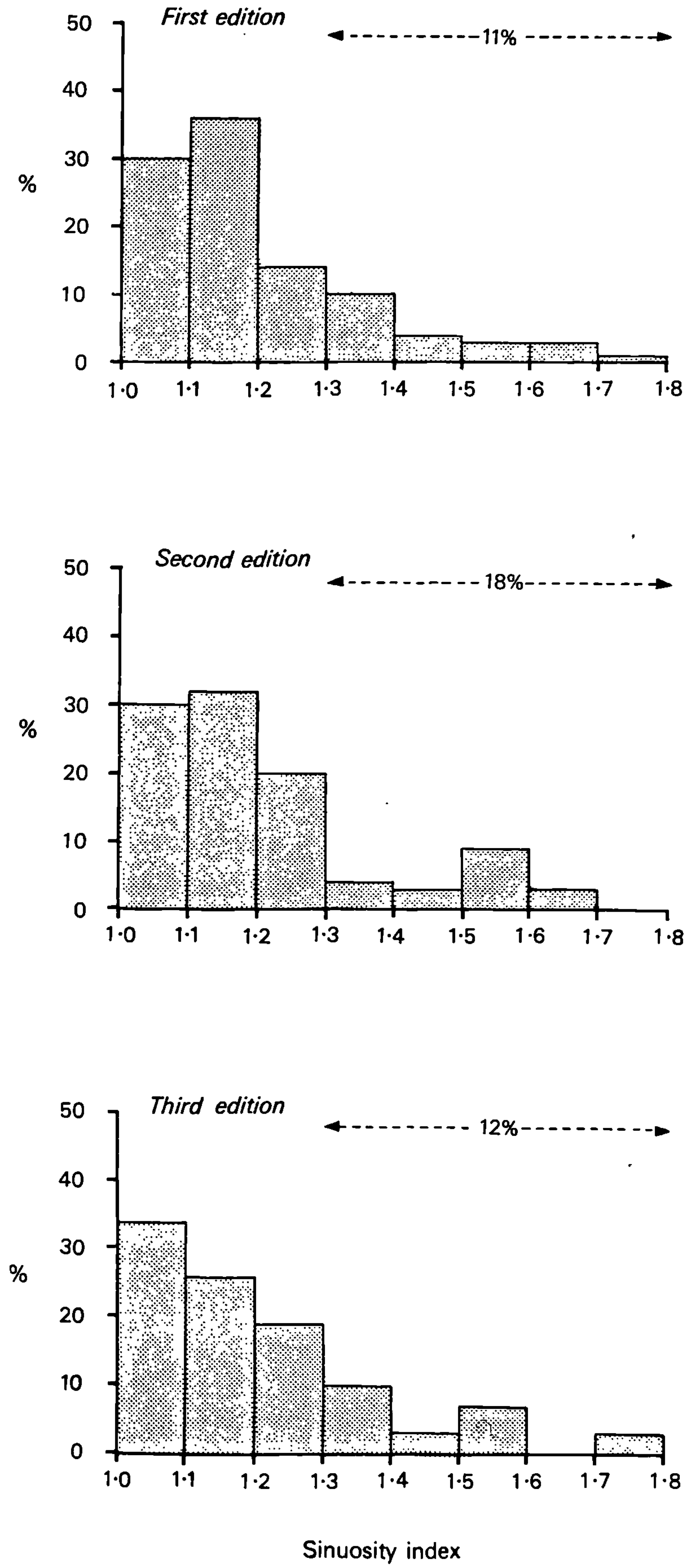
Table 4.6.1.(i)

Summary statistics for the activity indices sampled within the Dee study area

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
SIN 1 (1850)	1.08	1.14	1.24	0.16
SIN 2 (1900)	1.09	1.14	1.27	0.18
SIN 3 (1970)	1.09	1.13	1.27	0.18
Total sample				
BI 1 (1850)	1.00	1.02	1.12	0.12
BI 2 (1900)	1.00	1.05	1.13	0.13
BI 3 (1970)	1.00	1.01	1.11	0.11

Figure 4.6.1.(i)

Frequency of sinuosity index values within the Dee study area



Sampled sinuosity index values within the Dee study area for the three map dates

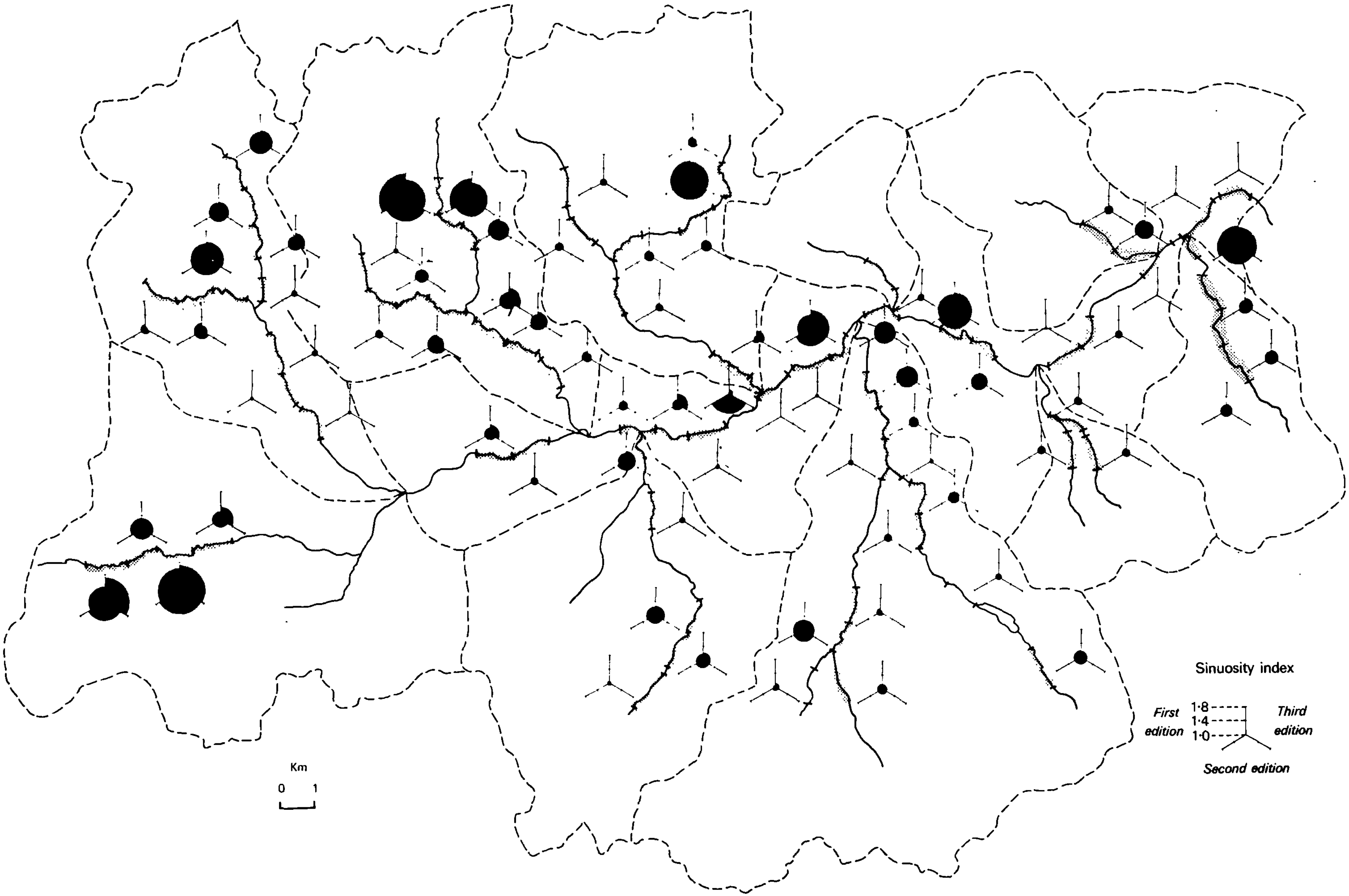


Figure 4.6.1.(iii)

Sampled sinuosity index values falling within the Clunie catchment for the three map dates

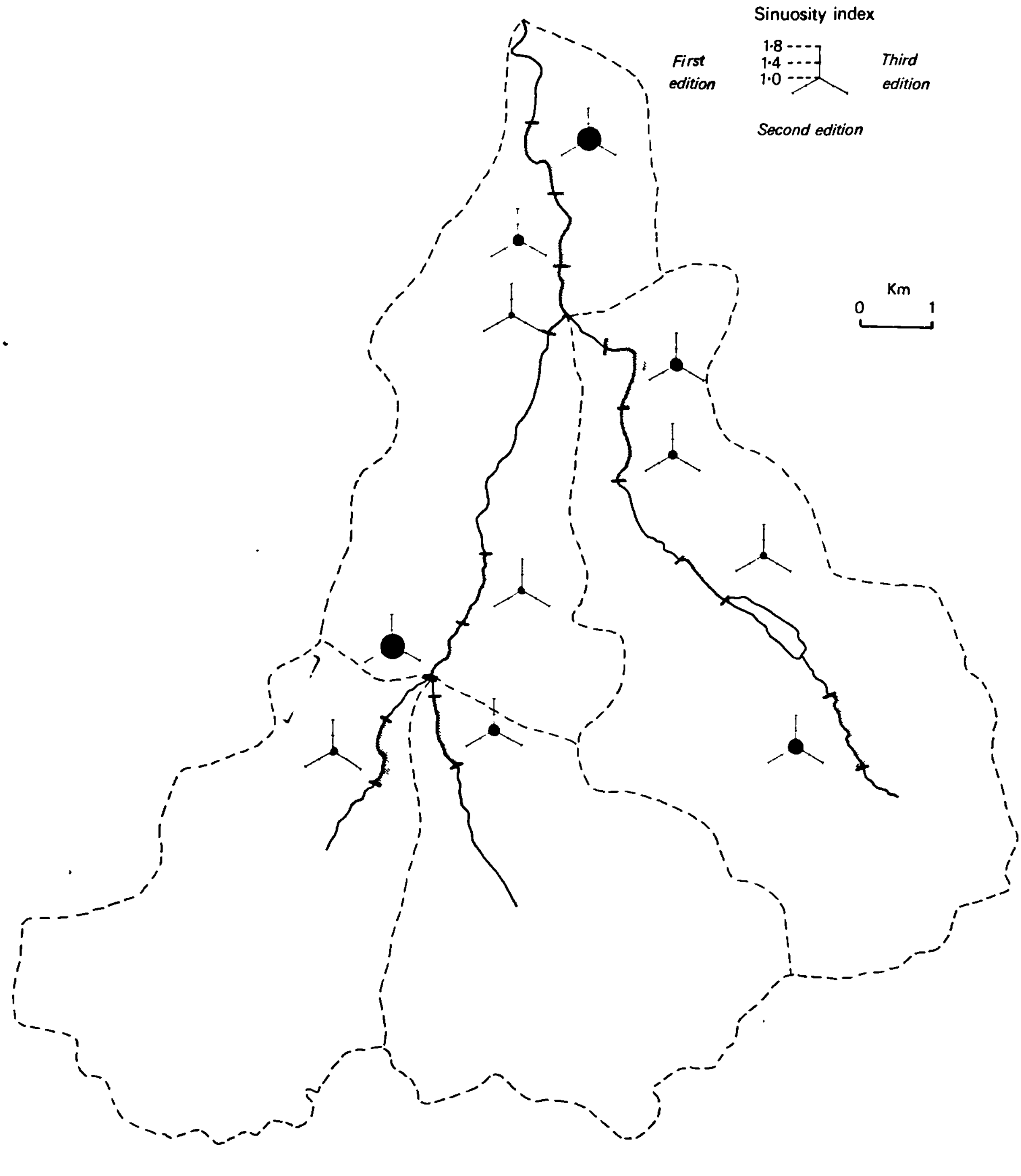
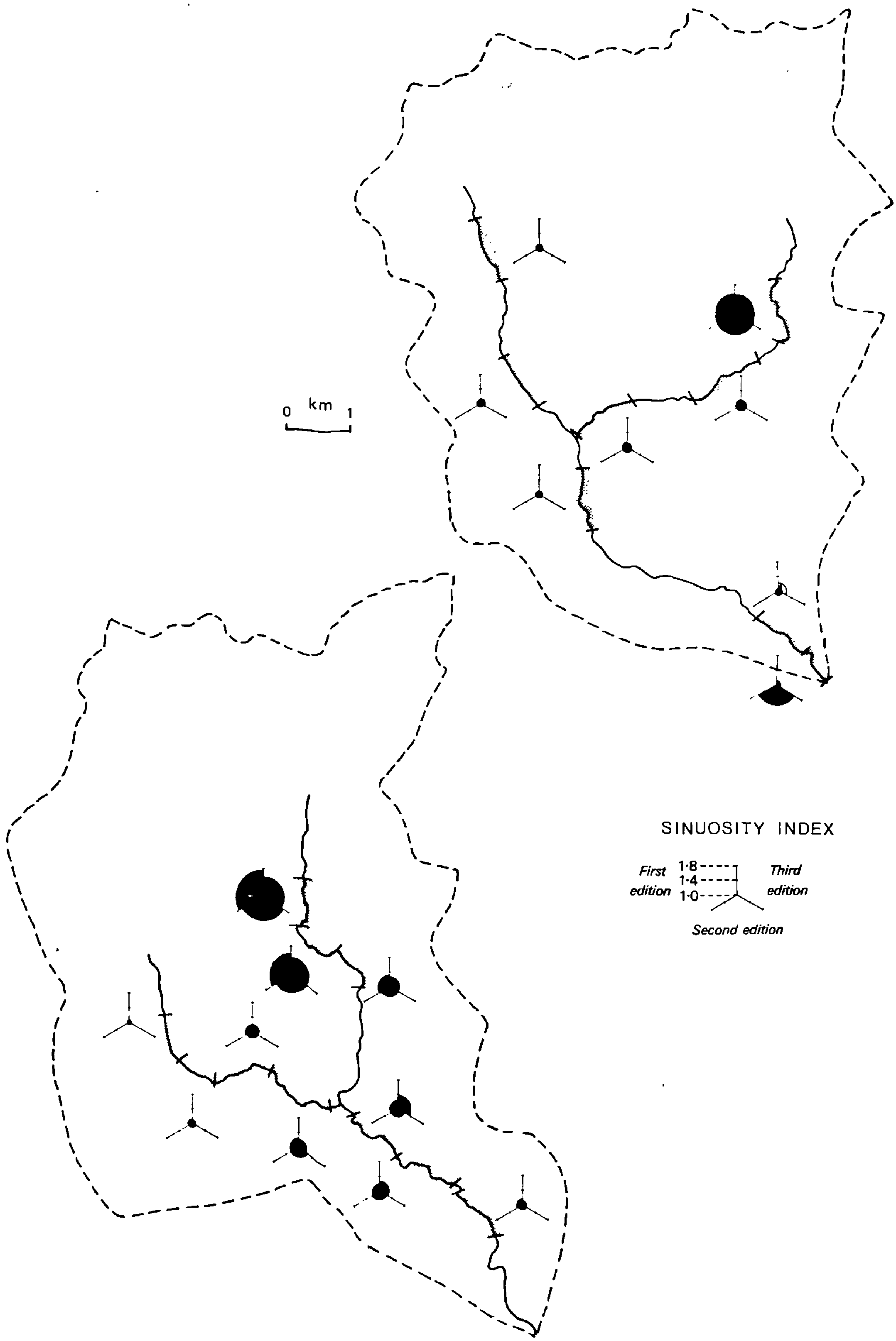


Figure 4.6.1.(iv)

Sampled sinuosity index values falling within the Lui and Quoich catchments for the three map dates



should be noted that not all the tributary confluence sites were sinuous (eg. the Lui above its confluence with the Dee where rock control was important) and there was great variation in sinuosity even where samples fell on adjoining reaches.

With the sample taken around 1869 (SIN1), the modal class (35.1%) was 1.10-1.19, with higher than average values existing at confluence sites such as on the Ey (SIN1=1.22; Sample 31) and Gelder (SIN1=1.59; Sample 73). However, it was in the tortuous meanders of the upland alluvial basins that the highest examples of sinuosity were found eg. on the upper Geldie (SIN1=1.63), the Derry (SIN1=1.75) and Glen Geusachan (SIN1=1.45) (Samples 13, 17 and 7 respectively).

The sample taken around 1902 (SIN2) had a similar frequency distribution to 1869 but there was an increase in the 1.20-1.29 class (13.5 to 20.3%) and in the 1.50-1.59 class (from 2.7 to 8.1%). The % of the sample over 1.30 had increased from 11-18% (Figure 4.6.1.(i)). Again, it was the upland alluvial basins that had the highest sinuosities with 1.63 on a Glen Derry reach (Sample 17) and 1.68 on an upper Geldie reach (Sample 13).

The sinuosity around 1970 (SIN3) had a slightly different distribution with a modal class (33.8%) of 1.00-1.09. This could be due to artificial straightening of certain reaches as there was actually a decrease in those values in excess of 1.20. Alternatively, there may have been complex response taking place within the system, gradually reducing the sinuosity to a less sinuous equilibrium condition. Two reaches with sinuosity > 1.70 occur in the tortuous meanders of the

Figure 4.6.1.(v)

Frequency of braiding index values within the Dee study area

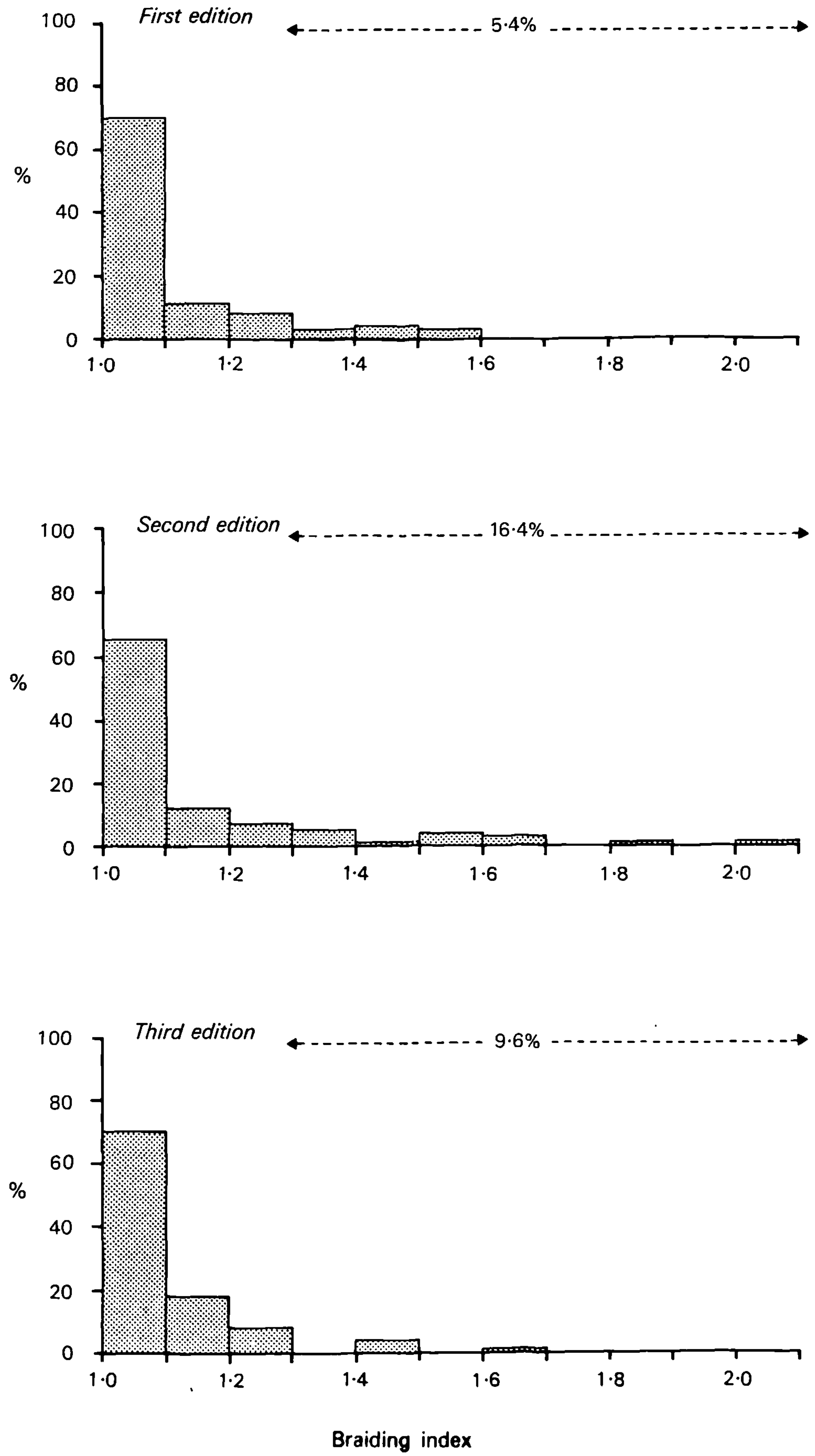


Figure 4.6.1.(vi)

Sampled braiding index values within the Dee study area for the three map-dates

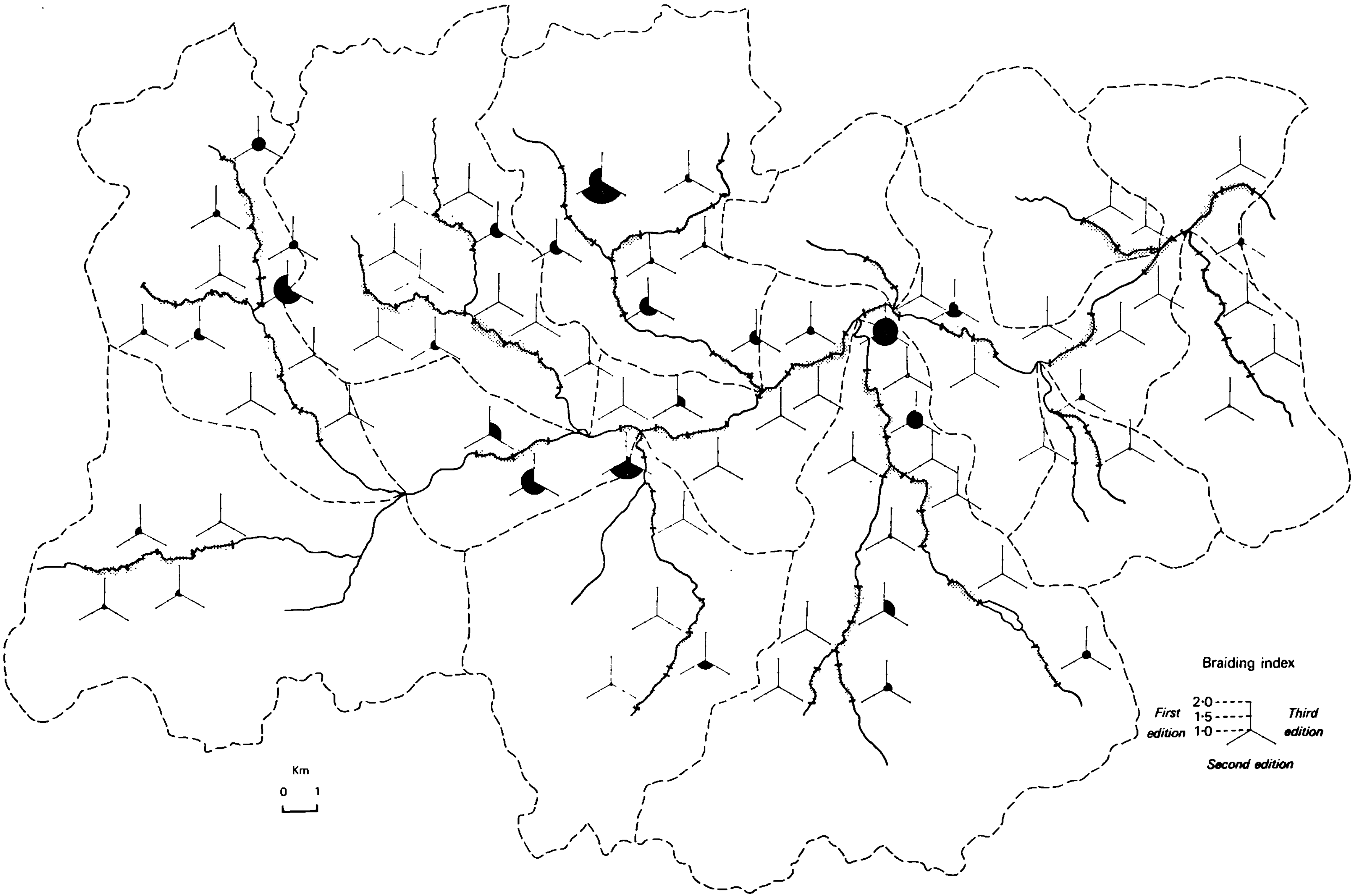


Figure 4.6.1.(vii)

Sampled braiding index values falling within the Clunie catchment for the three map dates

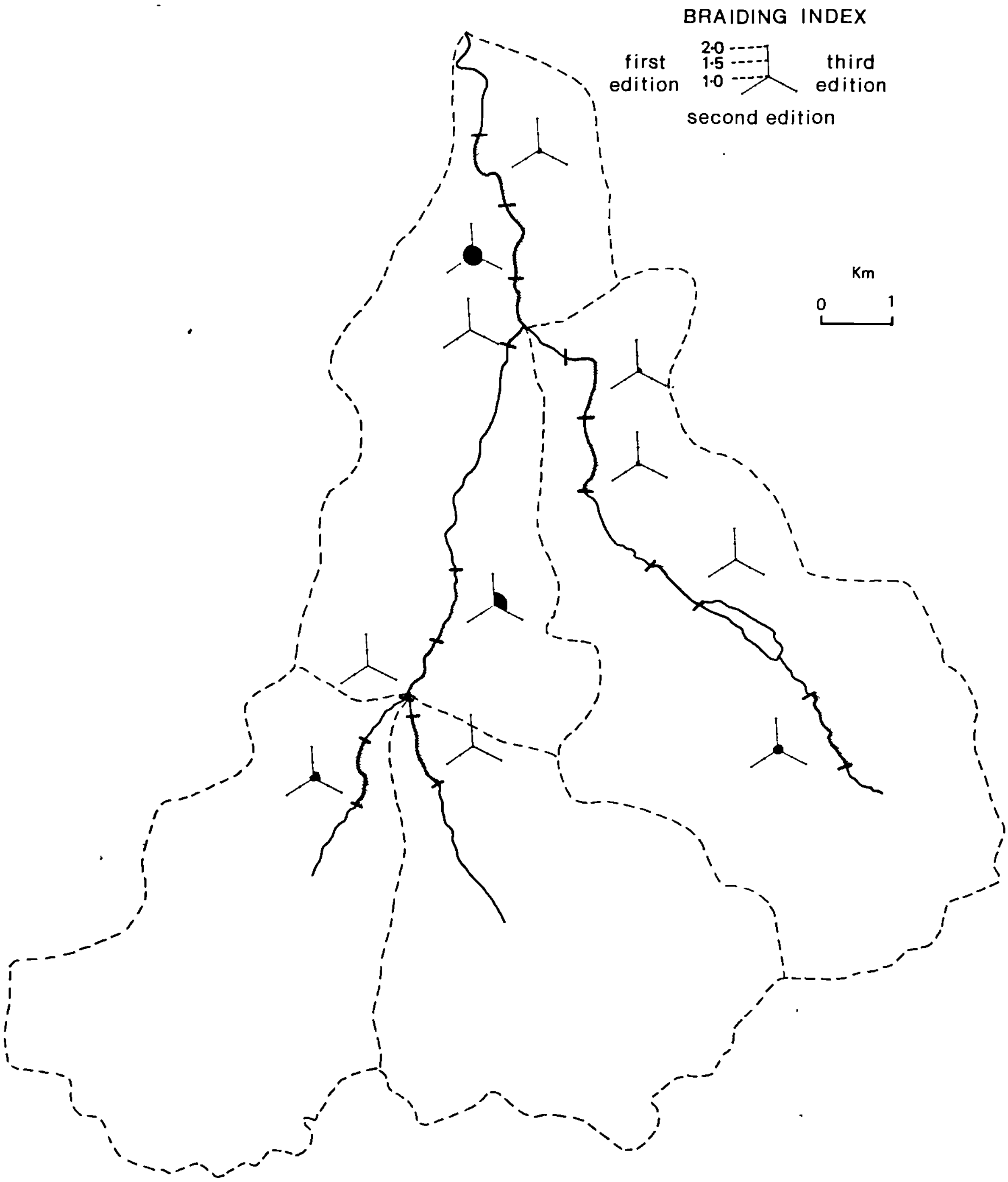
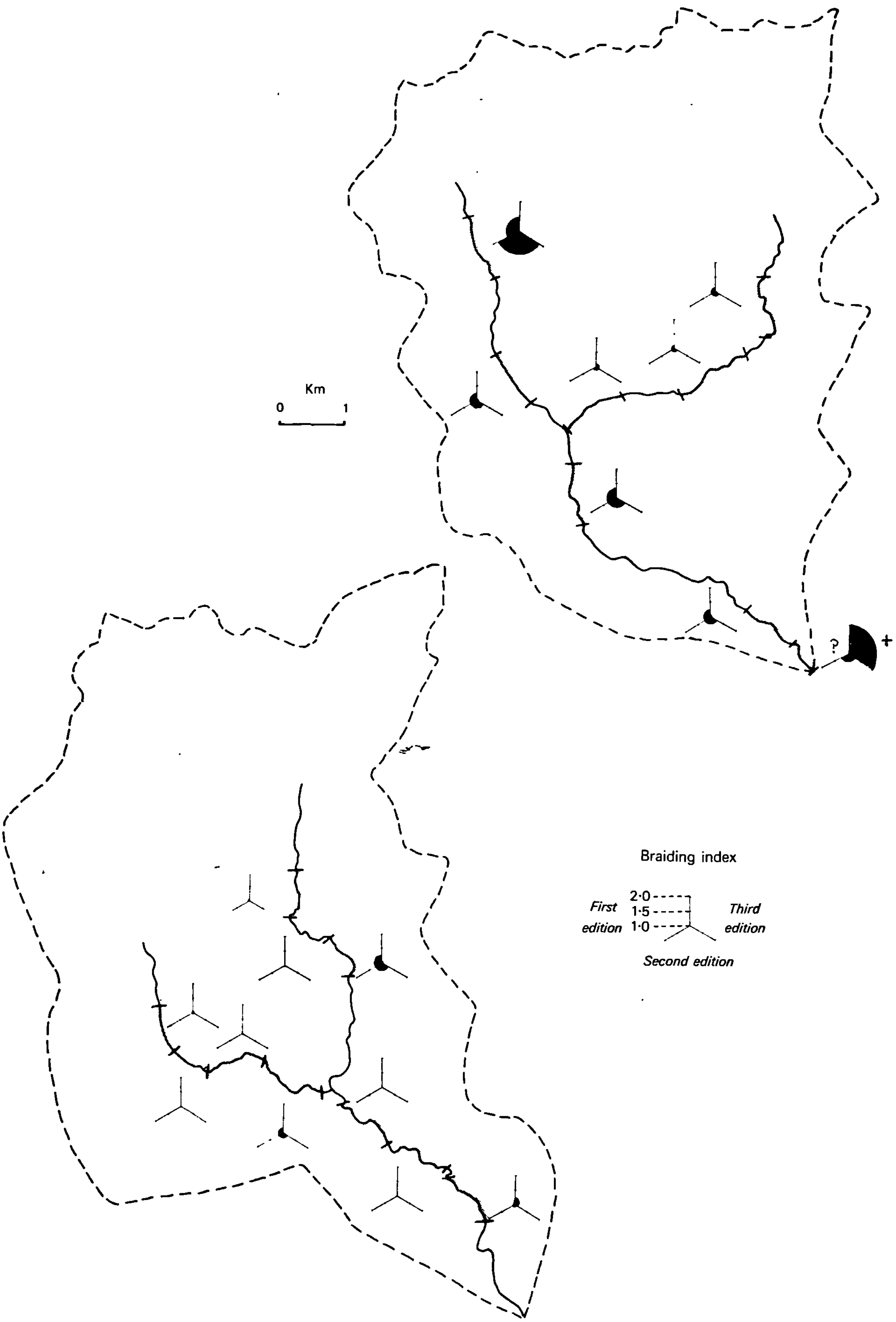


Figure 4.6.1.(viii)

Sampled braiding index values falling within the Lui and Quoich catchments for the three map dates



upper Geldie (Samples 11 and 13). Other reaches with notable high values were on the mainstream Dee (SIN3=1.52; Sample 45), on an irregular reach above the Clunie confluence (SIN3=1.32; Sample 56) and at an irregular meander bend near Clunie Cottage (SIN3=1.51; Sample 60; Dee study reach 8).

4.6.1.2 Spatial variations in braiding index (BI)

The basic summary statistics for the braiding index, for degree of energy dispersion, at each of the three map dates are collated in Table 4.6.1.(i). The distribution of braiding index classes at each map-date are shown in histogram form in Figure 4.6.1.(v), and the spatial distributions within each catchment are shown for each map edition in Figures 4.6.1.(vi) to (viii). As can be seen from Table 4.6.1.(i), the median value of the braiding index for the first edition sample (1869) was 1.02 with a IQR of 0.12. No bars or only very localised features (BI=1.00-1.09) were indicated on 71.2% sampled river segments and only 5.4% had a BI > 1.20. In terms of inter-tributary variation, some catchments (eg. the Ey) had very few bars within the sample. Possible explanations include greater confinement by high terraces or bedrock, and perhaps lesser availability and accessibility of fluvioglacial material to be reworked within the channel. In contrast, other catchments eg. the Lui and upper Dee had considerable spatial variation between closely adjoining reaches. This may suggest periodic sediment storage downstream, though a complete analysis of the total stream length population would be required for detailed study. Certain reaches stood out as having comparatively high BIs at this time, for example, the upper Dee had a value of 1.51 (Sample 4) and the Quoich confluence (Sample 42) had a highly disrupted pattern with a suspected high BI value (see Section 7.2.1).

Within the second edition sample (1902), the median value of 1.05 and a IQR of 0.13 indicated several high BI values and 16.4% of the sample had BI > 1.3 (cf. the first and third editions). The most extreme value by far (8.80) was found upstream of the Quoich confluence with the Dee, where a complex distributary fan with chaotic braided/ reticulate pattern occurred (Sample 42; see Section 7.2.1). Again other "above confluence" sites were characterised by high values, eg. Ey (1.68; Sample 31), Gelder (1.81; Sample 73). Other reaches with high BI values included upland confluence sites, associated with higher slopes than the lower tributary fans, where there were anabranches or major splits in the channel. For example, this occurred above the confluence of the upper Dee with the Geusachan Burn (Sample 4). Other high values occurred on the mainstream Dee where large stable bars yielded high values eg. on the Dee above the confluence with the Gleann an t-Slugain (Sample 57).

When studying BI3 values (1970), the most obvious feature was the range from 1.00 to 1.61, around a median of 1.01. A large number of sites thus had very low BI indices. There were however exceptions; a site on the mainstream Dee, where a wandering channel split around a large, stable bar, generated a value of 1.47 (Sample 57). Two other sites, one on the mainstream Dee (Sample 15) and one on the upper Clunie (Sample 49), displayed BI3 > 1.4. Certain catchments stood out because islands were either non-existent or occasional eg. the Gelder, the upper Clunie, and the upper Geldie. However in most catchments, there was an interspacing of sites with varying BI values. Often fairly high values of braiding index were followed downstream by sample reaches with much lower values eg. on the Lui Water. This suggested high variability in local controls, particularly the availability of material and erodibility of banks, again emphasising the importance of local sediment storage.

4.6.2 Basic range of typology characteristics

4.6.2.1 Importance of quantitative catchment characteristics

When the activity indices at each map date were correlated with the catchment parameters, the following observations were made. The sinuosity index and braiding index samples for each map-date were correlated with drainage basin area, average height and and ratio of maximum width of flood plain (MAXWID), but only those obtaining significant results are discussed here. Although the significance and the strength of the association (r^2) may vary as a powerful predictor, the distribution of points on the scattergram can tell us in qualitative terms something of the nature of these parameters influences. A significant correlation (r) with a sample size of 70 was 0.23 (0.05) and 0.30 (0.01).

Sinuosity was not significantly correlated with contributing basin area at any of the three map-dates. Trends appeared cyclical rather than linear or curvilinear. The reaches with smallest contributing catchment areas tended to have a low sinuosity but there was a sudden increase with only a small rise in drainage basin area when the upper reaches opened up abruptly into alluvial basins. For example, high sinuosities of 1.57-1.75 occurred over the three map-dates in Glen Derry (Sample 17), with a contributing area of 14.1 km^2, and 1.63-1.72 on a reach in the upper Geldie (Sample 13), with a contributing area of 18.4 km^2. However, there was not a simple decline in sinuosity with an increase in drainage basin area. Although there did seem to be a decline to a certain extent, some samples with a drainage basin area in excess of 300 km^2 again showed high sinuosity values eg. Sample 45 (SIN1=1.39-1.52; drainage basin area of 360.1 km^2) and Sample 60 (SIN1=1.46-1.51; drainage basin area of 481.2 km^2). It seemed that downstream distance reflected principally the macro-scale control of the localised intensity of glacial erosion. Longitudinal profiles (Figure 3.2.1.(iii)) showed lower slope basins, interspaced by rock-controlled reaches which must act as local baselevels eg. on the Clunie.

There was no clear positive or negative relationship between braiding index with ($\log_{10}$) contributing basin area, with low braiding indices occurring throughout the range of drainage basin area. At one end of the scale, there were very few high braiding indices with catchment areas in excess of 400 km^2 ie. the mainstream Dee rarely appeared to have undergone large-scale sedimentation with bars, implying that sediment was either stored within the channel margins or frequently flushed through the system. Only two sample sites appeared to represent sediment stores (Sample 57; Sample 60, Section 7.2.8). At the other extreme, there was a considerable range in BI for lower catchment areas. Contributing area did not seem very significant in influencing distribution of the range of BI values, however the envelope of maximum braiding index did seem to increase with area. Similar results were found for average height, again reflecting the dominant control of extensive glacial erosion.

When sinuosity was regressed against ($\log_{10}$) ratio of channel width to maximum possible width of floodplain (MAXWID), a significant positive correlation (r=0.23-0.25) was found on samples for all three map-dates. Reasonably high sinuosities in excess of 1.50 associated with confined meanders, eg. within rock-controlled reaches, occurred with minimal floodplain width. However, there was a general tendency at all three map-date for maximum sinuosity to increase as available active area increases.

Of the three quantitative variables, the most important seemed to be MAXWID when studying the distribution of BI values. BI3 and BI2 showed significant positive correlations with available width (especially BI2; r=0.32), while at BI1 the relationship was less well defined. The BI3 and BI1 channel planforms seemed to make less use of the available floodplain area for either planform migration or planform expansion. A threshold may have to be exceeded for floodplain reutilisation, as occurred by BI2. In contrast BI3 and BI1 may represent periods of comparative channel relaxation. Obviously, the flood history must be studied (Chapter 5) to assess the varying periodicities of stresses applied to the system. These correlations were still relatively low but are higher than those associated with either height or area. Width of available active area would therefore appear more important than position within the drainage basin, although none of these parameters represents a powerful predictor of. the Tweed study area.

4.6.2.2 Frequencies within the qualitative classifications

The frequency distributions of the qualitative parameters are shown in Figure 4.6.2.(i). In terms of upstream availability of sediment, the reaches with no evidence of sediment proximal to the channel were dominant, with 59.5% (third edition), 68.9% (second edition) and 73% (first edition). All three editions had some localised upstream sediment availability, with 9-10% of the samples having large supplies. In terms of local sediment availability, the majority of samples had no sediment proximal to the channel, implying a limited local sediment supply. However, again there were individual samples that had large amounts of available sediment.

The dominant vegetation types are shown in Figure 4.6.2.(i), with the most frequent floodplain and bank vegetation being heather/heath (category 4). This indicated a general stability of floodplain areas. However, the number of vegetation covered bars with anything more than rough-grassland was limited at all three map dates, implying a general instability and periodic reworking of bar forms within the channel system. The distribution of channel pattern types approximated a symmetrical distribution at all three map-dates but with a slight skew to the left and a tailing off to the right. Most channels fell in categories (2) sinuous/ straight to (5) irregular (Figure 4.6.2.(i)). The incidence of islands had a far more skewed pattern, with 87.8 to 95.9 % in categories 1 (no islands) to 3 (frequent islands).

Figure 4.6.2.(i)

Frequencies within each channel typology classification for the Dee study area (for detailed key, see Table 4.2.(i))

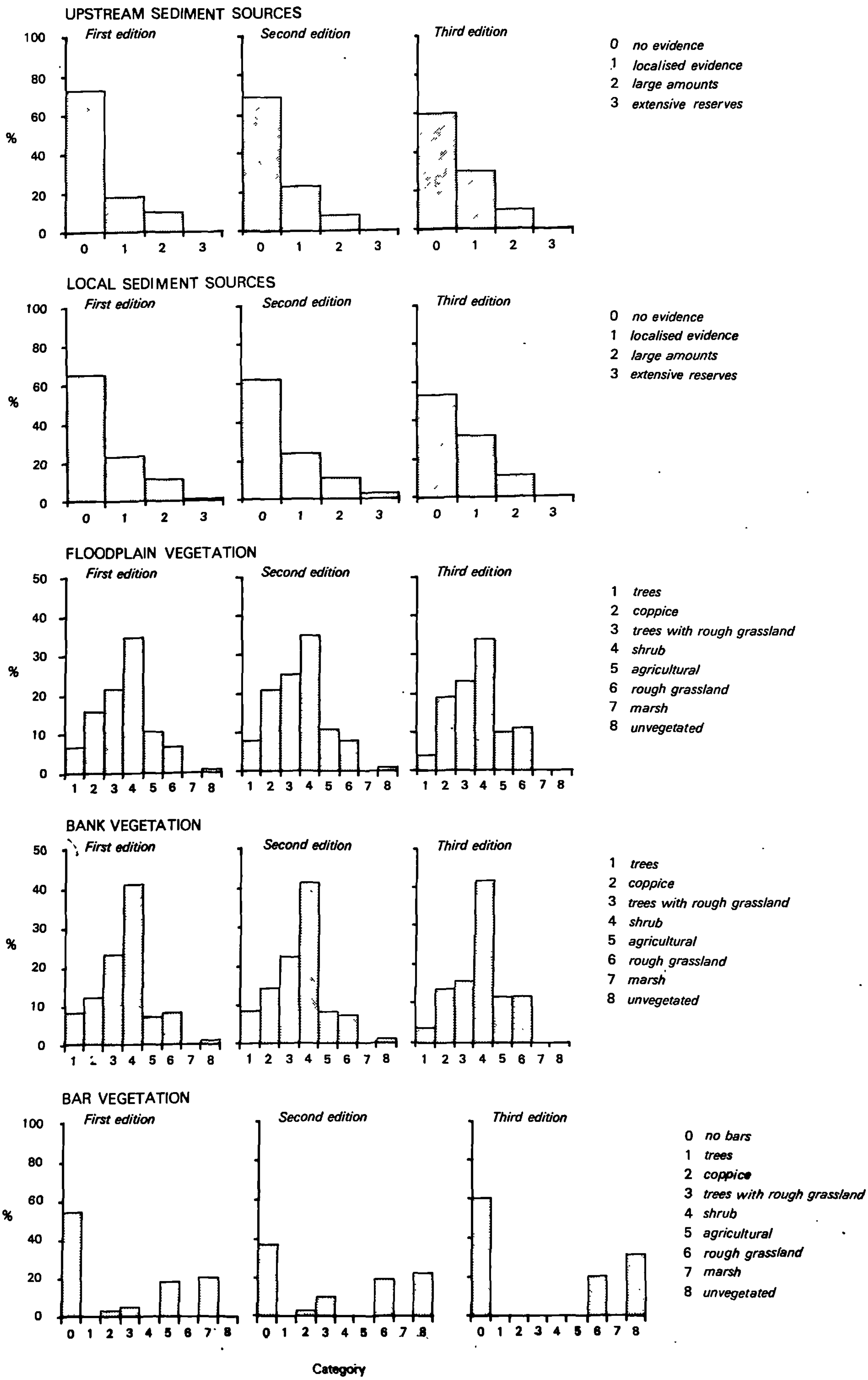
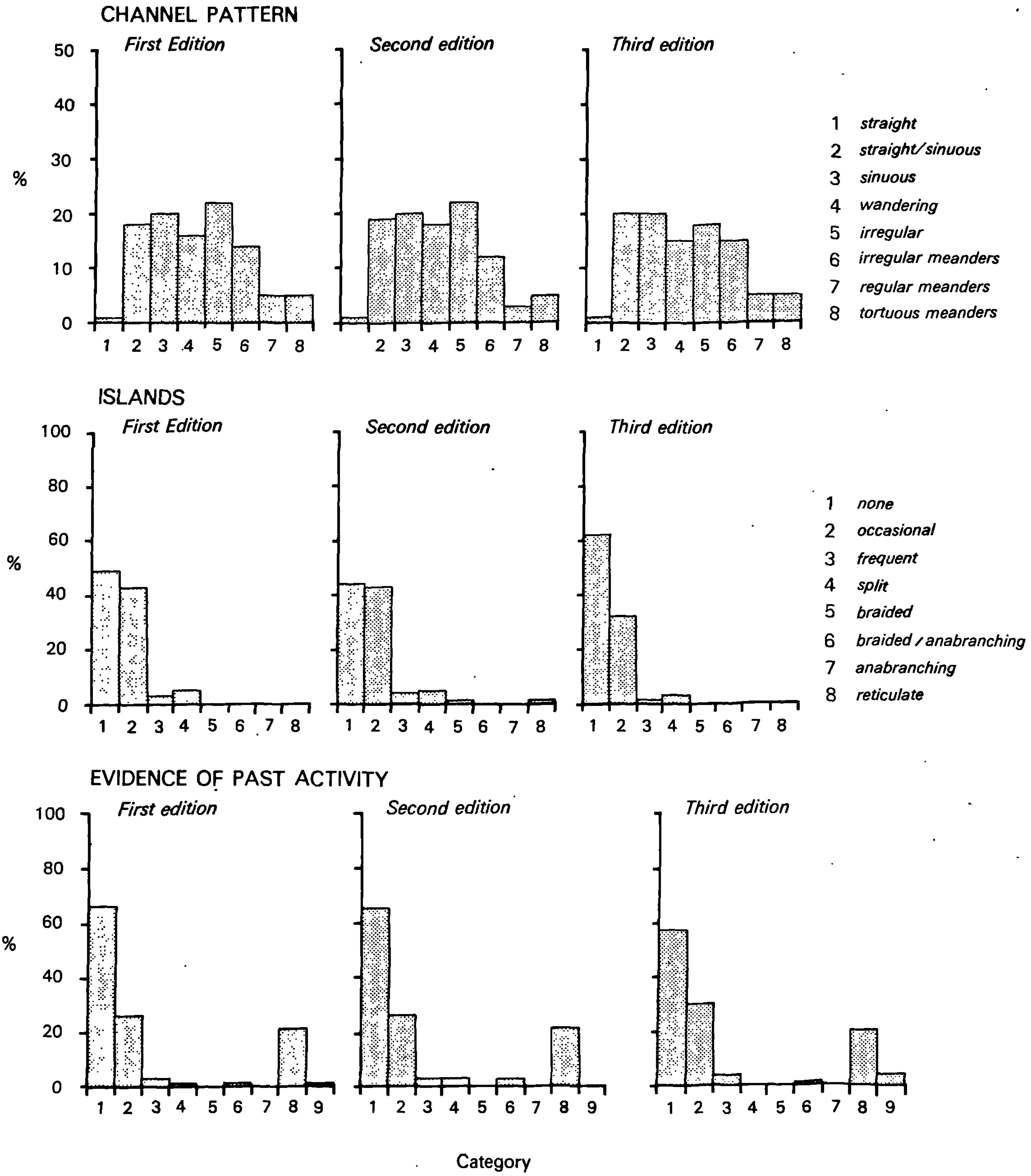


Figure 4.6.2.(i) cont.

Frequencies within each channel typology classification for the Dee study area



1 *no evidence*
2 *unvegetated gravels*
3 *progressions and cutoffs*
4 *irregular channel activity*
5 *avulsion*
6 *old palaeochannels*
7 *marsh*
8 *combination of 2 and 6*
9 *evidence of older palaeochannels*

Although a large proportion of sites (57 - 66%) had no indication of any lateral activity, there were examples of a variety of past palaeofeatures. This absence may in certain cases be because past features have been reworked. Examples where palaeofeatures did exist were the palaeomeanders on the lower Clunie and the mainstream Dee above Braemar. These were relicts of meander patterns present at the time of Roy's map (1750) and earlier. Comparisons with Roy's map yielded other planform changes. For example, the Clunie/ Dee confluence was split, sinuous and tree-lined in 1750 and on 1734 estate plans (RHP 3491) indicated a highly active reworking of neighbouring agricultural land. This was in contrast with the stable, rather artificial appearance of the first edition planform of 1869 (Figure 6.2.(iv)). Other examples are detailed in the case studies in Chapter 7.2. Contrasting features found to be much more stable over the studied timespan, with no obvious palaeofeatures, included the major island on the Dee (Sample 57) above its confluence with the Gleann an t-Slugain, which has existed without major modification since at least 1734 (Figure 6.2.(iv) and 7.2.3.(i)).

Map-based information about the temporal frequency of flooding was sparse. Incidence of flooding was not indicated on either the first or third edition samples, however on the second edition, it has been considered frequent enough to be indicated twice, namely on the low-lying ground beside the Quoich confluence fan (Sample 42; see Dee study reach 7.2.1) and on the mainstream Dee beside Clunie Cottage (Sample 60; see Dee study reach 7.2.8). Both sites will be discussed in further detail with the aerial photograph analysis (Chapter 7).

4.6.3. Temporal variation in channel pattern

4.6.3.1. Modes of channel adjustment

A variety of different types of planform adjustment and change were found within the Dee study area. These ranged in scale from slight local migration to examples of avulsions, progressions and cutoffs. Change can thus be subdivided into those which result in an increase, decrease or maintenance in the width of the active area. They can also be subdivided into extra-channel and intra-channel changes ie. changes that occur outwith the former active area on to previously undisturbed floodplain (or with no evidence of recent disturbance), in contrast to a channel switch within the present active area. The different categories of change are outlined in Table 4.6.3.(i) and from this range, a series of examples have been selected from the digitised maps, as seen in Figure 4.6.3.(i). It should be noted that if a channel shifts abruptly (eg. in response to a major flood event) then the active area would be expected to increase suddenly. Restabilisation of the floodplain area, which has been abandoned post-flood, is a much slower process.

The following categories were associated with an increase in width of active area (Table 4.6.3.(i), Section A). Category 1 occurred when the channel reach underwent a complete disruption of channel planform, from a dominantly undivided to braided pattern eg. the Quoich and Ey confluences (Samples 42 and 31). Such a change was however transient. Categories 2 to 4 involved flow onto previously unused or well restabilised parts of the floodplain or the reuse of former channel alignments, so that flow either became divided or completely diverted

Table 4.6.3.(i)

Categories of channel pattern change found within the Dee study area

(A) Channel pattern change associated with an increase in width of active area.

1: Channel metamorphosis of a transient nature

2: Migration/ avulsion of channel across a fan

3: Major avulsion outwith the present active area on unused floodplain

4: Minor avulsion outwith the present active area on unused floodplain

5: Major intra-channel switch/ division within the present active area

6: Minor intra-channel switch/ division within the present active area

7: Increase from single channel widening with intra-channel bars.

8: Meander progression, rotation and translation

9: Formation and subsequent reworking of medial bars within channel.

(B) Channel pattern changes associated with a reduction in width of active area.

1: Channel metamorphosis of a transient nature

2: Large scale reduction from a reticulate planform to a much less divided channel form

3: Large scale reduction from multichannelled/ split channel to a dominantly single channel planform.

4: Small scale local reduction in the number of channels

5: Channel with frequent medial bars to channel with fewer or no bars

6: Large scale reduction in channel sinuosity through major cutoffs: (a) neck (b) chute

7: Small scale reduction in channel sinuosity through minor cutoffs: (a) neck (b) chute

(C) Maintenance of an equilibrium channel dimension

1: Steady lateral migration of channel across floodplain

2: Progression across the floodplain maintaining initial irregular form

Figure 4.6.3.(i)

Examples of categories of channel pattern change sampled within the Dee study area

Sample 42

Sample 31

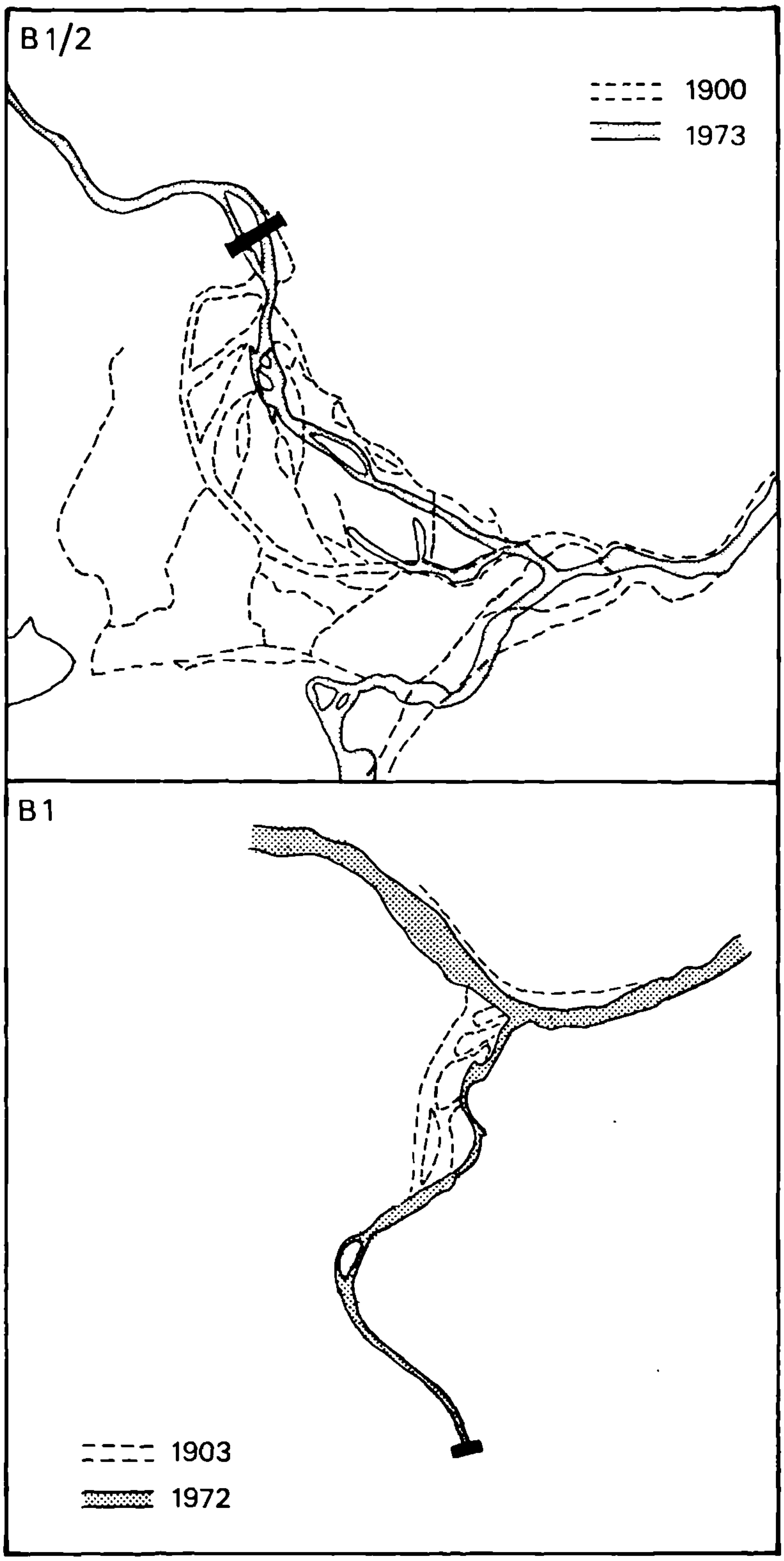
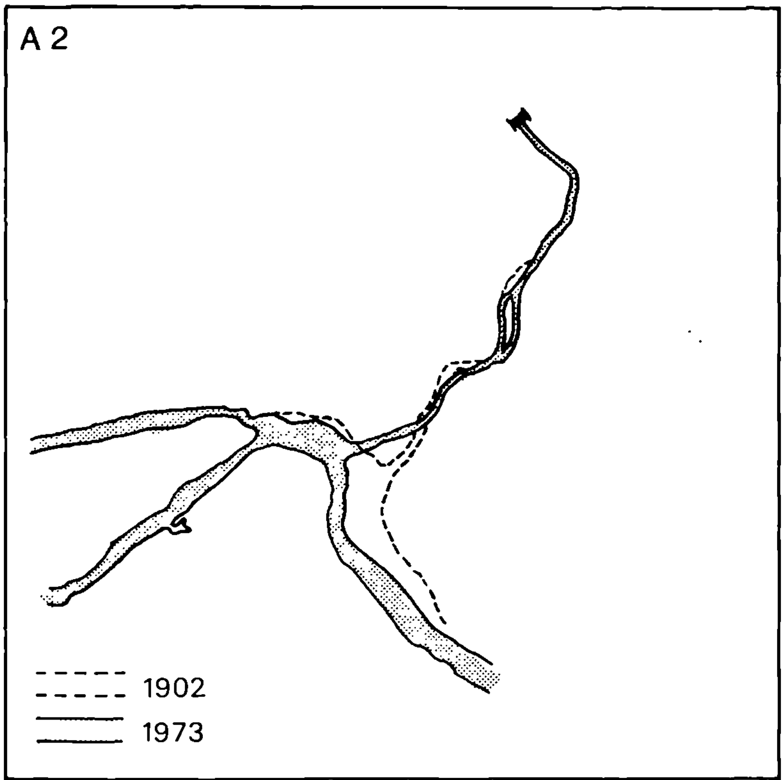


Figure 4.6.3.(i) cont.

Examples of categories of channel pattern change sampled within the Dee study area

Sample 59



Sample 33

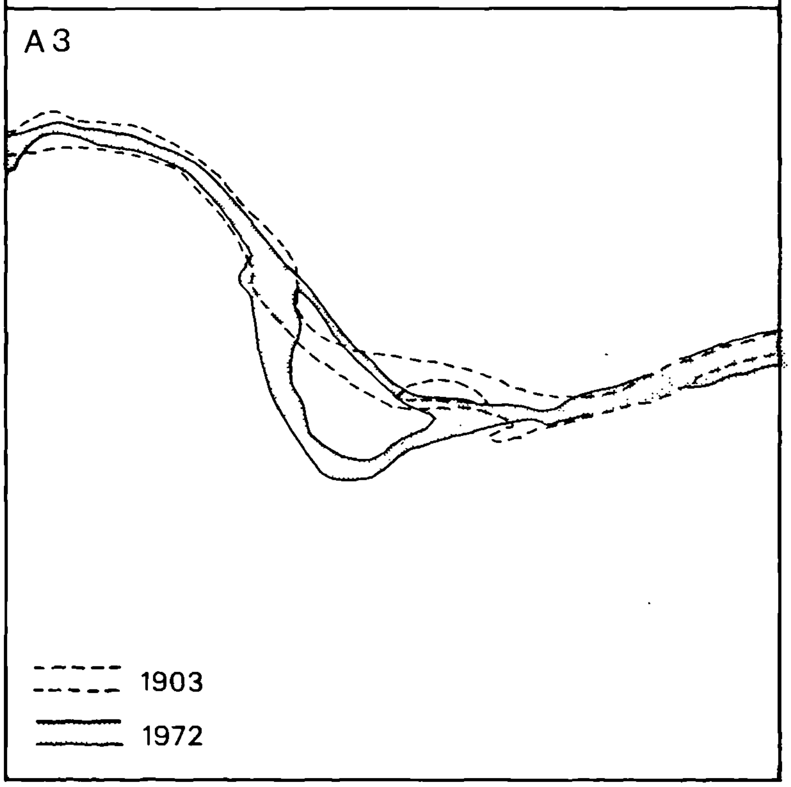
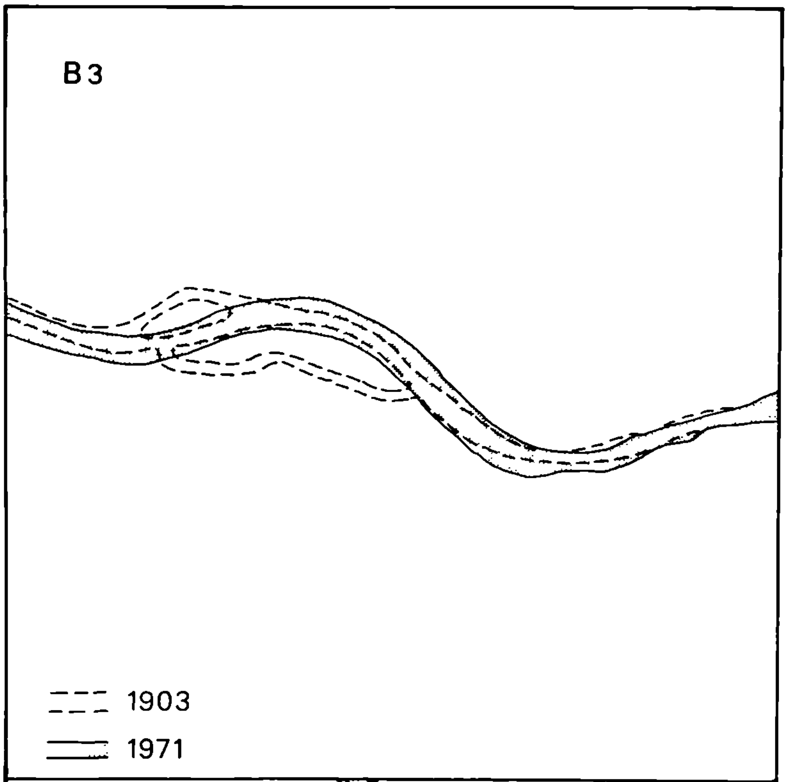


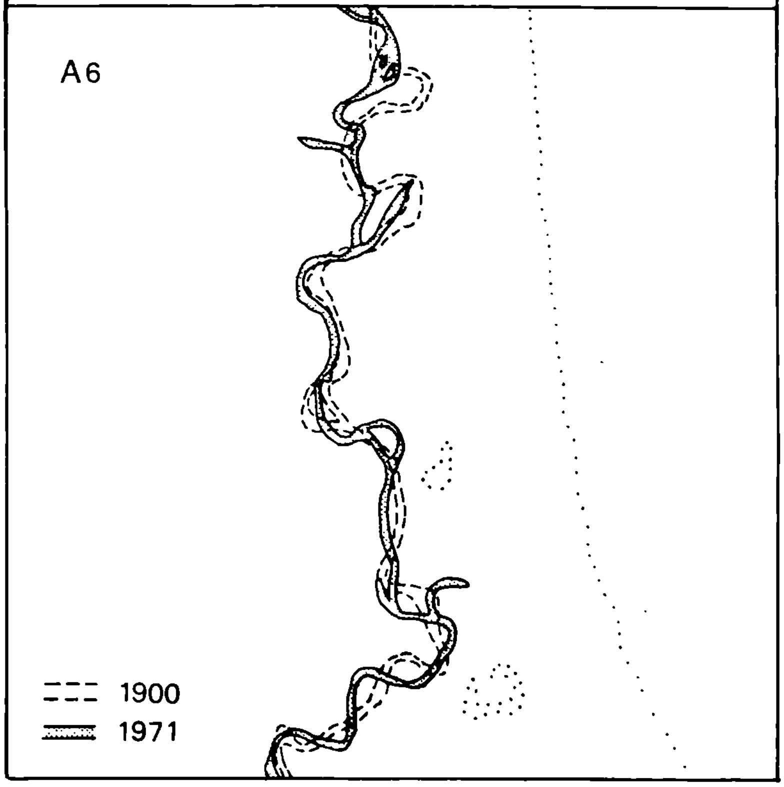
Figure 4.6.3.(i) cont.

Examples of categories of channel pattern change sampled within the Dee study area

Sample 15



Sample 17



through a new or reworked avulsion eg. the Gleann an t-Slugain confluence (Sample 59; Figure 4.6.3.(i)). Categories 4 and 5 involved the removal of a plug of sediment from either an old flood channel or an old bar surface, allowing the cleaning out and reactivation of an old channel. This however occurred principally within the confines of the active area. Redistribution of sediment within a reach as well as downstream sediment transfer may produce a new channel planform. As in Categories 2-4, flow became divided or switched to a new route. Category 7 occurred when sediment derived from localised erosion of banks was reworked into bars locally or downstream of source. This may involve the movement of plugs of sediment, stored in different bar forms, downstream. Categories 8-10 were associated with more sinuous planforms (categories 5-8 in the channel pattern typology) eg. Figure 4.6.3.(i), but mainly through a spatially isolated occurrence.

The following categories were associated with a decrease in width of active area (Table 4.6.3.(i), Section B). Category 1 occurred when a channel underwent a complete disruption (temporary or otherwise), which involved a change from a dominantly braiding to undivided planform (eg. Sample 42 at the Quoich confluence). Examples of categories 3-5 are shown in Figure 4.6.3.(i). Categories 6 and 7 were associated with sinuous meander bends, however sinuosity did not seem to have to be very high before such change could occur eg. the chute cutoff on the mainstream Dee (Sample 33).

Maintenance of an equilibrium channel dimension frequently occurred (Section C). In both these cases (Table 4.6.3.(i)), the dimensions of channel planform remained broadly similar. Maintenance of planform appeared to be associated with lower process rates. However, major discharge events did tend to reactivate channel side gravels or cause local channel-side sedimentation even on more stable reaches, if only temporarily.

4.6.3.2 Rates of change as indicated by sinuosity index (CHASIN)

When the SIN3, SIN2 and SIN1 samples were subjected in paired form to the Wilcoxon test, there was found to be no significant difference between the different map-dates indicating that overall channel sinuosities have remained similar. However, it was clear from the changing shape of the frequency distributions that individual reaches had undergone positive or negative adjustments over both time intervals and this may be of geomorphic significance. There was for example an increase in sinuosity values > 1.3 between the first and second edition (from 11 to 18%) (see Figure 4.6.1.(i)). This however represented principally an increase in main channel sinuosity around bar features rather than an increase in meandering planforms.

Values were plotted in a similar way to the "moment in time" sinuosity values, to ascertain the spatial distribution of these temporal changes (Figure 4.6.3.(ii) to (iv)). Sinuosity values were also plotted against time for each catchment to reduce clutter on one diagram

Figure 4.6.3.(ii) Sampled changes in sinuosity index within the Dee study area for the three map dates

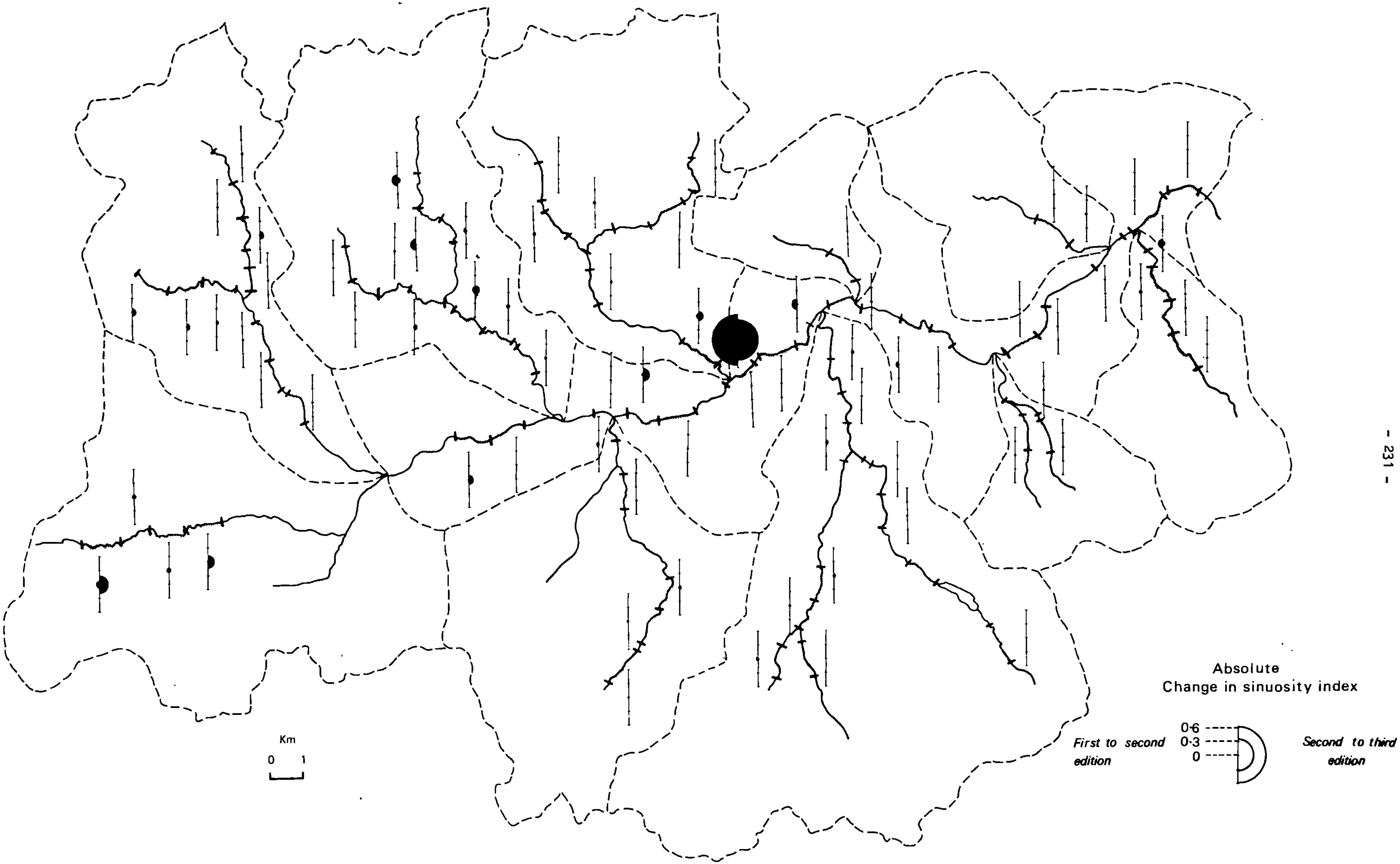


Figure 4.6.3.(iii)

Sampled changes in sinuosity within the Clunie catchment for the two intermap periods

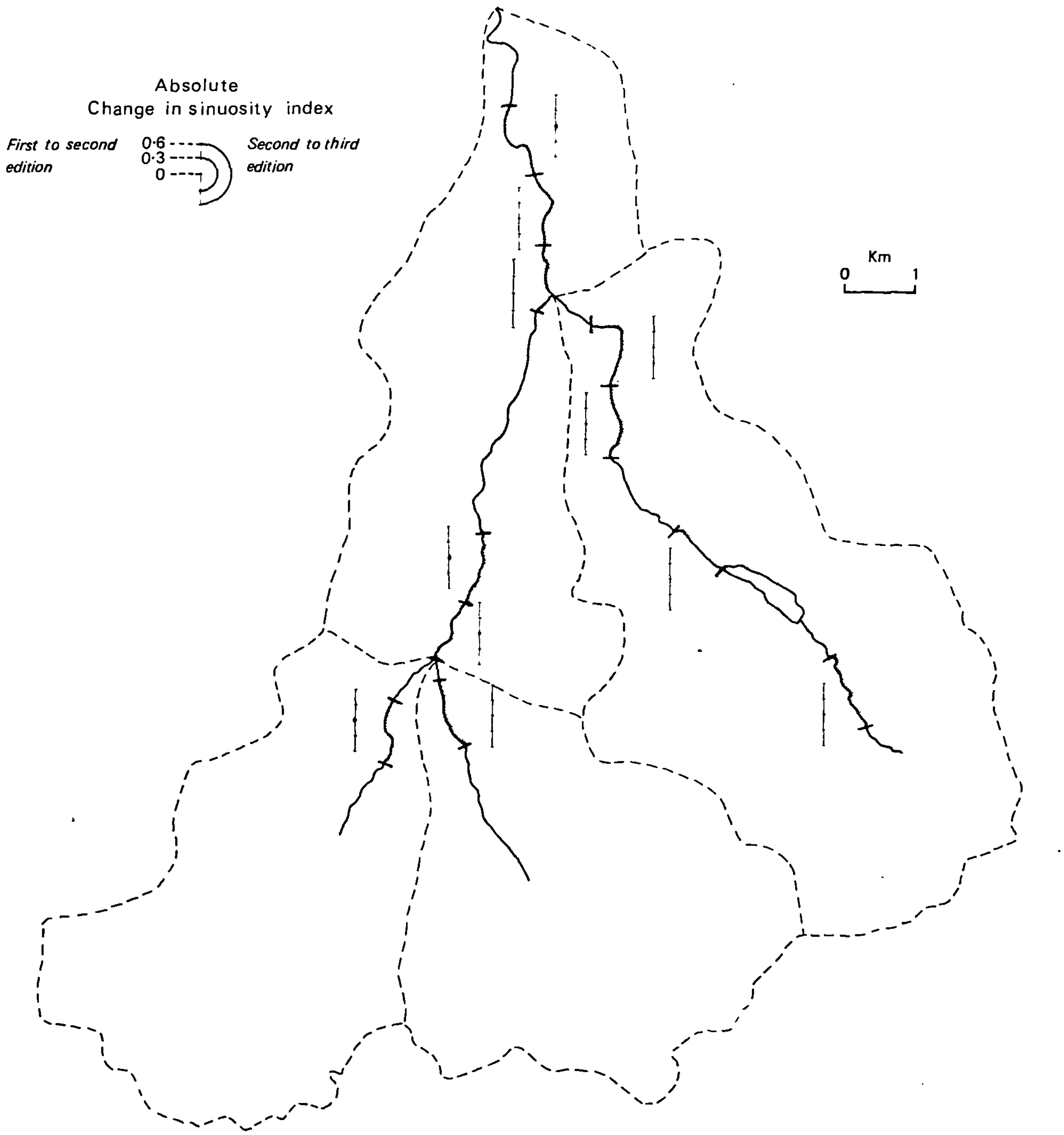
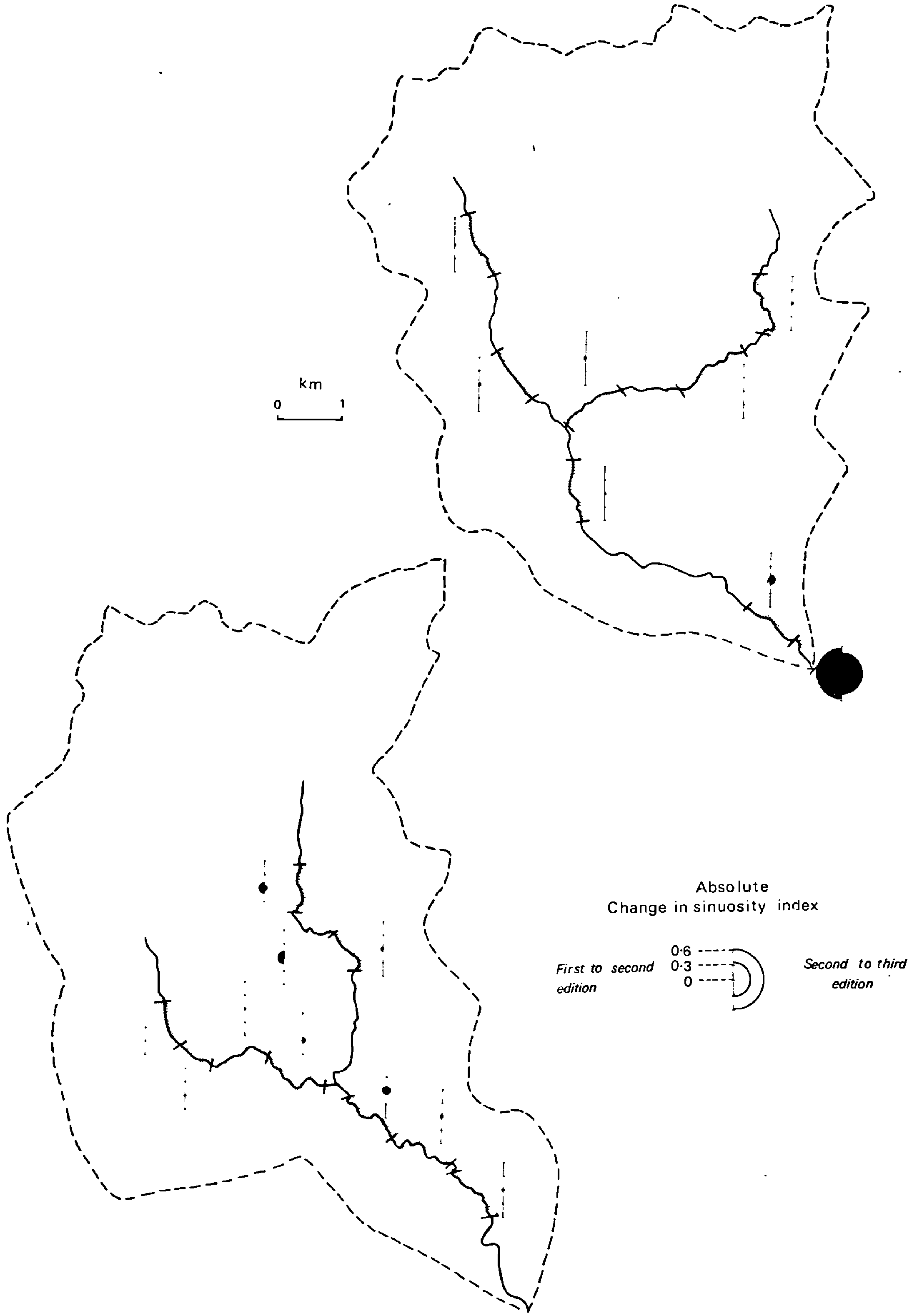


Figure 4.6.3.(iv)

Sampled change in sinuosity index within the Lui and Quoich catchments during the two inter map periods



(Figures 4.6.4.(v) to (ix)). This total sample diagram did have some value however, as it seemed to indicate two separate populations of sinuosity values and associated rates of change. The lower group from 1.00 to approximately 1.35 could alter their sinuosity but seemed to do so within these defined limits, implying controls on higher thresholds that were not easily exceeded. The second group had much higher values of sinuosity (1.4-1.8) and were associated with larger changes within these limits, suggesting much lower thresholds for change. With the exception of the Quoich confluence site (Sample 42), which underwent dramatic change, these 2 categories were mutually exclusive.

Studying the extremes of the distribution, between the first (1869) and second (1900) editions, only three sites had a total change in sinuosity in excess of + or - 0.10. One of these was in the alluvial basin of upper Glen Derry with sinuosity values reducing from 1.75 and 1.63 from 1866 to 1900 (Sample 17). Another reach above the confluence of the Quoich with the Dee (Sample 42), probably had an even larger adjustment (seen in Section 7.2.1.(i)). Only one site had an increase in sinuosity and that was on the mainstream Dee before the Clunie confluence (Sample 45) when sinuosity increased from 1.39 to 1.52 in 34 years. Most of the change occurred within well-defined limits rather than by major disruption and the majority of reaches only recorded sinuosity changes of +/- 0.05.

In terms of change in sinuosity between the second and third editions (a period ranging from 66-70 years), there were only 6 sample sites which had changes > +/- 0.1. The highest reduction in sinuosity (1.56 to 1.11 in 69 years) was again on the Quoich above its confluence

Figure 4.6.3.(v)

CLUNIE CATCHMENT: SINUOSITY INDEX CHANGE

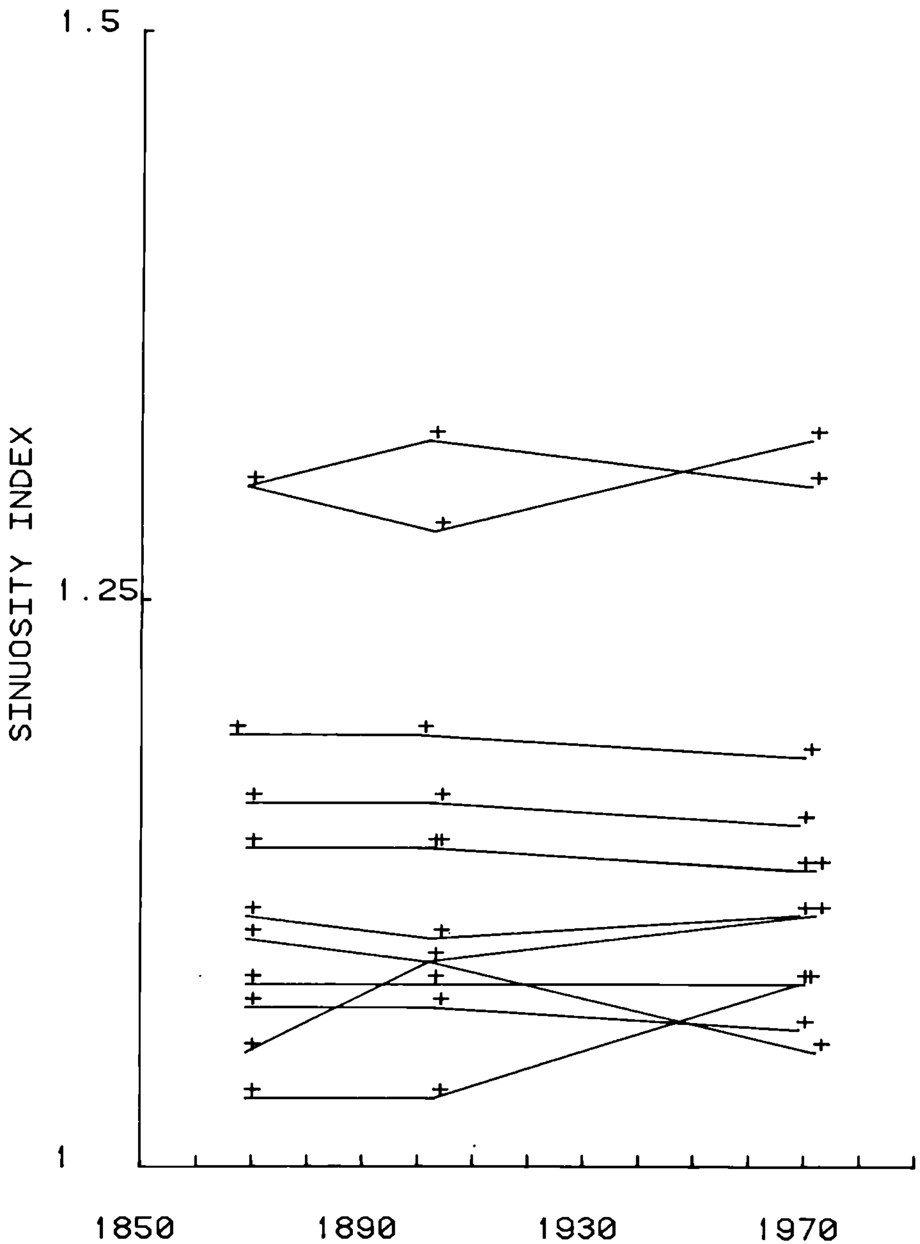


Figure 4.6.3.(vi)

EY: SINUOSITY INDEX CHANGE

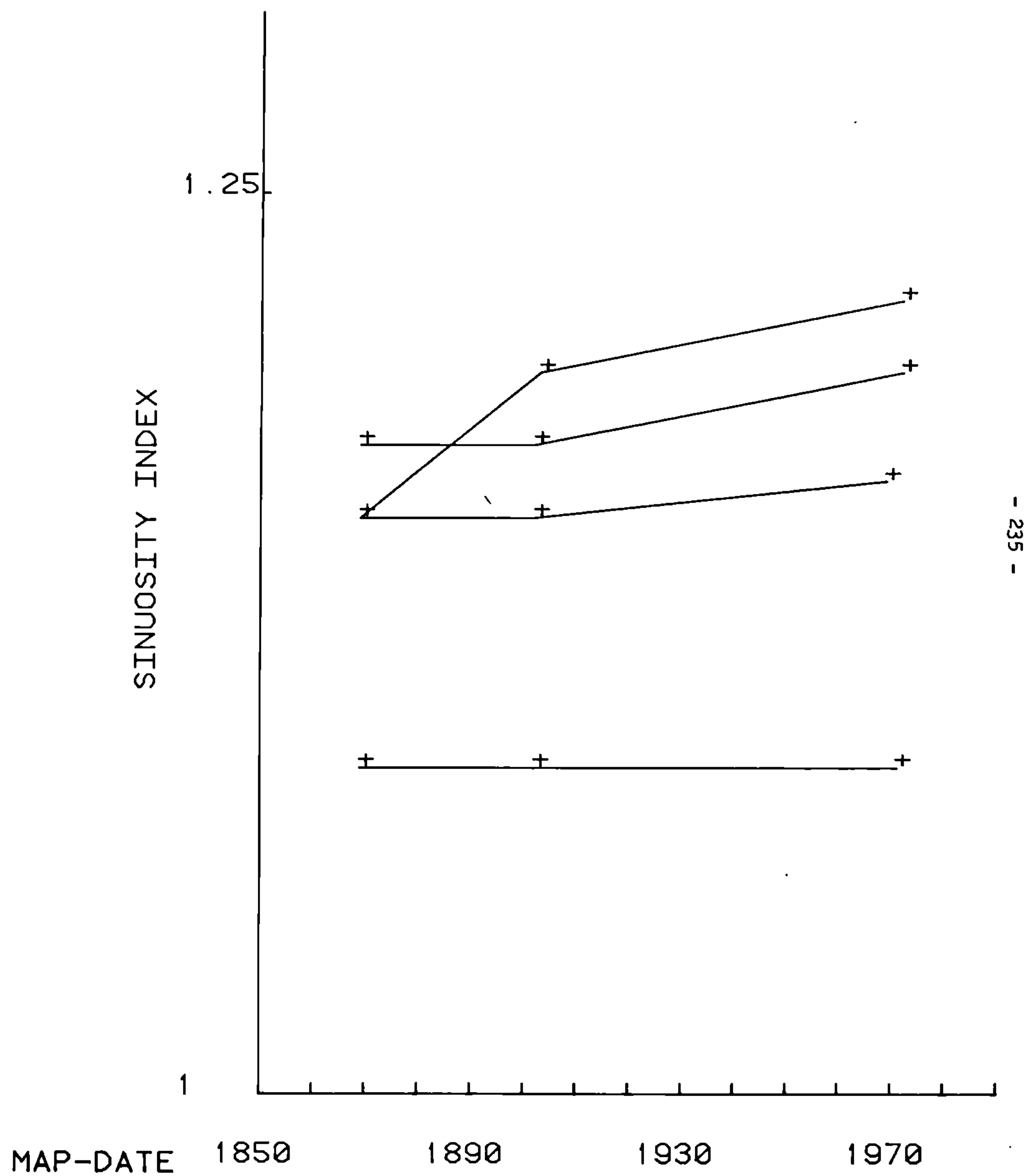
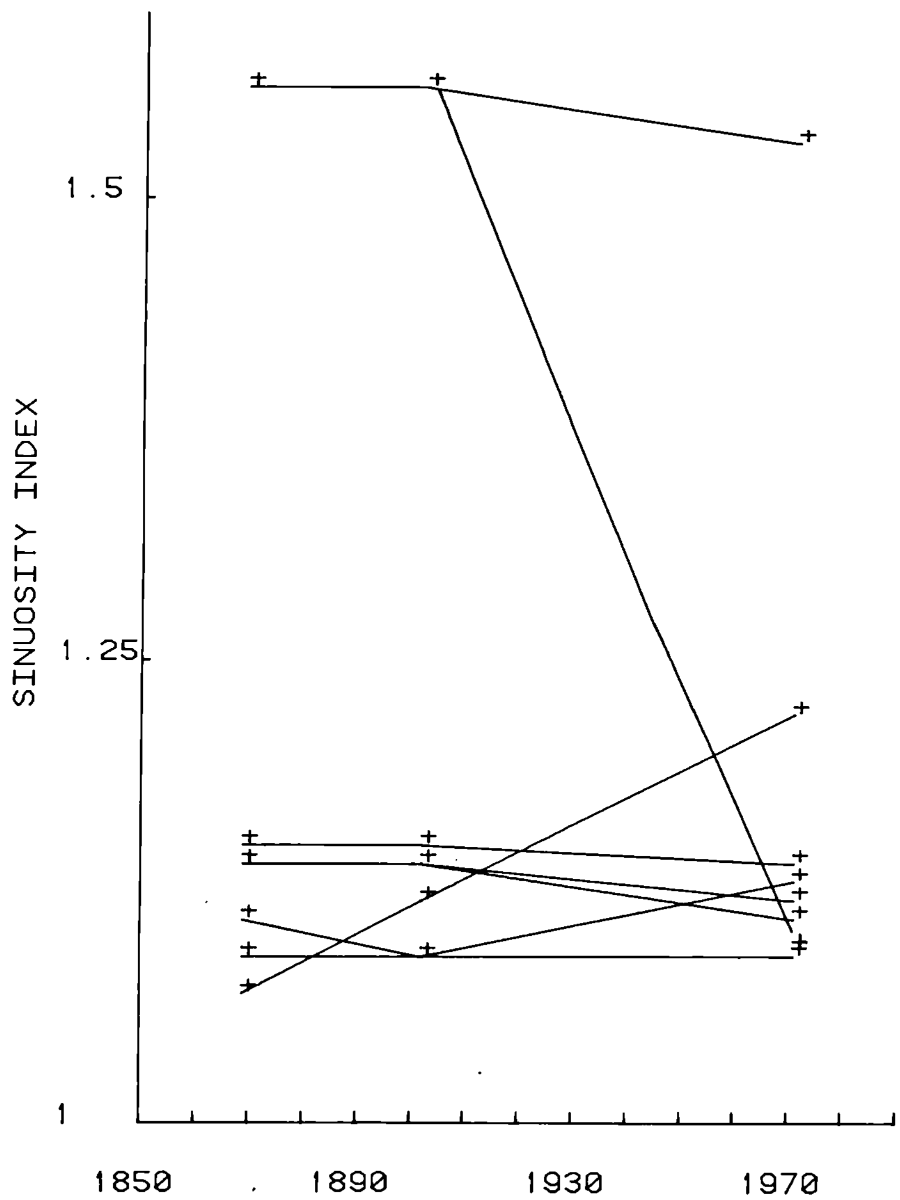


Figure 4.6.3.(vii)

QUOICH:SINUOSITY INDEX CHANGE



MAP-DATE

Figure 4.6.3.(viii)

LUI:SINUOSITY INDEX CHANGE

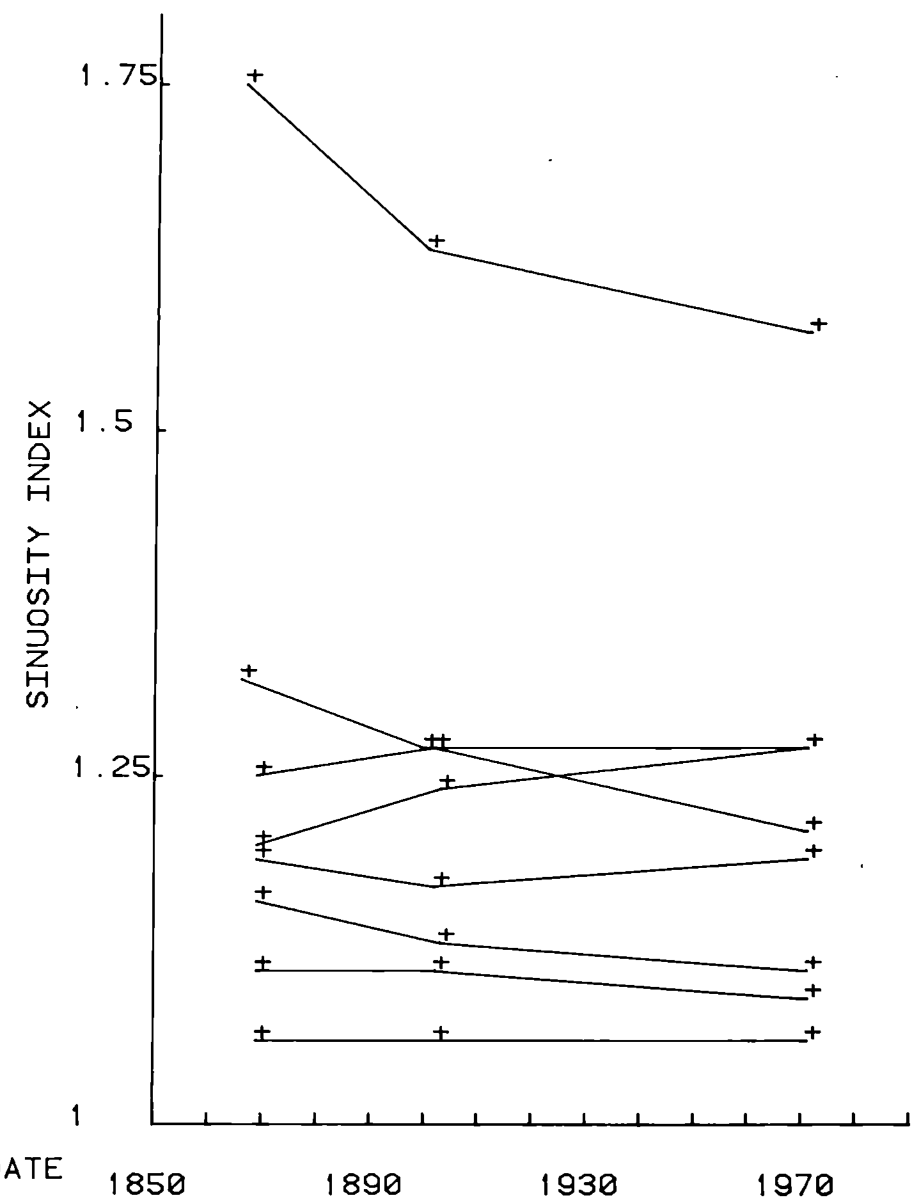
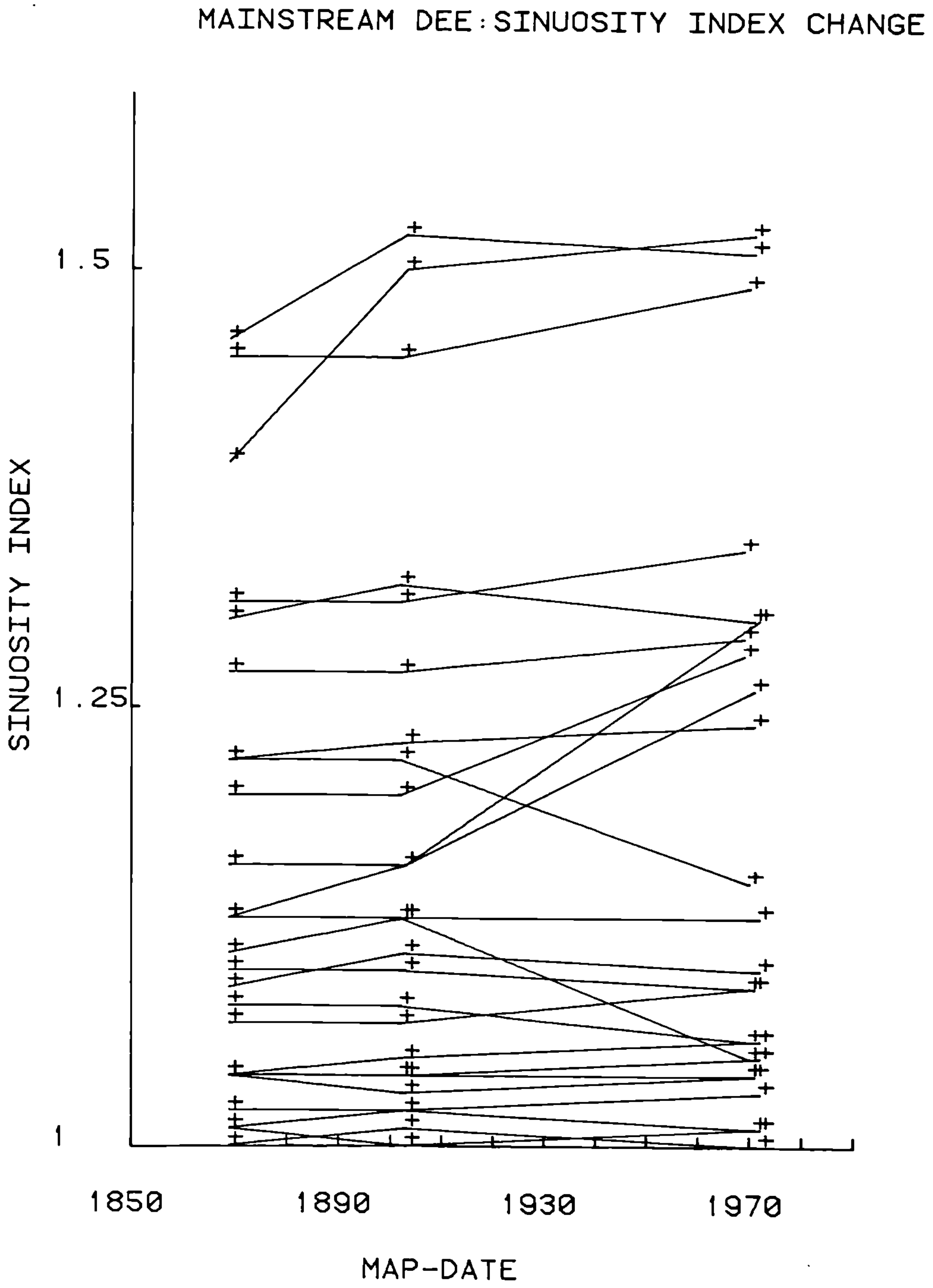


Figure 4.6.3.(ix)



with the Dee (Sample 42). Five sample sites had a positive increase in sinuosity > 0.10 (see Figure 4.6.3.(ii)); two being located on the mainstream Dee, two on the upper Geldie and one on the upper Quoich. In terms of the differential behaviour between reaches along the same tributary (eg. the Lui or Clunie Water), sinuosity could remain the same, increase or decrease on closely adjacent sites. Unfortunately, the sampling strategy did not unequivocally allow study of the relationship between all reaches within a catchment, though disparities did occur on samples that fell near on the same river. At the other extreme, 32 out of the 75 sampled reaches registered no change in sinuosity between the second and third editions. This either reflects confinement by bedrock control or high terraces, or else that the planform subsequently recovered to its original form between the two map surveys, despite the channel undergoing small scale disruption.

For comparisons between the two inter-map periods, the change in sinuosity was expressed as an annual rate and in certain cases made absolute (ie. direction of change ignored) so that results did not cancel out. Absolute change and corresponding average annual change covered a wide range of values (Table 4.6.3.(ii)). These suggest that slightly larger rates of annual change occurred between the second and third edition, though it was possible that in certain reaches over the 30 year timespan between map-dates 1 and 2, sinuosity fluctuated several times and only a fraction of the accumulated total was recorded. There were certainly a larger number of reaches that changed over 1900-1970 (39% reduced sinuosity; 46% increased sinuosity; see Figure 4.6.3.(ii)), though the sinuosity changes that did occur within the sample over both intermap periods were on average relatively small.

Table 4.6.3.(ii)

Rates of change in activity indices within the Dee study area

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
CHASIN1-2	0.00	0.00	0.01	0.01
CHASIN2-3	-0.01	0.00	0.02	0.03
AVSIN1-2	0.00	0.00	0.0003	0.0003
AVSIN2-3	-0.0002	0.00	0.0003	0.0005
BRAID1-2	0.00	0.00	0.01	0.01
BRAID2-3	-0.07	0.00	0.00	0.07
AVBRAID1-2	0.00	0.0004	0.0017	0.0017
AVBRAID2-3	-0.0011	0.00	0.00	0.0011
LSMAX1-2	0	5	15	15
LSMAX2-3	9	15	26	17
AVLSMAX1-2	0.0	0.1	0.5	0.5
AVLSMAX2-3	0.1	0.2	0.4	0.2
LSAV1-2	0	3	5	5
LSAV2-3	4	7	10	6
AVLSAV1-2	0.0	0.1	0.2	0.2
AVLSAV2-3	0.1	0.1	0.1	0.0

How important is initial sinuosity in relation to these subsequent changes, independent of direction? When absolute annual changes in sinuosity between the second and third editions (ABAVSIN2-3) were correlated with initial sinuosity at 1900, a positive correlation coefficient of 0.32 was gained, which indicates a highly significant correlation (Table 4.6.3.(iii)). The same pattern followed if ABAVSIN1-2 was correlated with the initial sinuosity in 1869, with an even stronger, highly significant, correlation coefficient of 0.67. It seems the higher the initial sinuosity at the commence of observation, the greater the likelihood of change in any direction. This implies that high sinuosity reaches had lower thresholds for sinuosity change. It also suggests that more sinuous reaches with their associated higher MAXWIDs have the ability to adjust their sinuosity, whereas less sinuous reaches have higher thresholds to exceed before sinuosity can be increased.

4.6.3.3 Rates of change as indicated by braided index (BRAID)

Median and IQR values for the average annual change in braided index reported a slight increase between map-dates 1-2 (1869-1902) and a reduction between 2-3 (1902-1970; see Table 4.6.3.(ii)). When BI values were submitted to the Wilcoxon test, highly significant differences were found between the first and second map editions (Z = -2.9) and second and third map editions (Z = -2.7). But the first and third editions comparison did not yield a significant difference implying similar overall levels of braiding at 1869 and 1970 after some disruption during

Table 4.6.3.(iii)

The correlation (r) between subsequent activity index change and initial activity index

(a) Importance of initial sinuosity index

	ABAVSIN2-3	ABAVSIN1-2
SIN 1	****	0.67+
SIN 2	0.32+	****

(b) Importance of initial braiding index

	ABBRAID2-3	ABBRAID1-2
BI 1	****	0.77+
BI 2	0.08	****

+ significant at 0.01 level

the intervening period. Chapters 5 and 6 will assess whether the climatic inputs and catchment variables were average over that period (ie. if some internal threshold was crossed) or if inputs were atypical of present day processes and some external threshold was exceeded. We may thus be studying a recovery to the system, after a geomorphic "shock".

When the values of BI1, BI2 and BI3 are plotted against map-date, the following patterns appeared (Figure 4.6.3.(x) to (xiv)). Again to reduce clutter, diagrams were subdivided by catchment and those samples which showed no change omitted. On the mainstream Dee for example, the increases in braiding that took place between 1869 and 1900 seemed small in comparison to the substantial decreases that took place between 1900 and 1970, again suggesting some long-term adjustment in the system to a major mobility of sediment. The only catchment that seemed to have an overall increase in BI between the second and third edition was the Clunie (see Figure 4.6.3.(x)). Thus, the variation in the inter-spacing of major flood events on different tributaries must be assessed.

When studying spatial patterns in the changes in BI between the first and second edition within the whole Deeside study area, it can be seen that there was considerable variation, with both no change and highly positive and negative changes recorded (Figure 4.6.3.(xv)). It must be noted that a large number of reaches (69%) remained stable in terms of braiding index, whilst in other more "change susceptible" areas, a variety of types of change occurred. High positive increases in BI occurred at 4 confluence sites along by the Dee, eg. at the Ey (+0.68; Sample 31) and Gelder (+0.36; Sample 73) confluences. An

Figure 4.6.3.(x)

CLUNIE CATCHMENT: BRAIDING INDEX CHANGE

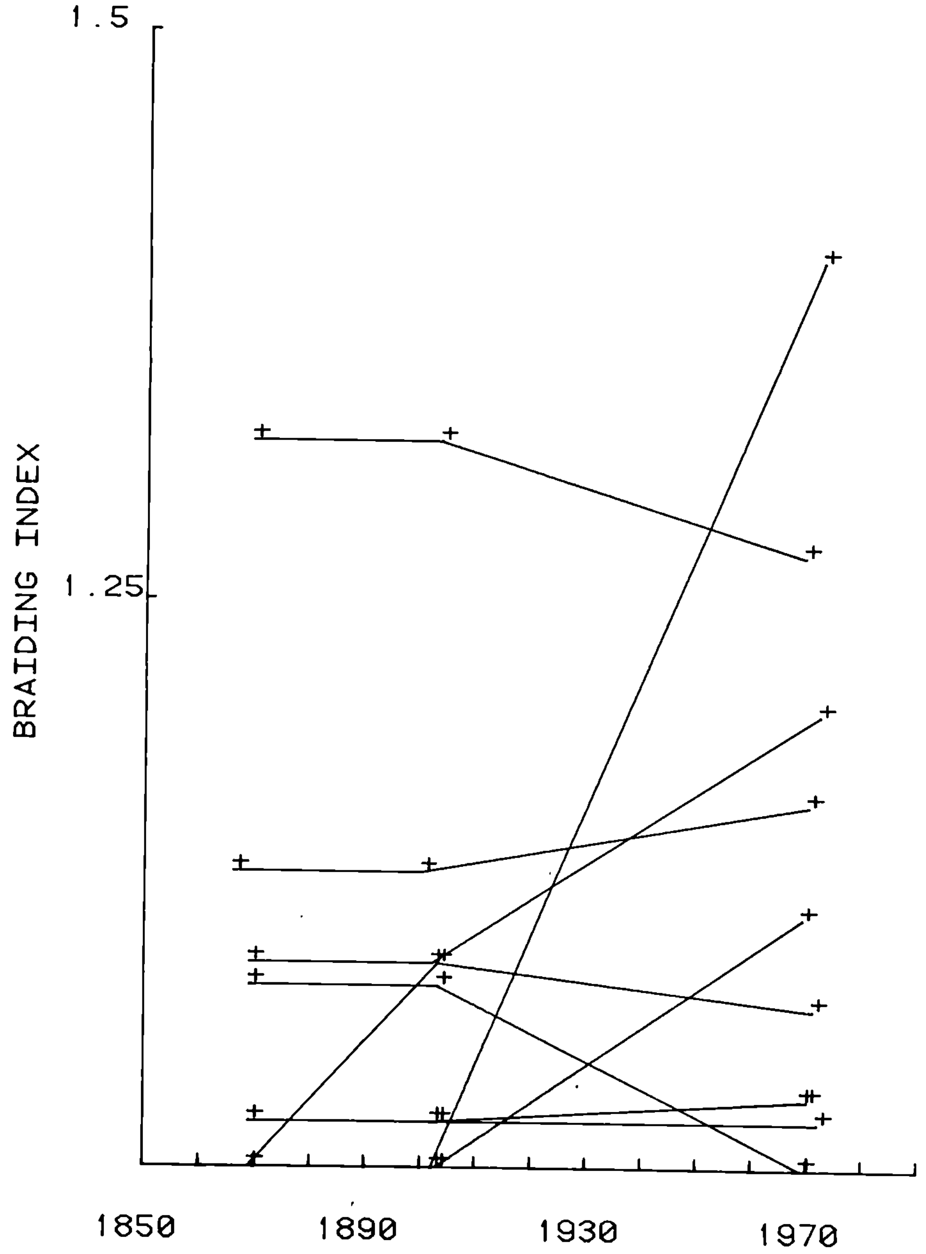


Figure 4.6.3.(xi)

EY: BRAIDING INDEX CHANGE

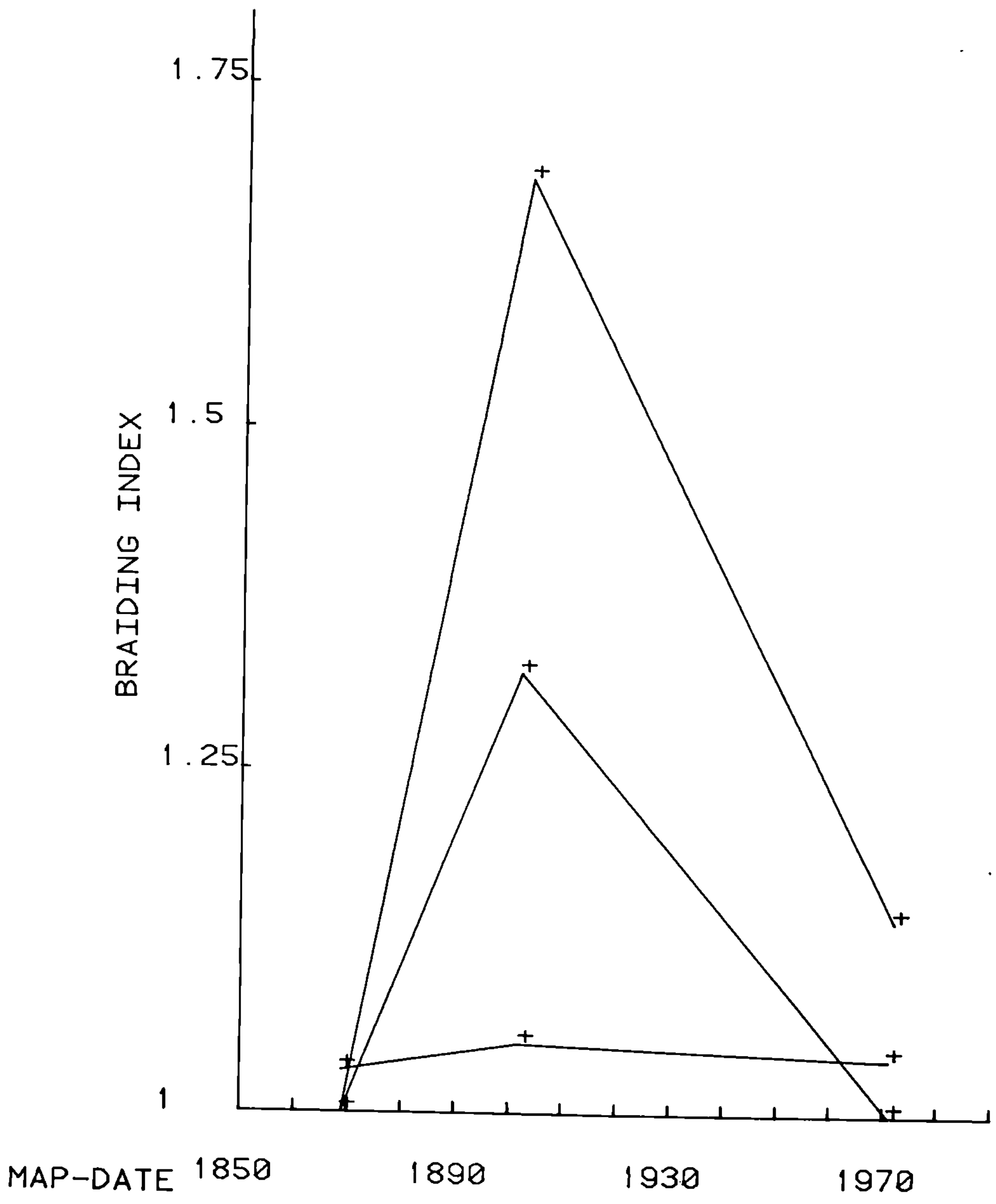


Figure 4.6.3.(xii)

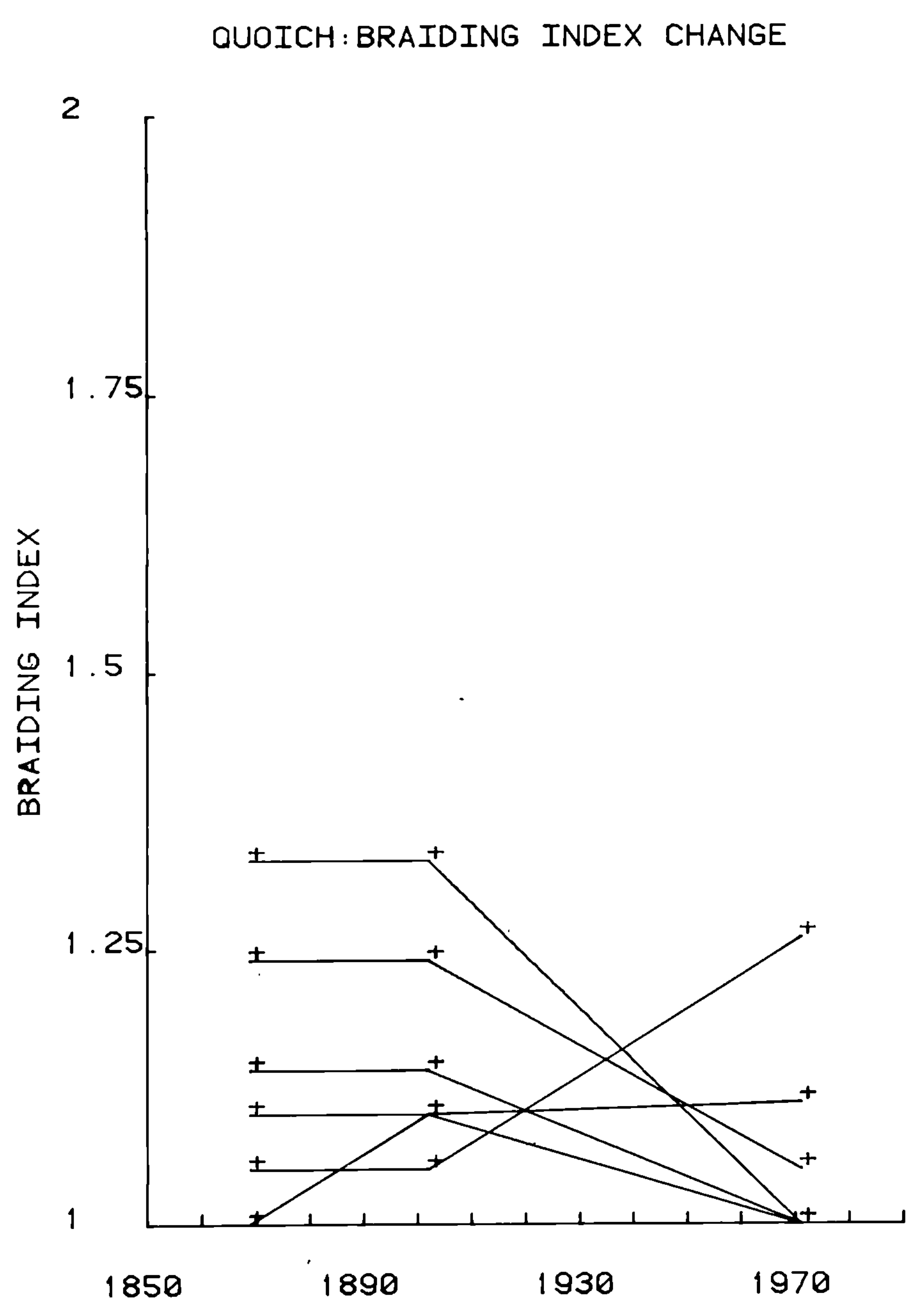
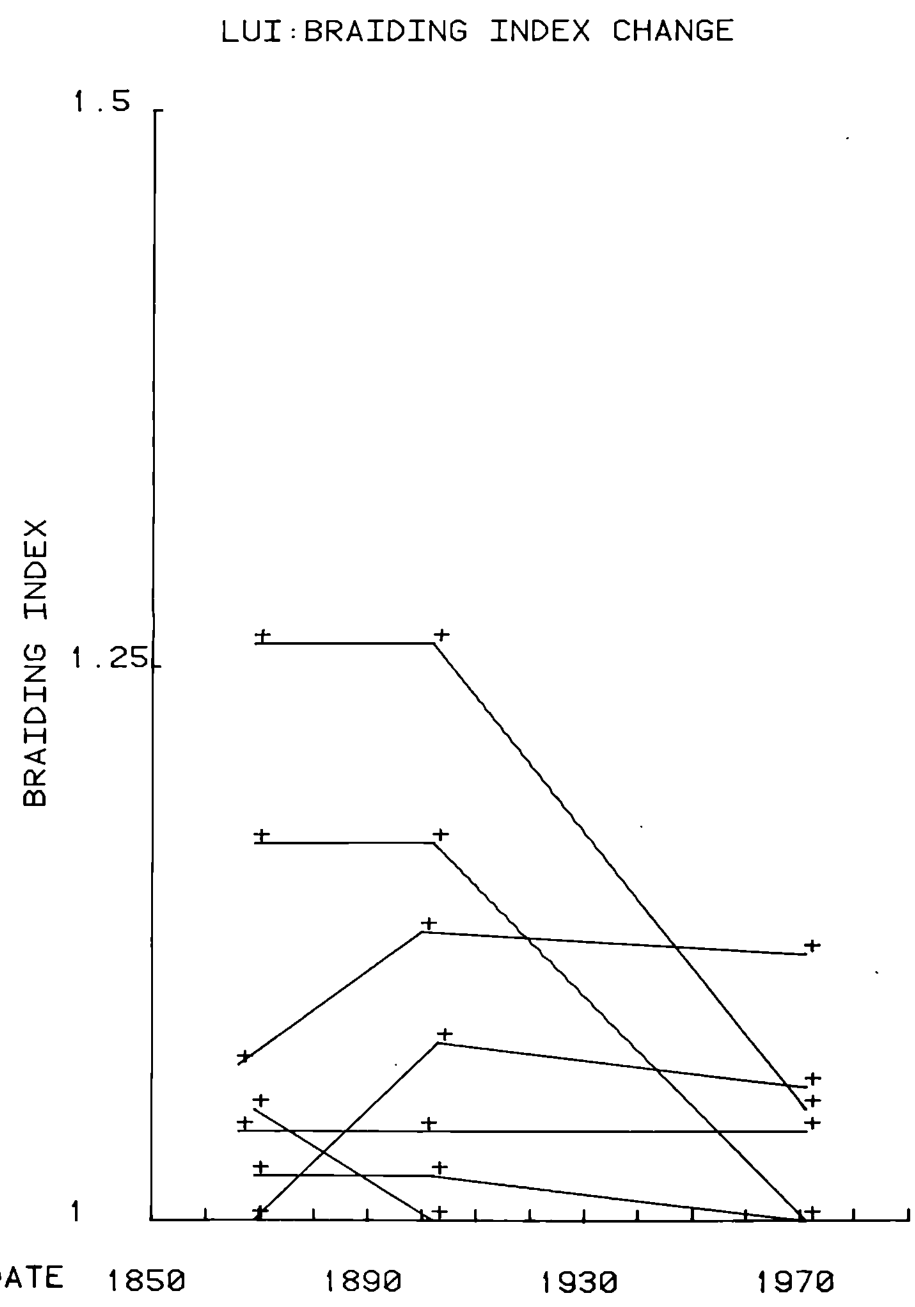


Figure 4.6.3.(xiii)



MAP-DATE

Figure 4.6.3.(xiv)

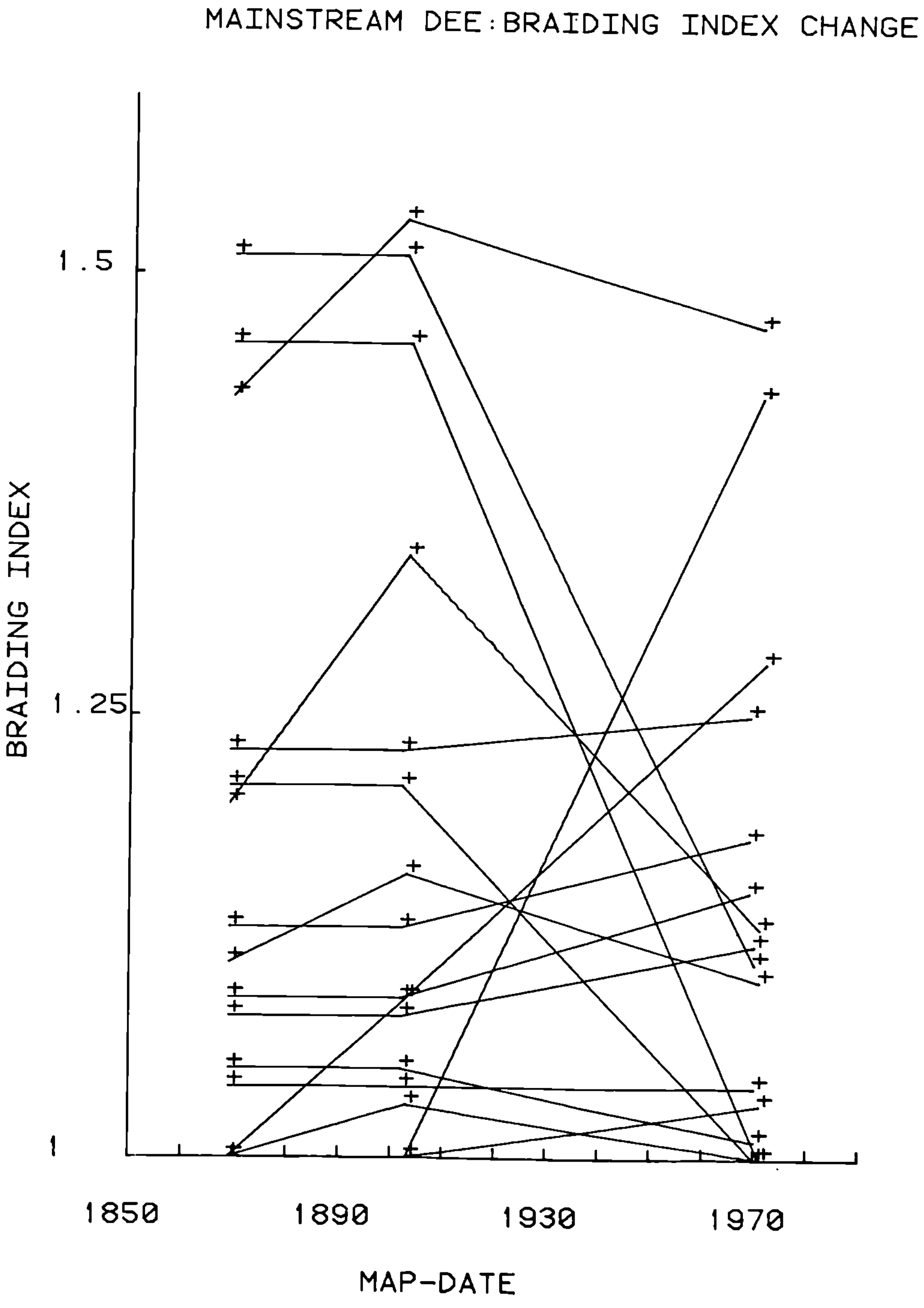


Figure 4.6.3.(x)

Sampled change in braiding index within the Dee study area during the two intermap periods

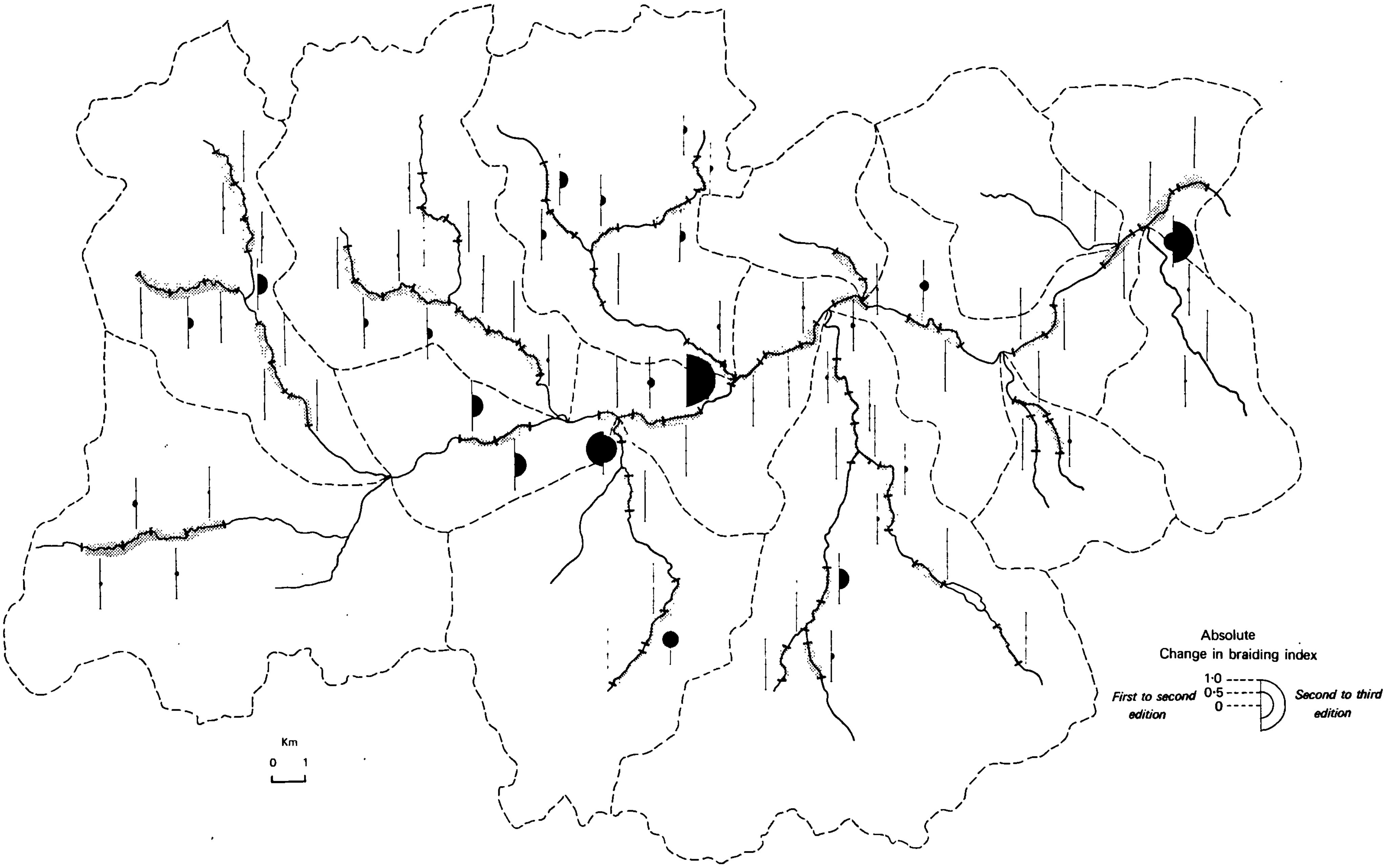


Figure 4.8.3.(xvi)

Sampled change in braiding index within the Clunie catchment during the two intermap periods

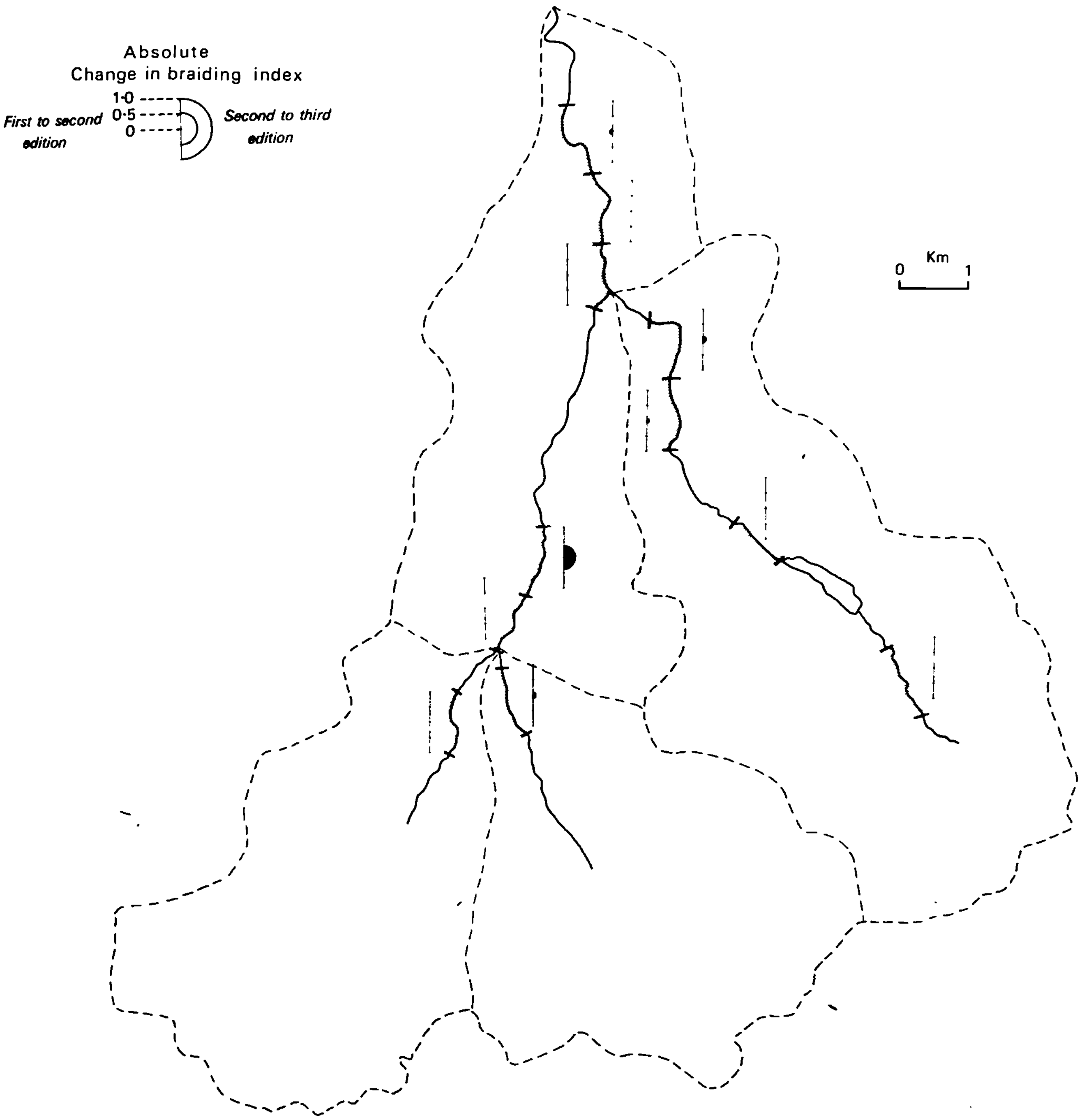
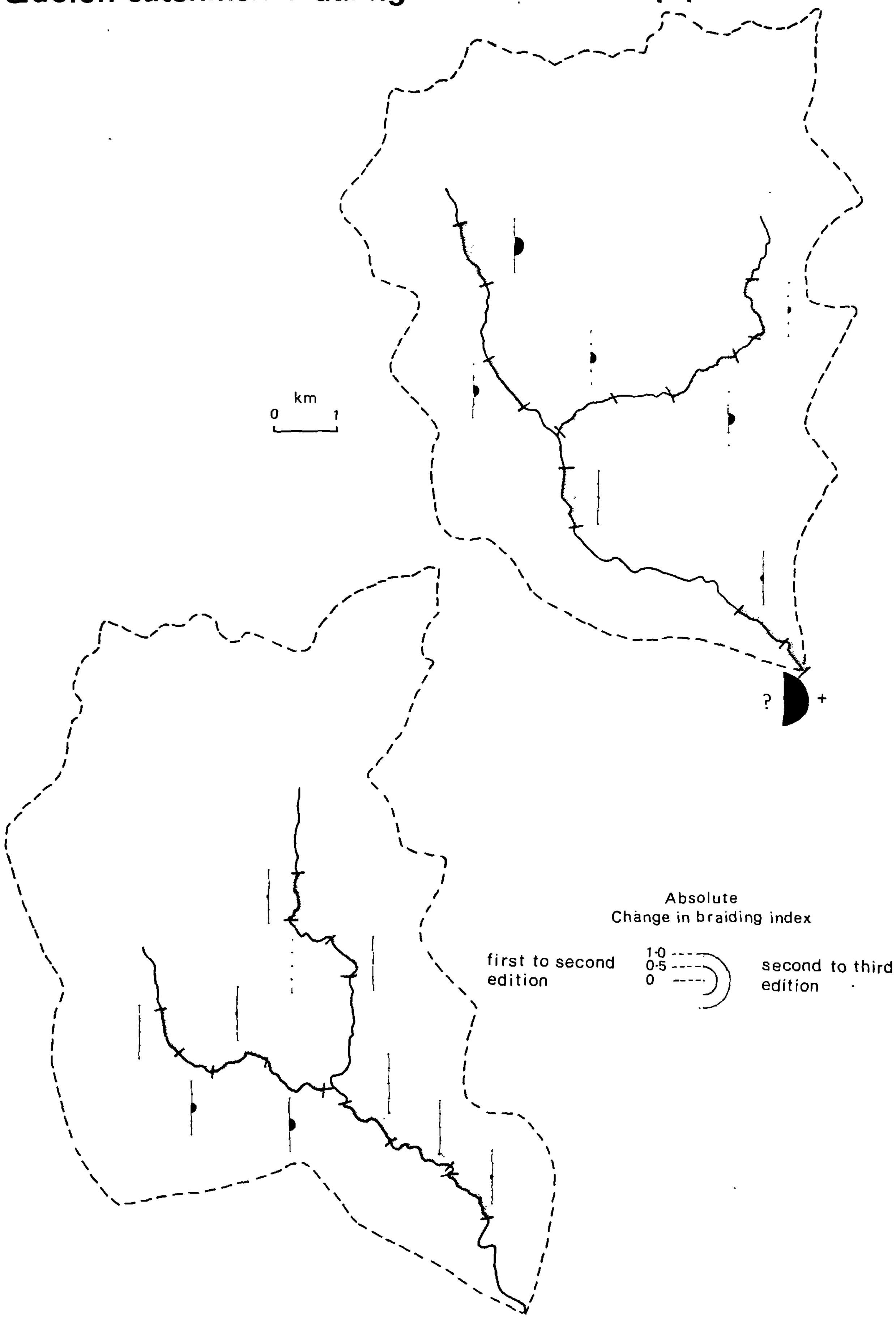


Figure 4.6.3.(xvii)

Sampled change in braiding index within the Lui and Quoich catchments during the two intermap periods



exception was the extreme case of the Quoich confluence, which had a large reduction in braiding index rendered unmeasurable since it was indicated as a distributary fan in the first edition, with channels too numerous to draw. The other highly positive or negative changes in BI occurred in the wide alluvial basins of the upper Ey (+0.32; Sample 28) and the Geldie (-0.14; Sample 12). In contrast, of the samples that fell in the Clunie basin, only one registered a change in braiding index between the first and second editions, implying a comparative stability over that period. This was in extreme contrast to between the second and third editions when a positive change in BI > 0.10 occurred in 4 out of 12 samples that fell within the Clunie catchment.

There was much change evident between the second and third map editions (1902-1970), with both negative and positive planform change. However, it appeared that many of the channels, whose braiding indices slightly increased between 1869 and 1900, decreased by 1970 (eg. Figure 4.6.3.(xiv)). Whether this represents a long-term adjustment to a change in planform controls remains to be assessed. This pattern showed up in an extreme way in two samples from the Ey catchment, especially above the confluence site with the mainstream Dee (Samples 31 and 28) and on the upper Geldie (Sample 11).

How important is the initial braiding index in determining the subsequent likelihood of this change? When ABBRAID2-3 was correlated with BI2, there was a very highly significant correlation of 0.77 (see Table 4.6.3.(iii)). This could imply once a channel crossed the threshold to a multi-barred planform, it continued to develop in a similar manner or fluctuate around a braiding equilibrium. Channel segments with a less braided or unbraided initial planform were less likely to change their BI. In contrast, a much lower correlation (0.08) was found between ABBRAID1-2 with BI. This implies that both reaches with no bar forms and those with BI > 1.00 at 1869 were equally likely to alter their BI. There is known to have been disruption of previously stable reaches.

4.6.3.4 Spatial variations in the rates of lateral shift

When carrying out the lateral shift measurements, several sites had to be rejected due to an absence of suitable or acceptably accurate fixed points. The range of values obtained for lateral shift of the mid-channel over time was considerable (Table 4.6.3.(ii)). The median total maximum shift between the first and second map editions (1869-1900) was 5 m but the IQR was large (15 m). 14 sample sites recorded no change at all along the reach, whilst at the other extreme, the highest maximum shift of 82 m (or 2.5 m yr^{-1}) occurred on the low level, gravel fan of the Gleann an t-Slugain, above its confluence with the mainstream Dee (Sample 59). This was clearly an exceptional case as the next highest was only 36 m (or 1.1 m yr^{-1}) and it must be evaluated whether this shift can be related to a single flood or sequence of floods. In terms of the total average shift over the whole reach as calculated by systematic cross-sections, the mean change was 3 m with a maximum of 37 m again recorded on the Gleann an t-Slugain. However, 82.6% of the sample recorded a maximum annual lateral shift of < 0.5 m yr^{-1} and average annual rates of lateral shift were only 0.1 m yr^{-1}.

When lateral shift was studied on the upper Clunie between the second to third editions, total maximum and total average shifts of 130 m (1.9 m yr^{-1}) and 15 m (0.01 m yr^{-1}) respectively were recorded at one sample as it changed its channel description from a "sinuous" to "wandering" channel, with "occasional" islands as opposed to none (Sample 49). Other high values included a total maximum lateral shift of 123 m (1.8 m yr^{-1}) on the mainstream Dee (Sample 33), caused by the chute cutoff of a meander bend. Both these sites recorded changes considerably larger than any others in the sample, due to the type of planform adjustment involved. The total average lateral shift between the second and third edition (AVLSAV2-3) over the reach had a median value of 7 m but 87% of the sample had a annual maximum lateral shift < 0.5 m. When both AVLSMAX and AVLSAV were compared for the two time-periods (Table 4.6.3.(ii)), very little difference can be identified in terms of median values but the IQR was larger between the first and second editions.

4.6.4.1 Importance of quantitative parameters

Rates of change in activity indices were correlated with contributing drainage basin area, average height and ratio of channel width to maximum valley width (MAXWID), however only those with a significant correlation will be discussed here. Rates of change had no significant correlation with area or height.

The only quantitative variable to show a significant relationship with activity was MAXWID, but within the Dee study area, the value of MAXWID tended to be comparatively small. Although there was not a positive correlation between AVSIN2-3 and MAXWID, the <u>maximum</u> rate of change seemed to increase with increase in MAXWID over this timespan. If however ABAVSIN1-2 was studied, a highly significant correlation of 0.44 was found. Again maximum change generally increased with available width of floodplain. However, many rivers which had a potentially wide floodplain seemed to have changed within much lesser, defined limits. Possible explanations must include channel incision, terrace features, and more stable banks. However, palaeofeatures within many floodplain areas, at a range of scales from Glen Derry to the mainstream Dee, showed that a larger proportion of the floodplain had been reworked in the past (pre-1750).

When the relationship of annual rates of average lateral shift between first and second edition (AVLSAV1-2) to available active area was studied, a highly significant positive correlation (r=0.63) was found, ie. as available active area increased so the rates of lateral shift increased. This occurred both in maximum and average terms, implying that the rivers were using a large proportion of the available floodplain area. This positive pattern was not so pronounced between the second and third editions (r=0.44) where although some channels had large available MAXWID ratios, the floodplains had not been so extensively reworked.

When change in braiding index was correlated with ($\log_{10}$) MAXWID, again high significant correlations were obtained, especially with ABBRAID1-2 (r=0.46). This implied width of available active area was being utilised, frequently allowing bar development.

4.6.4.2 Importance of qualitative parameters

It was necessary to assess the consistency of the typology classifications over the three map-dates. When parameters with each classification were cross-tabulated to see if there was any difference between the distribution of values at different map-dates, the following results were gained.

A strong relationship existed between floodplain vegetation at all three map-dates indicating no major change in floodplain vegetation. Shifts that did occur involved stabilisation of the floodplain ie. a decrease in vegetation category. The same cross-tabulation was carried out on river bank vegetation and the highest association was between bank vegetation at the first and second edition, with greater deviation from the dominant axis between the second and third edition. With upstream availability of sediment again the similarity of the samples at map-dates 1 and 2 must be noted though all three editions were highly related. This strength of relationship is clearer when studying local availability of sediment.

In terms of the persistence of channel pattern type, channel planform classification was relatively stable between the first and second map editions. However, between the second and third editions, 12 samples were reallocated to a higher channel pattern category (cf. 4 which were reallocated to a lower). In terms of the parameter "ISLANDS", the results were slightly different. Although this variable was related at each of the three map-dates, the strength of the relationship was clearly weaker between the second and third edition. There had however been a shift between categories at both inter-map periods. Between the first and second editions, there was an increase in the number of sites with islands, especially in sites with initially no islands. In contrast, 18 reaches which had "occasional bars" to "split channels" on the second edition had no bars by the third edition. Only 4 reaches without islands on the second edition had occasional islands on the third edition. This classification thus reflected some of the change in BI.

Activity rates were subdivided by the classifications within the channel typology to assess their relative importance, using the SPSS "breakdown" program (significant examples are given in Table 4.6.4.(i)). When channel activity rates were broken down by the availability of upstream sediment supply (UPSEDMT), the following results were obtained. When ABAVSIN2-3 underwent this analysis, category 1 (ie. localised incidence/s of sediment accumulation upstream of sample reach) had the highest ABAVSIN values (0.0009 yr^{-1}). Thus, large quantities of upstream sediment were not associated with change in sinuosity but rather with localised deposits eg. point bars. However,

Table 4.6.4.(i)

Selected results of the "breakdown" analysis with the Dee study area

(a) ABAVSIN1-2 broken down by UPSEDMT 1

UPSEDMT 1	Mean	SD	
0	0.0004	0.0006	
1	0.0007	0.0006	Sig = 0.001
2	0.0015	0.0015	

(b) ABAVSIN1-2 broken down by LOCSEDMT 1

LOCSEDMT 1	Mean	SD	
0	0.0004	0.0006	
1	0.0010	0.0010	Sig = 0.001
2	0.0004	0.0006	
3	0.0035	0.0000	

Table 4.6.4.(i) cont.

(c) ABAVSIN2-3 broken down by LOCSEDMT 2

LOCSEDMT 2	Mean	SD	
0	0.0004	0.0004	
1	0.0004	0.0006	Sig = 0.002
2	0.0016	0.0022	
3	0.0006	0.0004	

(d) ABBRAID1-2 broken down by LOCSEDMT 1

LOCSEDMT 1	Mean	SD	
0	0.0009	0.0033	
1	0.0011	0.0013	Sig = 0.03
2	0.0353	0.0936	
3	0.0018	0.0000	

Table 4.6.4.(i) cont.

(e) ABBRAID2-3 broken down by LOCSEDMT 2

LOCSEDMT 2	Mean	SD	
0	0.0013	0.0024	
1	0.0016	0.0021	Sig = 0.02
2	0.0163	0.0377	
3	0.0013	0.0016	

(f) ABAVSIN2-3 broken down by CHANPATT2

CHANPATT2	Mean	SD	
1	0.0001	0.0000	
2	0.0002	0.0002	Sig = 0.003
3	0.0006	0.0006	
4	0.0002	0.0001	
5	0.0004	0.0003	
6	0.0009	0.0009	
7	0.0001	0.0001	
8	0.0021	0.0030	

Table 4.6.4.(i) cont.

(g) ABBRAID1-2 broken down by CHANPATT1

CHANPATT1	Mean	SD	
1	0.0000	0.0000	
2	0.0001	0.0002	Sig = 0.01
3	0.0188	0.0686	
4	0.0005	0.0010	
5	0.0023	0.0053	
6	0.0014	0.0016	
7	0.0009	0.0015	
8	0.0005	0.0009	

Table 4.6.4.(i)

(g) AVLSMAX2-3 broken down by CHANPATT 2

CHANPATT 1	Mean	SD	
1	0.10	0.00	
2	0.36	0.47	
3	0.92	0.66	Sig = 0.03
4	0.68	0.63	
5	0.75	0.59	
6	1.01	0.52	
7	0.87	0.80	
8	1.43	0.03	

when the same analysis was carried out on ABAVSIN1-2, category 3 had a considerably larger mean value of 0.0015, as opposed to 0.0004 and 0.0007 respectively for categories 1 and 2. The category means had a highly significant difference and thus reaches with large reserves of upstream sediment were associated with the largest changes. The association may be that between 1869-1900, upstream sediment supply was reworked, flushed downstream and redeposited, contributing to both an increase in braiding index and an increase in main channel sinuosity cf. post-1900. When ABAVSIN1-2 was broken down by LOCSEDMT1, highly significant differences were found; the highest mean rates of change occurred when there were both local amounts and extensive gravels by and within the channel. Again, there were two main categories associated with change, namely sinuous to irregular channels associated with point and lateral bar deposits but also braided reaches, with abundant sediment supply, which had increased thalweg sinuosity around islands.

When ABBRAID2-3 and 1-2 were broken down by UPSEDMT, the highest absolute rate of change was associated with localised gravels (again associated with point and lateral bars), although the actual value was not significant. When the same occurred with LOCSEDMT categories, the highest mean values were found within category 2, ie. large reserves of local sediment, and there was a significant difference between the category means. Thus, high local sediment supply, whether reworked in situ or previously from upstream sources, was associated with higher rates of change in BI. Highest mean values of lateral shift were also significantly associated with large upstream sediment supply.

It was also necessary to study the relationship between activity change and vegetation. The absolute change in sinuosity between the two inter-map periods was broken down by category of hydrologically-significant floodplain vegetation (see Table 4.3.(i)). When ABAVSIN1-2 was broken down by floodplain vegetation at the first edition, the highest mean values were found when the floodplain was agricultural land but also surprisingly, when the vegetative cover was trees and shrub. The existence of groups of trees could accentuate the erosion problem, causing differential stability down a reach. The result also suggested that stream powers at these sites were sufficient under certain discharges to rip through wooded floodplain and rework the underlying floodplain material. Of course, unknown variables such as the age of the stands must be important. These vegetation categories were not however significantly different. In contrast, when ABAVSIN2-3 was broken down by floodplain vegetation at the second edition, the highest rates of activity are associated with rough-grassland and in one sample with marsh. The marsh sample was removed from the breakdown analysis, as there was only one sample and it was highly exceptional. There was no significant difference between categories, and thus floodplain vegetation was not in itself an important control.

When ABAVSIN2-3 was studied, high mean values of activity were found within the heather category ie. any stability provided little impact (see Table 4.6.4.(i)). A similar pattern occurred with bank vegetation. Stream powers must be sufficient to erode the banks despite the vegetative cover or else the underlying bank composition tends to be unstable. Change in BI showed no significant relationship with either floodplain or bank vegetation.

It was necessary to assess the relationship between activity rates and position of the reach within the "channel description" classification. When ABAVSIN2 was broken down by CHANPATT1 (ie. initial channel pattern), the highest mean rates of change in sinuosity were associated with the "irregular meander" and "tortuous meander" categories of the channel pattern classification though the categories were not significantly different. When ABAVSIN2-3 was broken down in the same manner by CHANPATT2, the highest rates of change were found in these two categories, however this time the F statistic was significant (see Table 4.6.4.(i)). Larger changes in sinuosity were associated with higher categories of channel pattern.

When ABBRAID1-2 was broken down by CHANPATT1, by far the highest mean absolute change in braiding index was found in the "sinuous" category (mean ABBRAID1-2 = 0.028) and the second highest was in the "irregular category"; categories were not however significantly different. In contrast, when ABBRAID2-3 was broken down by CHANPATT2, it was the tortuous meander category that had the highest mean rate of absolute change in BI, with the F statistic significant. Annual maximum lateral shift rates (1-2) were not significantly categorised by channel pattern type, with mean activity rates similar across the range. However, annual maximum lateral shift rates (2-3) broken down by CHANPATT2 had a significant F statistic, with high mean values of shift in "irregular" and "tortuous" meanders. In comparison, "straight" and "straight/ sinuous" planforms were characterised by lowest mean shift rates. Thus overall, highest rates of lateral activity were associated with sinuous, irregular and tortuous meanders.

4.7 Spey study area: Results of the map analysis

The results of the map analysis for the Spey study area are now presented. A reference map showing the locations of the 85 sampled grid squares within each of the major tributaries and on the mainstream Spey, is found in Figure 4.7.(i). A general location map, detailing catchments within the study area, is found in Figure 4.7.(ii). Results will be discussed both for the total sample and, where appropriate, sub-divided for individual catchments eg. when one catchment showed a particular divergence from the general trend.

4.7.1 Basic range of activity indices

4.7.1.1 Spatial variations in sinuosity index

The basic summary statistics for the sinuosity index using the total sample are found in Table 4.7.1.(i) and the data presented in histogram form are shown in Figure 4.7.1.(i). It was notable that both the median and interquartile range were lower than for the Dee study area (see Section 4.6.1). Table 4.7.1.(i) also shows the sample parameters when the data were subdivided by catchment. Highest median values and IQR were found within the Nethy catchment at SIN3, with 1.23 and 0.25 respectively. However, the spatial distribution of sinuosity values are shown in Figures 4.7.1.(ii) to (vii), from which it was clear that there was considerable inter-tributary variation. For example, the Tromie was characterised by low median and IQR values at all sampled sites at all three map-dates (SIN < 1.22). Within the total sample, highest sinuosity values were found in the middle to lower reaches of

Sampled river channel segments within the Spey study area

Figure 4.7.(i)

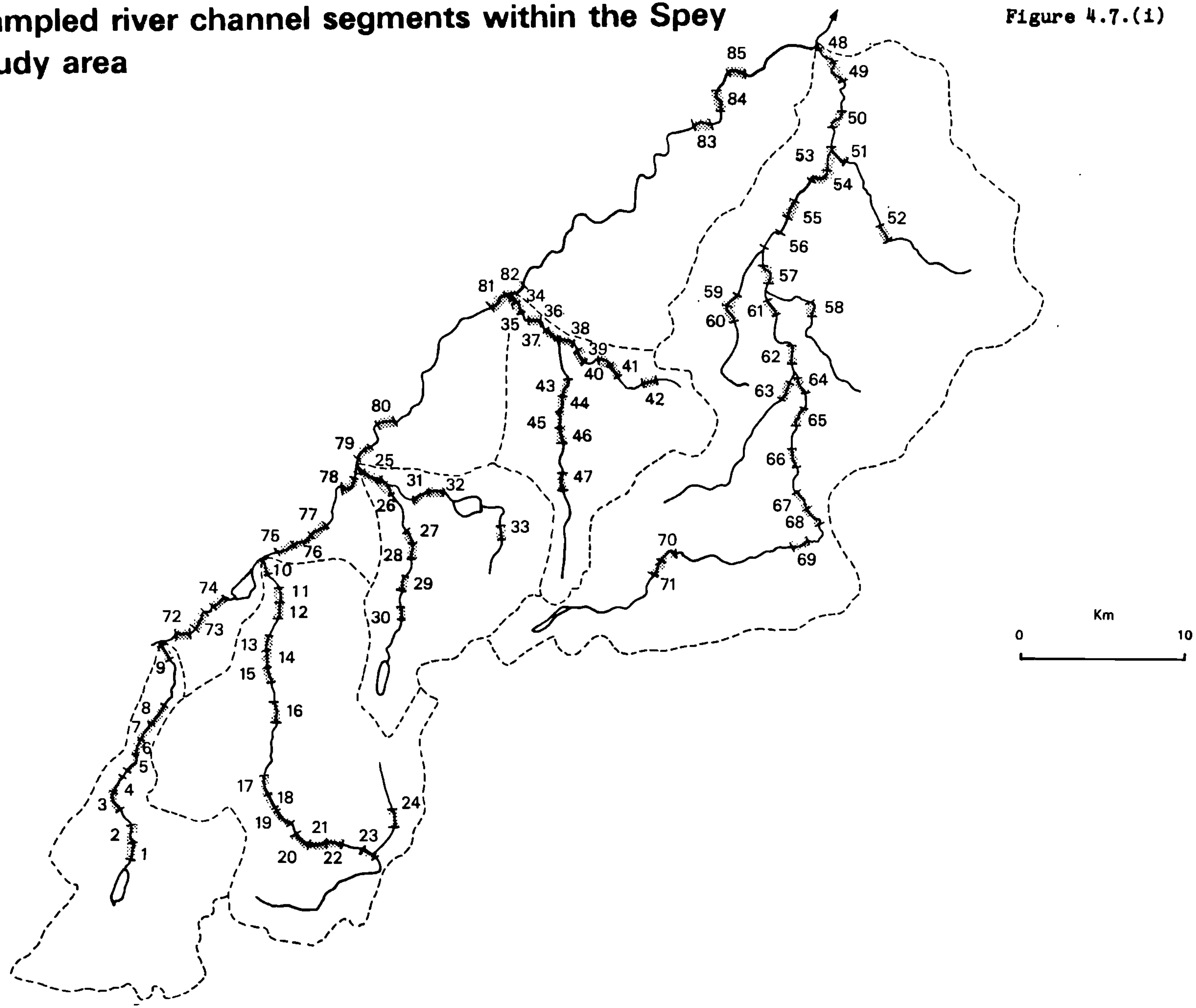


Figure 4.7.(ii)

General location map for the Spey study area

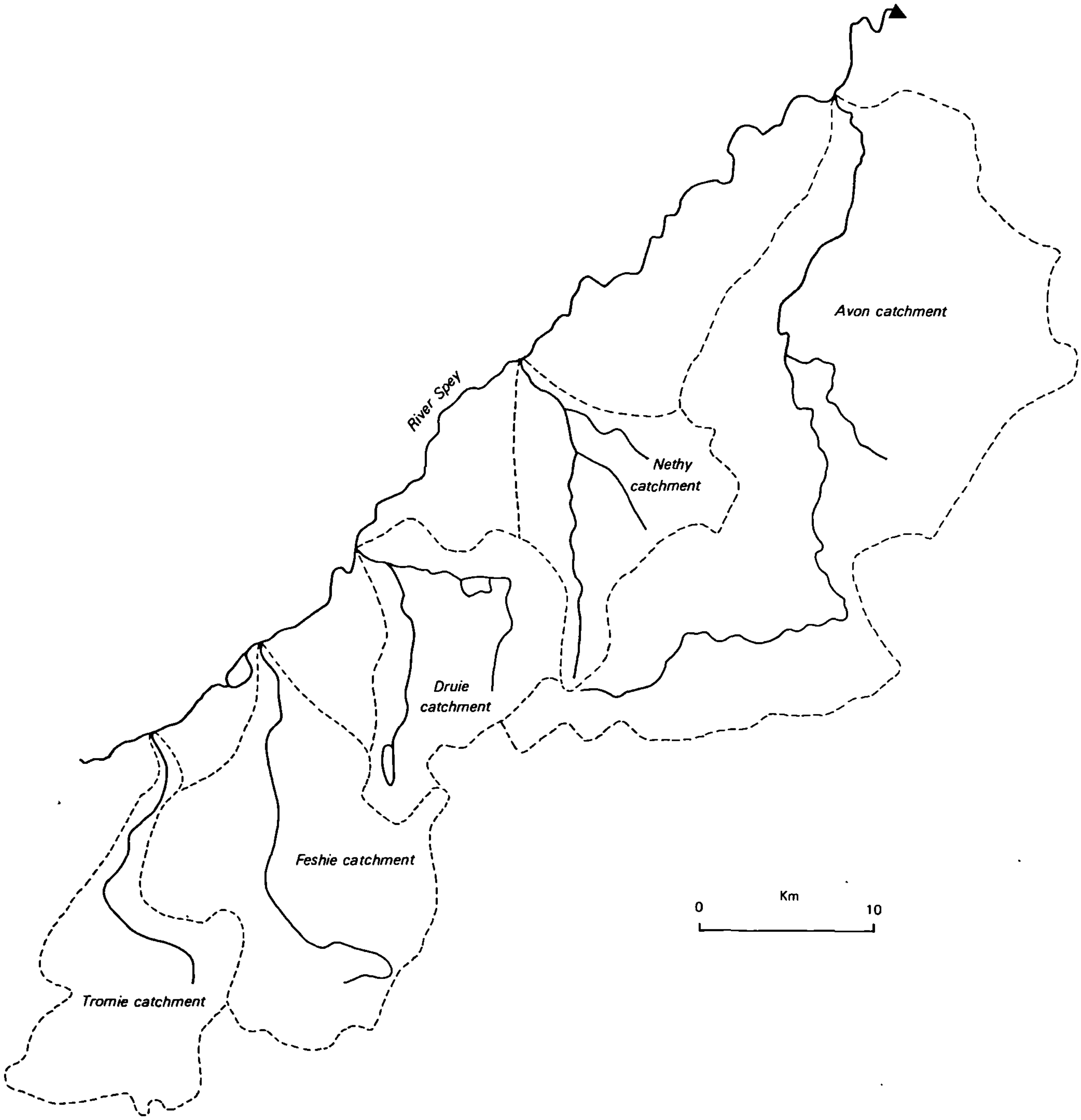


Table 4.7.1.(i)

Summary statistics for the sinuosity indices within the Spey study area

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
SIN 1 (1850)	1.07	1.12	1.20	0.13
SIN 2 (1900)	1.07	1.13	1.19	0.12
SIN 3 (1970)	1.07	1.12	1.20	0.14
Tromie				
SIN 1	1.03	1.10	1.15	0.12
SIN 2	1.04	1.09	1.12	0.08
SIN 3	1.05	1.07	1.12	0.07
Feshie				
SIN 1	1.08	1.12	1.22	0.14
SIN 2	1.11	1.13	1.17	0.06
SIN 3	1.10	1.13	1.19	0.09
Druie				
SIN 1	1.11	1.16	1.18	0.07
SIN 2	1.11	1.16	1.19	0.08
SIN 3	1.11	1.14	1.19	0.08

Table 4.7.1.(i) cont.

Parameter	25th Percentile	Median	75th Percentile	IQR
Nethy				
SIN 1	1.08	1.17	1.28	0.20
SIN 2	1.09	1.17	1.30	0.21
SIN 3	1.12	1.23	1.36	0.24
Avon				
SIN 1	1.08	1.14	1.22	0.14
SIN 2	1.06	1.16	1.21	0.15
SIN 3	1.06	1.13	1.20	0.14
Spey				
SIN 1	1.05	1.08	1.11	0.06
SIN 2	1.04	1.08	1.15	0.11
SIN 3	1.05	1.08	1.17	0.12

Figure 4.7.1.(i)

Frequency of sinuosity index values within the Spey study area

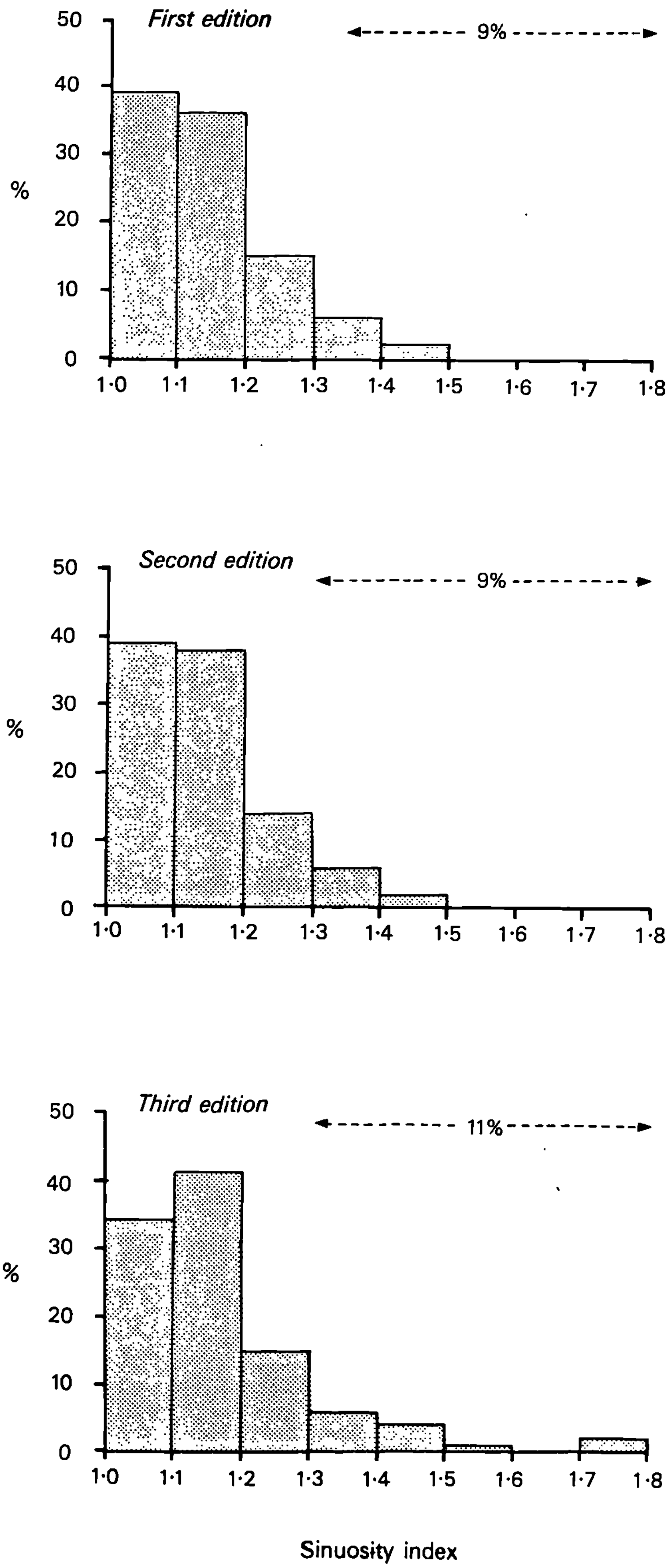


Figure 4.7.1.(ii)

Sampled sinuosity index values falling within the Tromie catchment for the three map dates

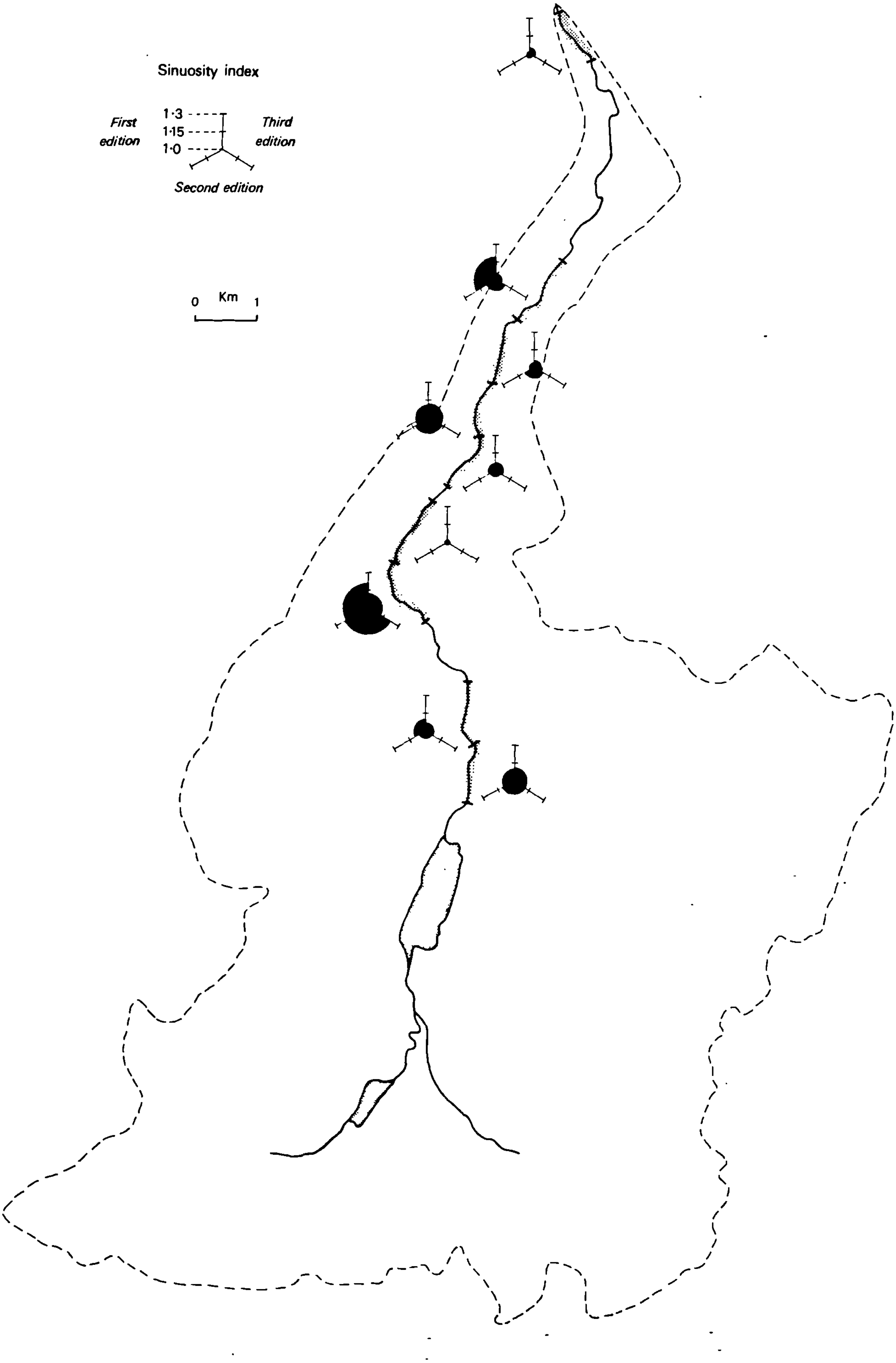


Figure 4.7.1.(iii)

Sampled sinuosity index values falling within the Feshie catchment for the three map dates

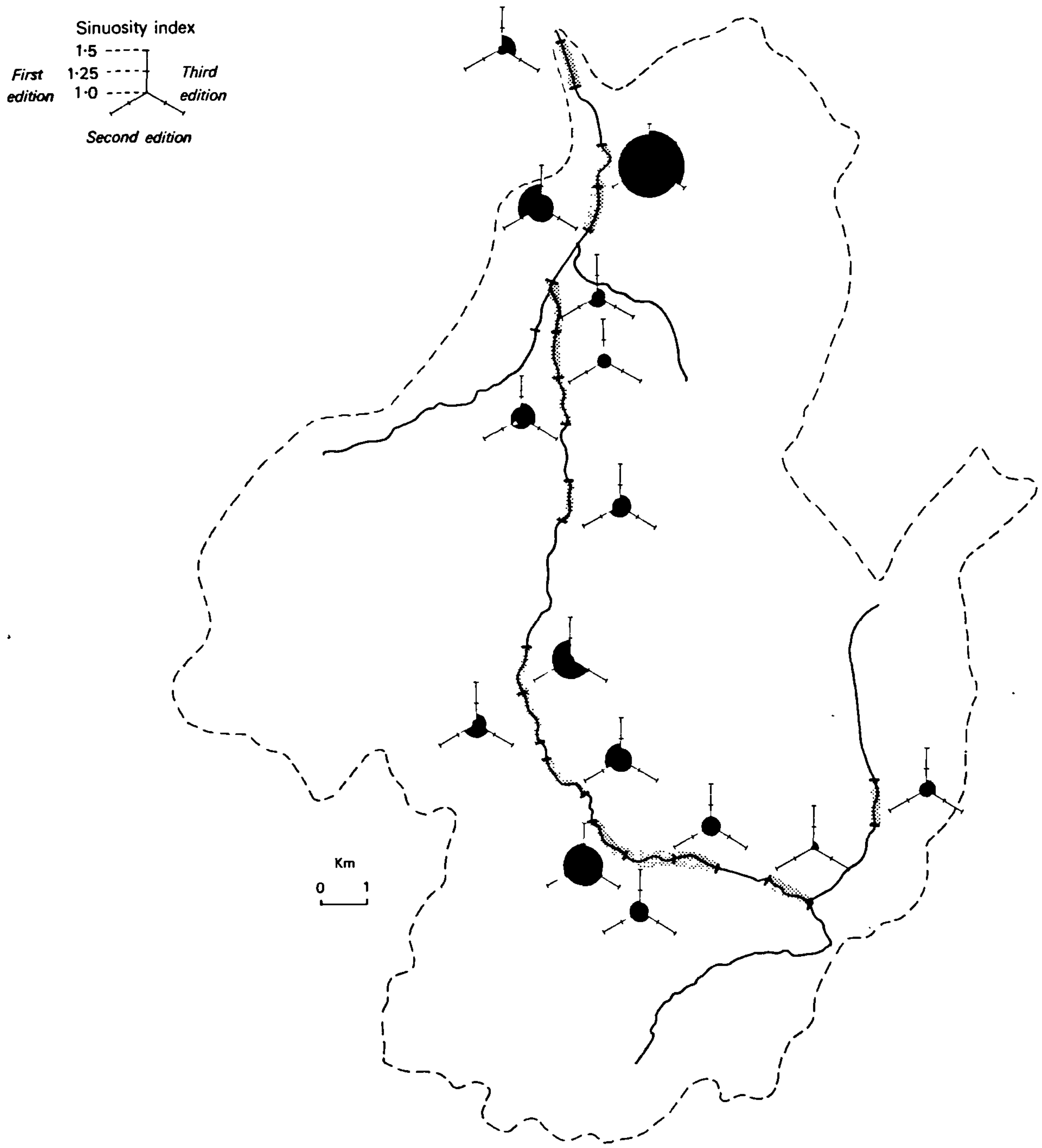


Figure 4.7.1.(iv)

Sampled sinuosity index values falling within the Druie catchment for the three map dates

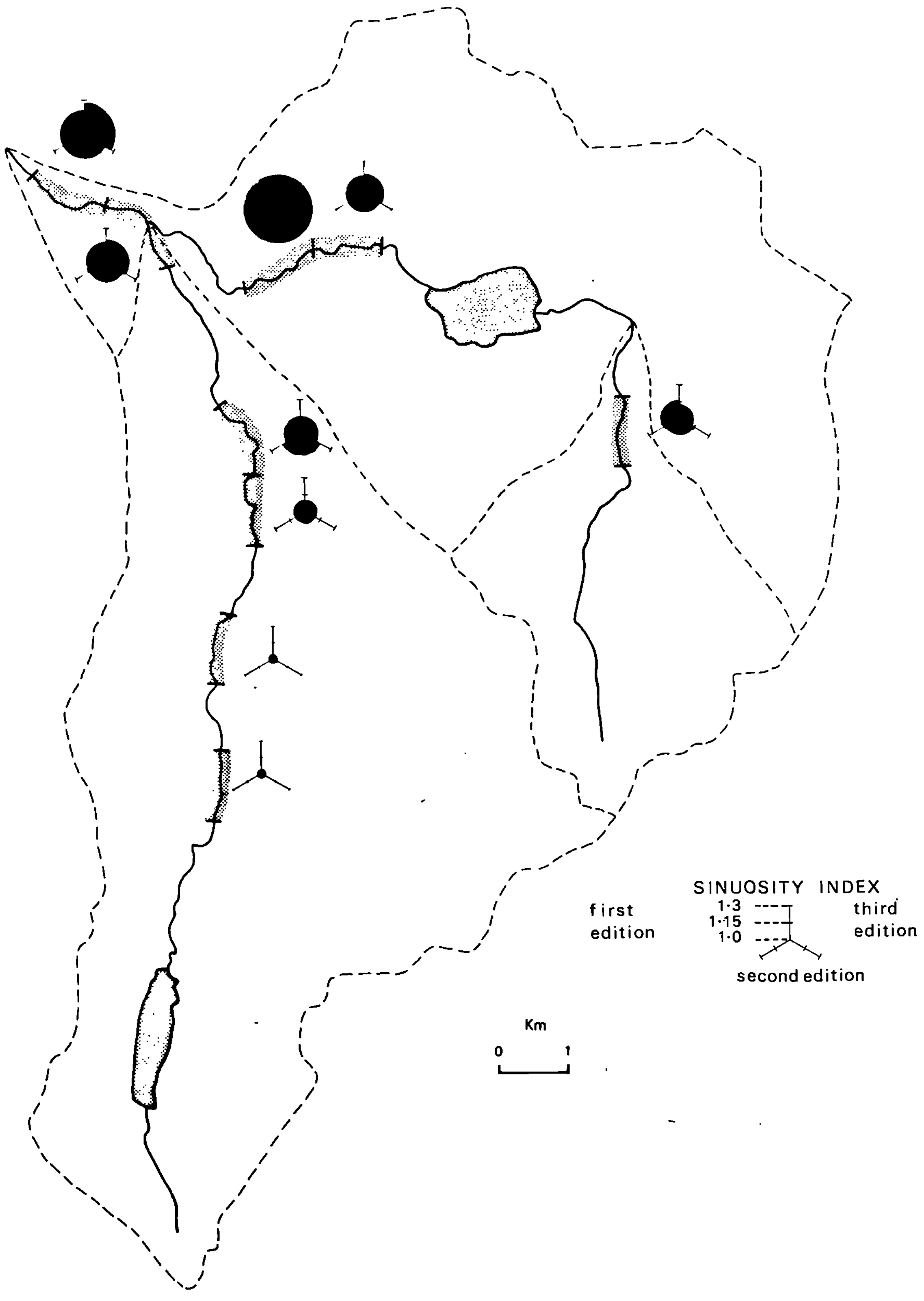


Figure 4.7.1.(v)

Sampled sinuosity values falling within the Nethy catchment for the three map dates

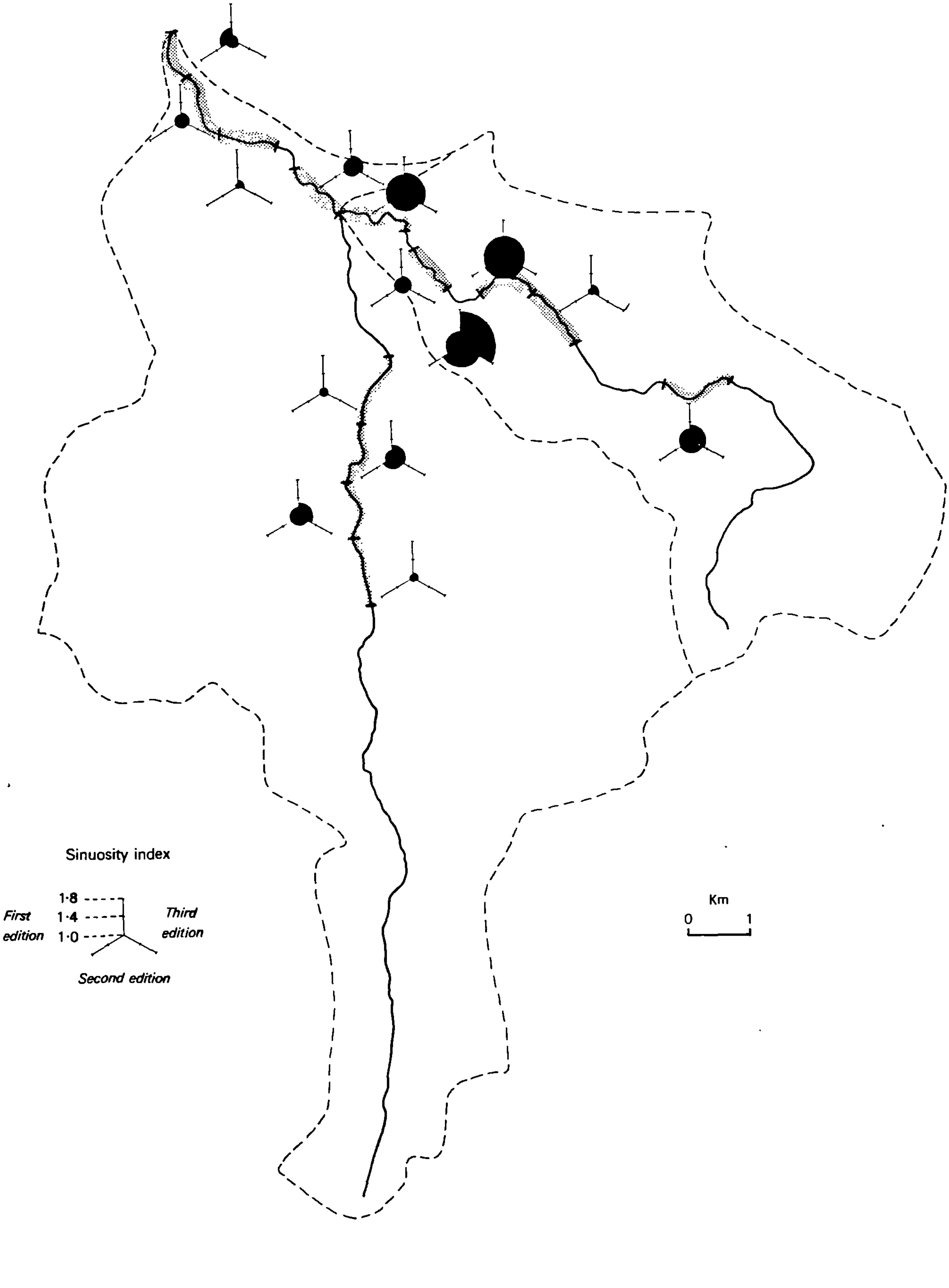


Figure 4.7.1.(vi)

Sampled sinuosity index values falling within the Avon catchment for the three map dates

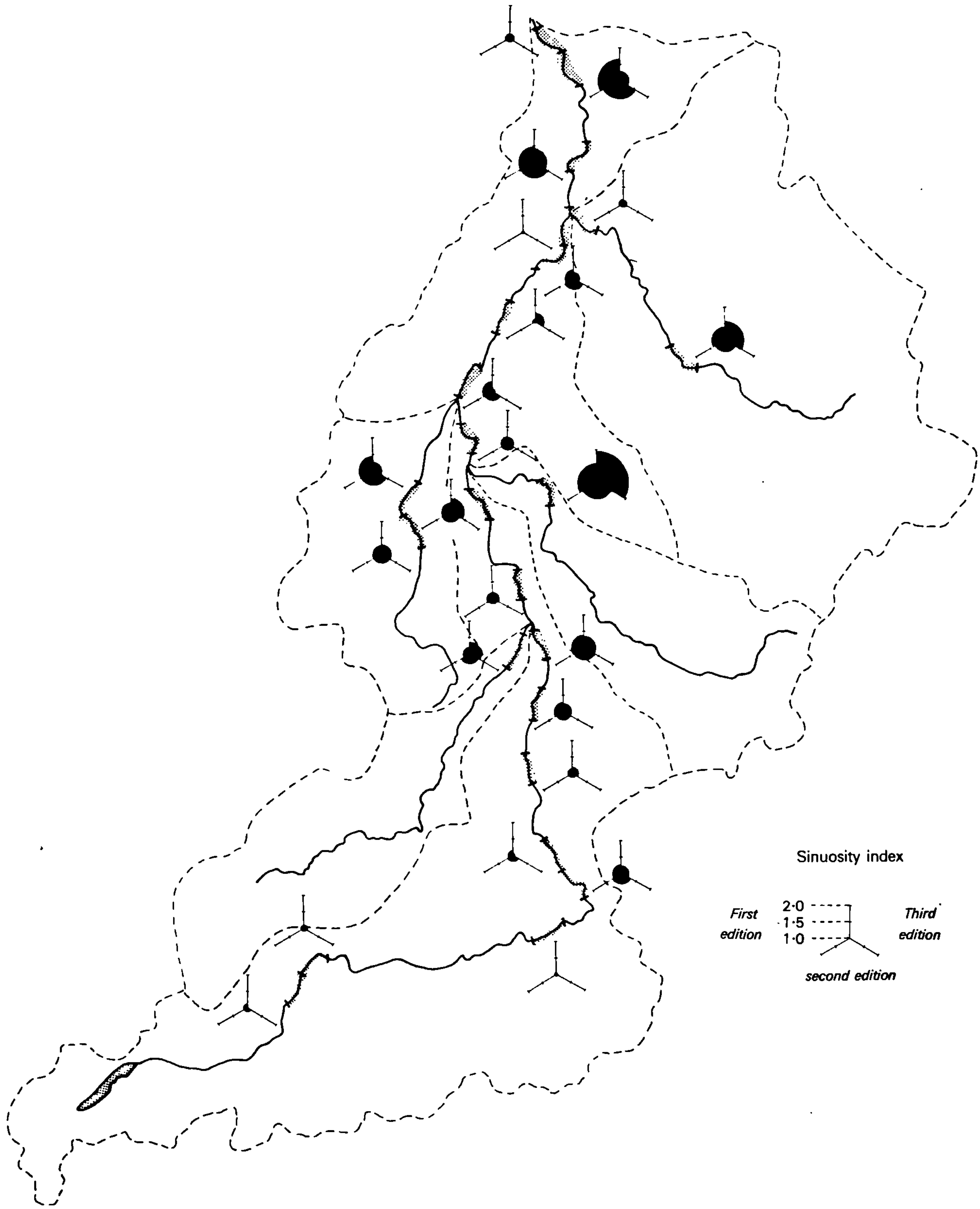
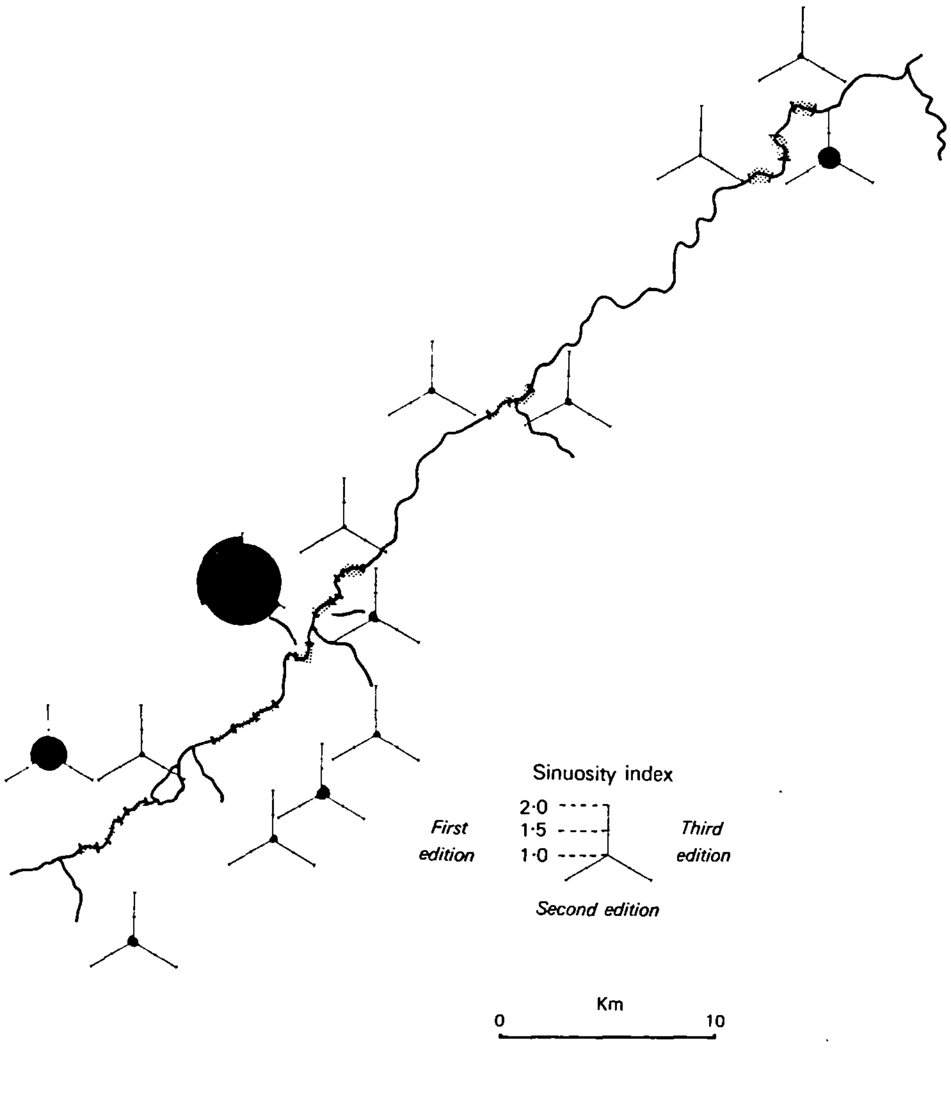


Figure 4.7.1.(vii)

Sampled sinuosity index values falling on the mainstream Spey for the three map dates



river (eg. Sample 38 on the Nethy (1.37-1.45) and Sample 11 on the Feshie (1.22-1.31)). In comparison to Deeside, the channel segments above confluence sites were not highly sinuous, usually below 1.20 (eg. the Tromie and Avon confluences). Whether or not this was their natural planform will be assessed later (Chapter 6). At the other extreme, the mainstream Spey was characterised by low sinuosity reaches (< 1.20 and which contrast with the higher sinuosities indicated on Roy's map in 1750). However, limited reaches of much higher sinuosity also occurred, for example, the regular meanders upstream of the Spey/ Druie confluence (Sample 78; SIN1= 1.78-1.93, Figure 4.7.1.(vii)).

4.7.1.2 Spatial variations in braiding index (BI)

The braiding index values were sampled for the entire Speyside data set and the values collated in Table 4.7.1.(ii) and presented in histogram form in Figure 4.7.1.(viii). Median and IQR values were much higher than for Deeside (cf. Table 4.6.1.(i)), with several reaches with BI > 2. However when these results were broken down by catchment, a rather different pattern emerged (Table 4.7.1.(i)). Highest median values occurred on the Nethy but highest IQR values occurred on the Feshie. The most interesting range fell in Glen Feshie (Figure 4.7.1.(x)), where especially for BI2 and BI3, there were incidences of very high values (maximum of 3.00-4.24; Sample 17) as well as other non-divided reaches. When all three study areas were compared, these were by far the highest values. Locally high values, with BI > 2, also occurred on sampled reaches on the Druie and Avon (Figures 4.7.1.(xi) and (xiii)).

Table 4.7.1.(ii)

Summary statistics for the braiding index within the Spey study area

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
BI 1	1.00	1.08	1.35	0.35
BI 2	1.00	1.09	1.34	0.34
BI 3	1.00	1.03	1.25	0.25
Tromie				
BI 1	1.00	1.15	1.33	0.33
BI 2	1.00	1.10	1.31	0.31
BI 3	1.00	1.00	1.09	0.09
Feshie				
BI 1	1.00	1.08	1.50	0.50
BI 2	1.00	1.03	1.41	0.41
BI 3	1.00	1.05	1.87	0.87
Druie				
BI 1	1.01	1.30	1.51	0.50
BI 2	1.02	1.24	1.67	0.65
BI 3	1.03	1.28	1.44	0.41

Table 4.7.1.(ii) cont.

Parameter	25th Percentile	Median	75th Percentile	IQR
Nethy				
BI 1	1.00	1.17	1.41	0.41
BI 2	1.00	1.23	1.39	0.39
BI 3	1.00	1.05	1.25	0.25
Avon				
BI 1	1.01	1.09	1.31	0.30
BI 2	1.00	1.10	1.27	0.27
BI 3	1.00	1.00	1.11	0.11
Spey				
BI 1	1.00	1.00	1.17	0.12
BI 2	1.00	1.00	1.18	0.18
BI 3	1.00	1.12	1.31	0.31

Frequency of braiding index values within the Spey study area

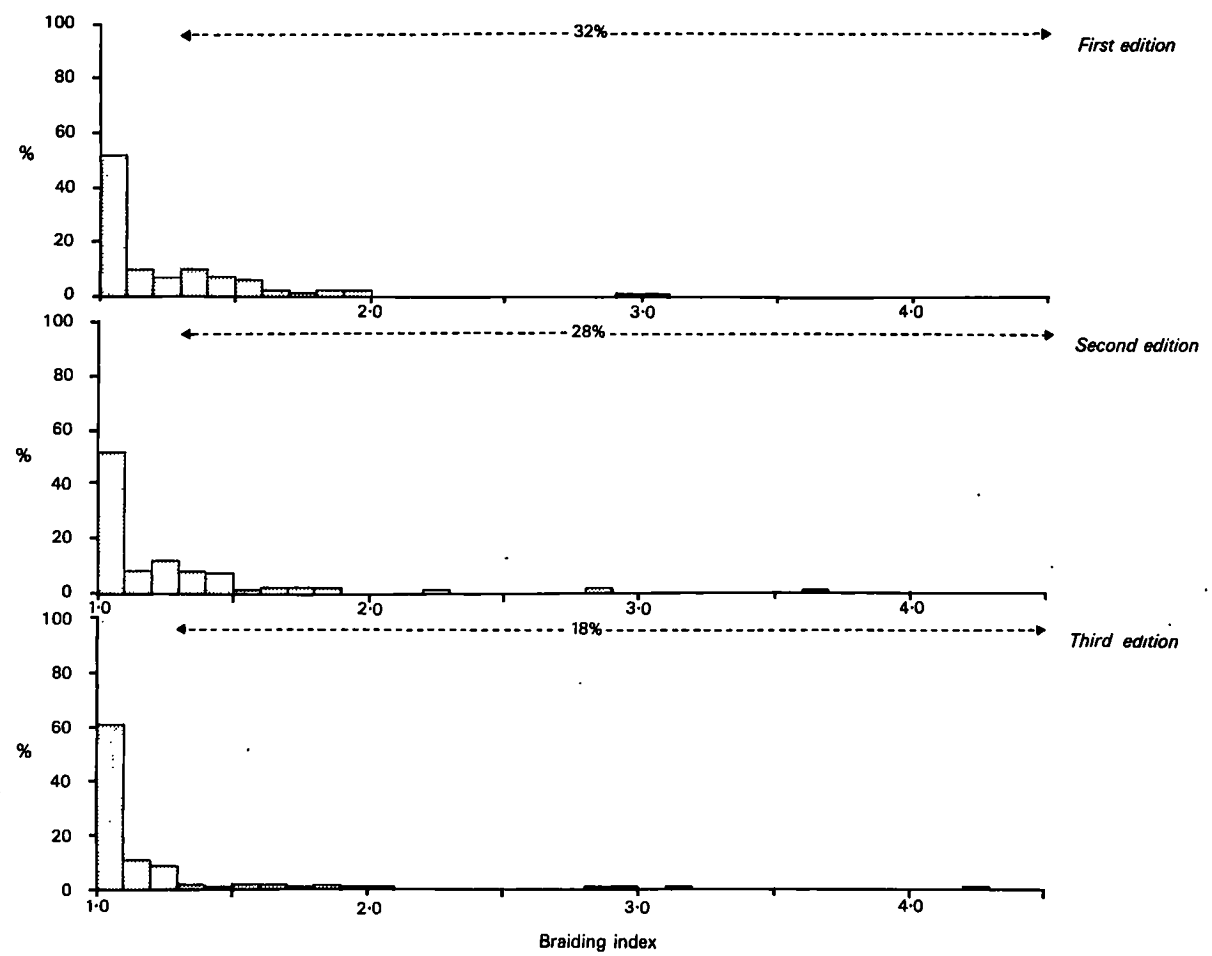


Figure 4.7.1.(ix)

Sampled braiding index values falling within the Tromie catchment for the three map dates

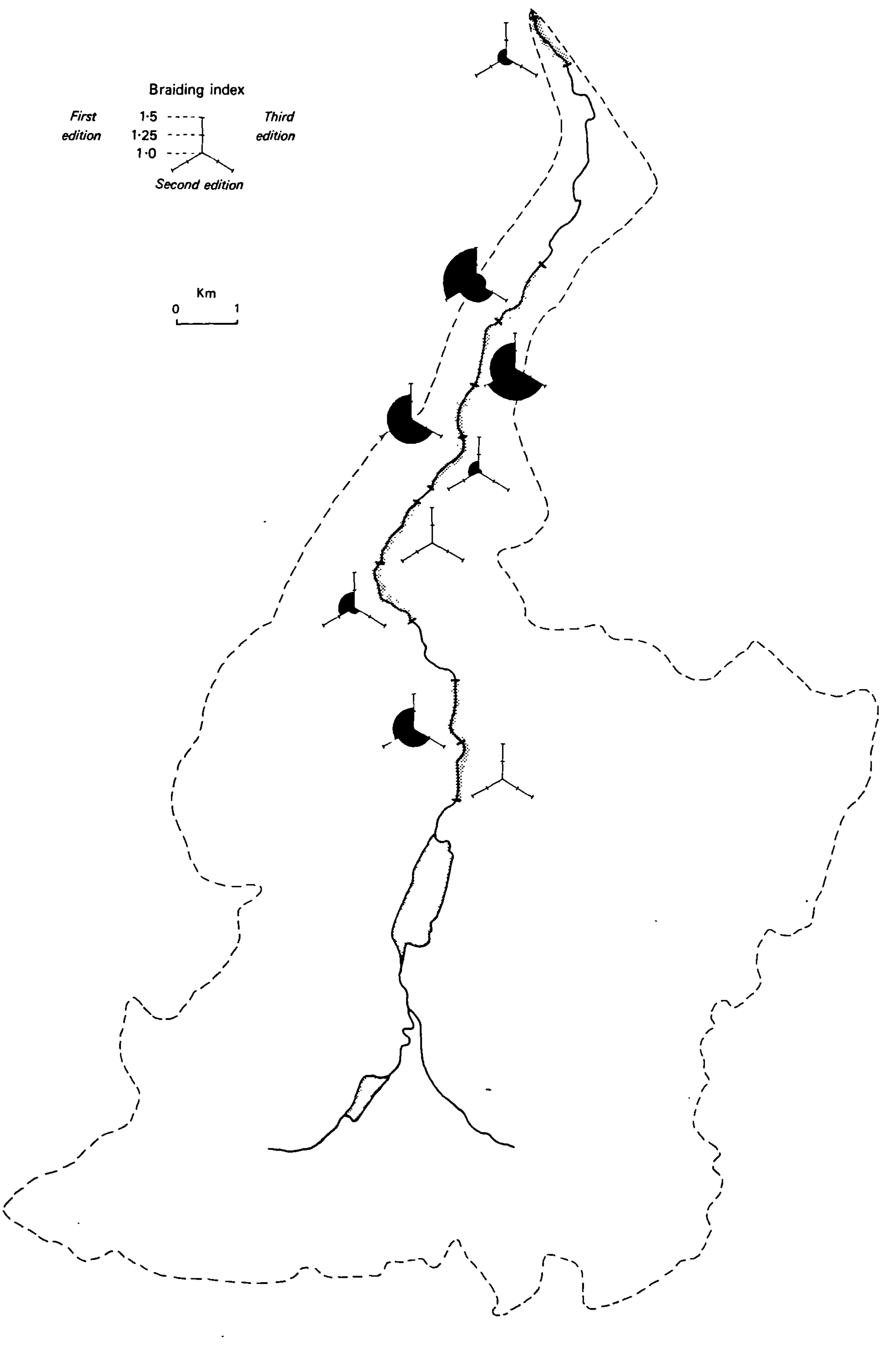


Figure 4.7.3.(x)

Sampled braiding index values falling within the Feshie catchment for the three map-dates

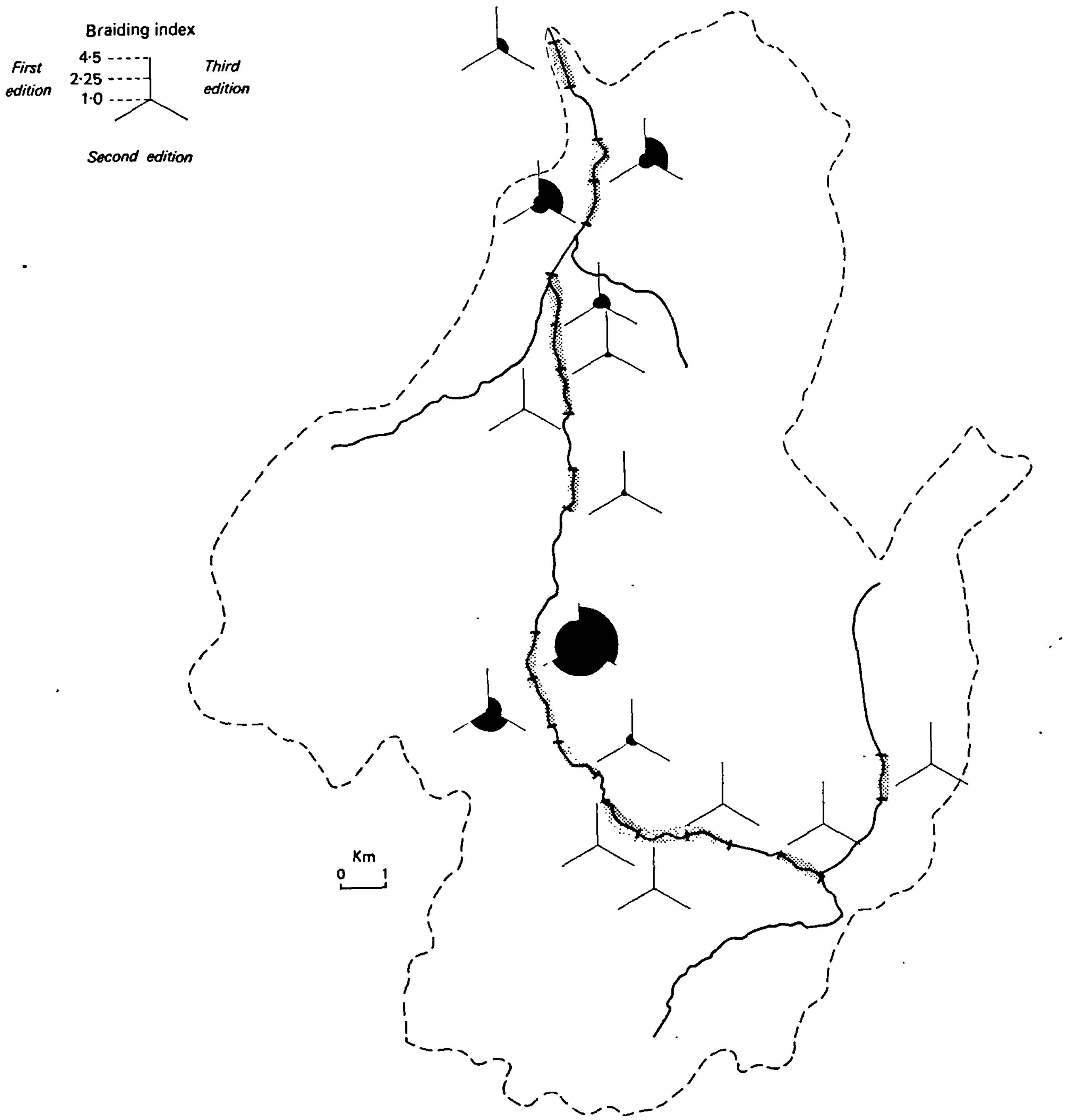


Figure 4.7.1.(xi)

Sampled braiding index values falling within the Druie catchment for the three map dates

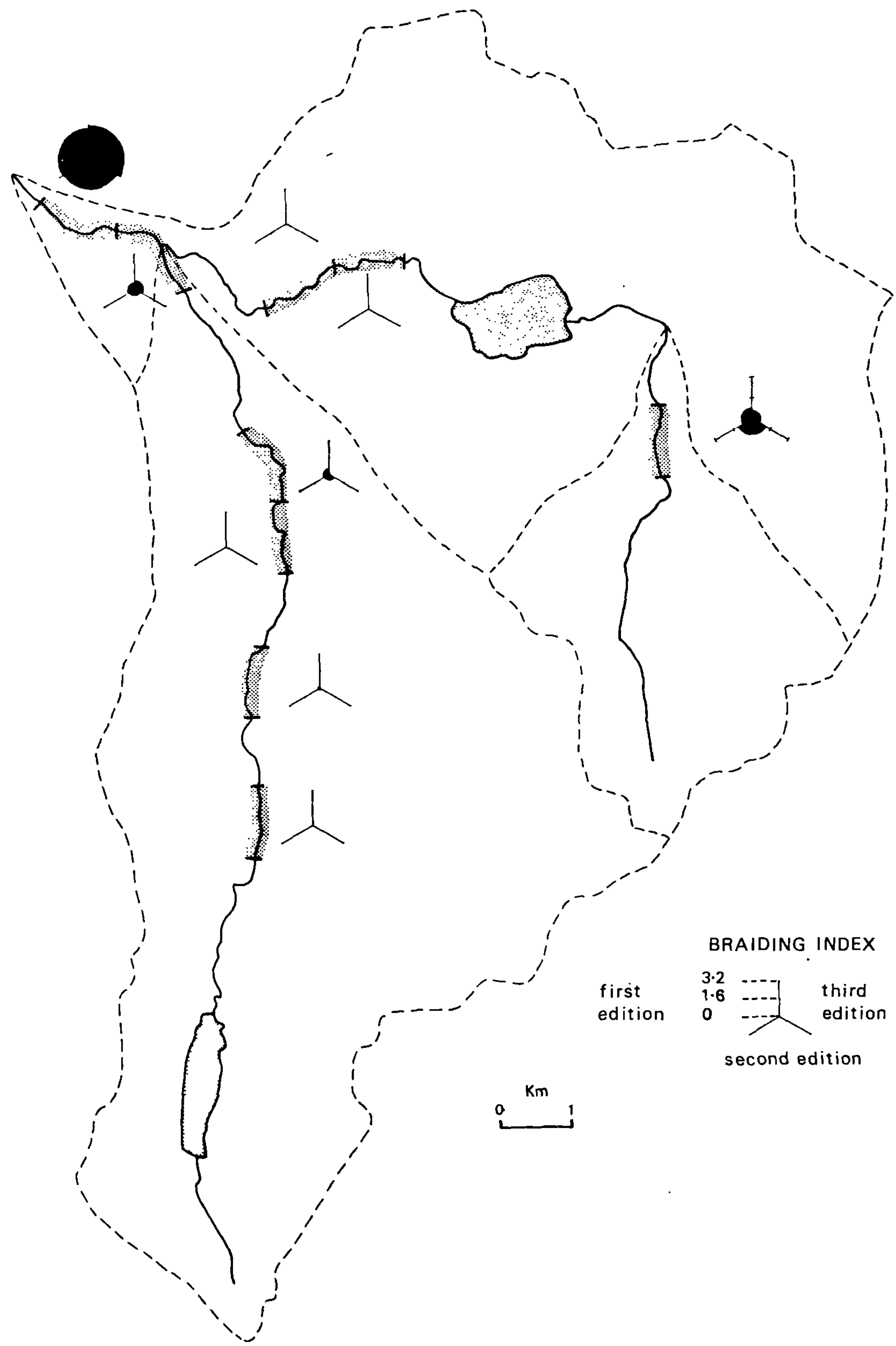


Figure 4.7.1.(xii)

Sampled braiding index values falling within the Nethy catchment for the three map dates

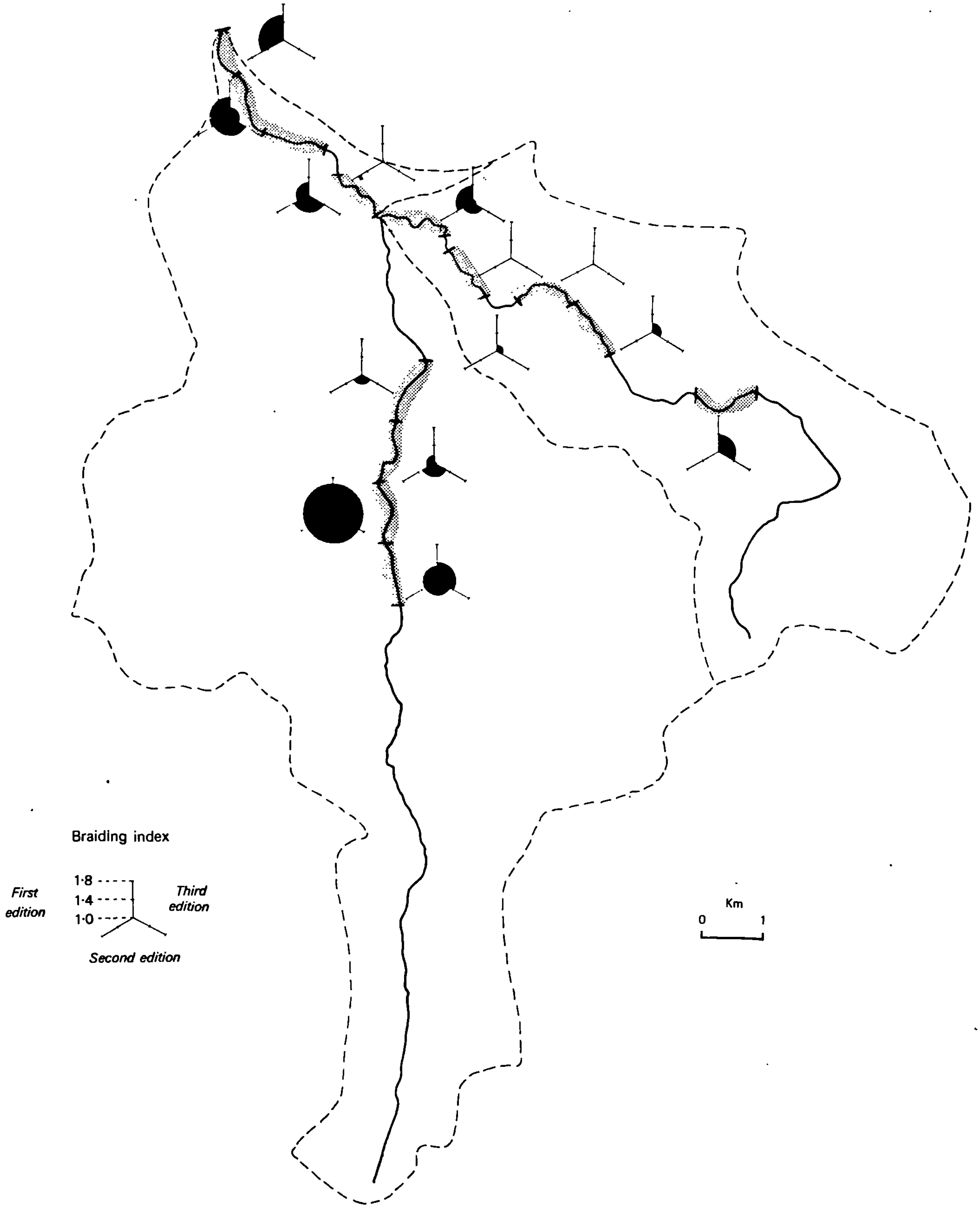


Figure 4.7.3.(xiii)

Sampled braiding index values falling within the Avon catchment for the three map dates

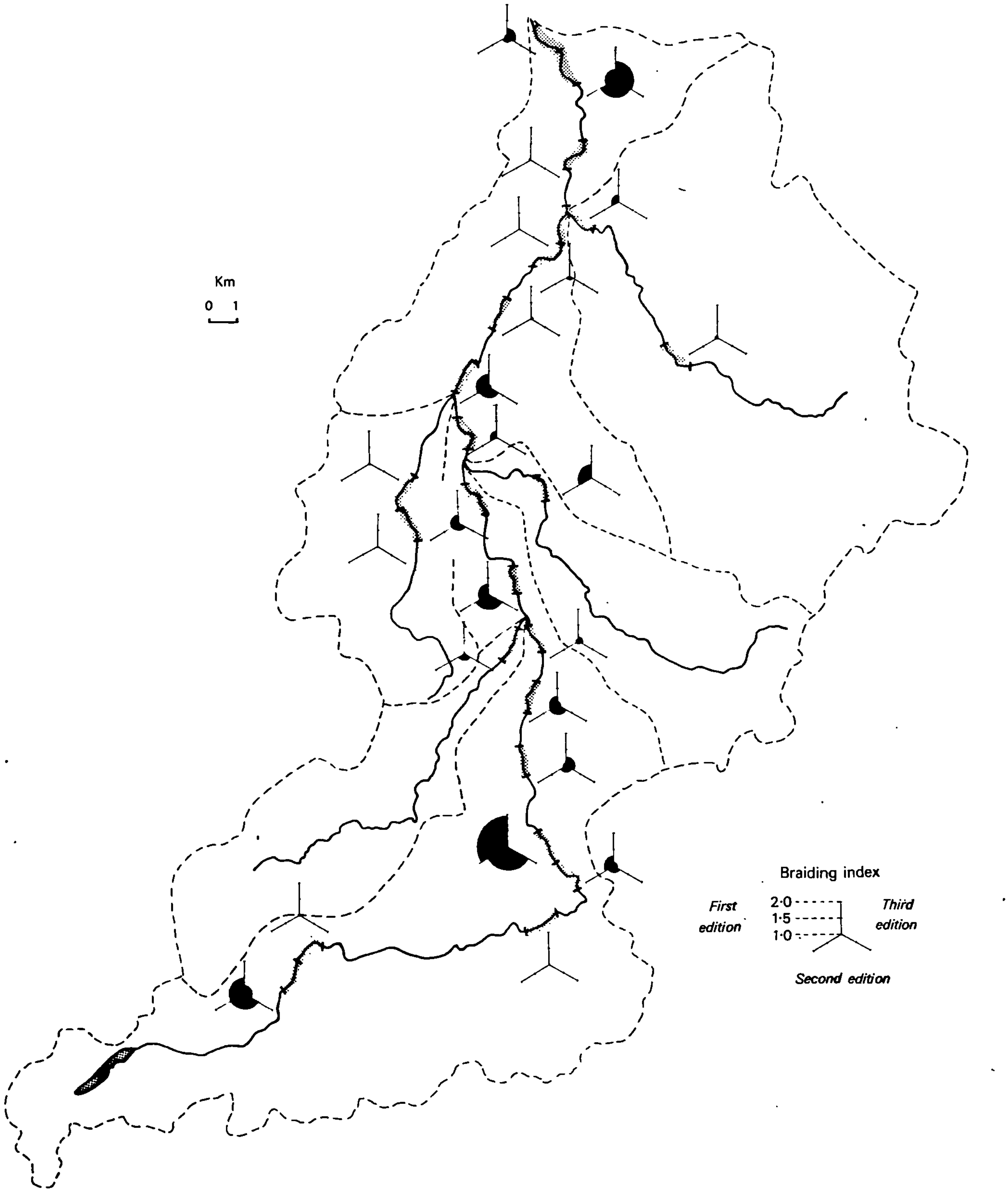
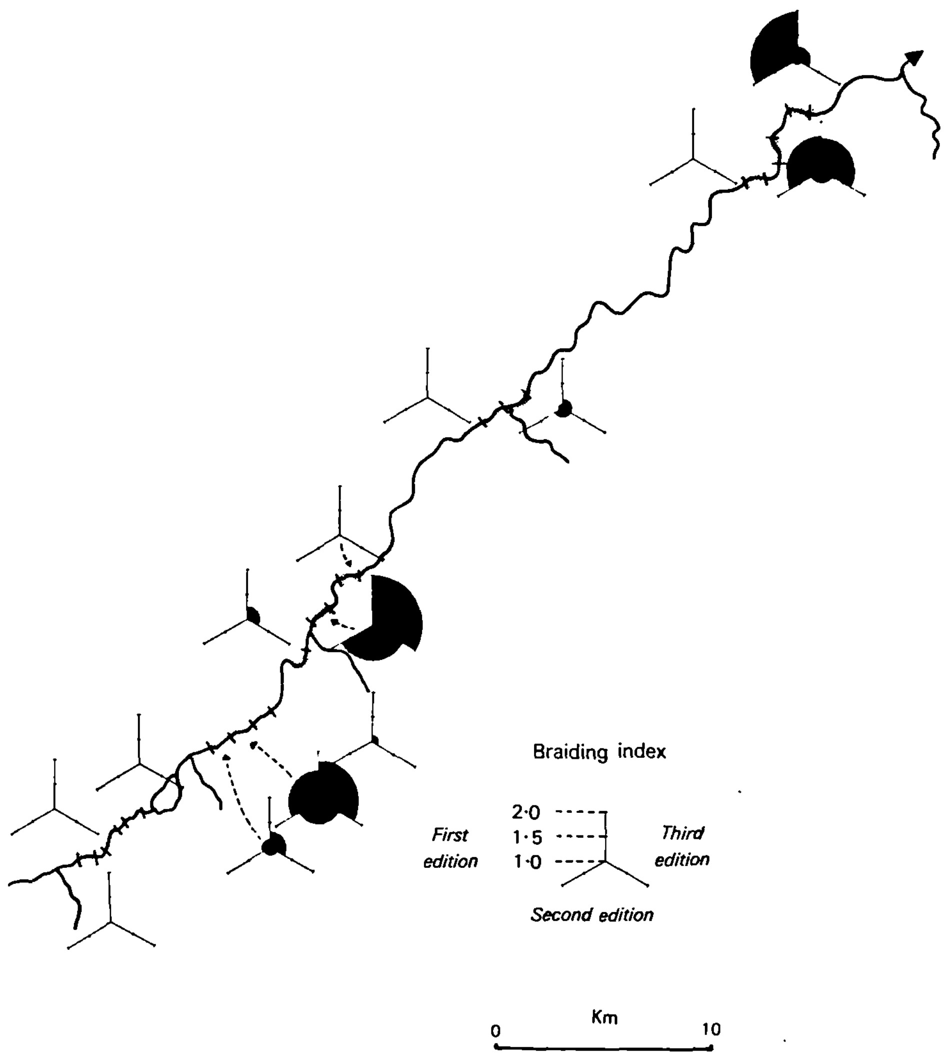


Figure 4.7.1.(xiv)

Sampled braiding index values falling on the mainstream Spey for the three map dates



When the braiding index values were plotted within the catchments, certain patterns were broadly evident, although these were subject to some exceptions. This highlights spatial similarities in controls within similar locations of different catchments, but also the delicate shift in balance of controls that may occur. High braiding indices occurred at or near confluence sites- on the Nethy (2.23 in BI2; Sample 34), Feshie (1.87 in BI1; Sample 10), and Druie (3.14 in BI1; Sample 25), but in contrast, the Tromie (BI=1.00-1.12; Sample 9) and Avon (BI=1.00; Sample 48) joined the Spey by single channels. Other occasional high values occurred on upland alluvial fans associated with mountain torrents eg. the Allt Mor fan (Sample 33). Finally, in certain catchments, samples which fell in the middle reaches of rivers also recorded locally high values indices eg. the Avon, Dorback Burn and Feshie (see Figures 4.7.1.(ix) to (xiv)), but not the Druie, as the Luineag was highly stable with few bars. The periodicity of these higher values varied from an almost regular but contained interspacing of split reaches on the Avon, to the major wandering braided reaches on the Feshie. There must be important differences in the slopes and availability of sediment within different catchments (see Figure 3.3.1.(i)).

On the mainstream Spey, long reaches had no bars (Figure 4.7.1.(xiv)) but these were interspaced with reaches of anabranching planform with higher values (eg. BI=1.46-1.79; Sample 76) which must represent sediment stores along the Spey. The reaches upstream of Loch Inch had no bar features but immediately after the Feshie and Druie confluences, there was considerably more sediment stored within the channel on the Spey. These must be related to large amounts of reworked fluvioglacial material flushed out into the mainstream Spey by its tributaries, especially by the Feshie (Figure 4.7.1.(xiv)).

4.7.2 Basic range of typology characteristics

4.7.2.1 Quantitative catchment characteristics

All the quantitative catchment characteristics (area, average height and MAXWID) were correlated with activity rates but only the relationships reported below were statistically significant. In the following analysis, a significant correlation (r) with a sample size of 80 was 0.22 (0.05 level) and 0.28 (0.01 level). When the importance of height was assessed, only the Feshie catchment had a significant negative correlation ie. sinuosity increased as height decreased. When the total sample was correlated with ($\log_{10}$) MAXWID, a significant positive correlation of r=0.22-0.24 was found ie. sinuosity was higher when area available to migrate was higher but there were plenty of reaches with high apparent MAXWID, which still maintained lower sinuosities. This must suggest confinement by terraces or bedrock close to the present channel level. Certainly, terrace sequences have been documented on the River Feshie (se Section 3.3.3.). None of these correlations was however a powerful predictor, suggesting more complex controls.

4.7.1.3 Range of catchment and reach characteristics

In terms of both upstream and local sediment, in 17% of the samples, there were both large and extensive amounts of unvegetated sediment by and within the channel (categories 2 and 3), especially in the first and second editions (Figure 4.7.2.(i)). There was a much lower % (28-30%) with "no evidence of sediment" proximal or within the channel, than within the Dee sample. This must relate to the increased volume of fluvioglacial material within the Spey study area (see Section 3.3.2) and hence greater availability/ accessibility for reworking and incorporation into the channel system.

When vegetative cover was studied, the most frequent category of both floodplain and bank vegetation was (3) trees and rough-grassland (cf. (4) heather within the Dee study area). Bar vegetation also indicated some bar stability, with some bars displaying a sere development (bar vegetation 1-3), associated trees indicating a bar surface which had not been reworked for several years.

Analysis of channel description indicated the most frequent category of "channel pattern" was "sinuous", with "wandering" and "irregular" also frequent. There were fewer samples falling in the "meandering" categories of the typology (6-8) (cf. the Dee and Tweed samples; Figure 4.7.2.(i)). This explains the generally low sinuosity index values. However, evidence of previous "activity" revealed a variety of palaeofeatures which indicated pre-1850 movement, and which in some cases could be correlated with the planform from Roy's map (1750). Differential vegetation cover also frequently indicated a past

Figure 4.7.2.(i)

Frequencies within each channel typology classification for the Spey study area (for detailed key, see Table 4.2.(i))

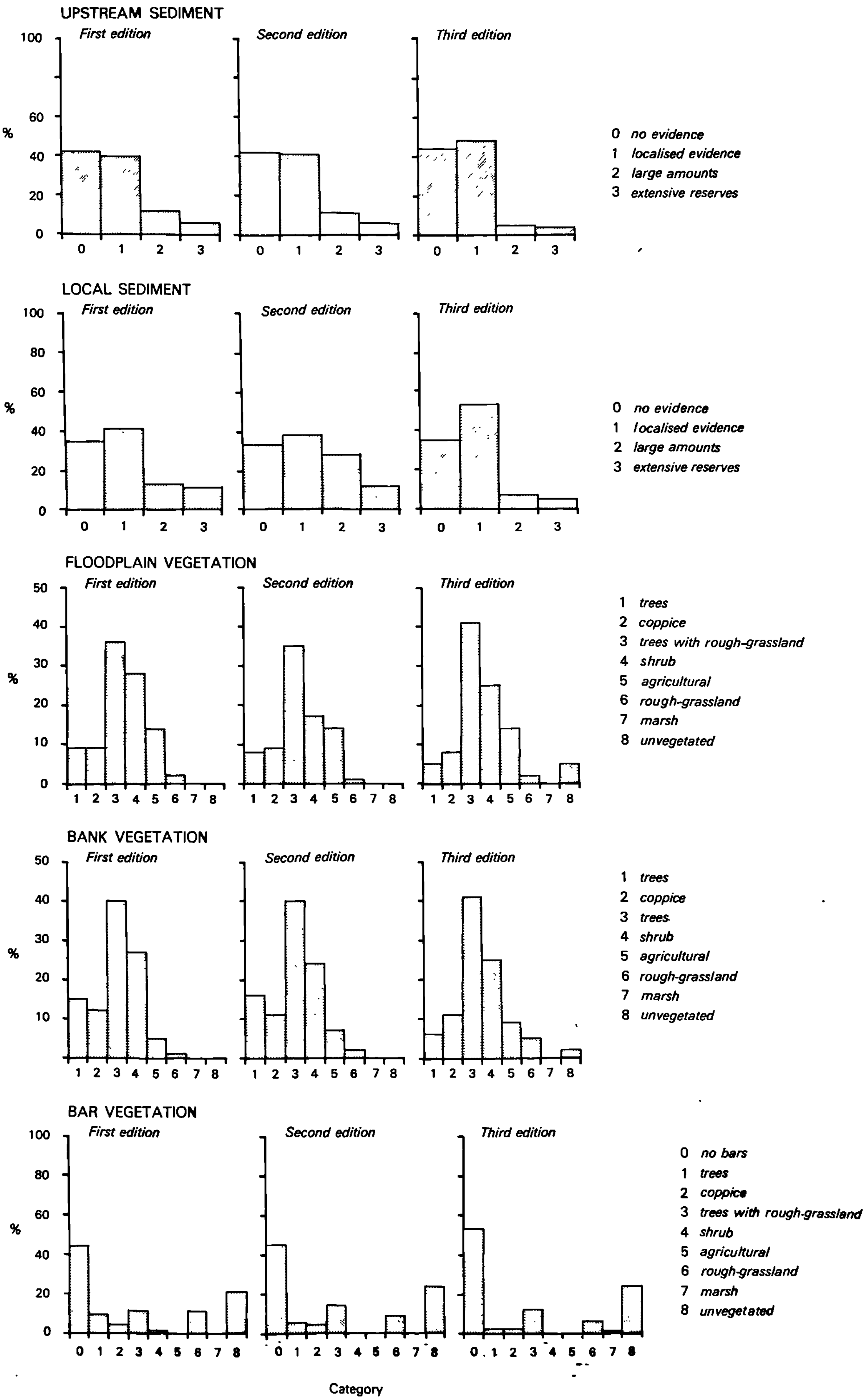
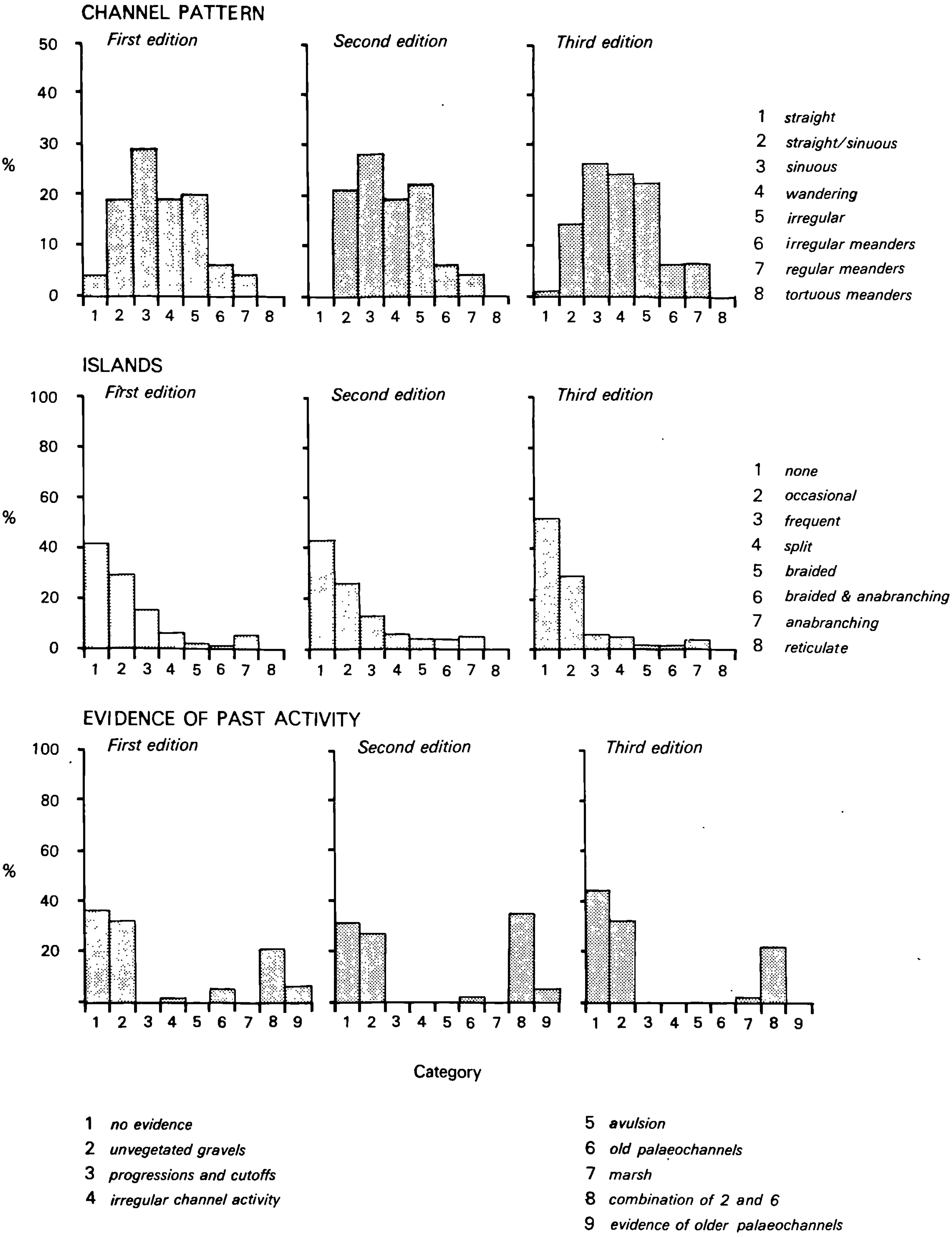


Figure 4.7.2.(i) cont.

Frequencies within each channel typology classification for the Spey study area

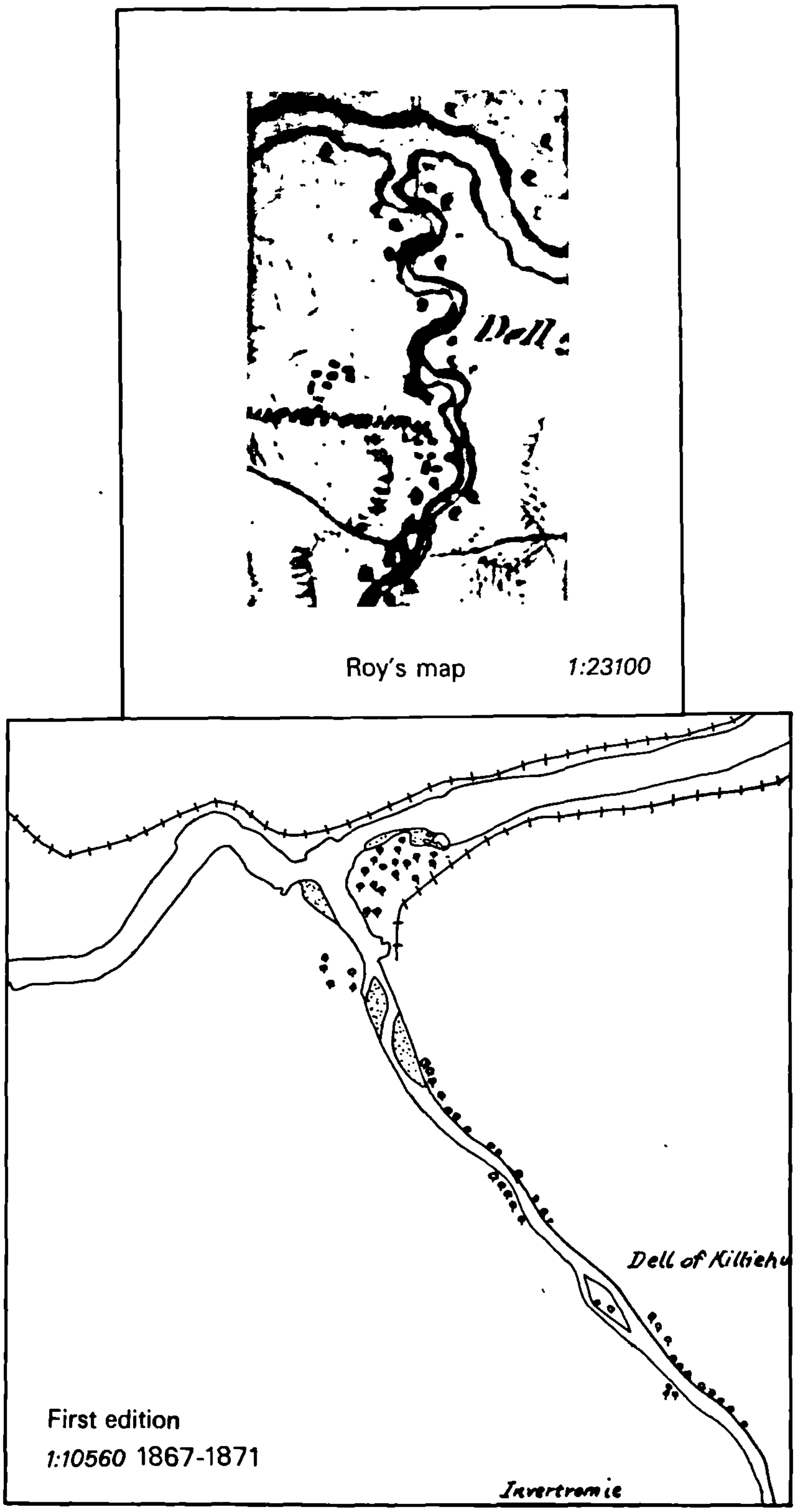


history of channel shift, especially on the Avon, where there was evidence of split planforms at several reaches pre-1850. For example, the Avon had several disused split flood anabranches, cut through large amplitude, irregular meander bends at periodic intervals downstream of Tomintoul eg. at Kilnmaichlie, which must date to at least one earlier flood event (Section 5.5.1). This suggested spatially recurrent forms of change may take place within the same catchment where planform controls show periodic similarity downstream, and where response to flood events is similar.

Roy's map provided other indications of planform change. For example, the Tromie confluence (Sample 9) was indicated as highly sinuous with regular meanders in 1750, while by 1867-1871, it was a straight/ sinuous channel (Figure 4.7.2.(ii)). It is interesting also to compare this with the contrasting braided planform above the Feshie confluence, as indicated on Roy's map. This again suggests considerable spatial and temporal variation in controls. High braiding indices within the middle Feshie catchment were found on Roy's map, indicating that a highly divided channel planform has existed in excess of a 250 year timespan (eg. Sample 17, the upper braided reach; see Figure 4.7.2.(iii)). Other map-based palaeofeatures included an oxbow cutoff on the mainstream Spey above Loch Inch (Figure 4.7.2.(iv)), which followed the parish boundary and occurred on the area of former lake bed, described in Chapter 3.3.3. Whether this abandonment was artificial or natural, it was difficult to tell but the meander was certainly tortuous, unlike any sample reaches on the present Spey. This length of channel formerly may have been similar to the tortuous meanders of Glen Derry (Deeside) but at a much larger scale, with low

Figure 4.7.2.(ii)

Planform change above the Tromie confluence



Roy's map *1:23100*

First edition
1:10560 1867-1871

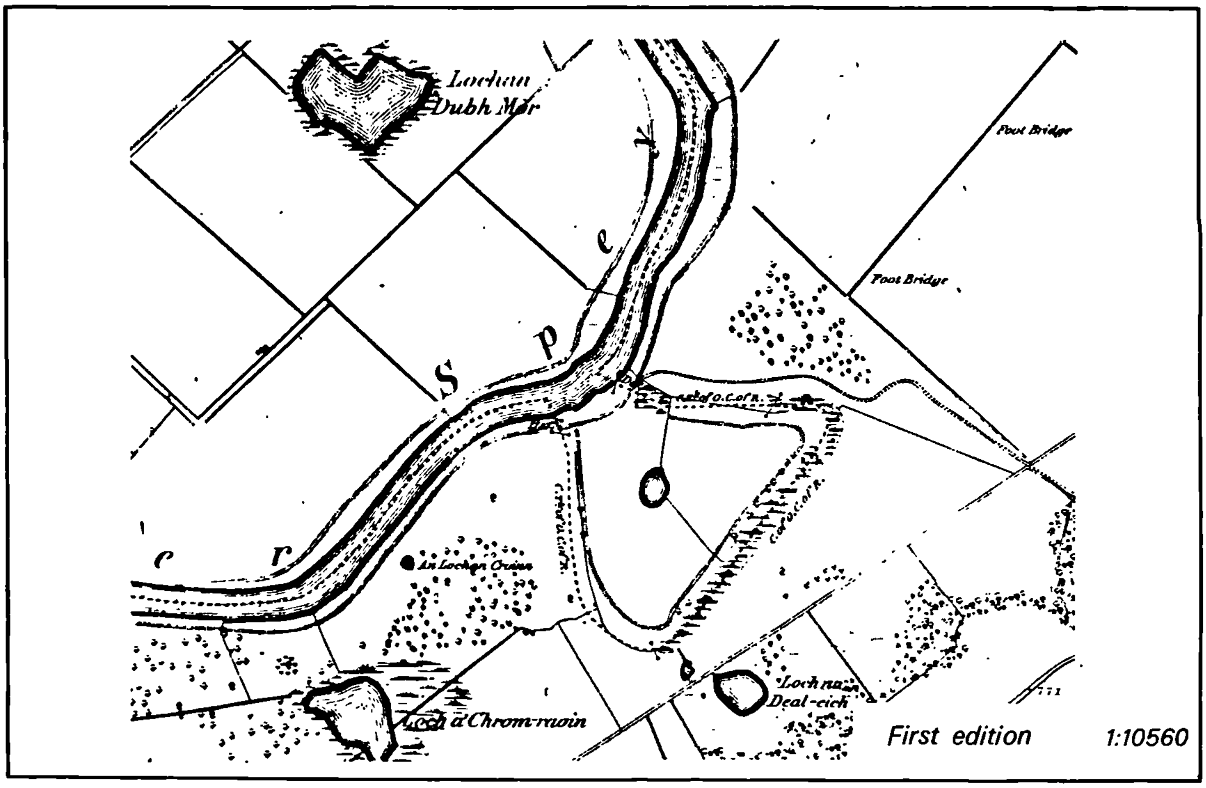
Figure 4.7.2.(iii)

Upper braided reach on the Feshie as indicated on Roy's map

Roy's map

1:23100

Figure 4.7.2.(iv)



Former tortuous meander on the River Spey above Loch Insh

valley slopes and Loch Inch acting as a local baselevel.

No sample sites had any annotation of "liability to flooding" at the third edition while the first and second editions had 3 and 5 respectively. Reaches above confluence sites seemed most susceptible to perception of a flooding hazard. The Nethy confluence was noted for its "liability to flooding" on both the first and second edition maps, 1871 and 1900 respectively. Similarly, the Feshie confluence was also susceptible to frequent inundation. Most frequent indication of a flood problem occurred on the mainstream Spey, especially above Loch Insh on the first and second edition maps.

4.7.3 Temporal variation in channel pattern

4.7.3.1 Modes of channel adjustment

A large range in magnitude of different modes of planform adjustment were found in the Spey catchments, but not the same range of form as found in upper Deeside, ie. there was a greater tendency for avulsion rather than cutoffs. As within the Dee study area, the categories of change could be subdivided into those which caused an overall increase or decrease in the width of the active area and those which maintain the active area in an equilibrium form. Table 4.7.3.(i) shows the types of planform change found at the sample sites and illustrative examples are shown in Figure 4.7.3.(i).

Figure 4.7.3.(i)

Examples of categories of channel pattern change sampled within the Spey study area

Sample 19

Sample 34

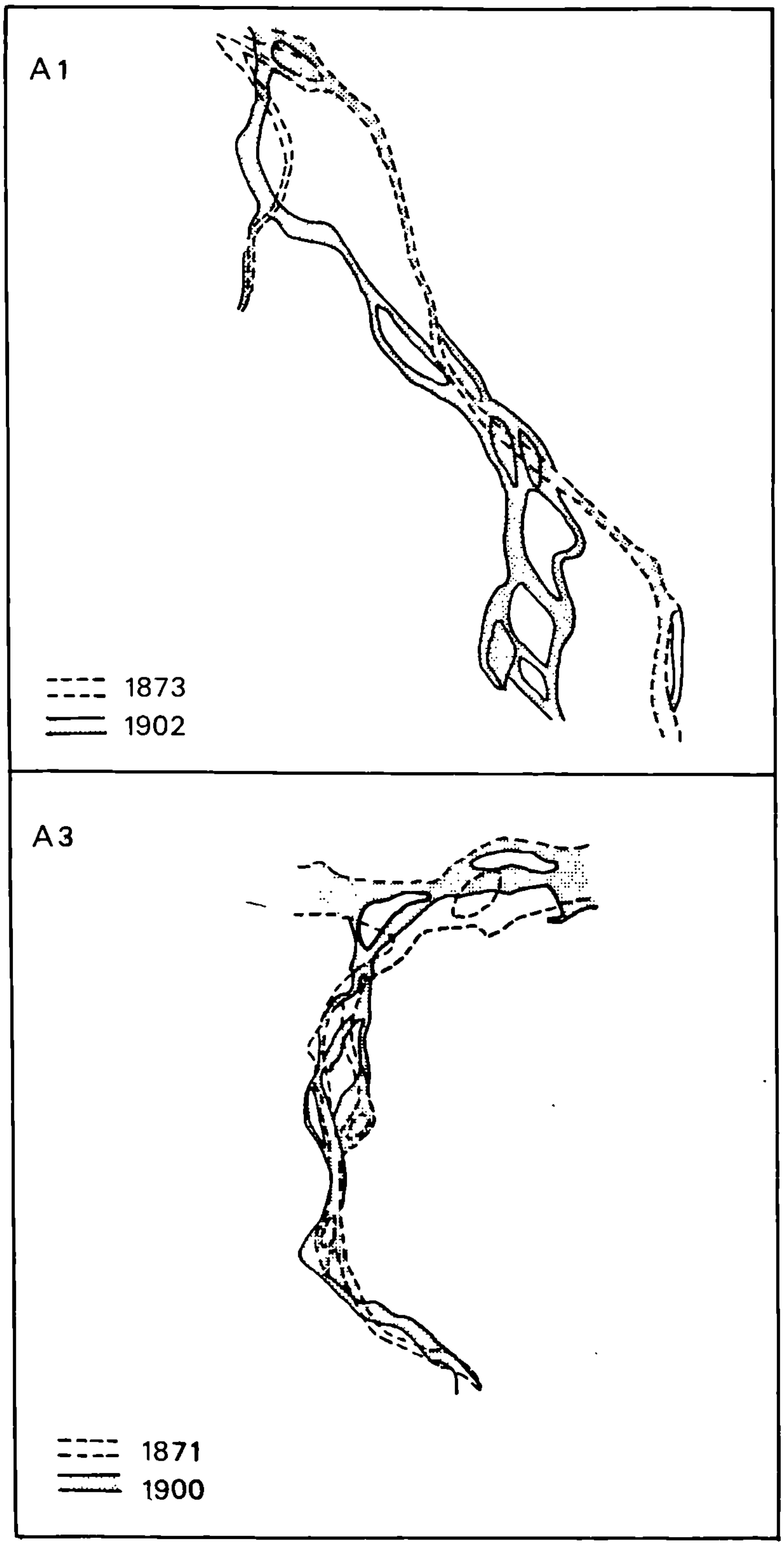


Figure 4.7.3.(i) cont.

Examples of categories of channel pattern change sampled within the Spey study area

Sample 62

Sample 79

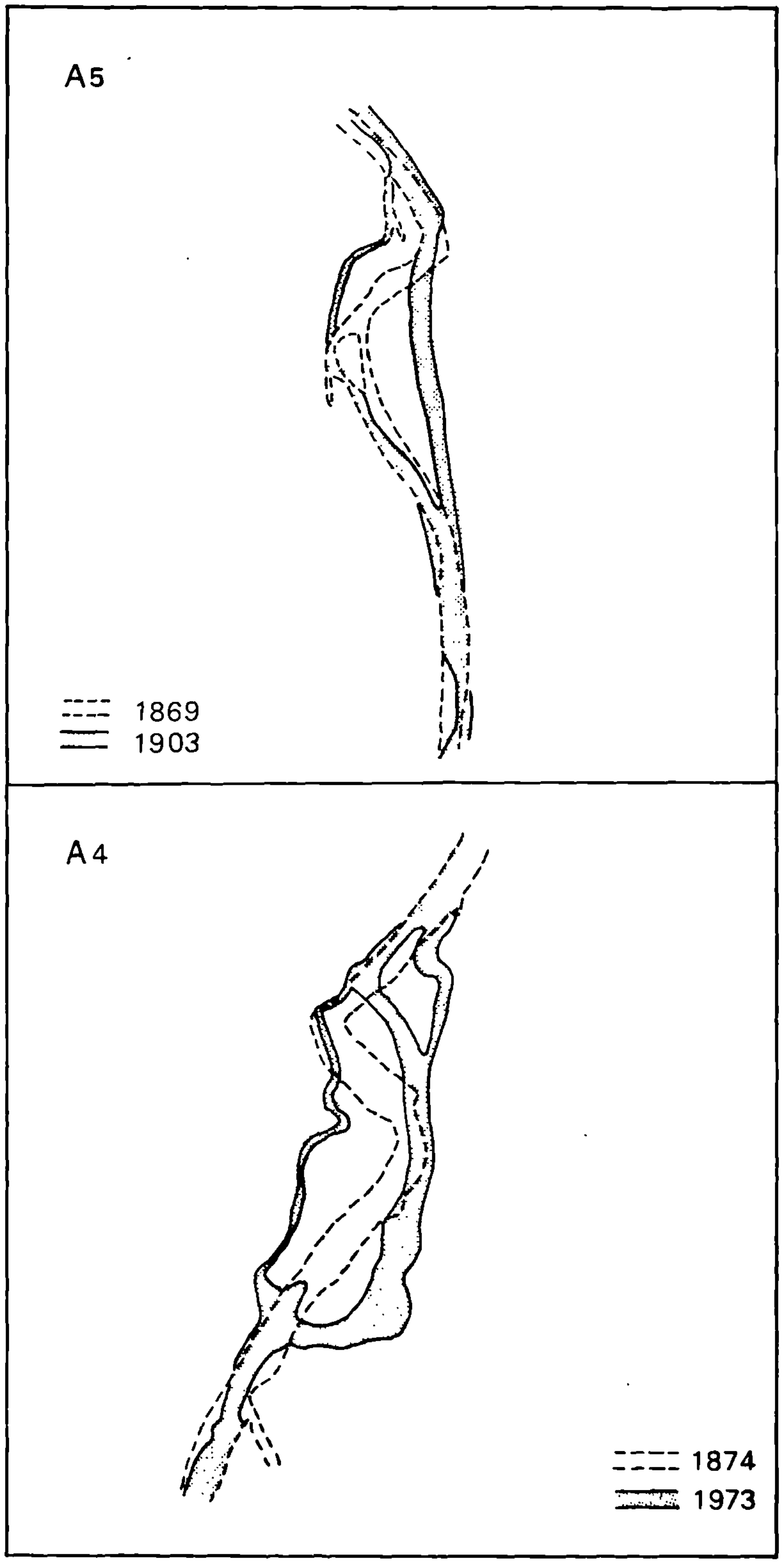
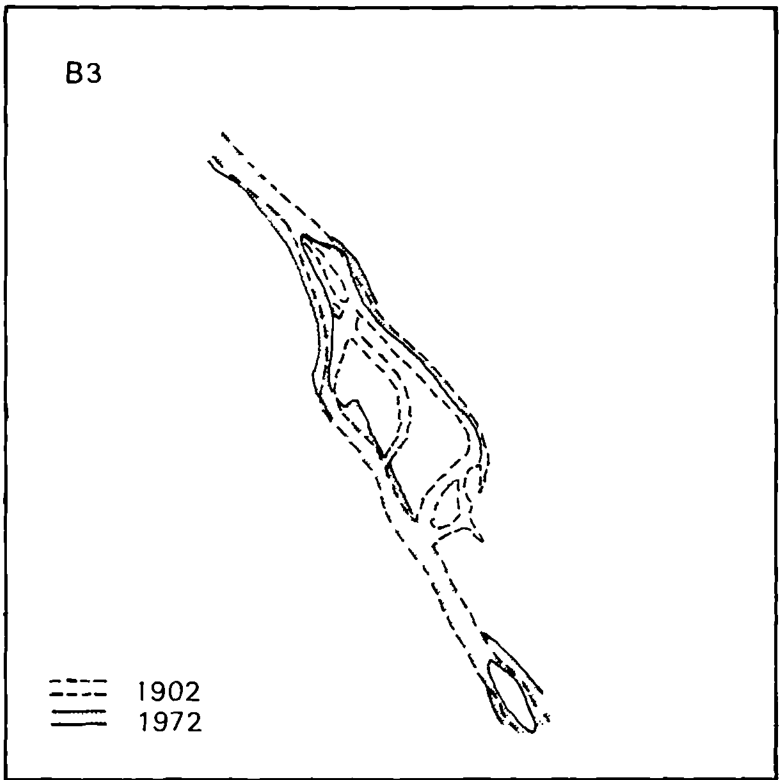


Figure 4.7.3.(i) cont.

Examples of categories of channel pattern change sampled within the Spey study area

Sample 67



Sample 19

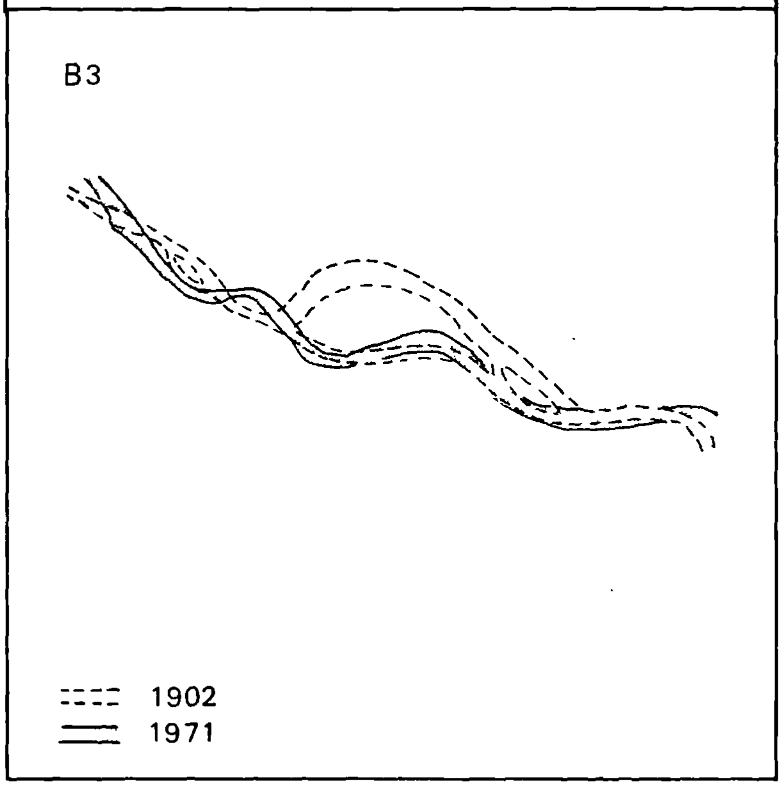


Table 4.7.3.(i)

Categories of channel pattern change found within the Spey study area

(A) Changes associated with an increase in width of active area

1: Permanent channel metamorphosis

2: Transient channel metamorphosis

3: Migration/ periodic avulsion across gravel fans

4: Complex major extra-channel avulsion

5: Single major extra-channel avulsion

6: Single minor extra-channel avulsion

7: Complex minor extra-channel avulsion

8: Major intra-channel avulsion with associated bank erosion

9: Minor intra-channel avulsion with associated bank erosion

10: Reworking/ trimming stabilised bars/banks into the present active area

(B) Changes associated with a reduction in width of active area

1: Temporary channel metamorphosis

2: Large scale reduction from multichannelled to a much less divided channel planform

3: Large scale reduction from split to single channel planform

4: Small scale local reduction in the number of channels

5: Large scale meander cutoffs

6: Small scale meander cutoffs

7: Cutting off of low amplitude meander bends

Table 4.7.3.(i) cont.

(C) Changes associated with a maintenance of width of the active area

1: Localised intra-channel avulsion

2: Slow lateral migration

Changes in channel planform involving an increase in active area are tabulated in Table 4.7.3.(i), Section A. Categories 1 and 2 occurred when the channel planform underwent a total large-scale disruption of earlier channel form eg. above the upper Feshie braided reach (Sample 18; 1873-1902), where channel pattern changed from "sinuous" to "wandering" and "islands" altered from "no bars" to "braided/anabranching". The important distinction between categories 1 and 2 was the possible reversibility of such change and the timescale over which it could take place ie. over a defined period of time, did channel planform retrograde to a pre- disruption form or had a major threshold been exceeded? In the example cited, a major threshold had clearly been exceeded and channel pattern 68 years later was still using a much larger proportion of its available active area.

Planform change of a less dramatic nature also occurred. Category 3 involved flow onto a previously unused or well re-established part of the floodplain eg. Nethy and Feshie confluences (Samples 10 and 34; see Figure 4.7.3.(i)) and Sections 7.3.1 and 7.3.2). This caused the flow either to become subdivided or completely diverted through a new avulsion; both resulted in an increase in the active area. Categories 4 to 7 involved increases in the braiding index through incorporation of floodplain outwith the former active area, at a variety of scales. Single major extra-channel avulsions occurred eg. on the upper River Feshie (Sample 19) and upper River Avon (Sample 64; Figure 4.7.3.(i)), while examples of complex, major, extra-channel avulsions occurred eg. on the mainstream Spey (Sample 79) and the Druie at Inverdruie (Samples 25/26). In contrast, categories 8 to 10 occurred when there was shifting around unstable bar features already within the active area eg. on the River Feshie at Lagganlia (Samples 11 and 12; see Section 7.3.4).

Changes in channel pattern associated with a reduction in active area are indicated in Table 4.7.3.(i), Section B. Channel metamorphosis associated with a reduction in channel planform occasionally occurred with a complete change in the character of the pattern (eg. Sample 76, Spey study reach 9 (see Section 7.2.9)). This might have been caused by a change in sediment supply or an alteration in the range of flows that would have previously reworked the planform. Other categories included a switch from a split to a single channel, a change which was frequently reversible at a later date. This occurred at a variety of scales for example on the Livet/Avon confluence (Sample 51), and Feshie (Sample 16; Figure 4.7.3.(i)). Chute and neck meander cutoffs were occasionally found at a variety of scales but only on localised reaches, such as on the mainstream Spey (chute) and Conglass Water (neck). As well as the more conventionally recognised neck cutoffs, there were also sweeping bypasses of low amplitude meander bends (Category 7), characteristic of the Avon. Flow frequently alternated between either or both channels.

Intra-channel avulsion could in some cases occur without major changes to the dimensions of the active area, as unstable material was reworked within its present limits. Change also took place at a much slower rate, with the maintenance of the dimensions of the former planform (Table 4.5.3.(i)), Section C).

4.7.3.2 Rates of change as indicated by sinuosity index (CHASIN)

As on Deeside, the median and IQR for change in sinuosity index for the two inter-map periods were similar (Table 4.7.3.(ii)), which suggested either that the channels were highly stable in terms of sinuosity or that changes in some reaches were cancelled by changes in the opposite direction by others. When the Wilcoxon test was carried out on sinuosity values for paired combinations of the three map dates (first/second, second/third, first/third), the Z values indicated that no significant difference between pairs occurred within the total sample. However, when the data were subdivided by catchment, the results were rather different. For example, the Nethy catchment showed a highly significant difference between the second and third and the first and third editions (Z= -2.79 and -2.48 respectively), but between first and second editions, no significant difference was recorded. This may be related to the disruption of previously naturally stable reaches or to the reworking of earlier channelisation, as will be assessed in Chapter 6.

Having determined that change had occurred at some sites, specific rates were studied. To allow the pattern of individual sites to be assessed, sinuosity change values were plotted against year of occurrence, to see if any overall pattern emerged for sites where activity had occurred (Figure 4.7.3.(ii) to (vii)). Activity data was also plotted within the catchments to assess any spatial patterns of change (Figure 4.7.4.(viii) to (xiii)). On the Tromie, for example, change in sinuosity was within well-defined limits; there being more change between 1873 and 1900, associated with a reduction in overall

Table 4.7.3.(ii)

Rates of change in sinuosity index within the Spey study area

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
CHASIN1-2	-0.02	0.00	0.01	0.03
CHASIN2-3	-0.02	0.00	0.04	0.06
AVSIN1-2	-0.0006	0.00	0.0003	0.0009
AVSIN2-3	-0.0003	0.00	0.0006	0.0009
Tromie				
CHASIN1-2	-0.025	0.01	0.015	0.040
CHASIN2-3	-0.035	0.00	0.015	0.040
AVSIN1-2	-0.0008	0.0003	0.0005	0.0013
AVSIN2-3	-0.0005	0.00	0.0002	0.0007
Feshie				
CHASIN1-2	0.00	0.00	0.03	0.03
CHASIN2-3	-0.03	0.00	0.04	0.07
AVSIN1-2	0.000	0.000	0.0010	0.0010
AVSIN2-3	-0.0004	0.000	0.0006	0.0010
Druie				
CHASIN1-2	-0.018	-0.005	0.03	0.048
CHASIN2-3	-0.035	-0.010	0.023	0.058
AVSIN1-2	-0.0006	-0.0002	0.001	0.0016
AVSIN2-3	-0.0005	-0.0001	0.0003	0.0008

Table 4.7.1.(ii) cont.

Parameter	25th Percentile	Median	75th Percentile	IQR
Nethy				
CHASIN1-2	-0.01	0.01	0.02	0.03
CHASIN2-3	0.02	0.05	0.08	0.06
AVSIN1-2	-0.0004	0.0001	0.0008	0.0012
AVSIN2-3	0.0002	0.0007	0.0011	0.0009
Avon				
CHASIN1-2	-0.02	0.00	0.00	0.02
CHASIN2-3	-0.05	0.00	0.03	0.08
AVSIN1-2	-0.0006	0.00	0.00	0.0006
AVSIN2-3	-0.0007	0.00	0.0004	0.0011
Spey				
CHASIN1-2	-0.03	0.01	0.00	0.03
CHASIN2-3	-0.01	0.01	0.04	0.05
AVSIN1-2	-0.0009	-0.0003	0.00	0.0009
AVSIN2-3	-0.0001	0.0001	0.0006	0.0007

Figure 4.7.3.(ii)

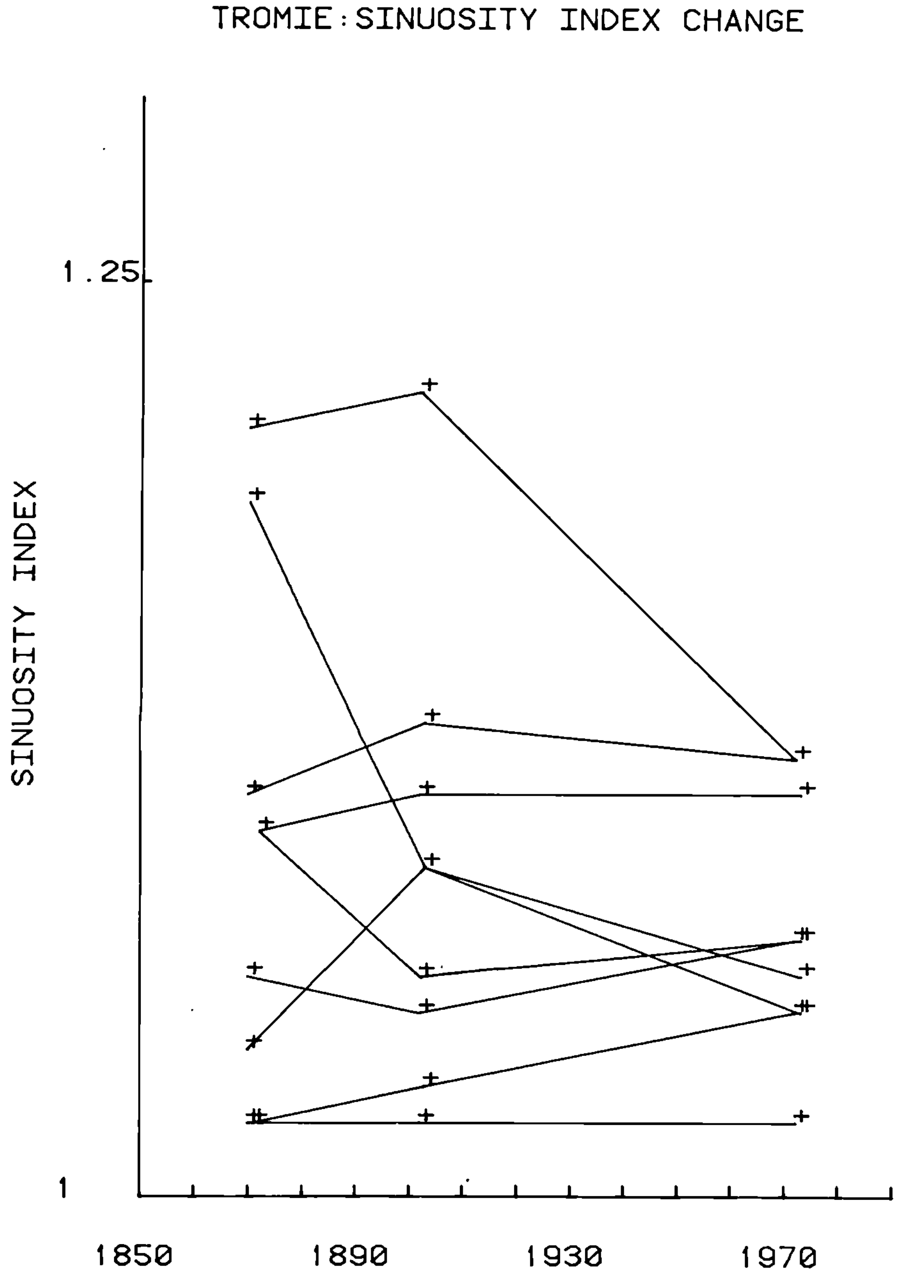
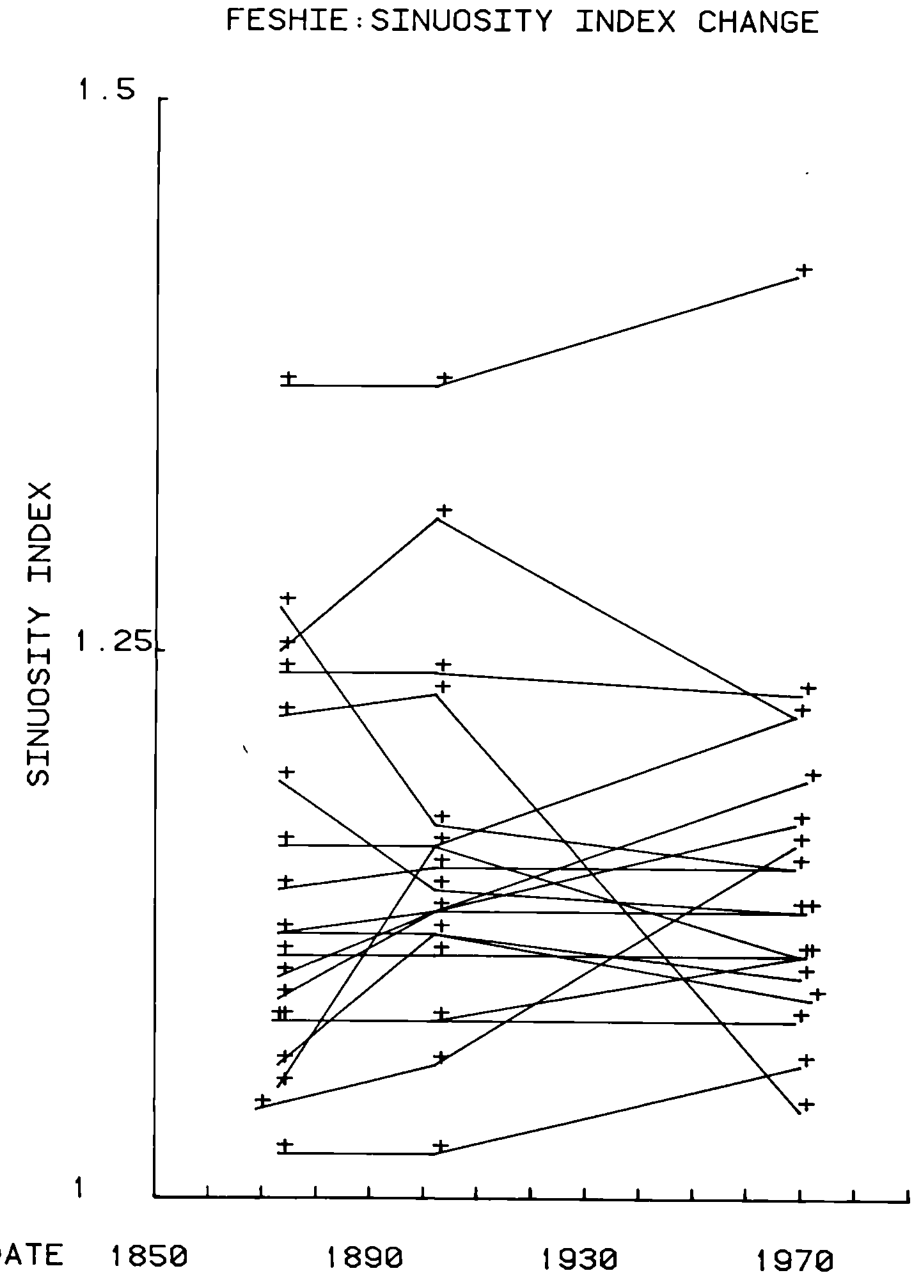


Figure 4.7.3.(iii)



MAP-DATE

Figure 4.7.3.(iv)

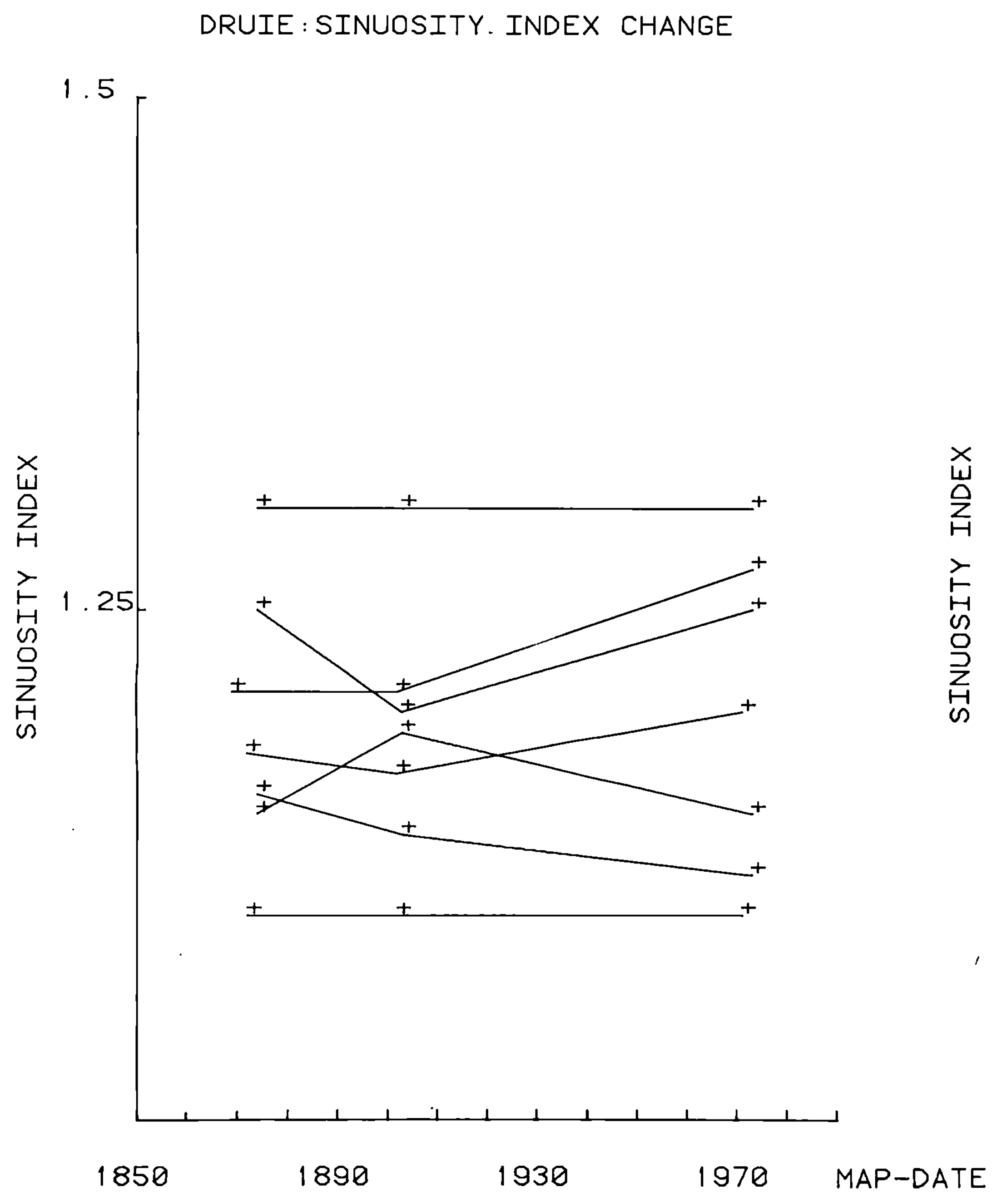


Figure 4.7.3.(v)

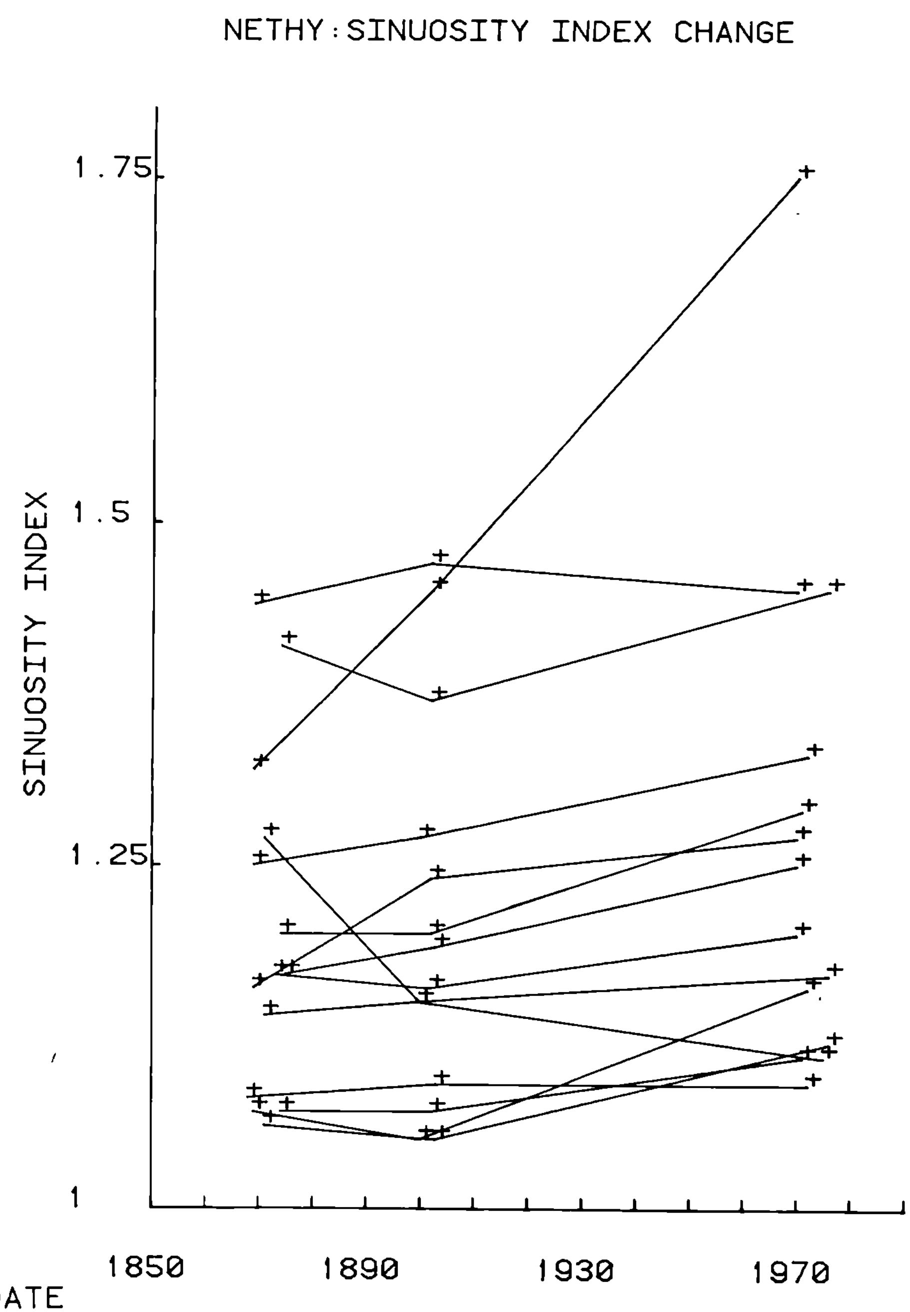


Figure 4.7.3.(vi)

AVON: SINUOSITY INDEX CHANGE

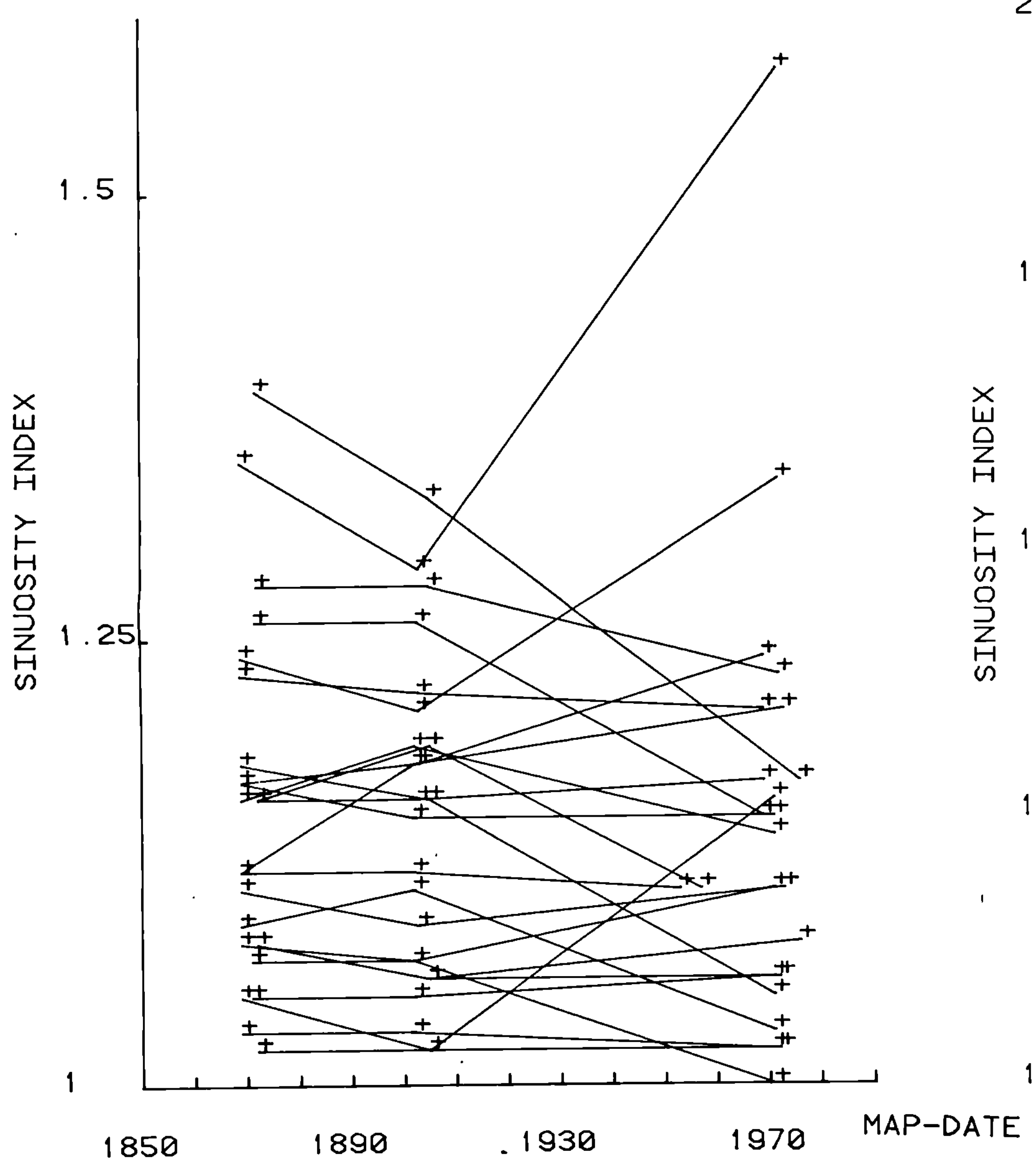


Figure 4.7.3.(vii)

MAINSTREAM SPEY: SINUOSITY INDEX CHANGE

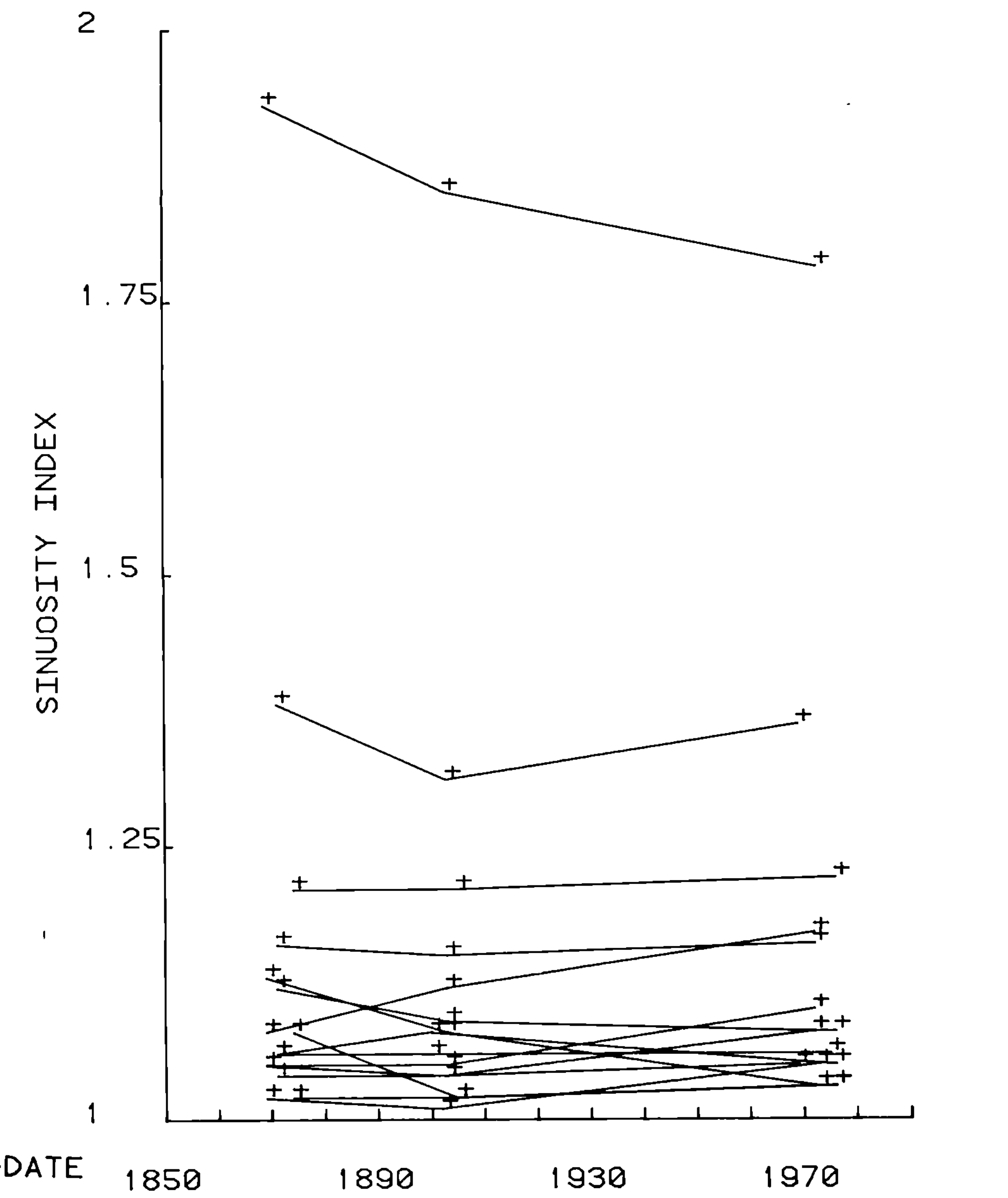


Figure 4.7.3.(viii)

Sampled changes in sinuosity index falling within the Tromie catchment during the two intermap periods

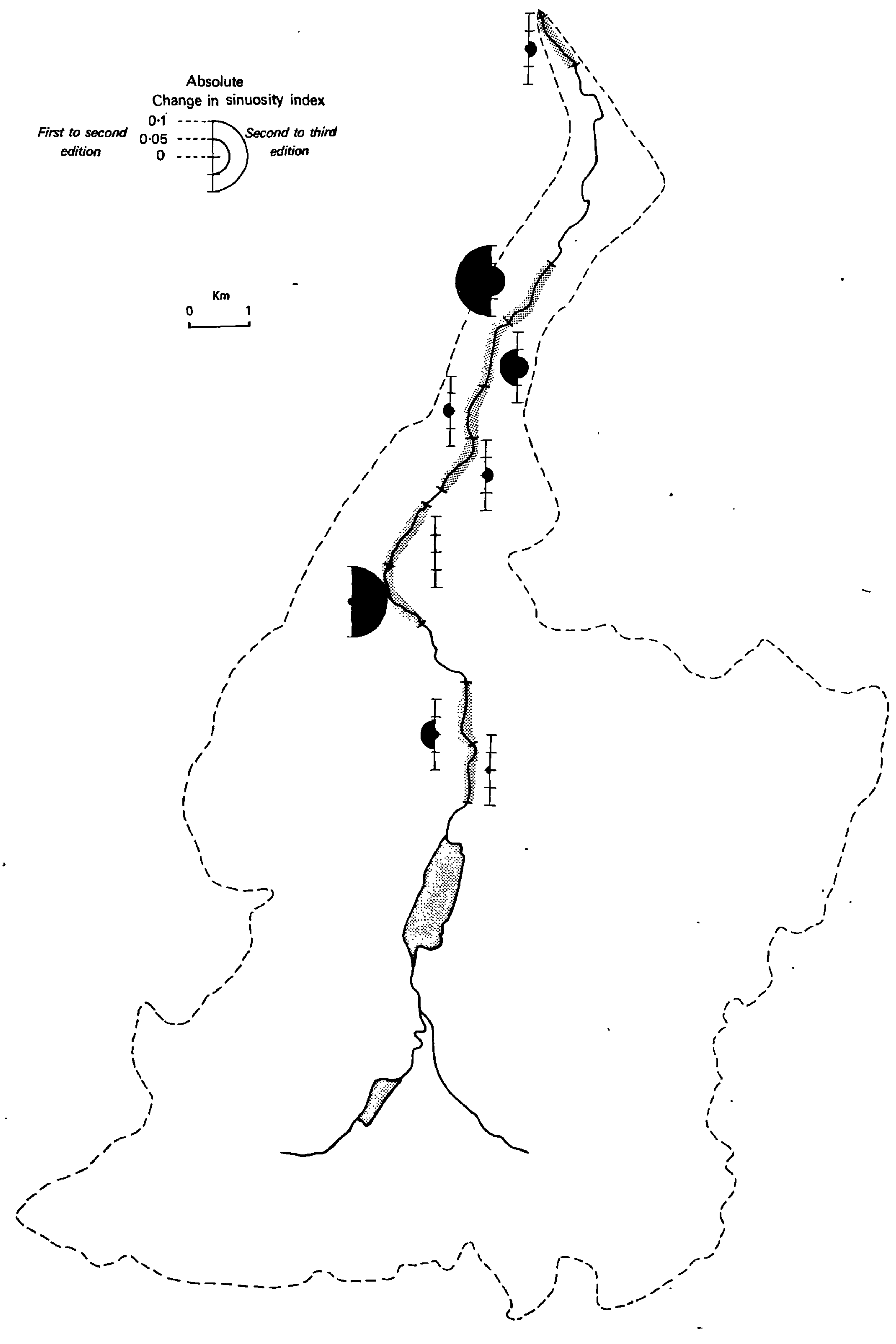


Figure 4.7.3.(ix)

Sampled changes in sinuosity index falling within the Feshie catchment during the two intermap periods

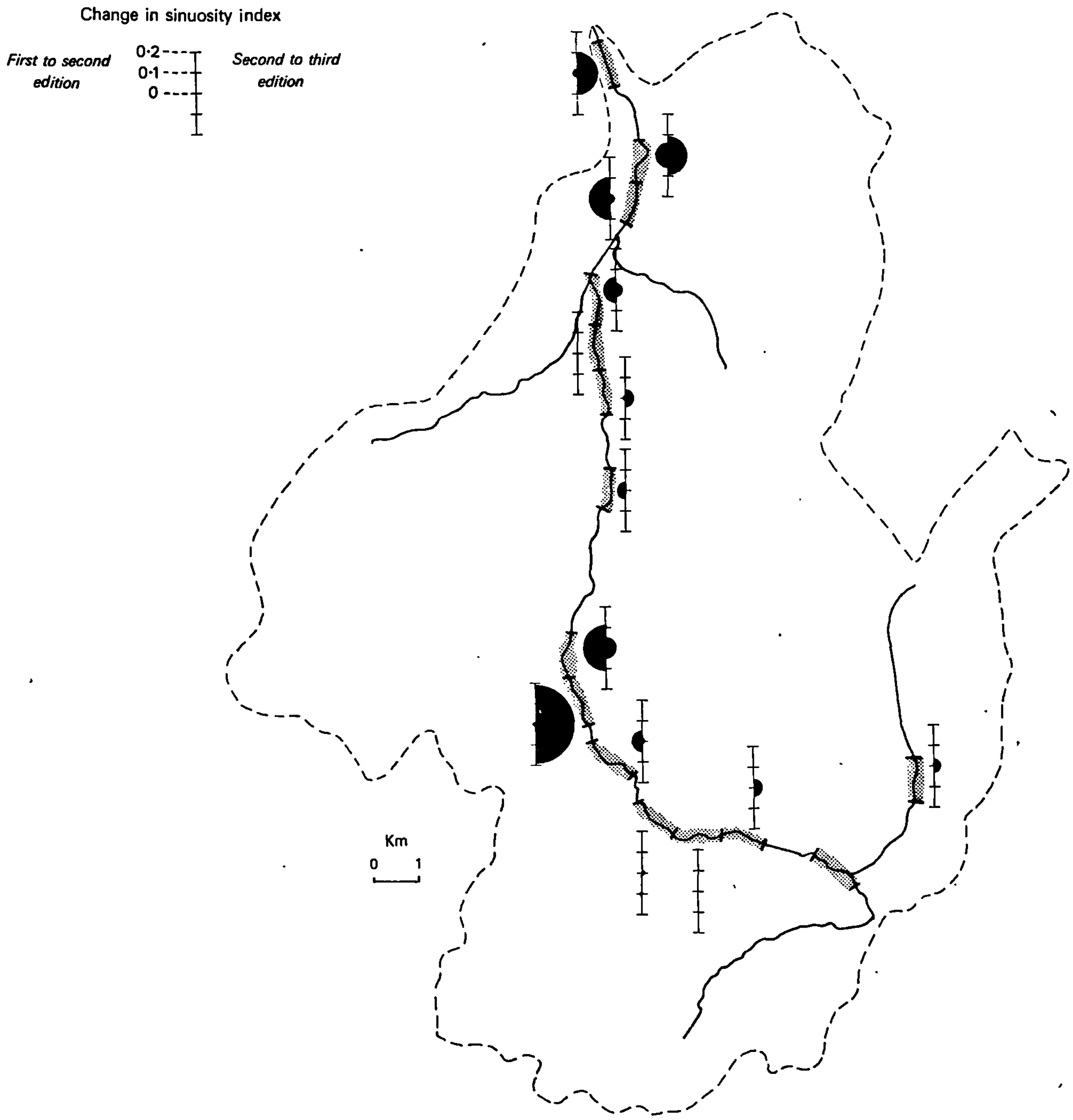


Figure 4.7.3.(x)

Sampled changes in sinuosity index value falling within the Druie catchment during the two intermap periods

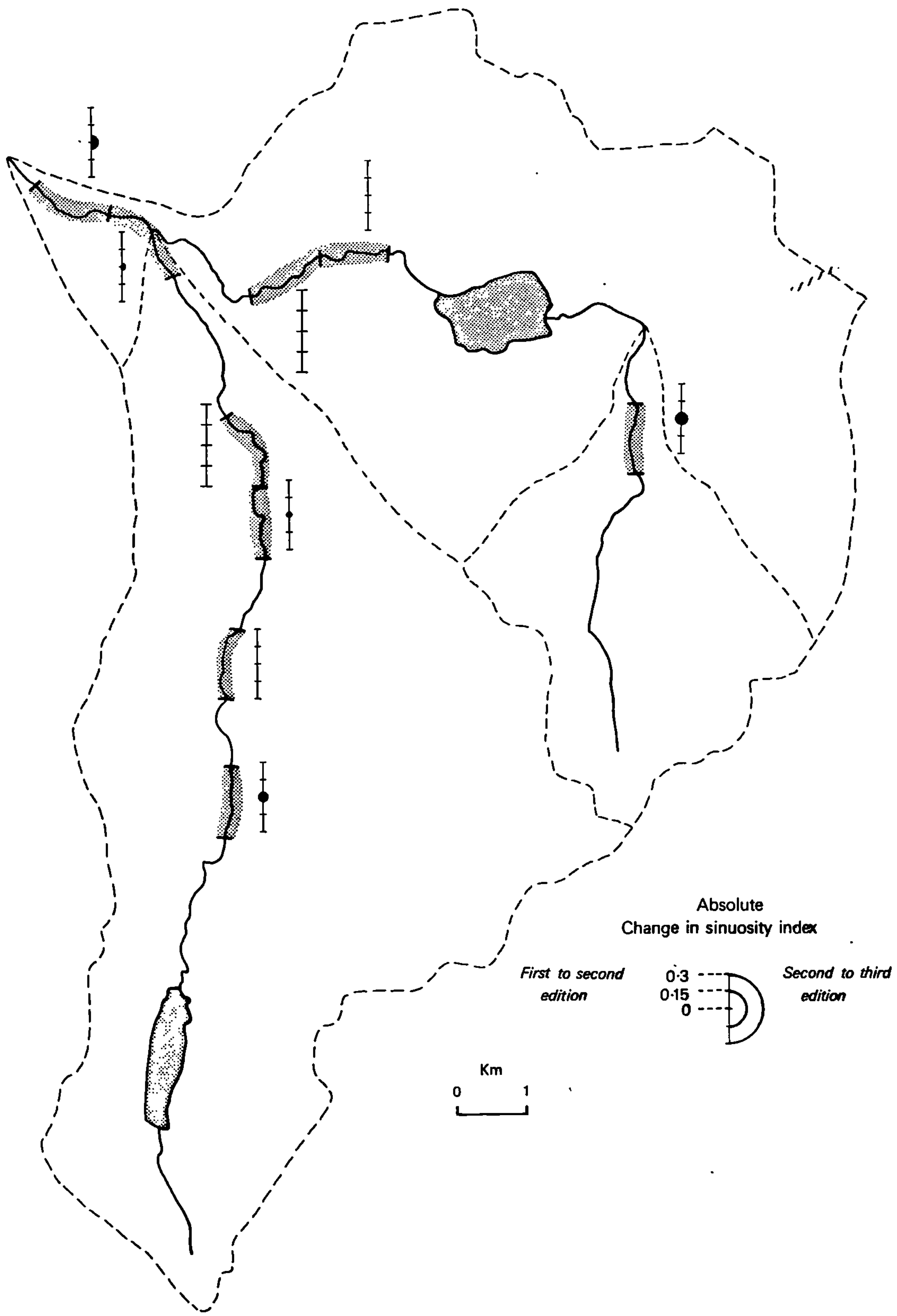


Figure 4.7.3.(xi)

Sampled changes in sinuosity index falling within the Nethy catchment during the two intermap periods

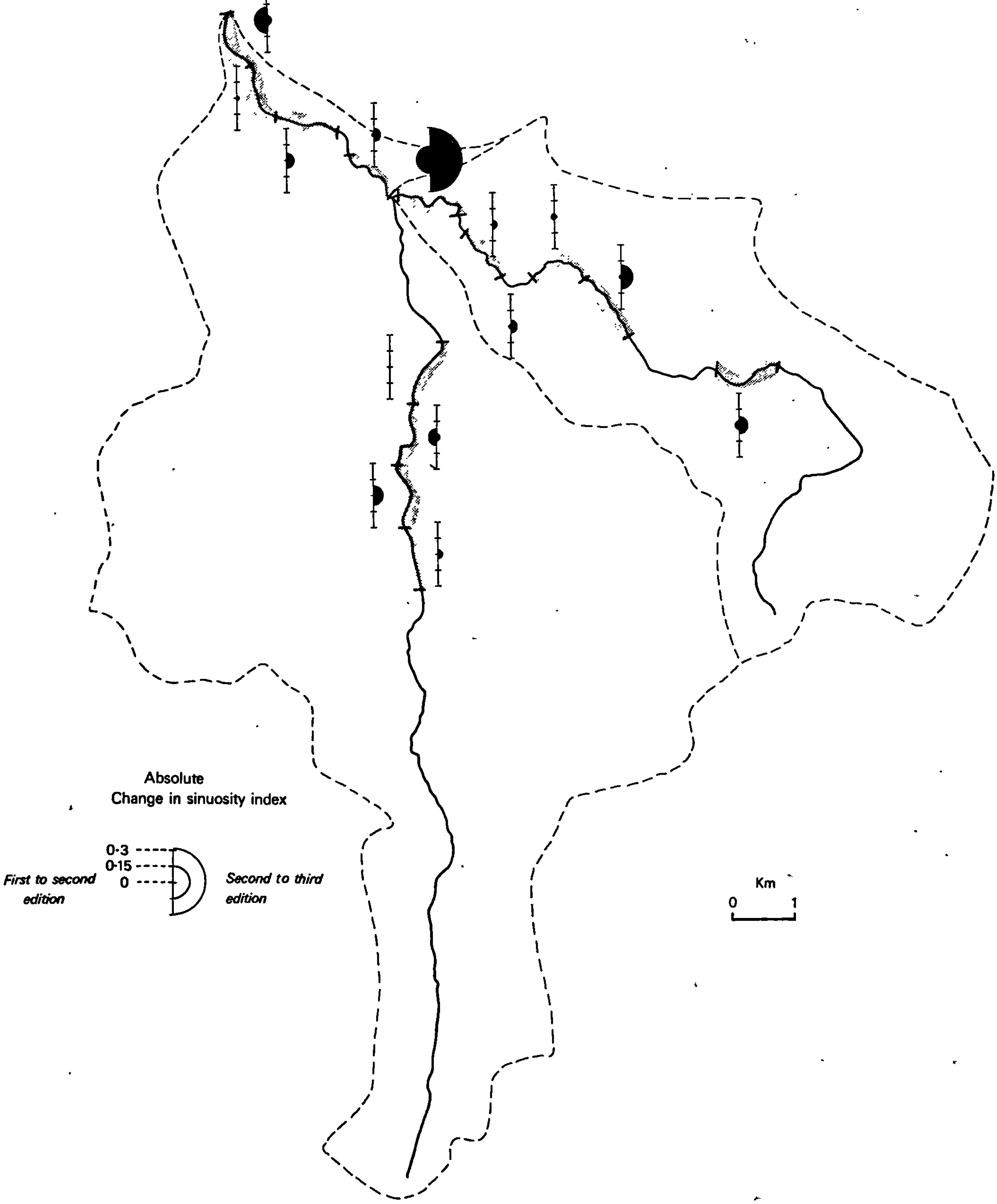


Figure 4.7.3.(xii)

Sampled changes in sinuosity index falling within the Avon catchment for the three map dates

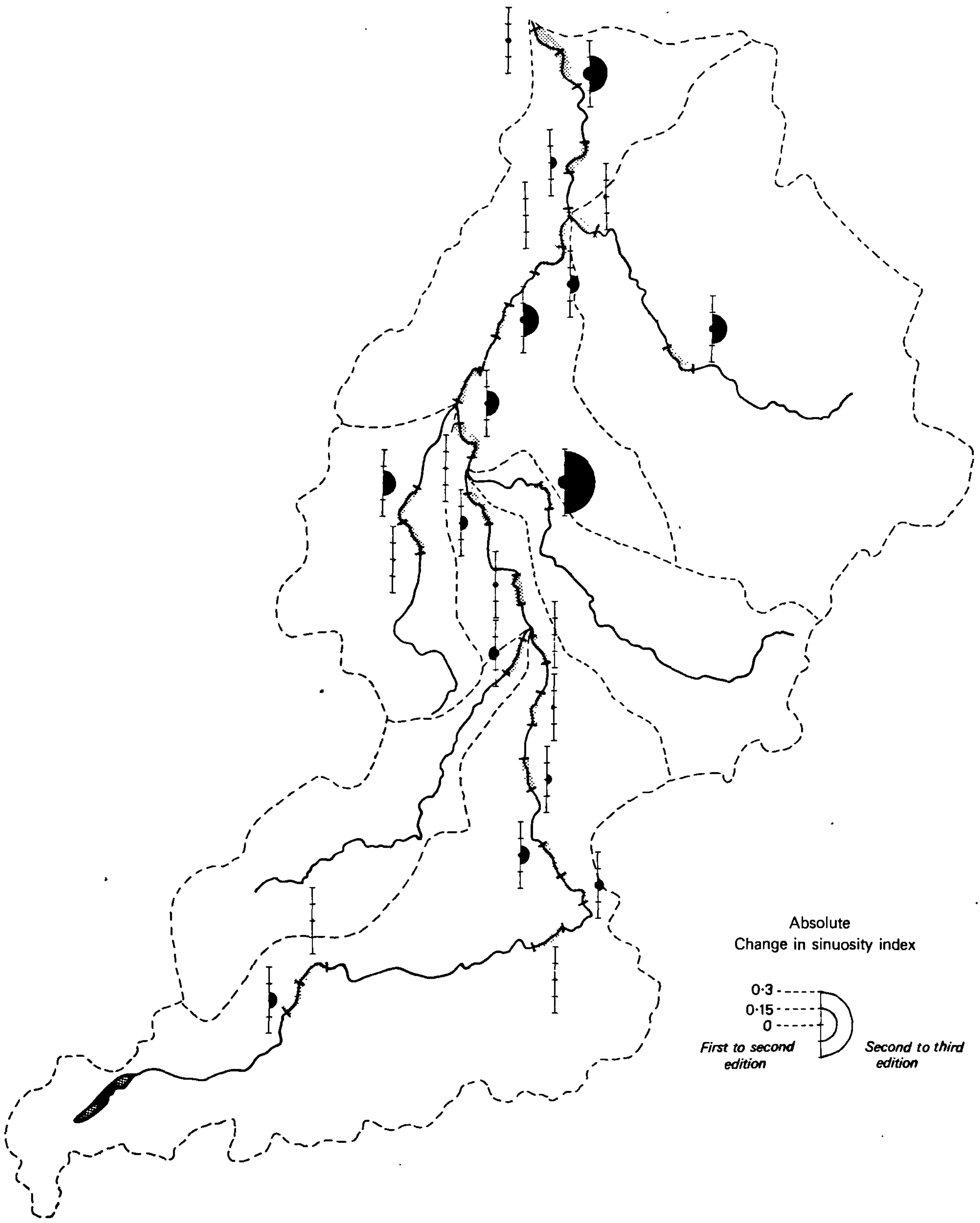
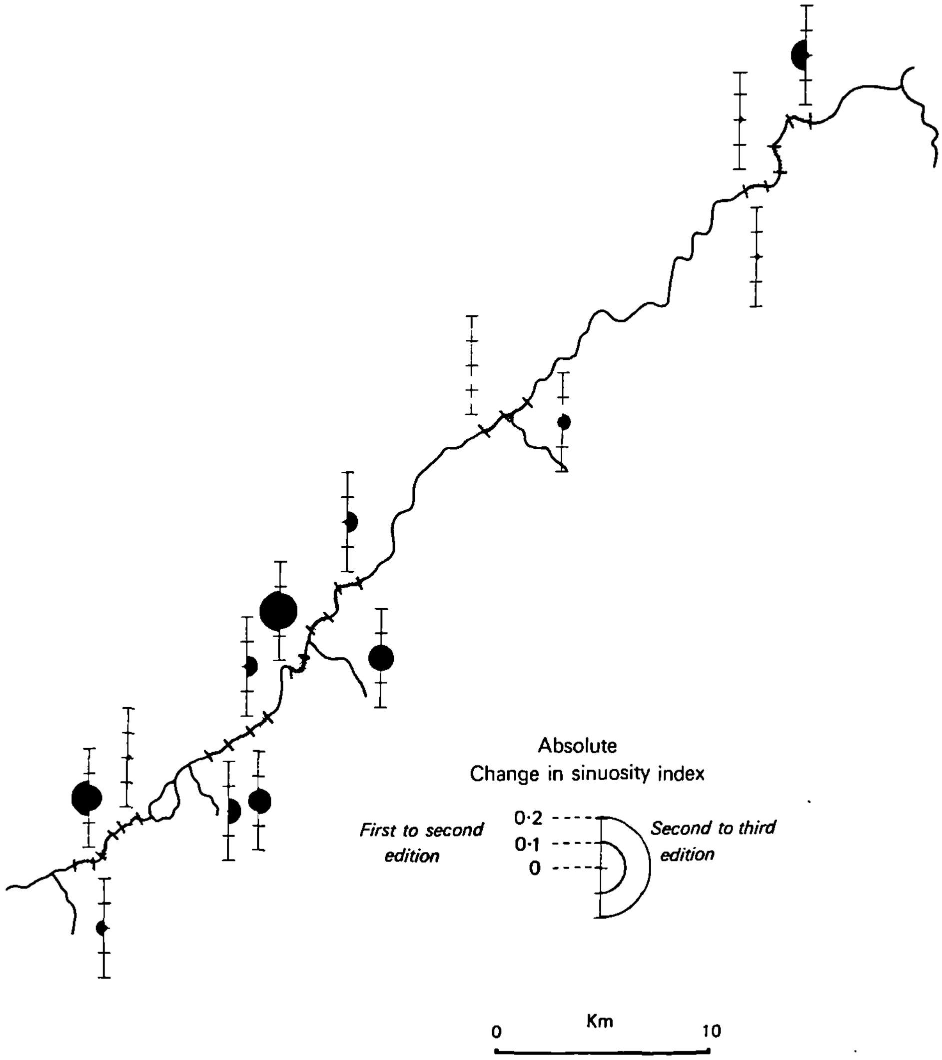


Figure 4.7.3.(xxiv)

Sampled changes in sinuosity index falling on the mainstream Spey during the two intermap periods



sinuosity (Figure 4.7.3.(ii)). This could be interpreted as a relaxation period after a major event but that may be over-simplifying the catchment's flood history. Other catchments were more chaotic in their patterns (Figures 4.7.3.(iii) to (vi)). For example, on the Avon, three sites had a major increase although the general pattern between the second and third edition was a reduction in sinuosity (see Figure 4.7.3.(vi)). It was important to note that whatever the general trend within a catchment, there were always other sample sites which behaved in a different manner. This suggests that different reaches may have different responses and rates of recovery to discharges of similar magnitude and recurrence interval.

But how important is initial sinuosity when absolute rates of change are studied? When ABAVSIN2-3 (ie. absolute annual change between second and third edition) was correlated with the initial sinuosity at 1900, a highly significant correlation coefficient of 0.41 was obtained (Table 4.7.3.(iii)). A similarly high value of 0.38 was found when ABAVSIN1-2 was correlated with initial sinuosity at 1850. This implied that it was the more sinuous channels that were more likely to alter their sinuosity, either positively or negatively. This is not surprising, in some respects one would have expected the correlation to be higher. However, the average initial sinuosities were low and thus, there were few examples of progressively developing meander loops.

Table 4.7.3.(iii)

The correlation (r) between subsequent sinuosity index change and initial sinuosity index

	SIN 1	SIN 2
ABAVSIN 1-2	0.38	-----
ABAVSIN 2-3	-----	0.41

The correlation (r) between subsequent braiding index change and initial braiding index

	BI 1	BI 2
Total sample		
ABBRAID1-2	0.17	------
ABBRAID2-3	------	0.58
Nethy		
AVBRAID1-2	0.53	-------
AVBRAID2-3	-----	0.84
Avon		
AVBRAID1-2	0.45	------
AVBRAID2-3	-----	0.76

4.7.3.3 Rates of change as indicated by braiding index (BRAID)

For the first edition (BI1) and second edition (BI2), the median and IQR values for the total sample were higher (1.08/1.09) than for BI3 (1.03), and thus the overall extent of braiding had changed over the 100 year timespan. Table 4.7.3.(iv) shows rates of change found for the whole sample. This pattern occurred in all catchments, except the Feshie, where there was a reduction from BI1 to BI2 and a large increase in the IQR to 0.87 with BI3 (IQR for AVBRAID2-3 was 0.0089; see Tables 4.7.3.(ii) and (iv)). Differences in the inter-spacing of tributary flood events must therefore be assessed in Chapter 5.

When the Wilcoxon test was carried out on the total sample of braided index values for pairs of the three map editions, no significant difference was recorded. However, when the Avon catchment was analysed separately, significant differences were found between BI2 and BI3 (Z=-2.20) and BI1 and BI3 (Z=-2.09). This suggested change had occurred between pre-1900 and post-1900 channel planforms and there was a considerable reduction in overall BI, inferring a relaxation of the channel system. In an extreme example, the Tromie underwent reductions in both median and IQR values of BI over both inter-map periods. Similarly, although the Nethy increased in median value between map editions 1 and 2, it then underwent a reduction between map editions 2 and 3 (Table 4.7.3.(ii)). However, this pattern cannot be generalised for all the tributary catchments. The IQR for BI in the Feshie catchment had increased from 0.50 to 0.87 between the first and third editions. The IQR for annual rates of change increased from 0.0055 (1-2) to 0.0089 (2-3). This suggested that channels had become

Table 4.7.3.(iv)

Rates of change in braiding index within the Spey sample

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
BRAID1-2	0.05	0.00	0.01	0.06
BRAID2-3	-0.12	0.00	0.06	0.18
AVBRAID1-2	-0.0015	0.00	0.0010	0.0029
AVBRAID2-3	-0.0019	0.00	0.0003	0.0018
Tromie				
BRAID1-2	-0.15	-0.02	0.05	0.20
BRAID2-3	-0.20	-0.08	0.01	0.21
AVBRAID2-3	-0.003	-0.011	0.0001	0.0031
AVBRAID1-2	-0.0045	-0.006	0.0014	0.0059
Feshie				
BRAID1-2	-0.01	0.00	0.15	0.16
BRAID2-3	-0.12	0.00	0.50	0.62
AVBRAID1-2	-0.0003	0.00	0.0052	0.0055
AVBRAID2-3	-0.0018	0.00	0.0071	0.0089
Druie				
BRAID1-2	-0.083	-0.04	-0.005	0.088
BRAID2-3	-0.03	0.085	0.245	0.275
AVBRAID1-2	-0.0028	-0.0014	-0.0002	0.0030
AVBRAID2-3	-0.0004	0.0012	0.0035	0.0039

Table 4.7.3.(iv) cont.

Nethy				
BRAID1-2	0.00	0.00	0.06	0.06
BRAID2-3	-0.23	0.00	0.09	0.32
AVBRAID1-2	0.00	0.00	0.0017	0.0017
AVBRAID2-3	-0.0033	0.00	0.0013	0.0046
Avon				
BRAID1-2	-0.06	0.00	0.08	0.13
BRAID2-3	-0.19	0.00	0.00	0.19
AVBRAID1-2	-0.002	0.00	0.002	0.004
AVBRAID2-3	-0.003	0.00	0.00	0.003
Spey				
BRAID1-2	-0.02	0.00	0.00	0.02
BRAID2-3	0.00	0.00	0.25	0.25
AVBRAID1-2	-0.0006	0.00	0.00	0.0006
AVBRAID2-3	0.00	0.00	0.0034	0.0034

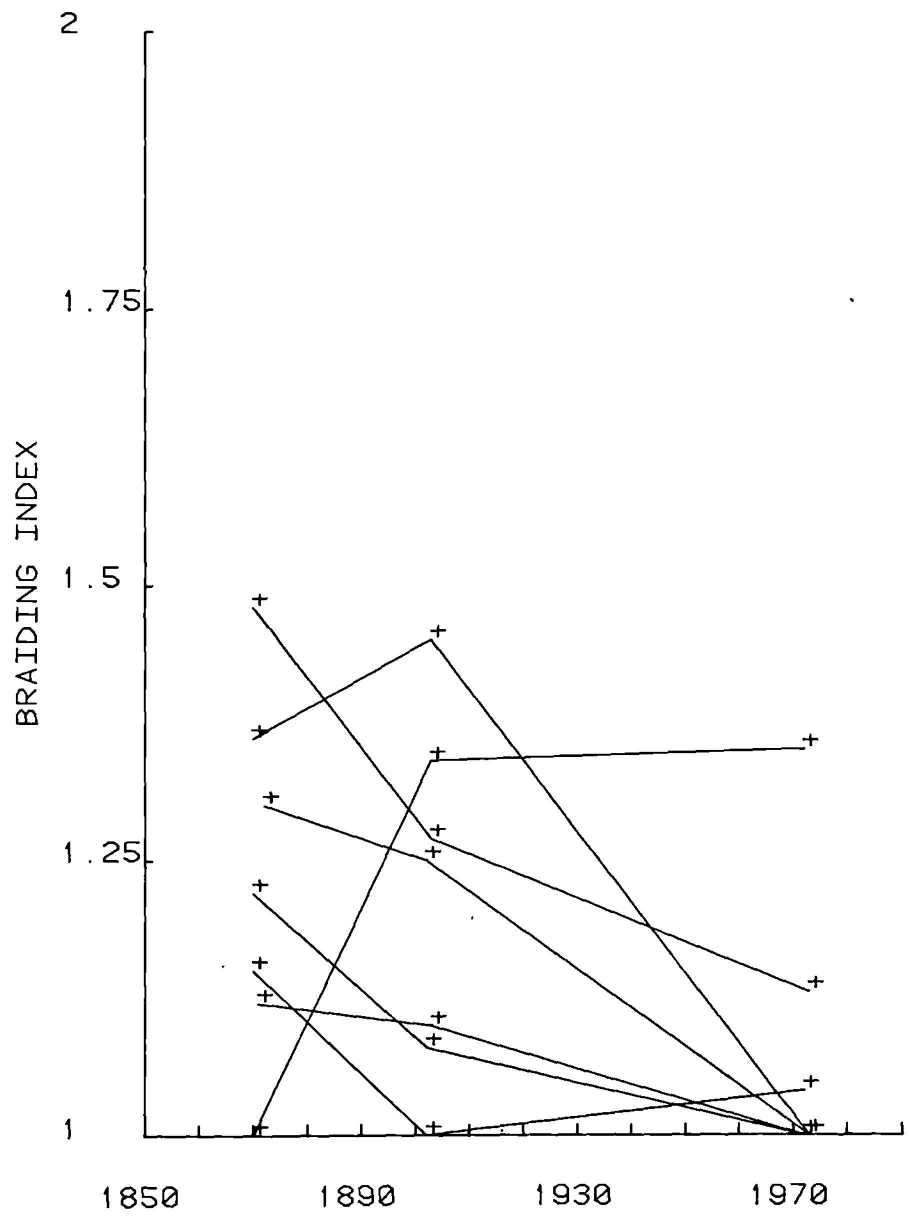
increasingly more disrupted and were not getting a chance to return to a more stable equilibrium form. Neighbouring catchments therefore had very different trends.

As well as considerable spatial variation in BI between catchments, there was also variation within catchments (Figures 4.7.3.(xx) to (xxiv)). The braiding index values plotted against year of occurrence are shown in Figures 4.7.3.(xiv) to (xix). For example within the Nethy catchment (Figure 4.7.3.(xxiii)), the Nethy confluence site (Sample 34) increased from 1.00 to 2.23 between 1871 and 1900 and the reduced again to 1.55 by 1975 (see Figure 4.7.3.(xxiii) and Section 7.3.2). In contrast on Dorback Burn, while there was very little change between map editions 1 and 2 (1869-1900), there were increases on three samples between map editions 2 and 3 (1900-1972). This was perhaps related to localised convective events affecting only part of the catchment or differing thresholds for change within different parts of the catchment, where planform controls are different.

Although there was no statistically significant difference between sampled BI values at successive dates within the total sample, nevertheless at selected sites on individual tributaries geomorphically significant change occurred. For example, although the Feshie did not statistically undergo significant change, several reaches underwent considerable planform alteration, the highest being the increase between from 1.18 to 2.80 between 1873 and 1902 on the upper Feshie (Sample 18; Figure 4.7.3.(xxi)). Clearly a major disruptive stress had taken place to the system and a major threshold crossed, causing channel metamorphosis. Consequently, between 1902-1970, there was a reduction

Figure 4.7.3.(xiv)

TROMIE CATCHMENT: BRAIDING INDEX CHANGE



MAP-DATE

Figure 4.7.3.(xv)

FESHIE CATCHMENT: BRAIDING INDEX CHANGE

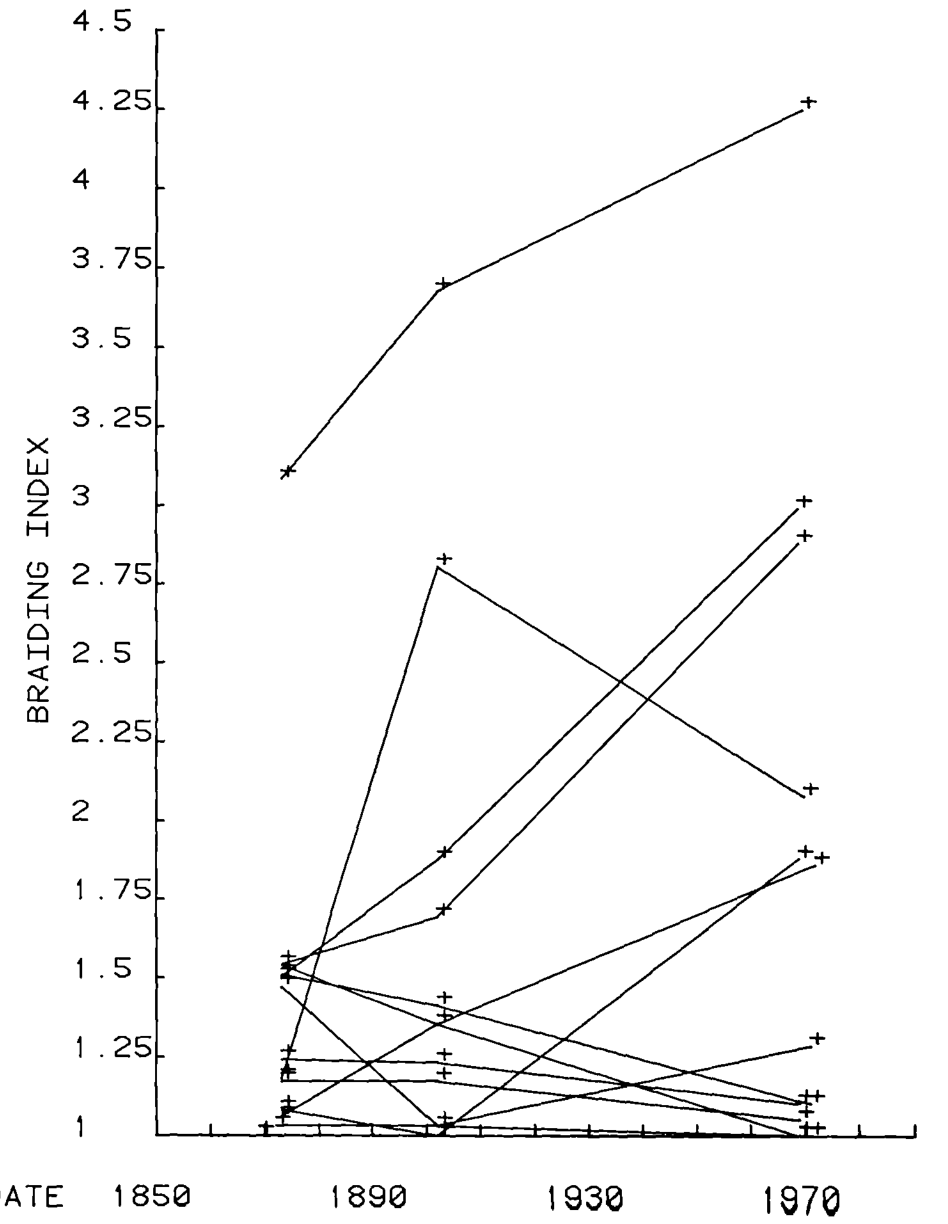


Figure 4.7.3.(xvi)

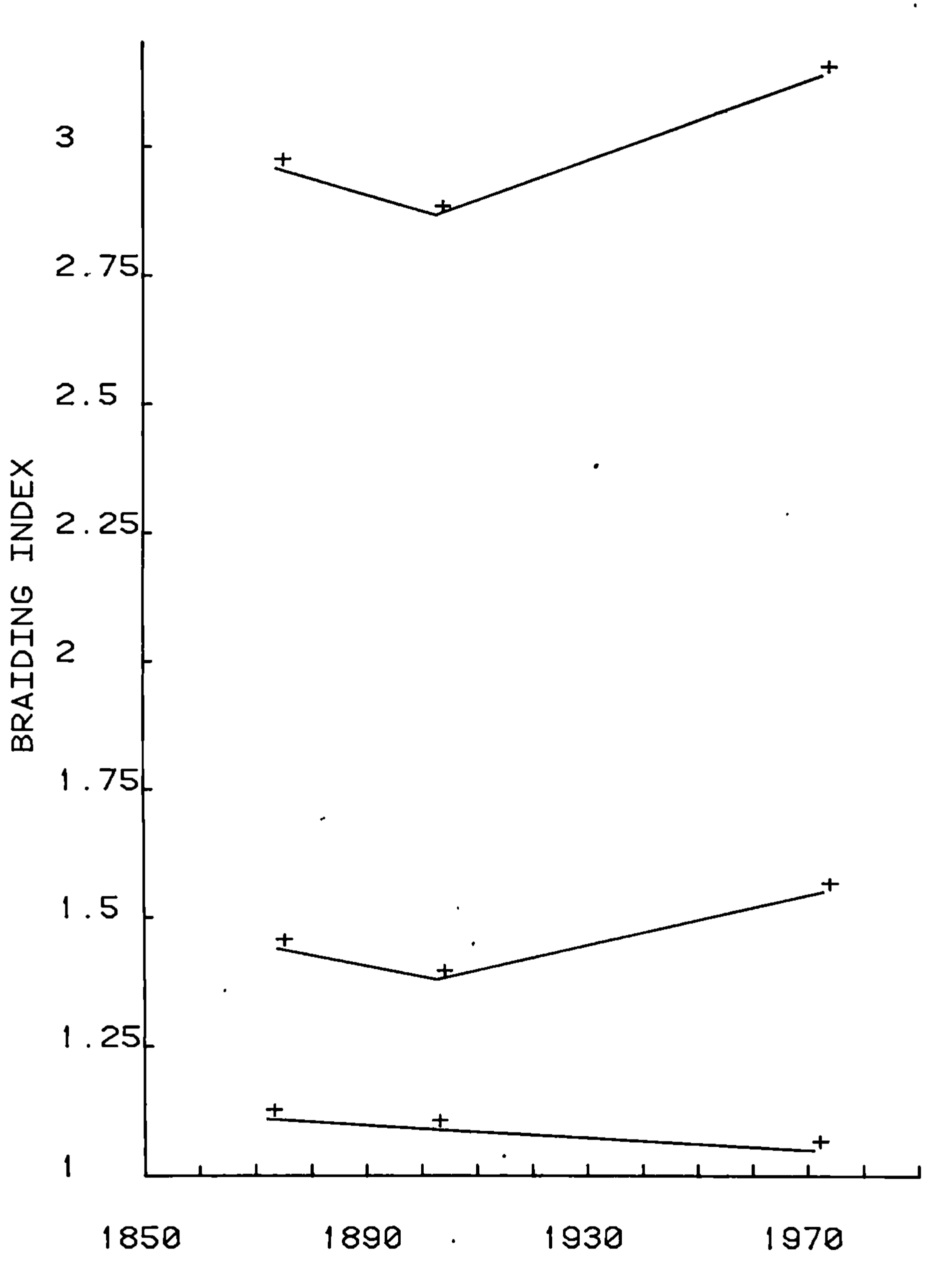
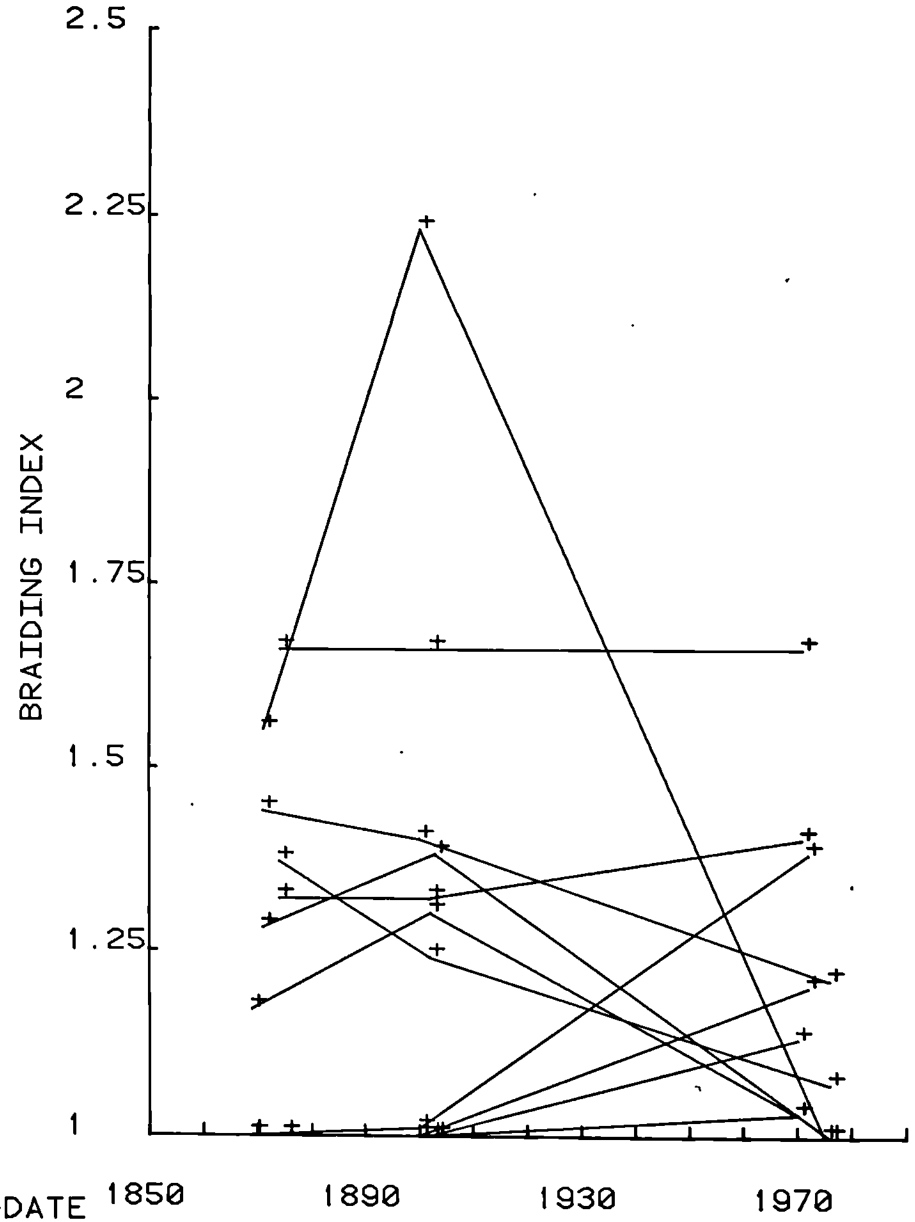


Figure 4.7.3.(xvii)



MAP-DATE

Figure 4.7.3.(xviii)

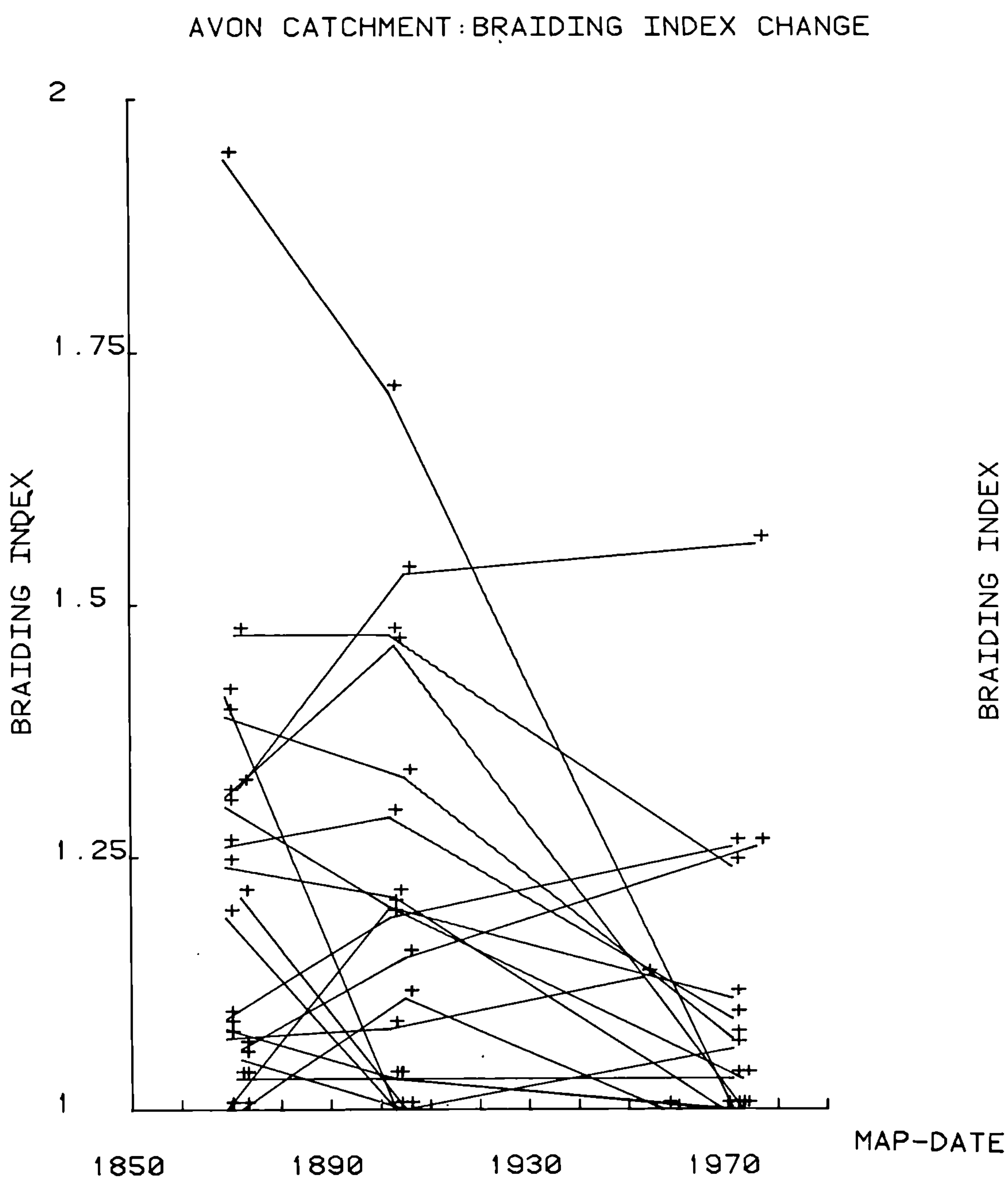


Figure 4.7.3.(xix)

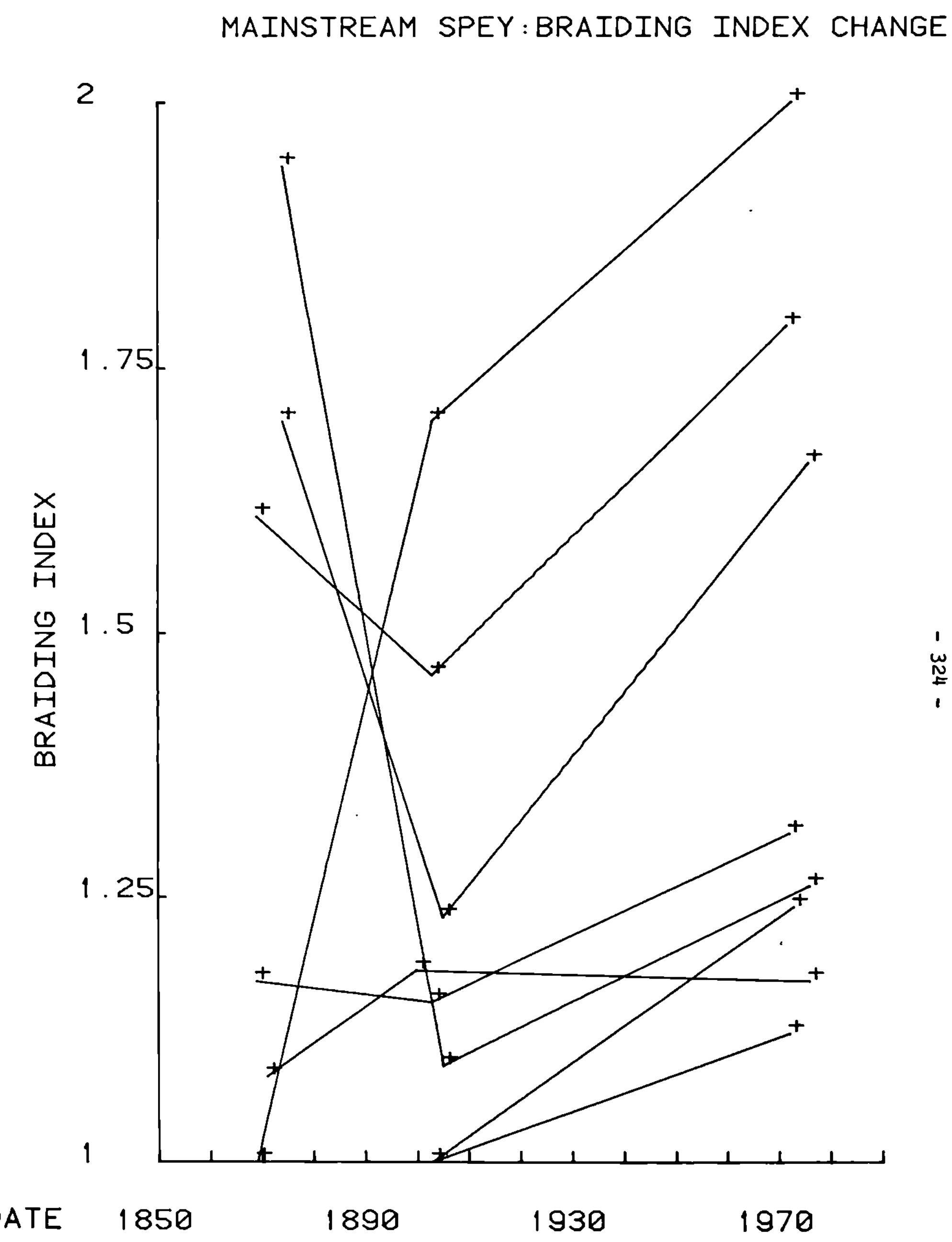


Figure 4.7.3.(xx)

Sampled changes in braiding index falling within the Tromie catchment during the two inter map periods

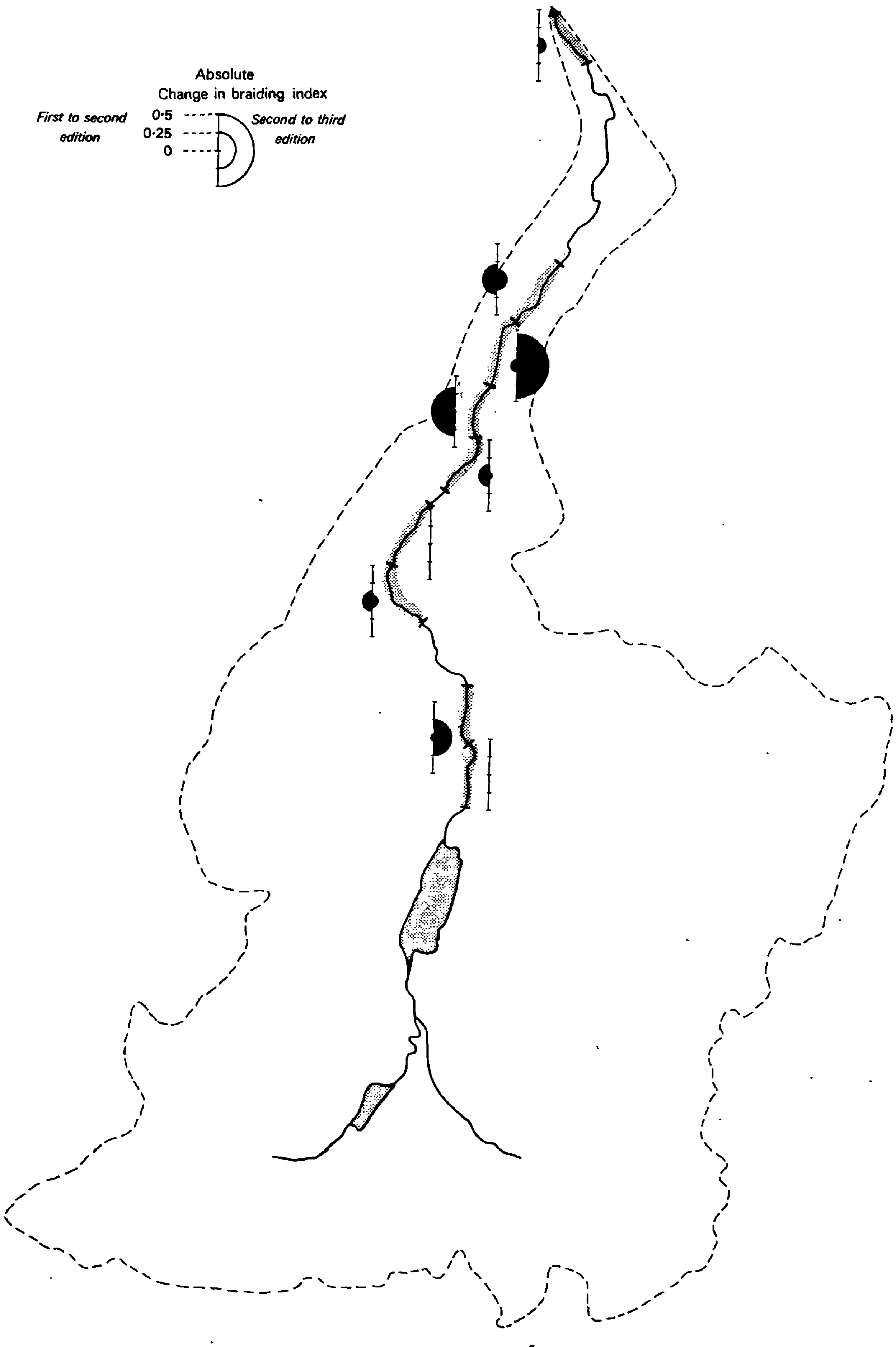


Figure 4.7.3.(xxi)

Sampled changes in braiding index values falling within the Feshie catchment during the two intermap periods

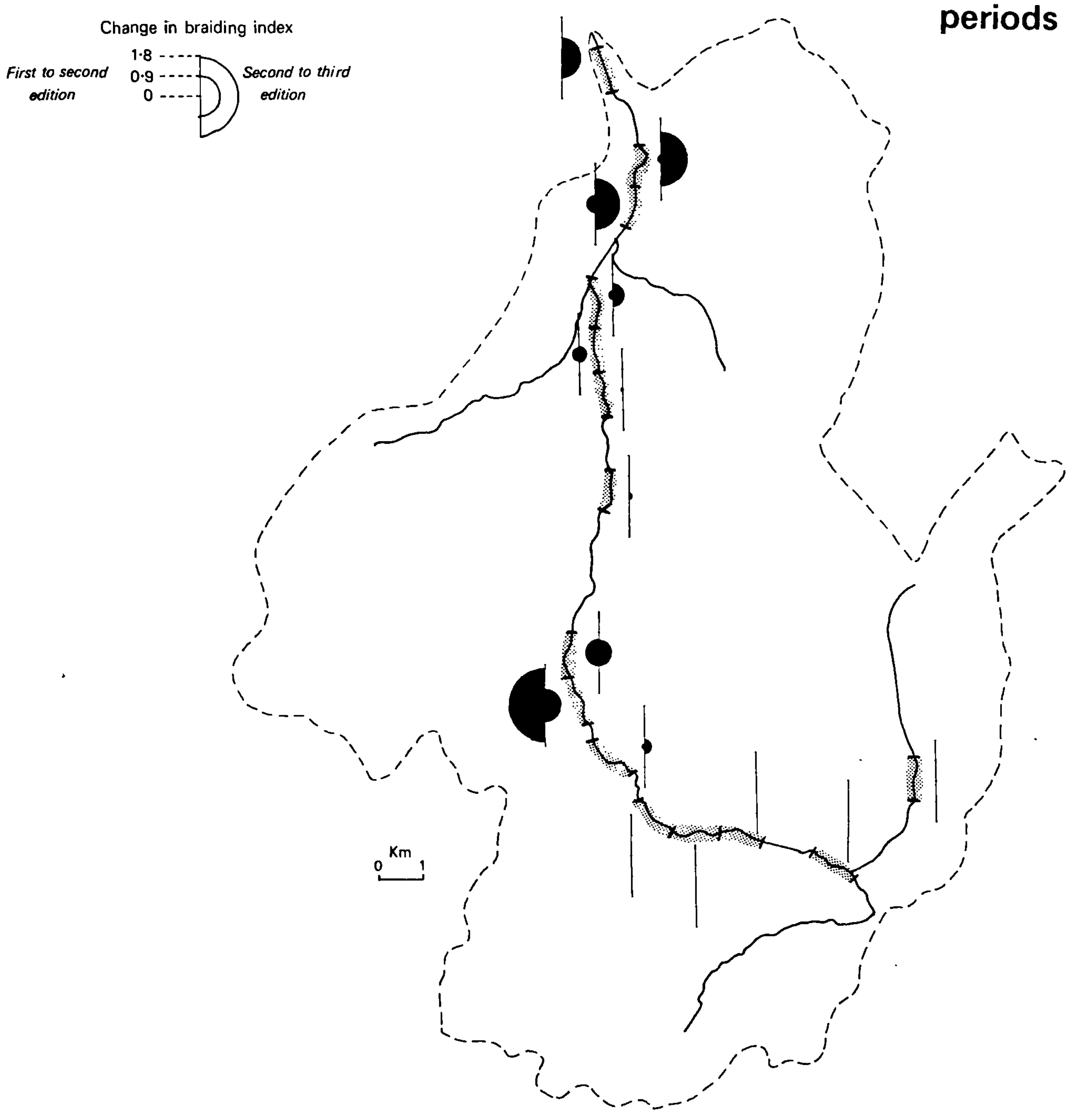


Figure 4.7.3.(xxii)

Sampled changes in braiding index falling within the Druie catchment during the two intermap periods

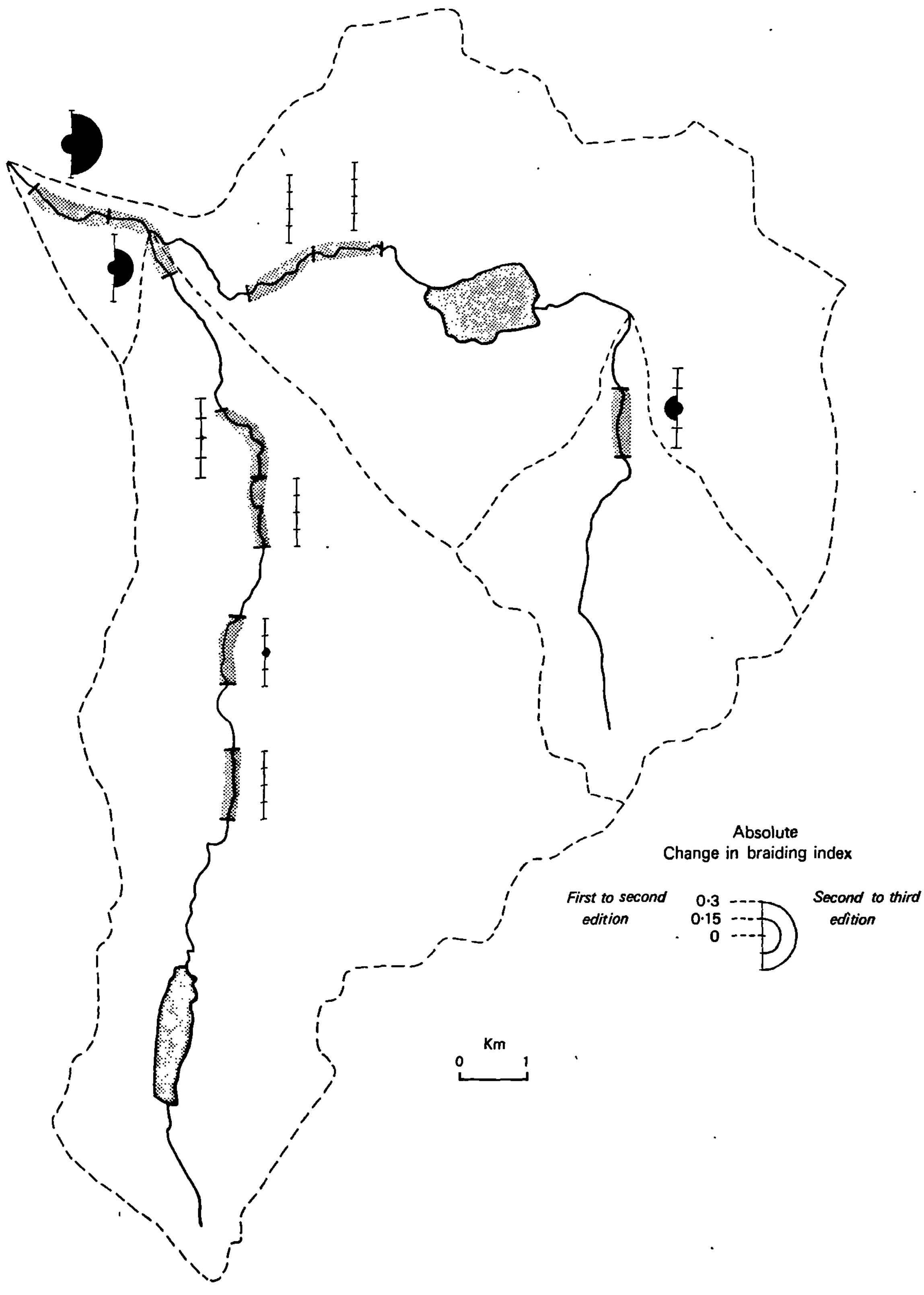


Figure 4.7.3.(xxiii)

Sampled changes in braiding index falling within the Nethy catchment during the two intermap periods

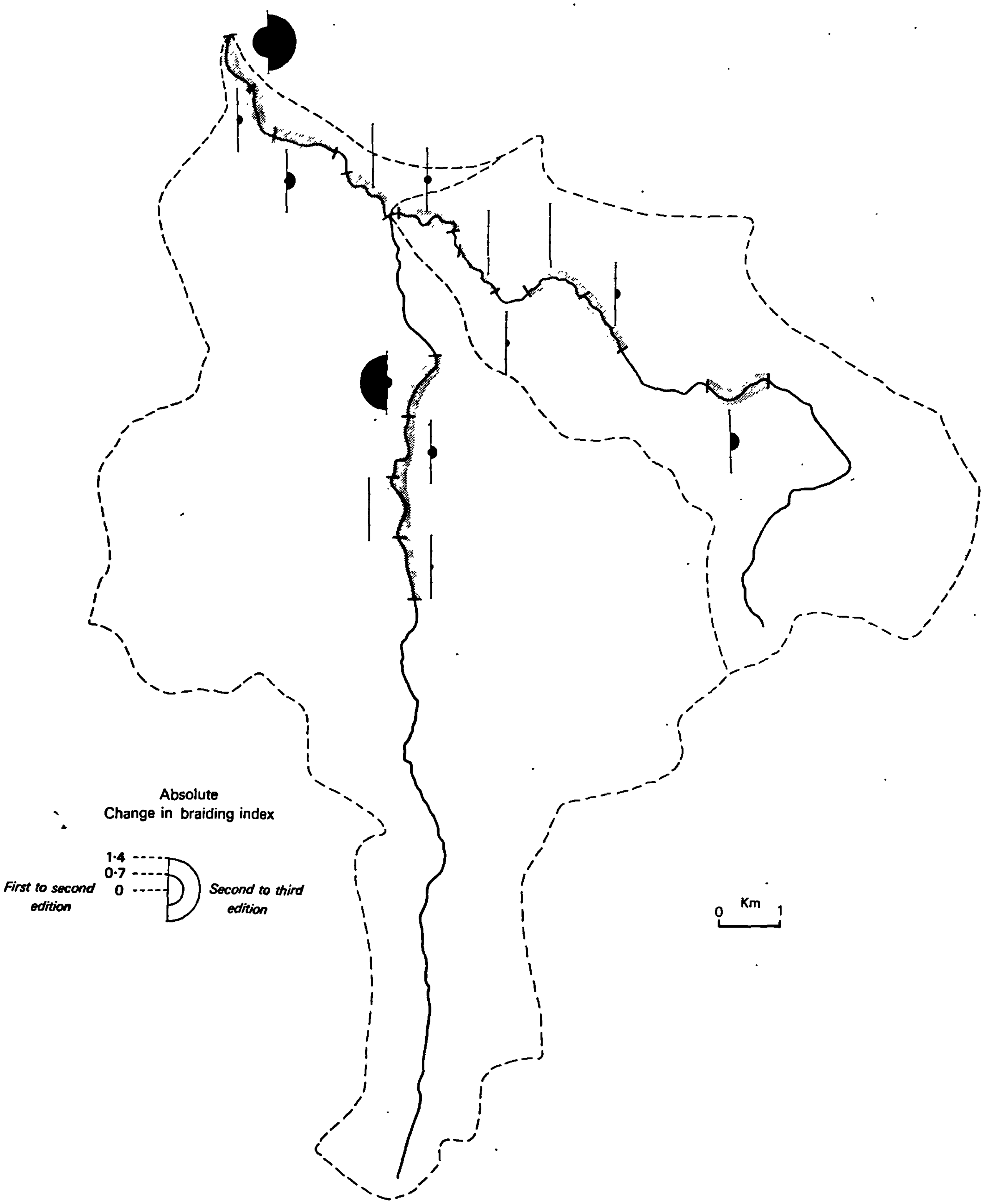


Figure 4.7.3.(xxiv)

Sampled changes in braiding index falling within the Avon catchment during the two intermap periods

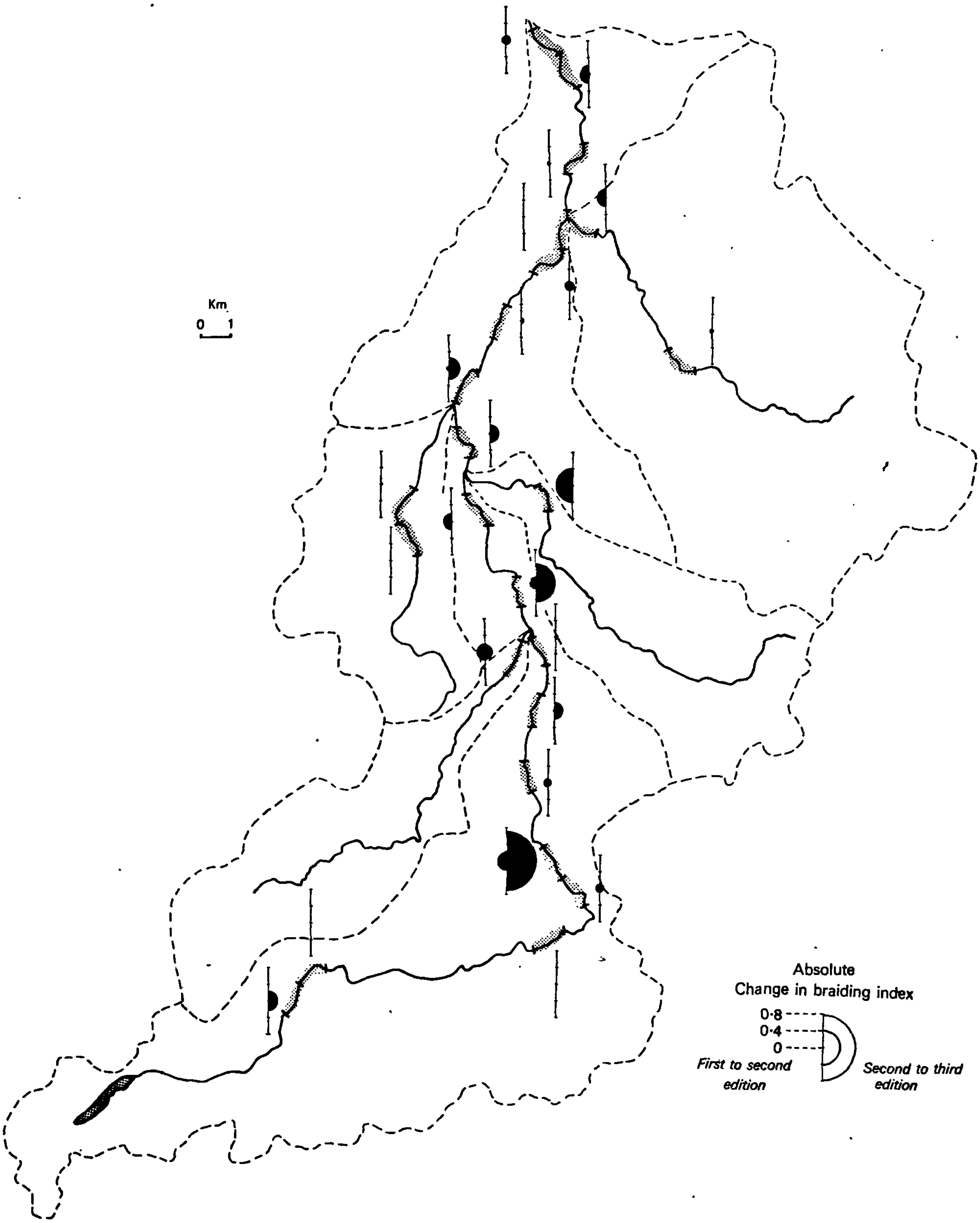
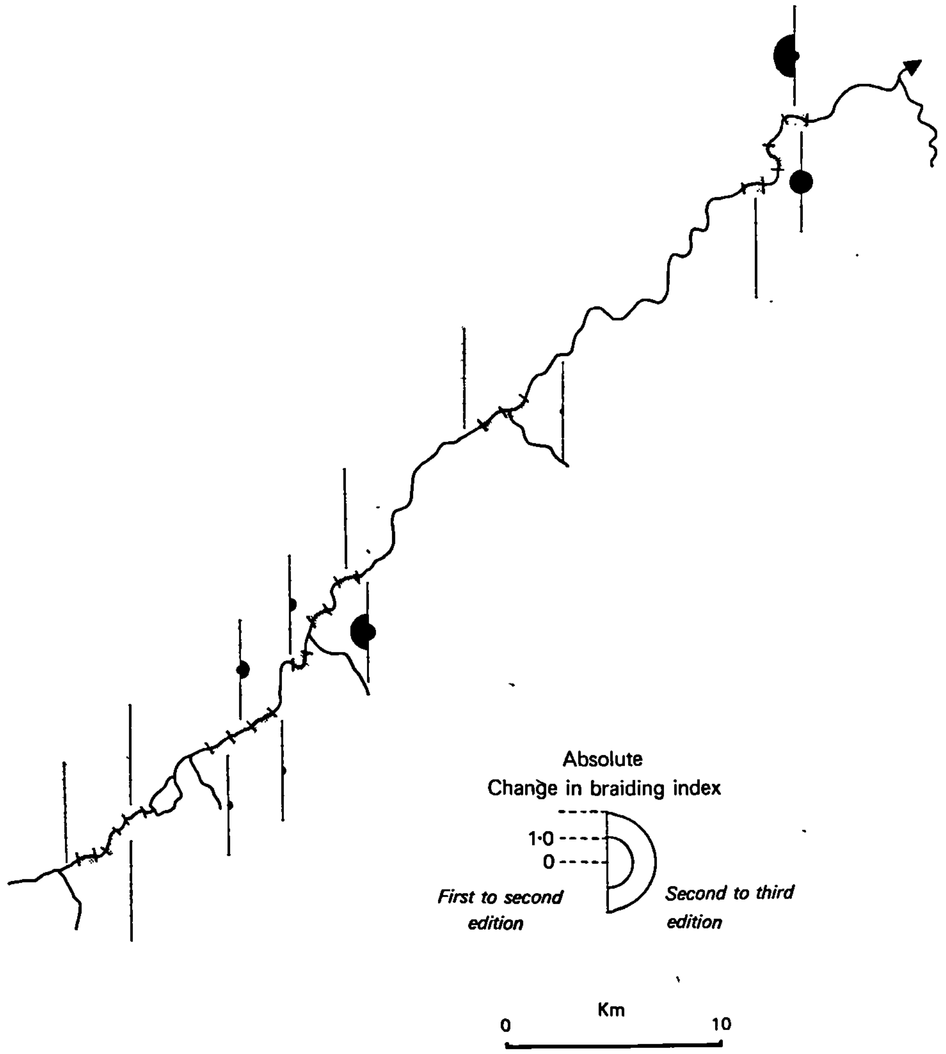


Figure 4.7.3.(xxv)

Sampled change in braiding index values falling on the mainstream Spey during the two intermap periods



in BI of -0.73, suggesting a degree of recovery to the system. The other site of major activity was the lower Feshie braided reach at Lagganlia where although there was a lesser increase in BI between 1873-1902, there was a considerable increase between 1902 and 1969. Thus, total increase in BI was from 1.50-2.98 and 1.54-2.87 on adjacent reaches from 1873 to 1969 (Samples 11 and 12), again suggesting an expanding channel planform. On the other tributaries as well, though change may not have been as great, sequences of interesting geomorphic change occurred. These are studied in detail in Chapter 7.3. However, of the reaches with bars, most had an increase in BI over the study period (1870-1965).

On the mainstream Spey, there were several reaches that remained without islands at any map-date. In contrast, the Spey over localised reaches gave the appearance of gradual sedimentation and stabilisation, especially on the sample sites downstream of the Feshie confluence (eg. Sample 79 which increased from 1.00 to 1.70 to 2.00 over 1869-1903-1973; Figure 4.7.3.(xxv)). There were exceptions however, such as the reach near Kinrara where there was a reduction in braiding index as bars had stabilised by joining on to the banks (Sample 76), causing an overall reduction in channel width and a reduction in BI.

When the importance of the initial braiding index was assessed in relation to likelihood of change, ABBRAID2-3 was highly significantly correlated with BI2 (r=0.58; see Table 4.7.3.(iii)). When individual catchments were assessed eg. the Nethy, this value (r) could be as high as 0.84. Major events pre-1900 may have activated new areas of change, whereas post 1900, many of these areas had merely been reworked (see

Chapter 5). This correlation was however reduced to 0.17 when ABBRAID1-2 was correlated with BI1, but certain catchments such as the Avon still had a significant correlations (r=0.45). Therefore change within the total sample between BI1 and BI2 was not necessarily at sites of initially high BI values ie. through new avulsions outwith the active areas, as indicated on the first edition.

4.7.3.4 Spatial variations in rates of lateral shift

As in Deeside, there were difficulties in obtaining fixed points on maps of upland areas, for the comparison of sequential lateral shift. These problems were overcome and at sites where measurements could be made, highly variable rates of lateral shift were recorded (see Table 4.7.3.(v)). The highest median values were found in the middle reaches of the River Feshie, which reworked its floodplain with great rapidity eg. Sample 18 shifted its channel by a maximum of 100 m and an average of 55 m (or an annual maximum rate of 3 m yr^{-1}) between the first and second editions. In comparison to such change, the shifts between the second and third editions, across the Feshie confluence (total maximum shift of 89 m; total average shift of 9 m) and the Nethy fan (total maximum shift 94 m; total average shift 46 m) were still large by Dee and Tweed standards. Typically, samples which had a large change between the first and second map editions also had large changes between the second and third editions.

Table 4.7.3.(v)

Rates of lateral shift within the Spey study area (m)

Parameter	25th Percentile	Median	75th Percentile	IQR
LSMAX1-2	18	29	71	53
LSMAX2-3	7	20	50	43
AVLSMAX1-2	0.5	0.9	2.2	1.7
AVLSMAX2-3	0.1	0.3	0.8	0.7
LSAV1-2	6	12	23	17
LSAV2-3	3	8	13	10
AVLSAV1-2	0.2	0.4	0.8	0.6
AVLSAV2-3	0.0	0.1	0.2	0.2

These high rates however only occurred at specific high activity sites. The Tromie catchment in comparison underwent only slow rates of shift (0.1 m yr^{-1}). Thus, rates of lateral shift were highly dependent on the associated forms of channel adjustment ie. avulsion as opposed to migration. A major avulsion can clearly cause a large lateral shift over a short time period. Many reaches however underwent no recordable change over the 100 year timespan and these were either confined by bedrock controls or terraces, or else had attained a stable equilibrium condition with a high threshold for change.

4.7.4.1 Importance of quantitative parameters

When attempts were made to assess whether sinuosity index change had any relationship with contributing basin area, average height or MAXWID, correlations were generally low and scatter high. In contrast to this general trend, both AVBRAID1-2 and 2-3 were significantly correlated with drainage basin area. There was however an important distinction with AVBRAID1-2 having a positive correlation ($r=0.18$) and AVBRAID 2-3 having a negative value ($r=-0.19$), though area clearly cannot be used as a powerful predictor.

Height in itself was not found to be an important parameter, however the highest samples over 400 m were generally characterised by low rates of change in sinuosity. Certain catchments indicated stronger trends. For example, BRAID2-3 on the Nethy was highly significantly correlated with height ($r=-0.70$) ie. lower heights were associated with high positive change. Height has implications in terms of the glacial history of a basin, with the drift cover becoming thinner with altitude. Thus, higher reaches will be more likely to be near bedrock contact eg. on the upper Avon.

The most important quantitative parameter was thus $\log_{10}$ MAXWID (ie. degree of confinement), which when correlated with ABBRAID2-3 yielded a highly significant correlation (r=0.39). The available floodplain width within which the channel could split, braid or anabranch was thus an important control, through change in the intensity of glacial erosion. ABBRAID1-2 had a lower but still significant correlation. Thus, position within the drainage basin was less important than available width of active area at the sample reach.

4.7.4.2 Importance of qualitative parameters

Consistency of the qualitative variables in the typology over the three map-dates was then assessed. When these parameters were crosstabulated to see if there was any noticeable difference in the distribution from the main diagonal at different map-dates, the following results were obtained. Activity rates were then "broken down" in terms of the various categories for each typology variable but only statistically significant results are reported here (see Table 4.7.4.(i)).

Table 4.7.4.(i)

Selected results of the "breakdown" analysis within the Spey study area

(a) ABAVSIN1-2 broken down by UPSEDMT 1

UPSEDMT 1	Mean	SD	
0	0.0004	0.0006	
1	0.0008	0.0008	Sig = 0.001
2	0.0016	0.0017	
3	0.0017	0.0013	

(b) ABAVSIN2-3 broken down by UPSEDMT 2

UPSEDMT 2	Mean	SD	
0	0.0003	0.0003	
1	0.0008	0.0008	Sig = 0.001
2	0.0014	0.0015	
3	0.0007	0.0004	

Table 4.7.4.(i) cont.

(c) ABBRAID1-2 broken down by LOCSEDMT1

LOCSEDMT 1	Mean	SD	
0	0.0009	0.0018	
1	0.0051	0.0079	
2	0.0108	0.0167	Sig = 0.01
3	0.0055	0.0070	

(d) ABBRAID2-3 broken down by LOCSEDMT2

LOCSEDMT 2	Mean	SD	
0	0.0003	0.0006	
1	0.0025	0.0025	Sig = 0.001
2	0.0036	0.0044	Sig = 0.008
3	0.0088	0.0056	

Table 4.7.4.(i) cont.

(e) ABBRAID2-3 broken down by CHANPATT2

CHANPATT2	Mean	SD	
2	0.0021	0.0032	Sig = 0.008
3	0.0022	0.0036	
4	0.0059	0.0056	
5	0.0016	0.0018	
6	0.0008	0.0010	
7	0.0020	0.0018	

There was little shift between categories over the period 1850-1900, ie. floodplain and bank vegetation remained relatively constant over time. Although the availability of upstream and local sediment did not show significant changes, there was some shifting at certain sample sites, with a slight increase in categories 2 and 3 (large and extensive amounts of sediment, respectively) between the first and second edition and considerable reduction between the second and third in the same categories.

When changes within channel description were considered, there were more "wandering" channel patterns and a reduction in "irregular" to "irregular meandering" between the first and second editions. By 1900, it appeared that there had been a switch back from "wandering" to these more sinuous categories. Thus, although there may be change at individual sites, the qualitative controls were not undergoing major changes in terms of the whole sample.

The association between the qualitative variables and rates of channel activity was then assessed, using the "breakdown" program and one way analysis of variance. When ABBRAID2-3 and 1-2 were broken down by availability of upstream sediment at both time intervals, there were significant differences between categories. The relationship was most pronounced with ABBRAID2-3 and highest mean values were found where there was an extensive supply of upstream sediment (category 3, see Table 4.7.4.(i)). When ABBRAID2-3 was broken down by local availability of sediment at map-date 2, there was again a highly significant difference, with the highest rates of change of BI (0.009 yr^{-1}) found where the most extensive local sediment occurred (category 3). The lowest rates occurred where there was no evidence of upstream channel sediment supply. When ABBRAID1-2 was broken down by local availability of sediment as indicated by the first edition maps, again a highly significant difference was recorded. Here, large amounts of sediment (category 2) were associated with the highest ABBRAID values (0.011 yr^{-1}). This suggests that there may be a threshold beyond which additional sediment availability made little difference.

When both ABAVSIN and ABBRAID were broken down by floodplain and bank vegetation, there was no significant difference between group means at either intermap period. No category stood out as having a particularly stabilising influence and thus vegetation in itself was thus not an important control (cf. Tweed sample). Only ABAVSIN2-3 showed a significant breakdown by bank vegetation categories. Highest rates of change were associated with rough-grassland (6) and agricultural landuse (5).

It was also important to assess the importance of channel pattern category in terms of changes in activity indices. When ABBRAID2-3 was broken down by CHANPATT2 (ie. initial channel pattern at the second edition), there was a highly significant difference between category means. Category 4 (wandering channels), as exhibited on the Feshie and Dorback Burn, yielded the highest mean rates of change (0.006 yr^{-1}; see Table 4.7.4.(i)). However, when ABBRAID1-2 was broken down by CHANPATT1, no particular category of channel pattern was significantly more susceptible to change over that period over 1870-1900 ie. change in BI occurred over a range of channel pattern categories. Change in sinuosity indicated no significant relationship with channel pattern category.

4.8 Tweed study area: Results of the map analysis

The results of the map analysis for the Tweed study area are now presented. Reference maps showing the location of the 85 sampled grid squares within the study area and the location of the tributary catchments, are shown in Figures 4.8.(i) and (ii), respectively.

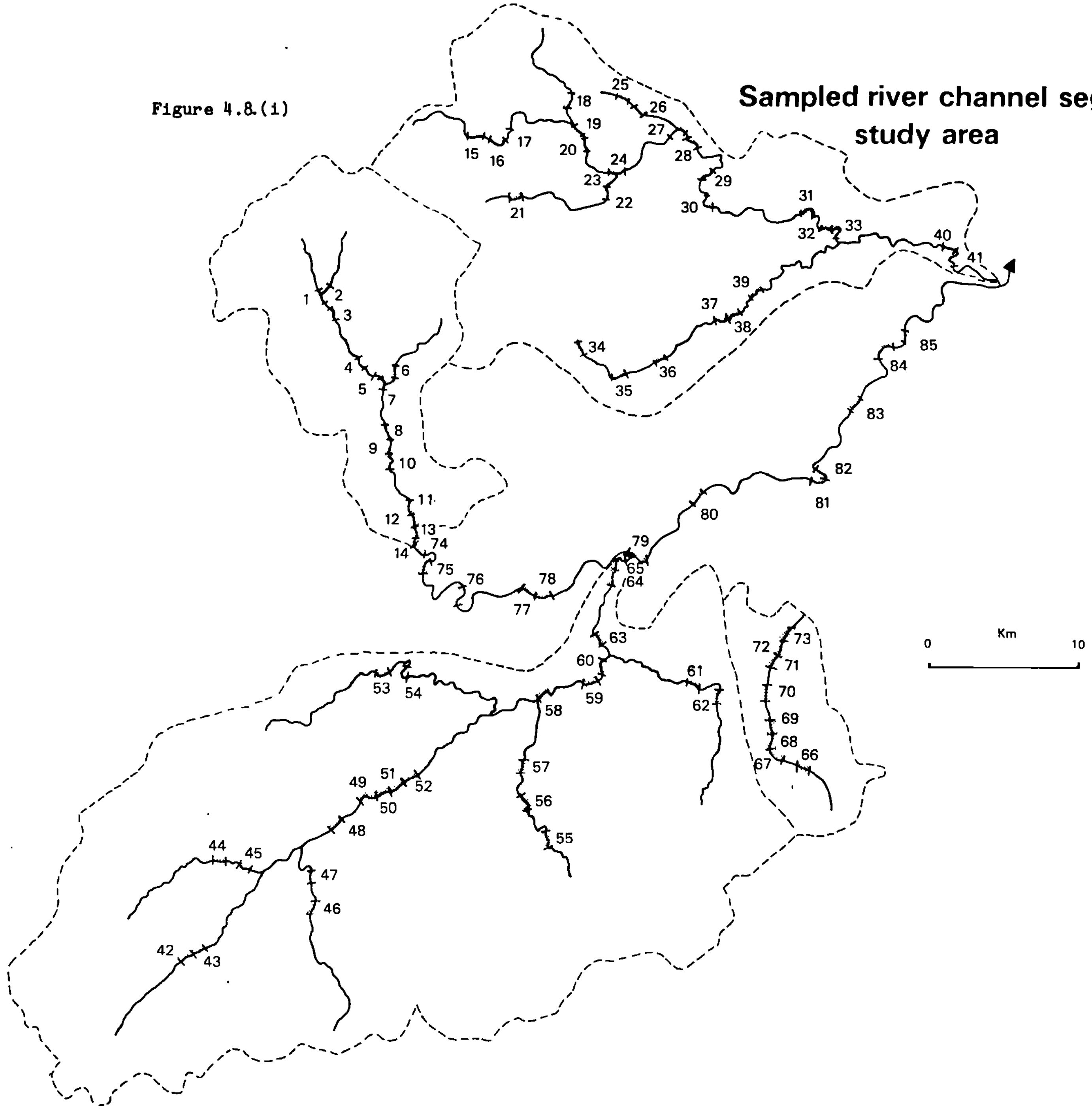
4.8.1 Basic range of activity indices

4.8.1.1 Spatial variations in sinuosity index

The basic summary statistics for the sinuosity index at each of the three map dates, for both the total sample and sub-divided by catchment, are shown in Table 4.8.1.(i) and the data in histogram form are shown in Figure 4.8.1.(i). 86-92% of the sites fell within the range 1.00-1.29 and thus the distribution was highly left skewed in favour of lower sinuosities at all three map-dates. The maximum values were however higher than for either of the other two study areas, associated with localised, confined meanders on the lower reaches of the tributaries eg. the Whiteadder (SIN=2.84; Sample 33). This was in contrast to the smaller amplitudes and higher frequencies associated with of the irregular to tortuous meanders on Derry Water (Dee study area). The data were also plotted on maps of the area, to assess their spatial distribution (Figures 4.8.1.(ii) to (vi)).

Figure 4.8.(i)

Sampled river channel segments within the Tweed study area



General location map for the Tweed study area

Figure 4.8.(ii)

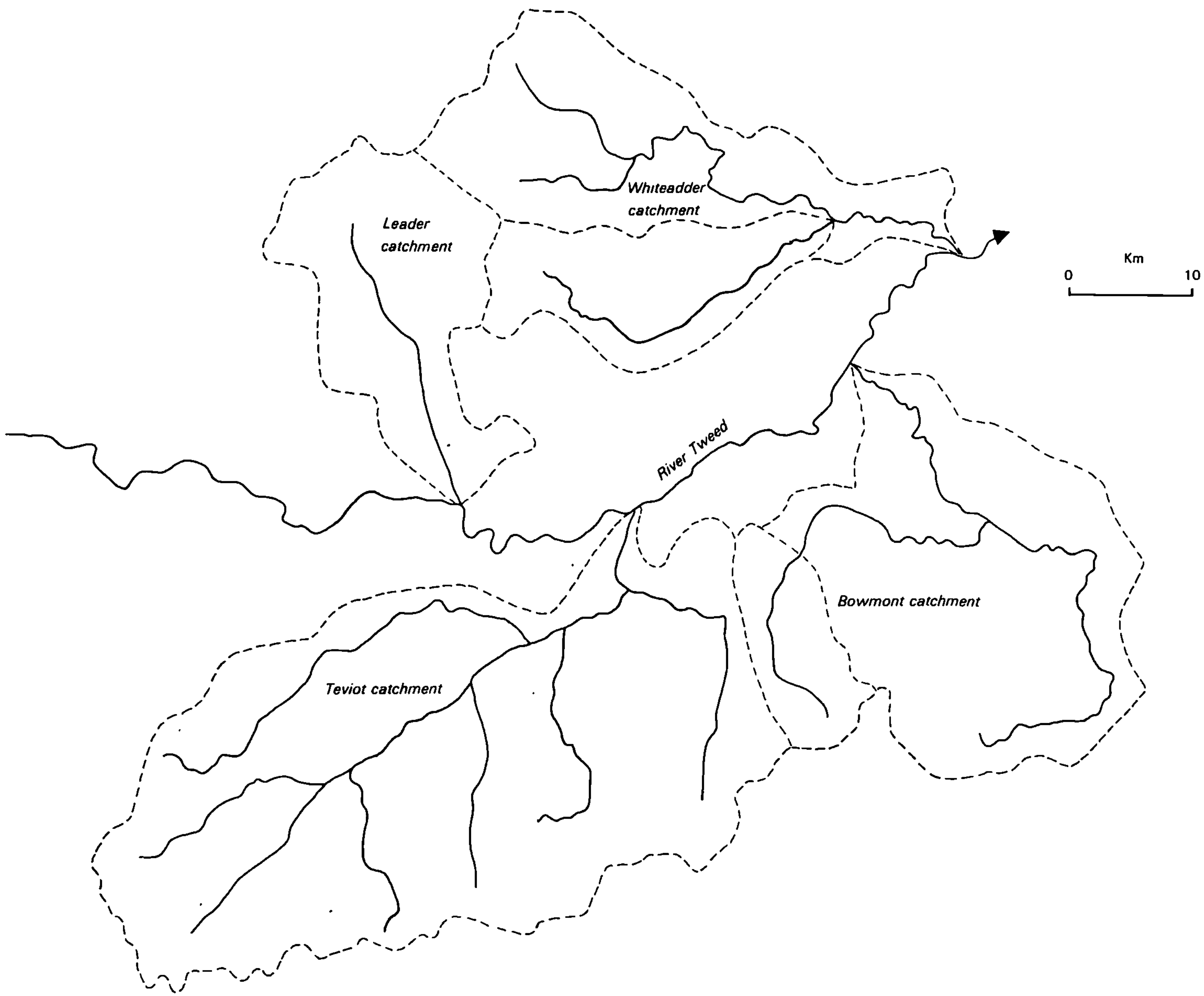


Table 4.8.1.(i)

Summary statistics for the sinuosity index within the Tweed sample

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
SIN 1 (1850)	1.08	1.15	1.22	0.14
SIN 2 (1900)	1.08	1.15	1.21	0.14
SIN 3 (1970)	1.07	1.14	1.20	0.13

Figure 4.8.1.(i)

Frequency of sinuosity index values within the Tweed study area

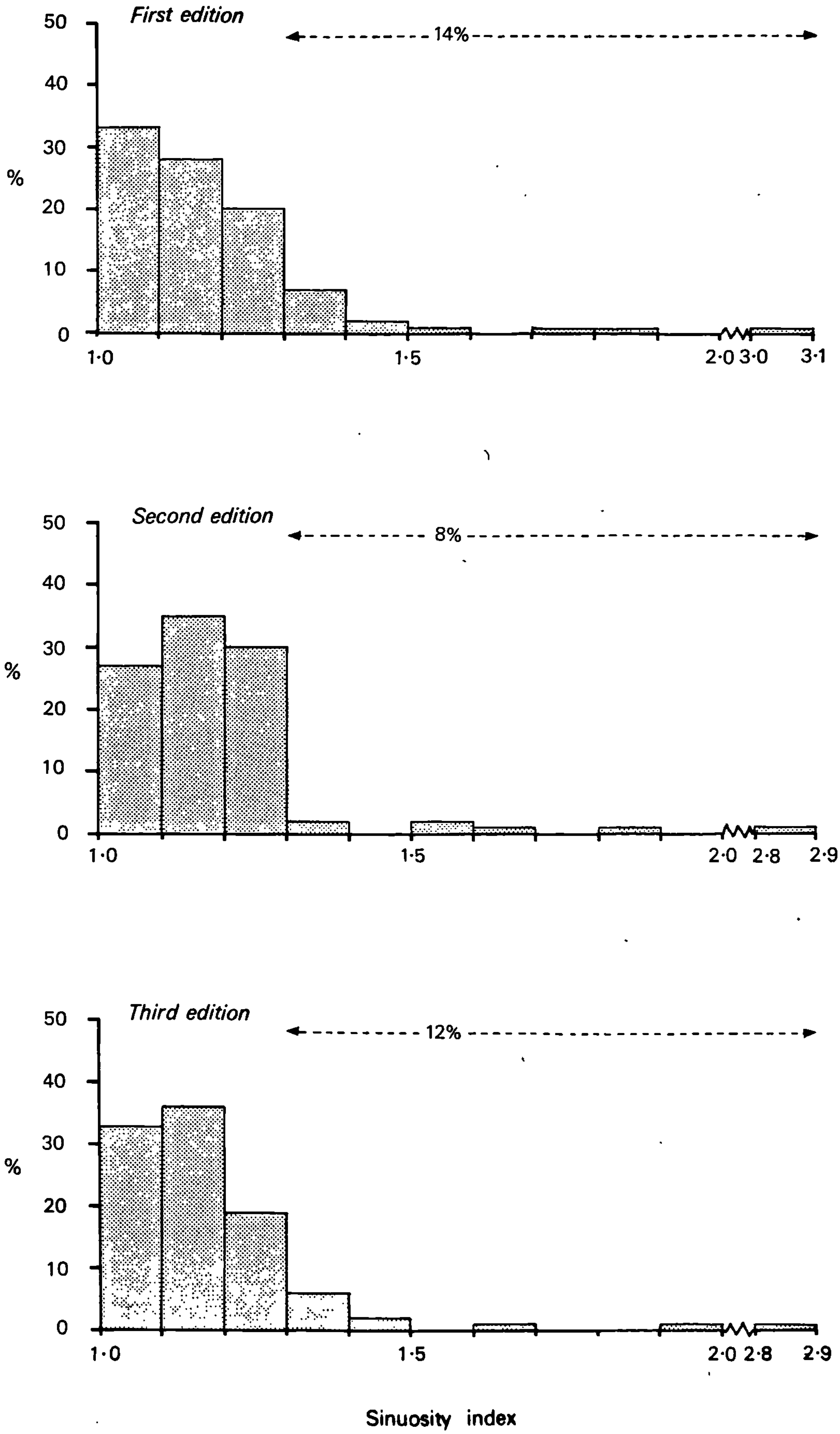
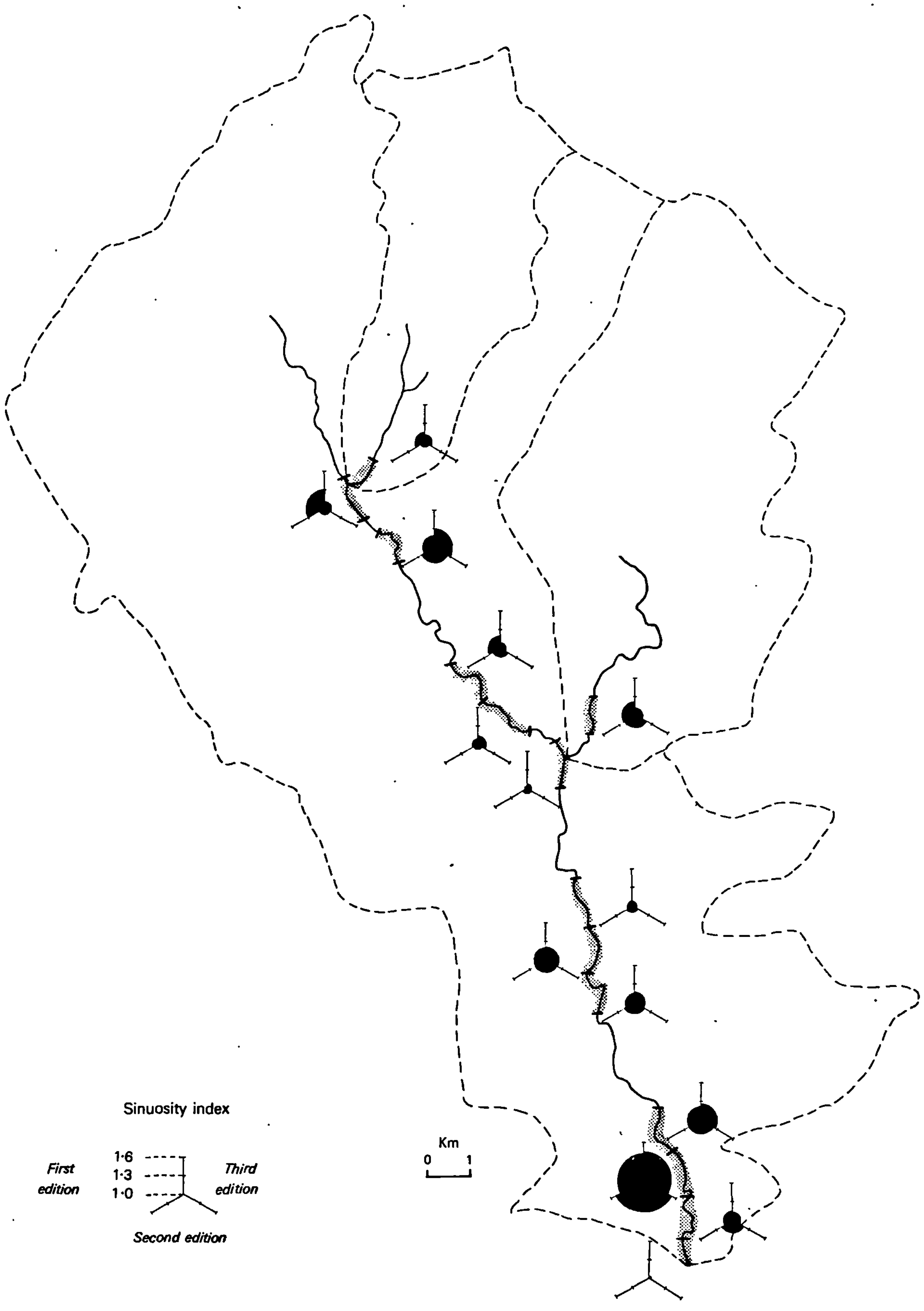


Figure 4.8.1.(ii)

Sampled sinuosity values falling within the Leader catchment for the three map dates



Sampled sinuosity index falling within the Whiteadder catchment for the three map dates

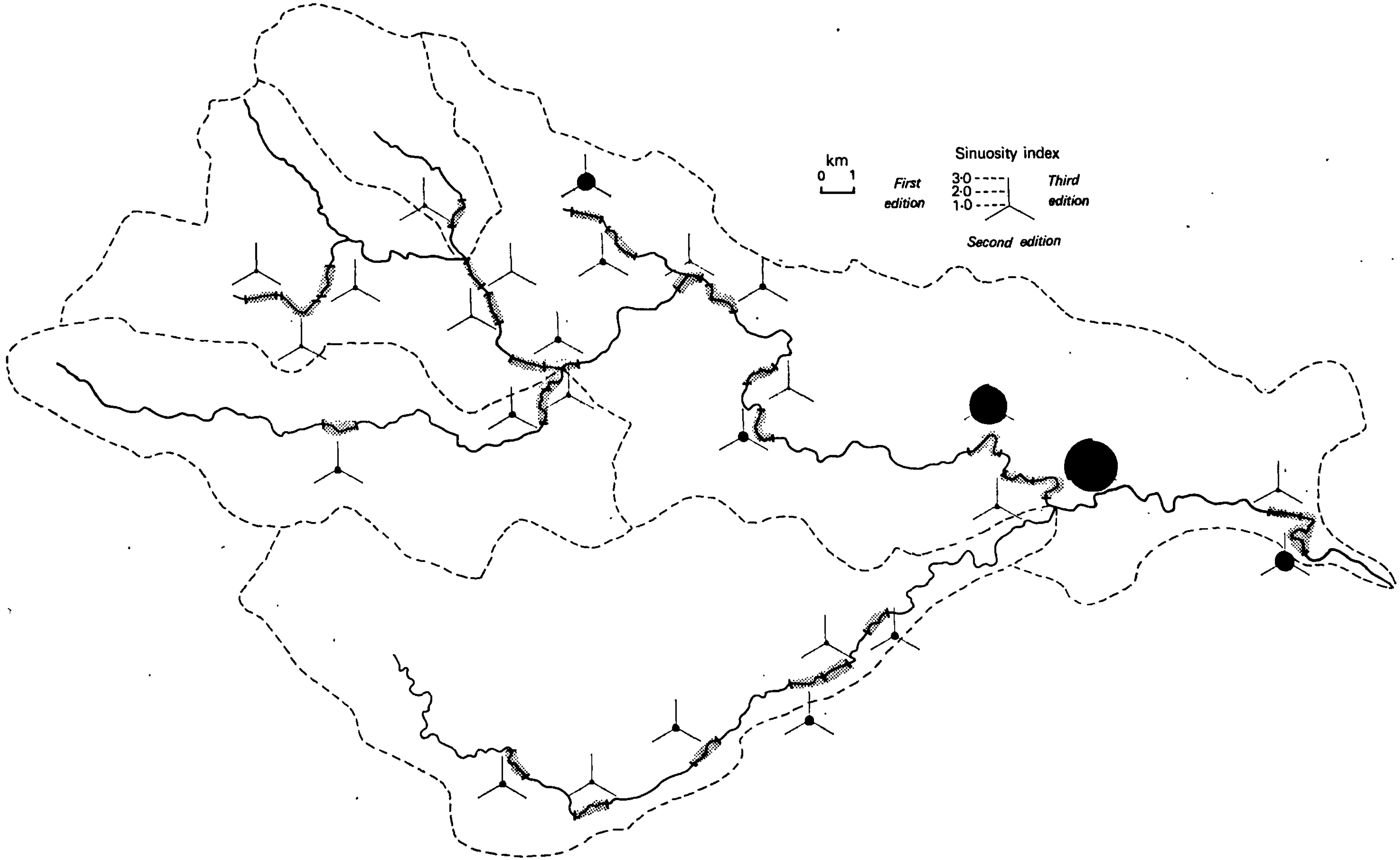


Figure 4.8.1.(iv) Sampled sinuosity index values falling within the Teviot catchment for the three map dates

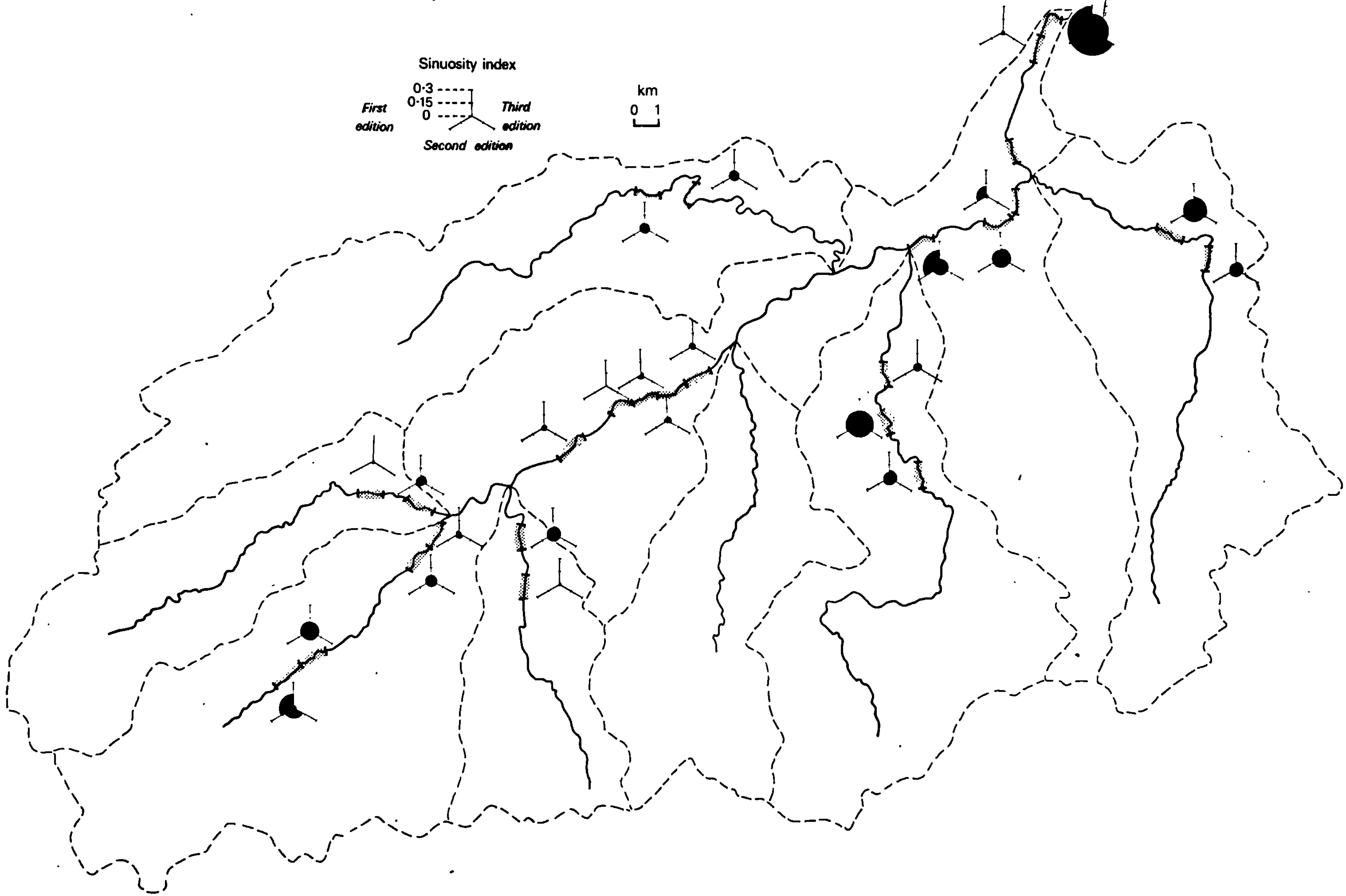


Figure 4.8.1.(xii)

Sampled sinuosity index values falling within the Bowmont catchment for the three map dates

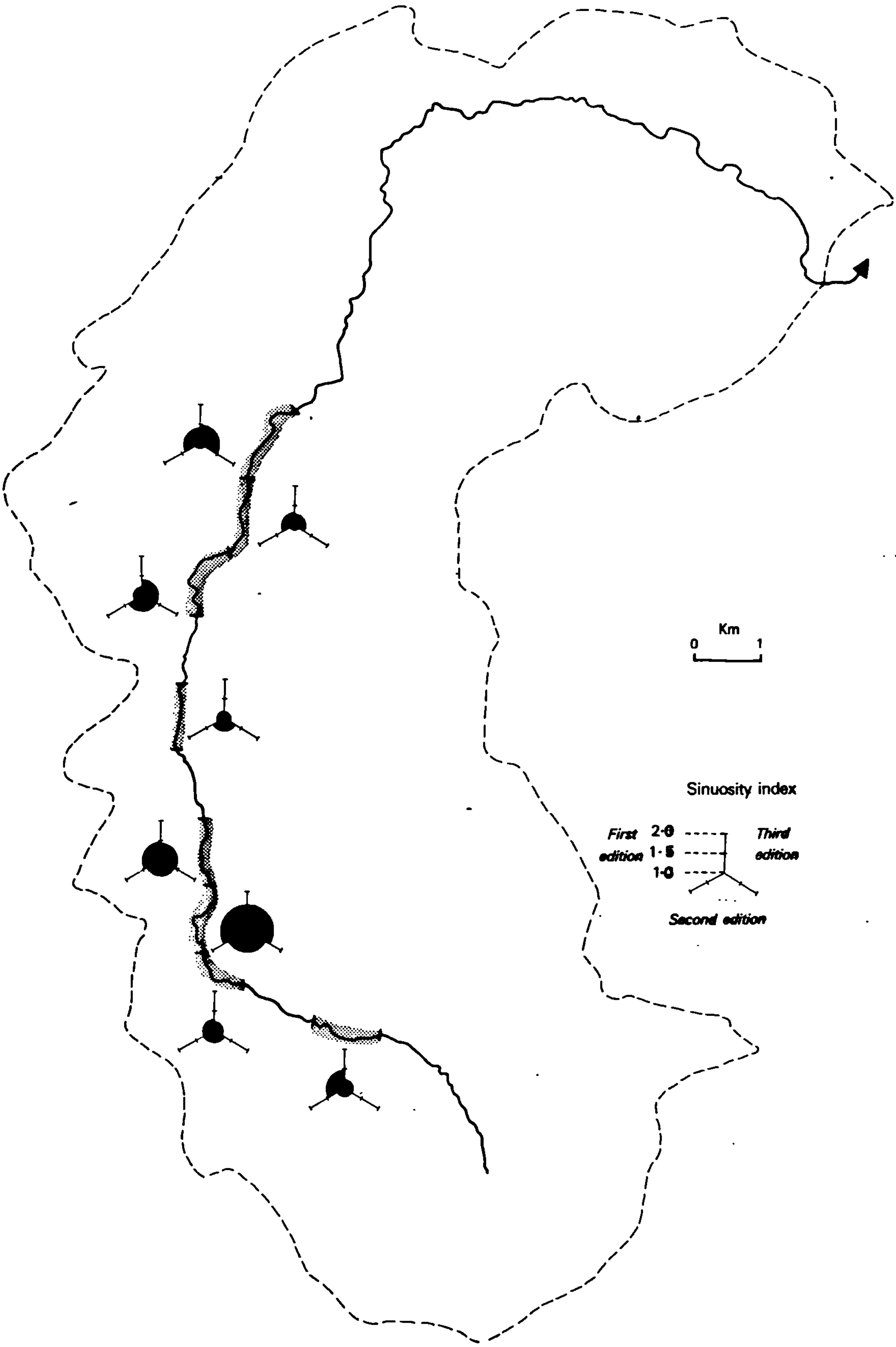
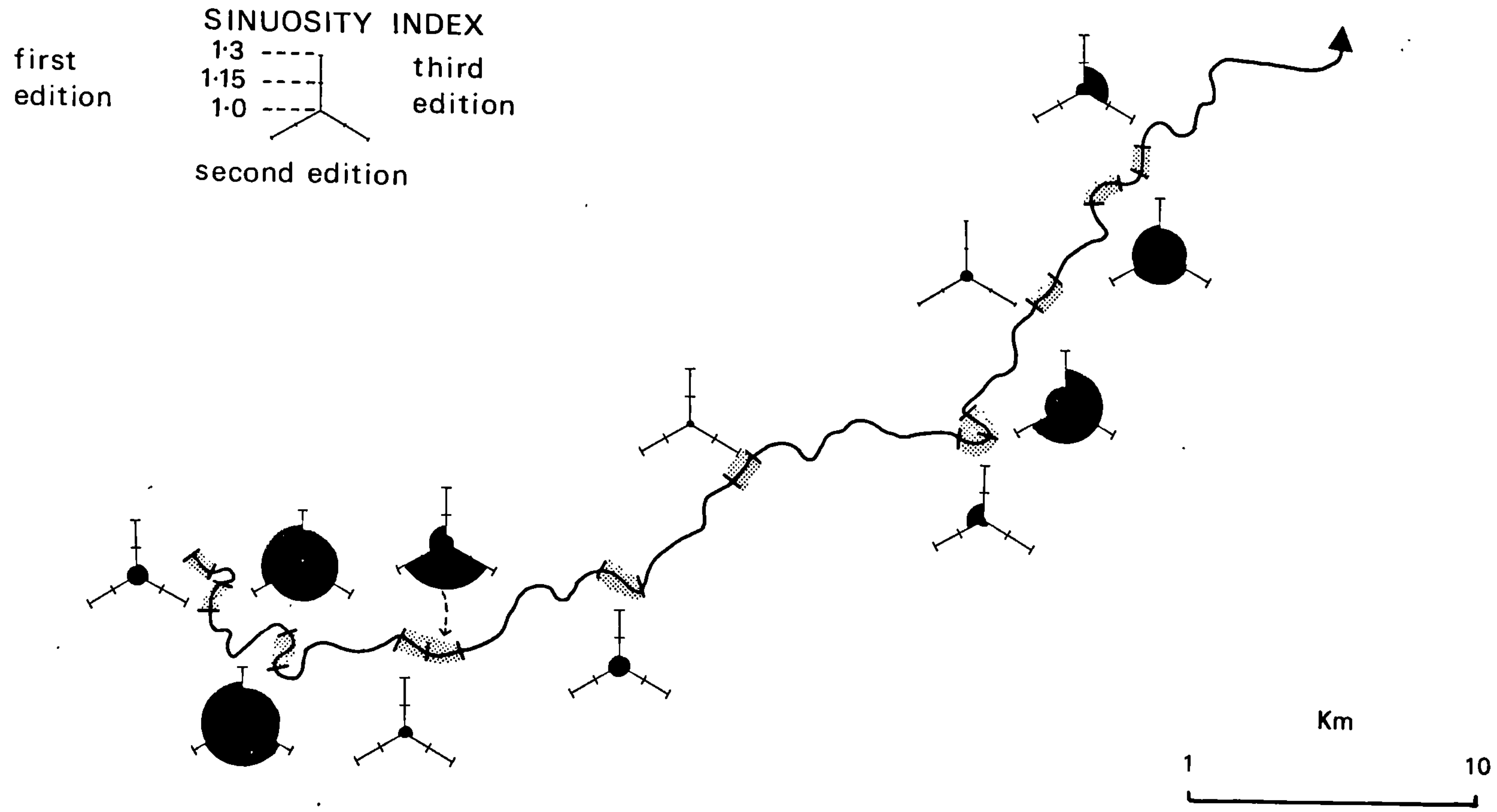


Figure 4.6.3.(iv)

Sampled sinuosity index values falling on the mainstream Tweed for the three map dates



Two separate river types were characterised by higher sinuosity. Within the lower tributary catchments, frequently higher sinuosity reaches represented localised entrenched or confined meanders, controlled by bedrock (Figure 3.4.2.(i)). Examples occur on the meander bends on the lower Leader confined by Upper Old Red Sandstone cliff faces (SIN1=1.42; Sample 12) and the lower Whiteadder associated with calciferous sandstone cliffs (eg. Sample 41 with SIN 1 and SIN 3 of 1.66 and 1.73 respectively). Other local controls, such as downcutting to bedrock (eg. mainstream Tweed above Kelso), cliff faces (eg. Jed Water and lower Leader), fluvioglacial deposits (eg. Bowmont Water) and Lateglacial terraces (eg. Teviot), all gave periodic rather than consistent floodplain confinement downstream (see Chapter 3). The same disruption to channel planform associated with extensive fluvioglacial deposits, as on Speyside, did not occur over a 250 year timespan. Occasional higher sinuosity reaches of > 1.20 , without any apparent confinement, were also interspaced by lower sinuosity reaches without any consistent periodicity eg. middle Leader, Bowmont and Whiteadder/Dye Water (SIN=1.22-1.33; Sample 22). These were associated with localised accumulations of material of either fluvioglacial (eg. at the former location of meltwater channels) or alluvial origin.

Tributaries were frequently locally more sinuous than the mainstream, eg. Ale Water in the Teviot catchment, but this was related to the location of drumlinoid deposits on the floodplain and bedrock control. This variation may also reflect channelisation and river training within the mainstream Teviot sample (Chapter 6). Each of the catchments had different patterns, with marked contrasts in the frequency of more sinuous reaches occurring between the Leader and Whiteadder and the Teviot and Bowmont, which must have varying degrees of bedrock/ terrace confinement and depths of fluvioglacial deposits. It must be noted that the sinuosity of the large scale features on the Tweed (eg. the meander bend downstream of Melrose) were frequently more tortuous than the sinuosity of individual grid-square meander segments, which is important in terms of the 1 km^2 sampling unit. For comparative purposes, the overall sinuosity of the complete meander bend was calculated as in excess of 2.00.

4.8.1.2 Spatial variations in braiding index

The basic summary statistics for the braiding index at each of the three map-dates, both for the total sample and sub-divided by tributary, are shown in Table 4.8.1.(ii). The median braiding index for the total sample was low in comparison with the Spey and Dee study areas, with 73-78% of the sample having BI equal to 1.00-1.09 (Figure 4.8.1.(vii)). This indicated that very little sediment was actually stored within the channel. Highest median values were found on the upper Bowmont water, especially at BI1 (1.39). However, these were clearly atypical for the rest of the study area and a value of >1.20 was rarer than in either Deeside or Speyside, again as a result of the much more limited supply of fluvioglacial material eg. upper Whiteadder basin (Chapter 3.4.2).

Table 4.8.1.(ii)

Summary statistics for the braiding index within the Tweed sample

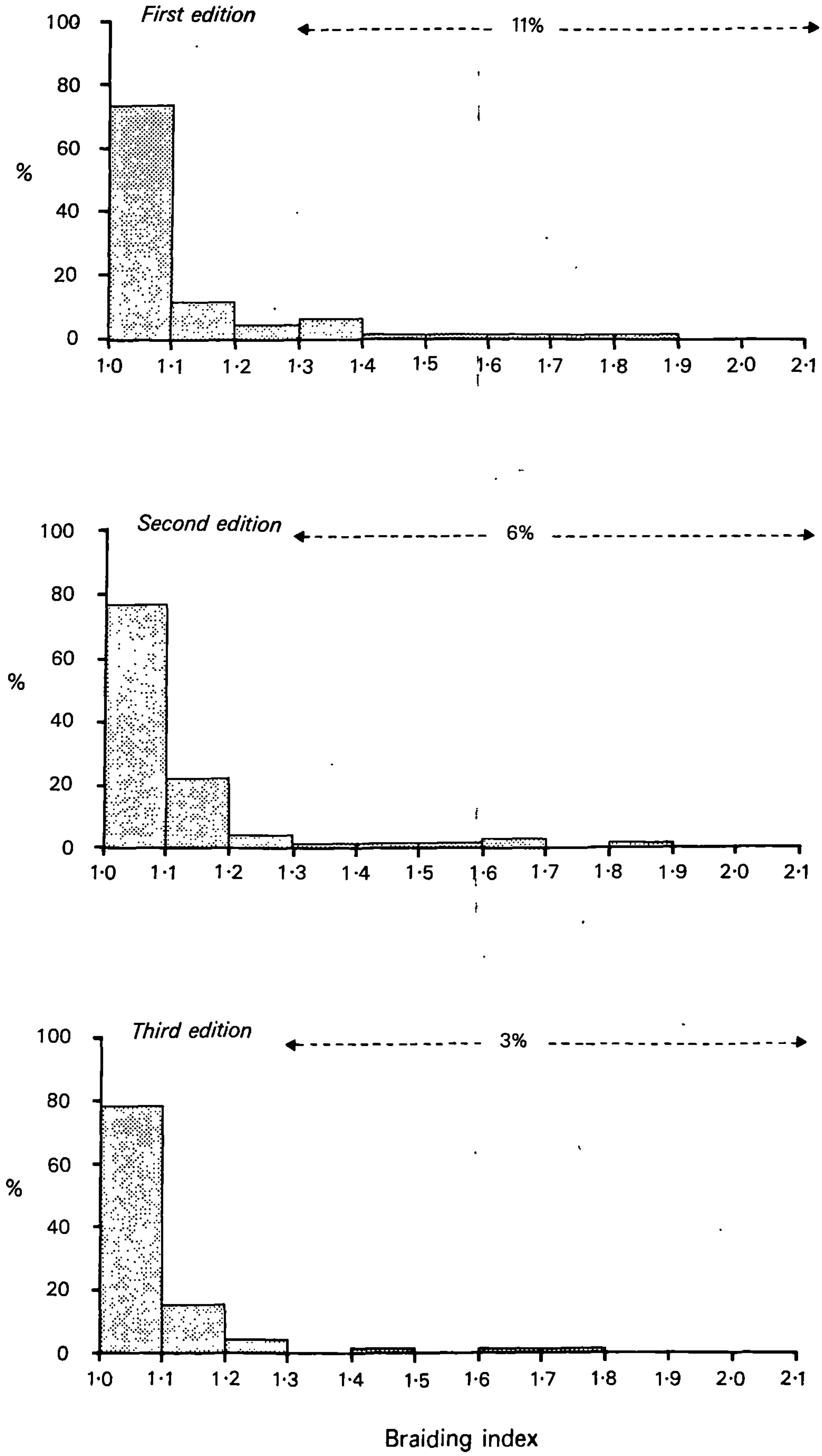
Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
BI 1	1.00	1.00	1.11	0.11
BI 2	1.00	1.01	1.08	0.08
BI 3	1.00	1.00	1.08	0.08
Leader				
BI 1	1.00	1.00	1.13	0.13
BI 2	1.00	1.03	1.09	0.09
BI 3	1.00	1.00	1.12	0.12
Whiteadder				
BI 1	1.00	1.00	1.04	0.04
BI 2	1.00	1.00	1.06	0.06
BI 3	1.00	1.00	1.06	0.06

Table 4.8.1.(ii) cont.

Parameter	25th Percentile	Median	75th Percentile	IQR
Teviot				
BI 1	1.00	1.00	1.08	0.08
BI 2	1.00	1.00	1.03	0.03
BI 3	1.00	1.00	1.01	0.01
Bowmont				
BI 1	1.22	1.39	1.70	0.48
BI 2	1.09	1.18	1.56	0.47
BI 3	1.00	1.04	1.20	0.20
Tweed				
BI 1	1.00	1.04	1.21	0.21
BI 2	1.00	1.05	1.49	0.49
BI 3	1.00	1.12	1.32	0.32

Figure 4.8.1.(vii)

Frequency of braiding index values within the Tweed study area



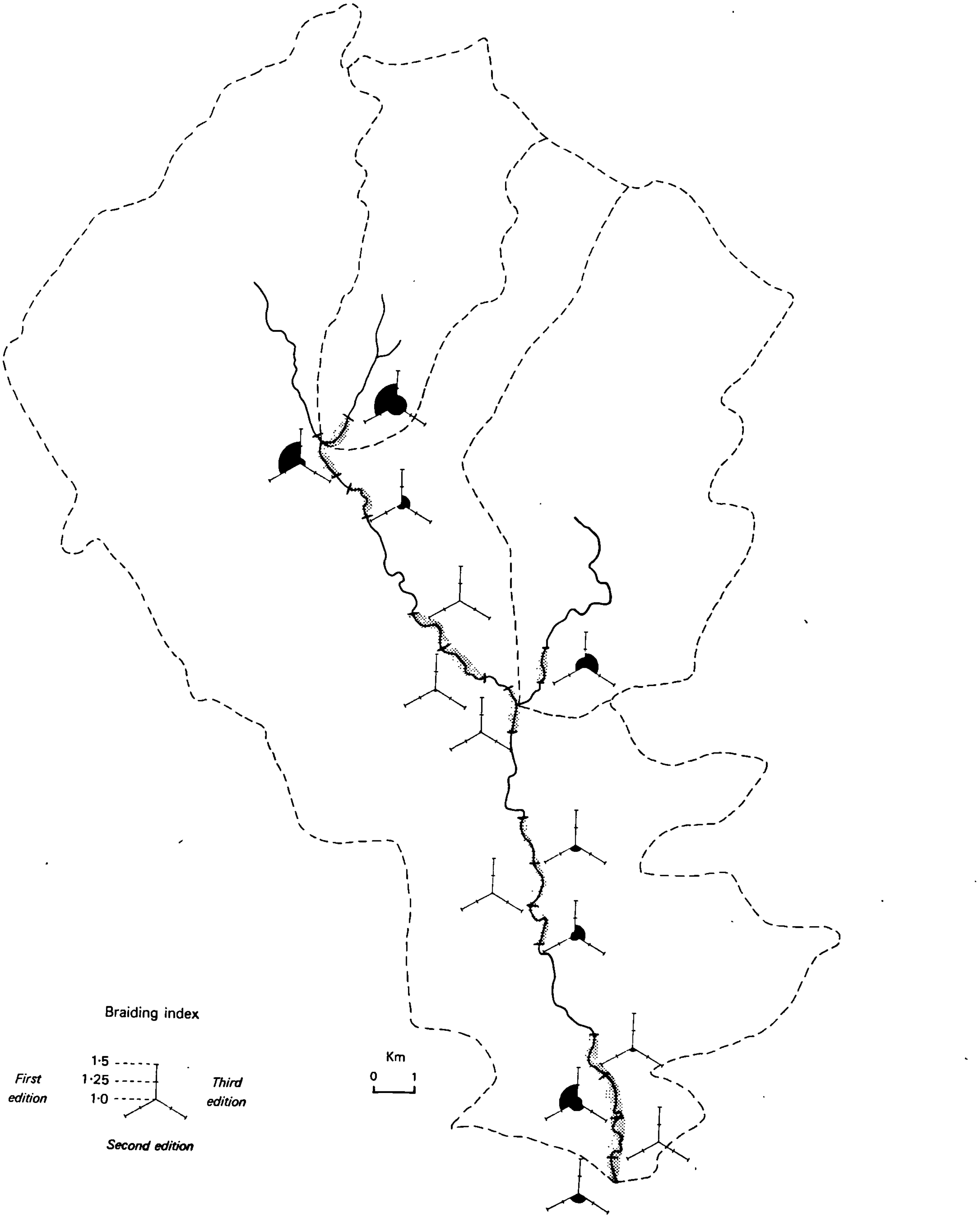
The spatial distribution of these BI values is shown in Figures 4.8.1.(viii) to (xiii). Confluence sites had characteristically lower values than the other study areas, reflecting an absence of active, low angle alluvial fans with frequent bars. The highest BI values within the Tweed study area occurred in three main areas. Firstly, high values occurred in the upper middle reaches of rivers eg. Bowmont (Sample 69) and Leader (Sample 1) but apart from the Bowmont, these were very localised in extent. Nowhere were there any values to compare with the persistently high BI values of the Feshie or Dorback Burn. The middle reaches of the smaller tributaries, eg. the Dye (where there were reserves of fluvioglacial material due to former meltwater activity), also had bar features within the channels. The mainstream Tweed was characterised by locally high BI values (eg. Sample 77; BI=1.33-1.69) at intervals downstream, reflecting the occasional incidence of readily erodible sediments and which represented both natural and man-made sediment stores. Finally, it was interesting to compare the high BI values of Bowmont Water with the much lower values on the Jed and Slitrig (tributaries of the Teviot), despite the fact that they both drained neighbouring upland areas of the Cheviot massif. Bedrock control was however much more important on the Teviot tributaries.

4.8.2 Basic range of typology characteristics

4.8.2.1 Quantitative catchment characteristics

Figure 4.8.1.(vii)

Sampled braiding index values falling within the Leader catchment for the three map dates



Sampled braiding index values falling within the Whiteadder catchment for the three map dates **Figure 4.8.1.(ix)**

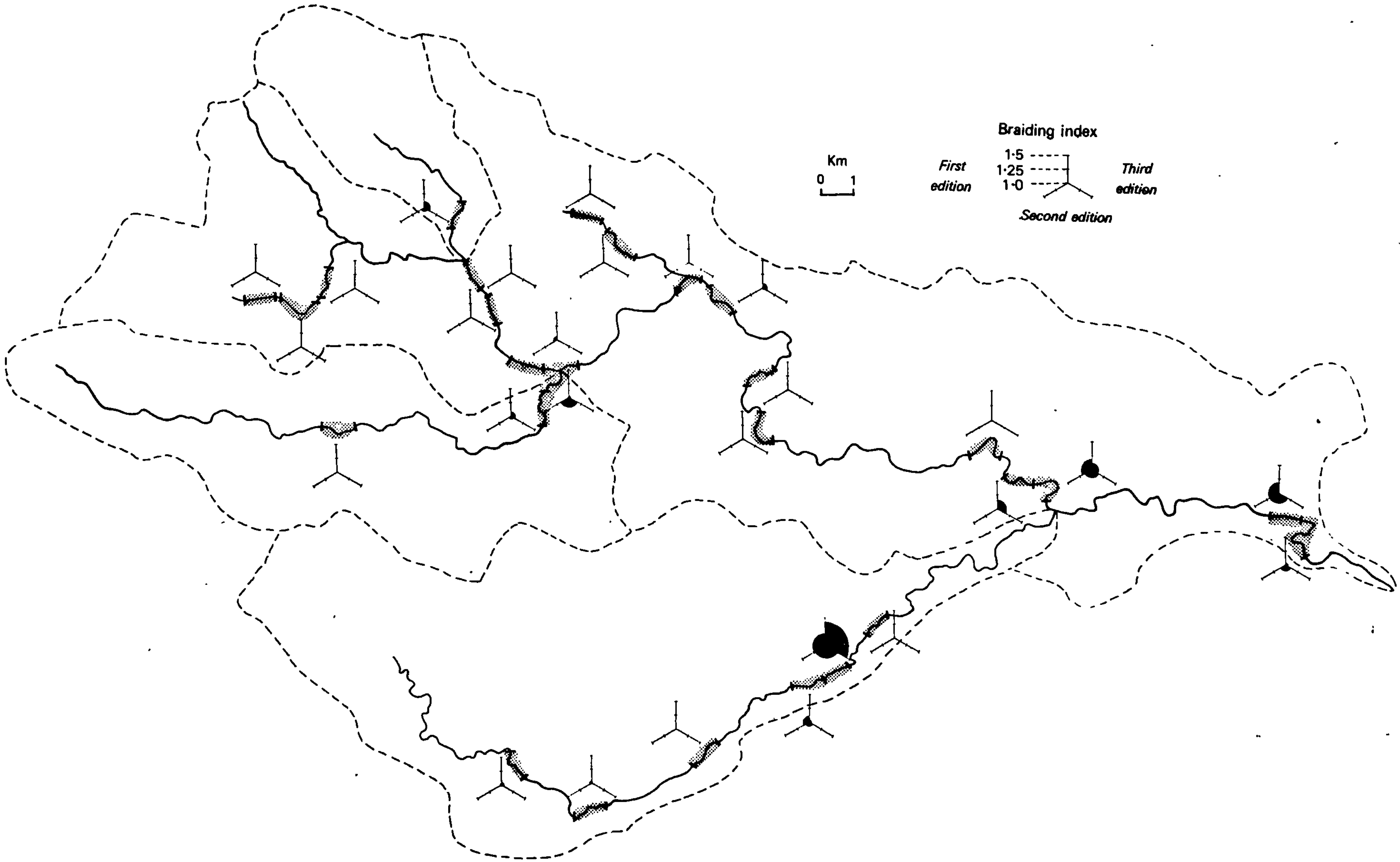


Figure 4.8.1.(x) Sampled braiding index values falling within the Teviot catchment for the three map dates

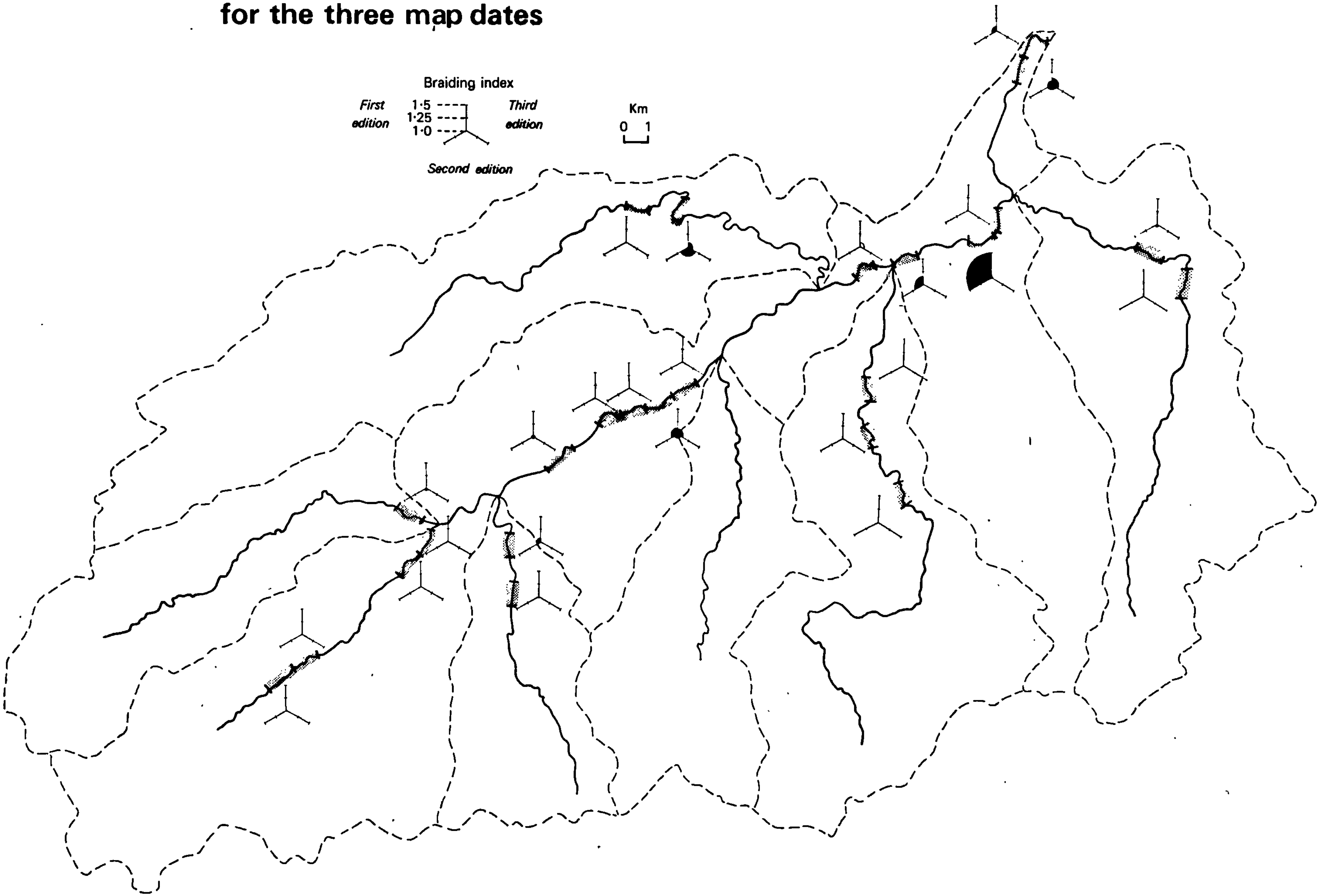


Figure 4.8.1.(xi)

Sampled braiding index values falling within the Bowmont catchment for the three map dates

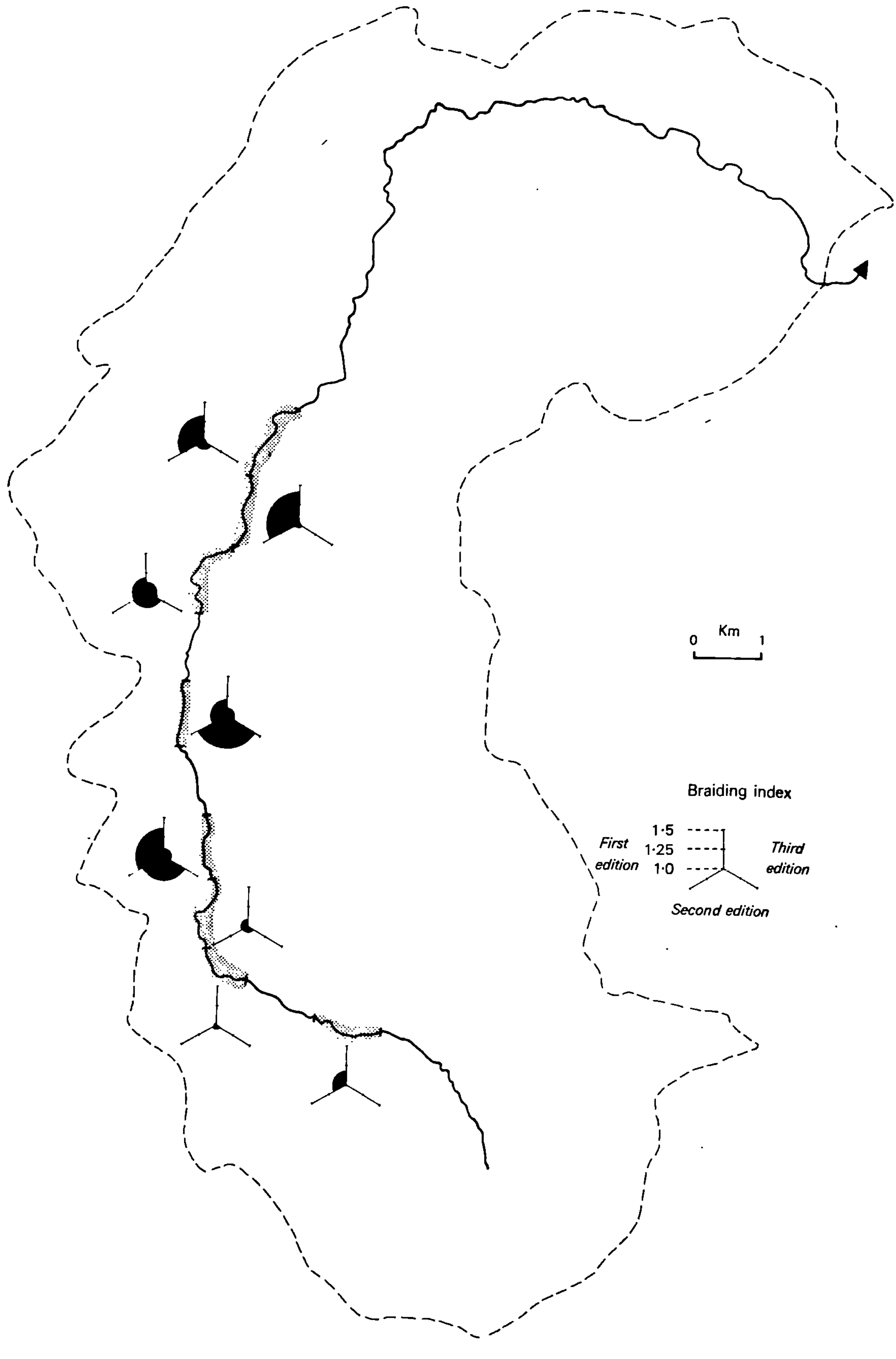
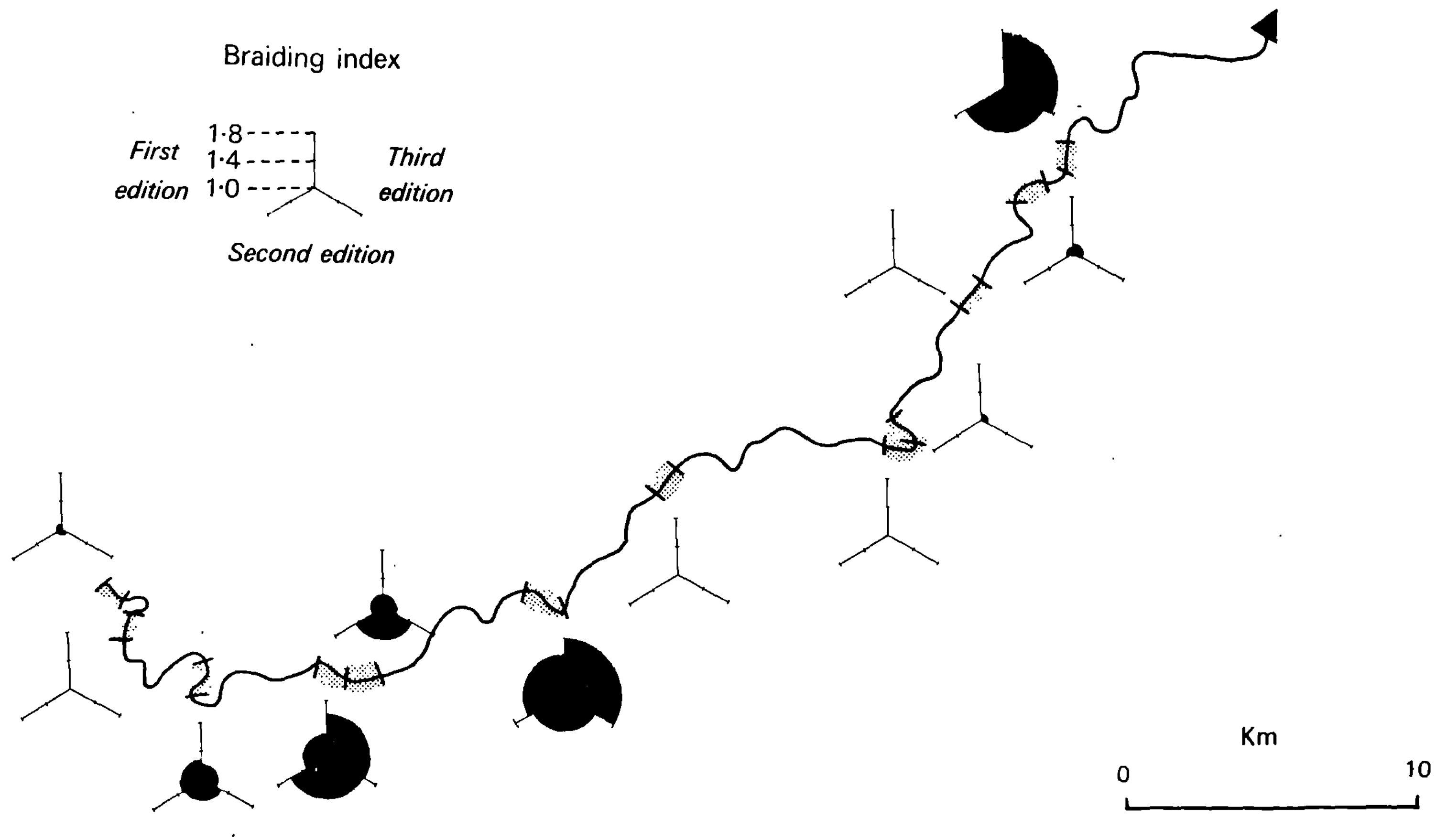


Figure 4.8.1.(xiii)

Sampled braiding index values falling on the mainstream Tweed for the three map dates



The contributing drainage areas were on average larger than for the Spey study area, ranging from 25.2 km^2 on the Cleekhimin Burn to 2400 km^2 on the mainstream Tweed and the average height of the sample sites was much lower than for the other study areas. In terms of MAXWID, 58% had the ratio of channel width to floodplain width < 20 and therefore, the Tweed sample was generally more confined than the Spey or the Dee. However, this did not take into account local obstructions not detectable from map evidence.

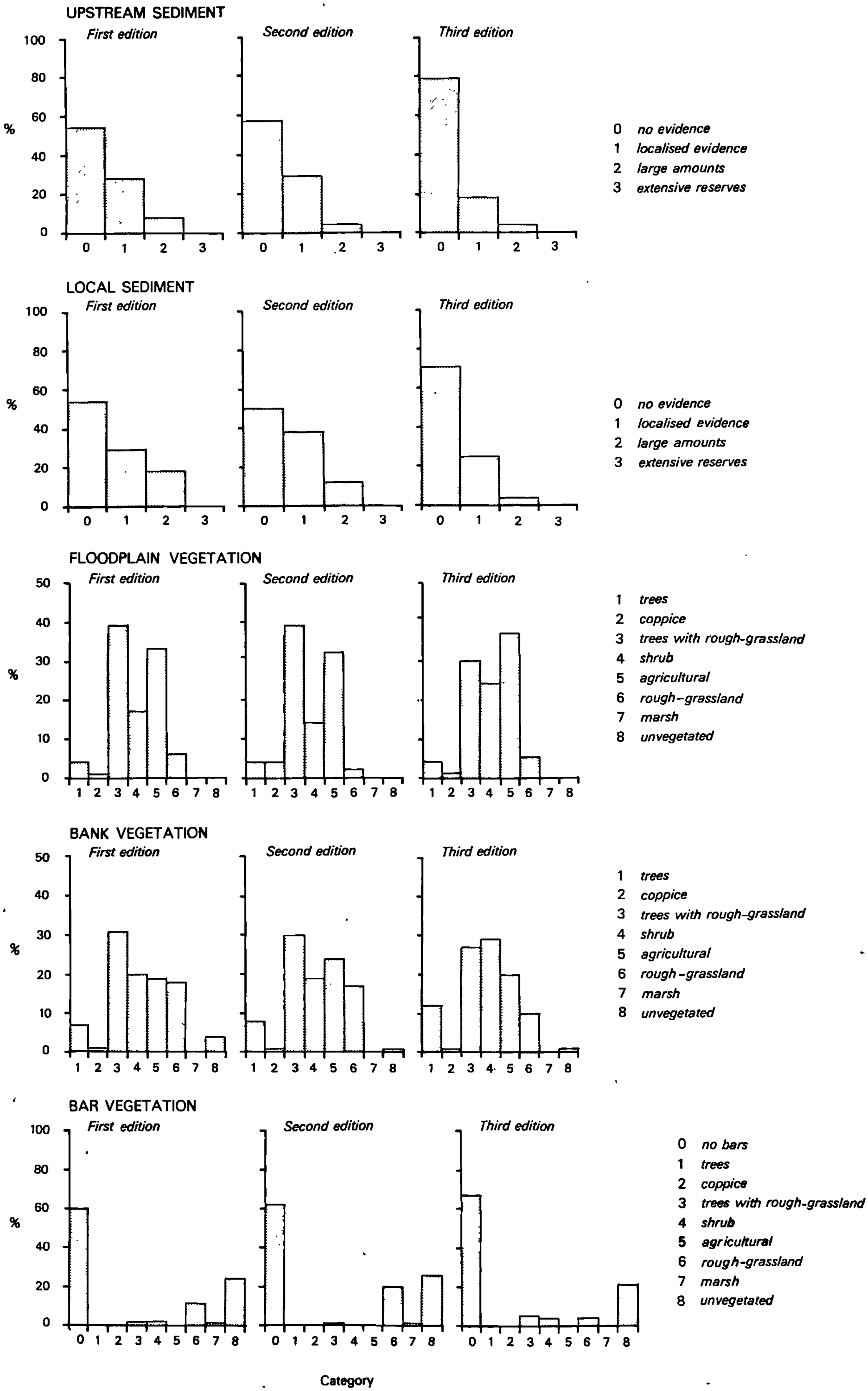
In the following analysis, a significant correlation (r) with a sample size of 80 was 0.22 (0.05 level) and 0.28 (0.01 level). When total sample sinuosity at each of the three map-dates was correlated with contributing area, a very low correlation was obtained and thus position within the drainage basin was not important. More sinuous channels were frequently associated with rock outcrops eg. on the Whiteadder and this occurred periodically along the total channel length where the river had cut down completely through fluvioglacial material, and thus being independent of area. In contrast, BI values showed highly significant positive correlations with area (r=0.31-0.54). A similar pattern occurred between BI values and basin height, with highly significant <u>negative</u> correlations (r=-0.26--0.59). This may relate to an intermittent increased availability of fluvioglacial material within the lower reaches. The only high correlation with MAXWID was BI1 with r=0.57, implying that in the 1850s, much more use was made of the available floodplain area. It was notable across the range of area, average height and MAXWID, there were a large number of samples which retained low values for both braiding and sinuosity.

4.8.2.2 <u>Frequency of qualitative catchment characteristics</u>

When the qualitative classification characteristics were presented as histograms (Figure 4.8.2.(i)), the following results were noted. Local and upstream sediment reserves were much more limited in extent than either in Speyside or Deeside, as indicated especially by the third edition. In terms of sediment availability, there were no sample sites at any map-date which had extensive reserves of sediment, within or proximal to the channel. In fact, 54-79% had no obvious upstream sediment reserves, while a similar percentage of sites had no evidence of local sediment stores. This had implications for the types of changes noted, with avulsion more important than the intra-active area shifting of the channel through unstabilised sediments. Dominant floodplain and bank vegetation were trees and rough-grassland (3) and agricultural land (5) (Figure 4.8.2.(i)). Channels stabilised by trees along the majority of the reach accounted for 40% of the sample, although whether the trees developed because reaches were inherently stable, or vice versa, must be assessed. Bars with trees as stabilising vegetation were uncommon though some bars were permanent enough to have established rough-grassland vegetation (4-11 % of sample).

Channel pattern was nearly symmetrical around the middle categories in comparison with the other study areas (Figure 4.8.2.(i)), with more reaches falling within the middle range of categories 3 and 5. Categories with the highest frequency were thus "sinuous" and "irregular", with 24-32% and 20-24% respectively of the total sample. However, "wandering" rivers (associated with the highest rates of change within the Spey study area) were much rarer than in the other two study

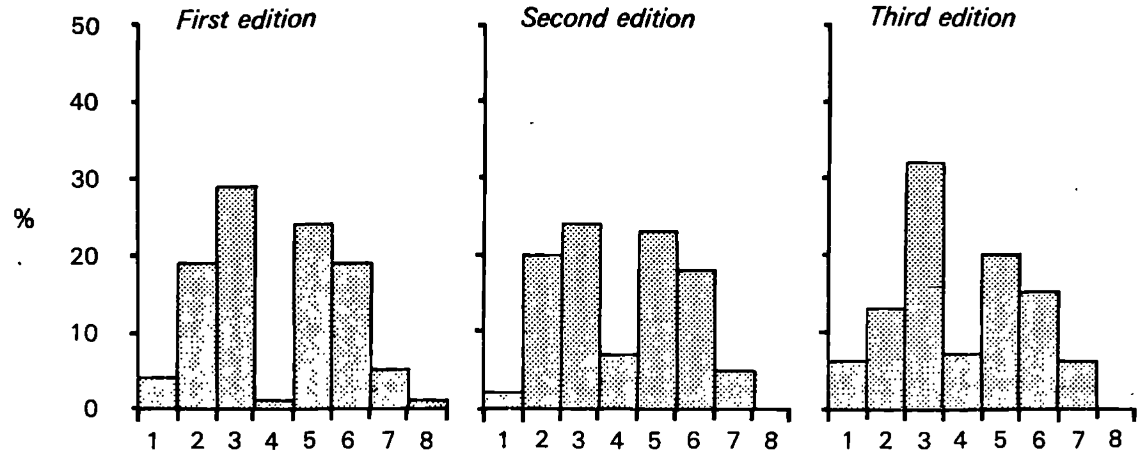
Frequencies within each channel typology classification for the Tweed study area
(for detailed key, see Table 4.2.(i))



Frequencies within each channel typology classification for the Tweed study area

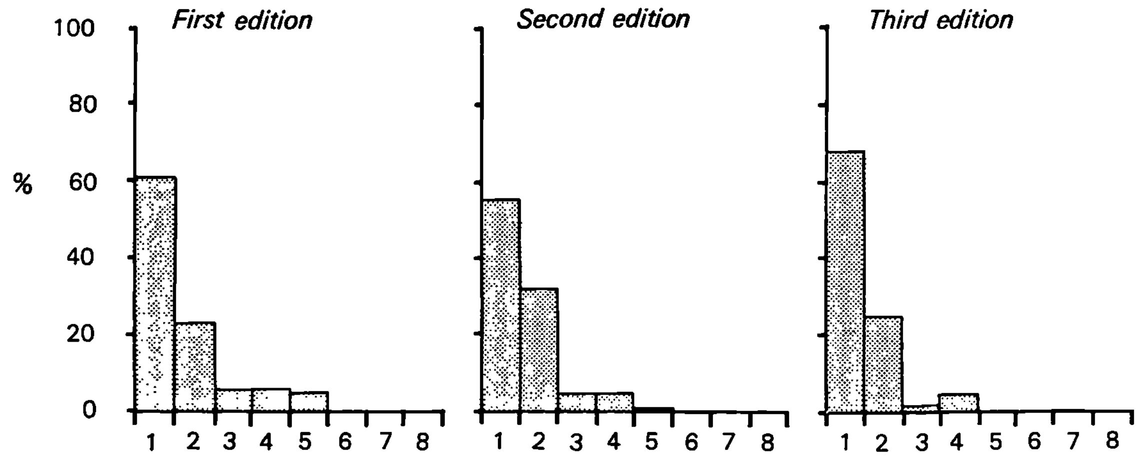
(for detailed key, see Table 4.2.(i))

CHANNEL PATTERN

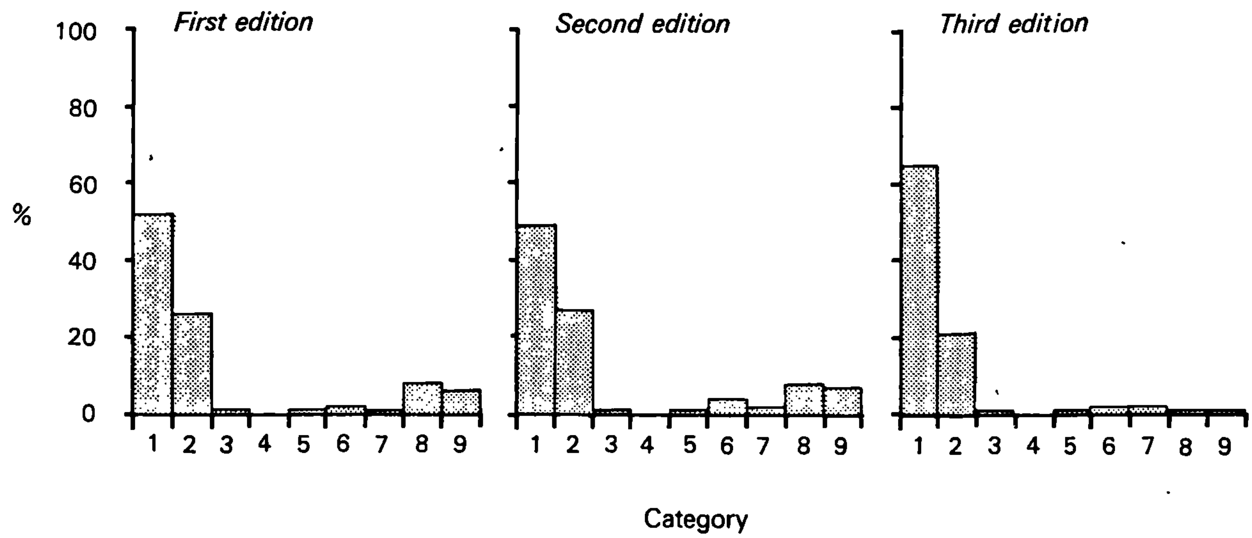


1 *straight*
2 *straight/sinuous*
3 *sinuous*
4 *wandering*
5 *irregular*
6 *irregular meanders*
7 *regular meanders*
8 *tortuous meanders*

ISLANDS



1 *none*
2 *occasional*
3 *frequent*
4 *split*
5 *braided*
6 *braided and anabranching*
7 *anabranching*
8 *reticulate*



1 *no evidence*
2 *unvegetated gravels*
3 *progressions and cutoffs*
4 *irregular channel activity*
5 *avulsion*
6 *old palaeochannels*
7 *marsh*
8 *combination of 2 and 6*
9 *evidence of older palaeochannels*

areas. The highest category of the "islands" classification attained was "braided" but this involved only 5 sample sites out of all three map-dates. By far the majority (57-65%) had no islands indicated ie. no major sediment storage within the channel, again in marked contrast to the Spey and Dee study areas.

In terms of "evidence of activity", a large number of first edition sites had evidence of earlier pre-1850 activity (Figure 4.8.2.(i)), especially in terms of higher sinuosities. For example, the Teviot on Roy's map was portrayed as being much more sinuous than in 1850 with vegetation patterns and palaeofeatures confirming this. Clearly along certain reaches, there had been considerable lateral shift, eg. at the Hassendean/ Teviot confluence (middle reaches of the Teviot), where the former site of the Hassendean chapel is now in the middle of the Teviot river bed. The Teviot also appeared to have had many more occasional bars along its course in the 1750s in comparison with the first map edition (1858). Thus, a localised reach on Borthwick Water was formerly braided but by 1863, had a sinuous planform (Figure 4.8.2.(ii)). Bars were frequently common at or below confluence sites as on the Jed and along the mainstream Teviot, for example near Denholm, which suggests more sediment storage within the channel in the 1750s (Figure 4.8.2.(iii)). It is impossible to assess whether these were stable vegetated bar features or less permanent unvegetated deposits, which were frequently reworked.

Figure 4.8.2.(ii)

Formerly braided reach on the Borthwick Water

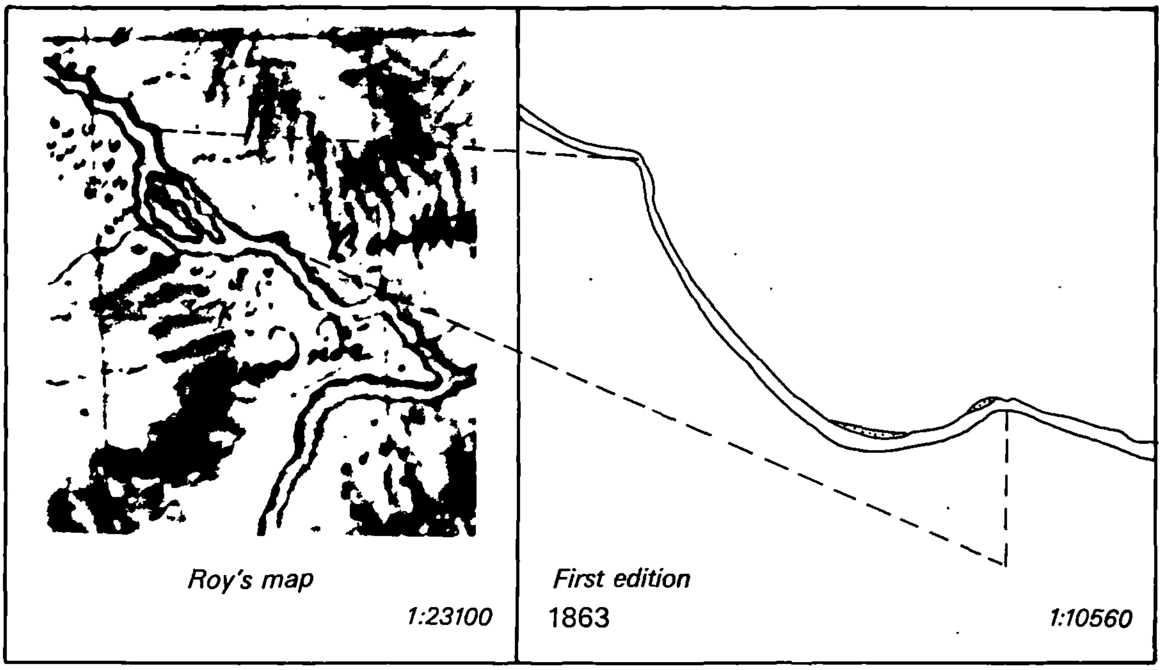


Figure 4.8.2.(iv)

Channel change below the Boonreigh/ Leader confluence on the Leader

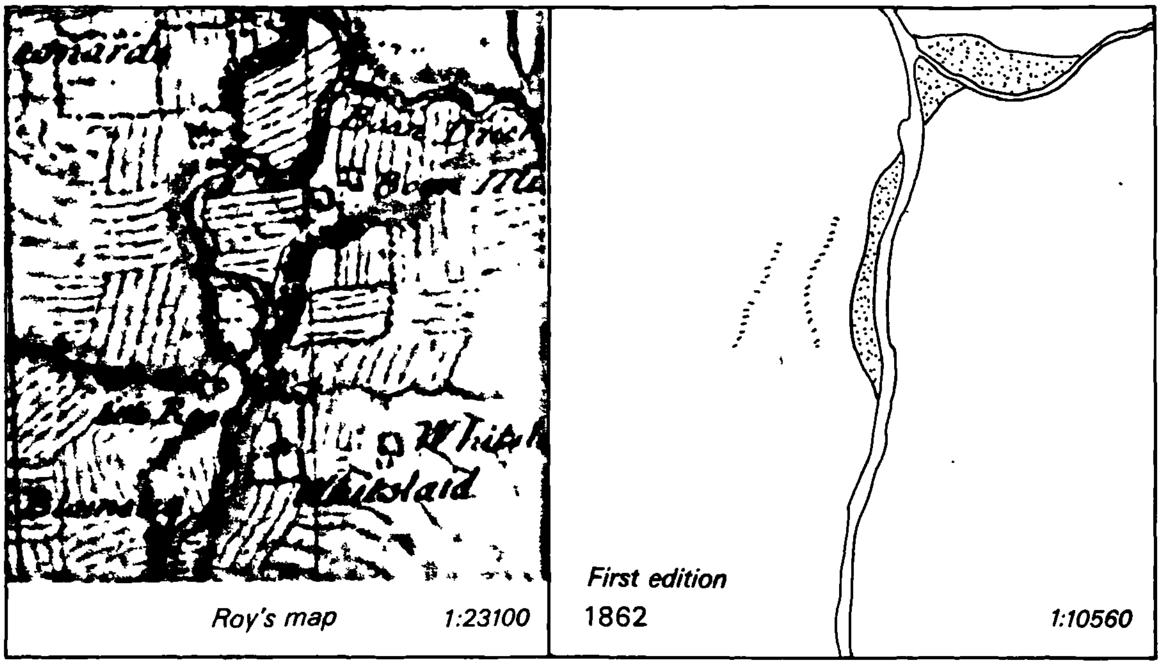
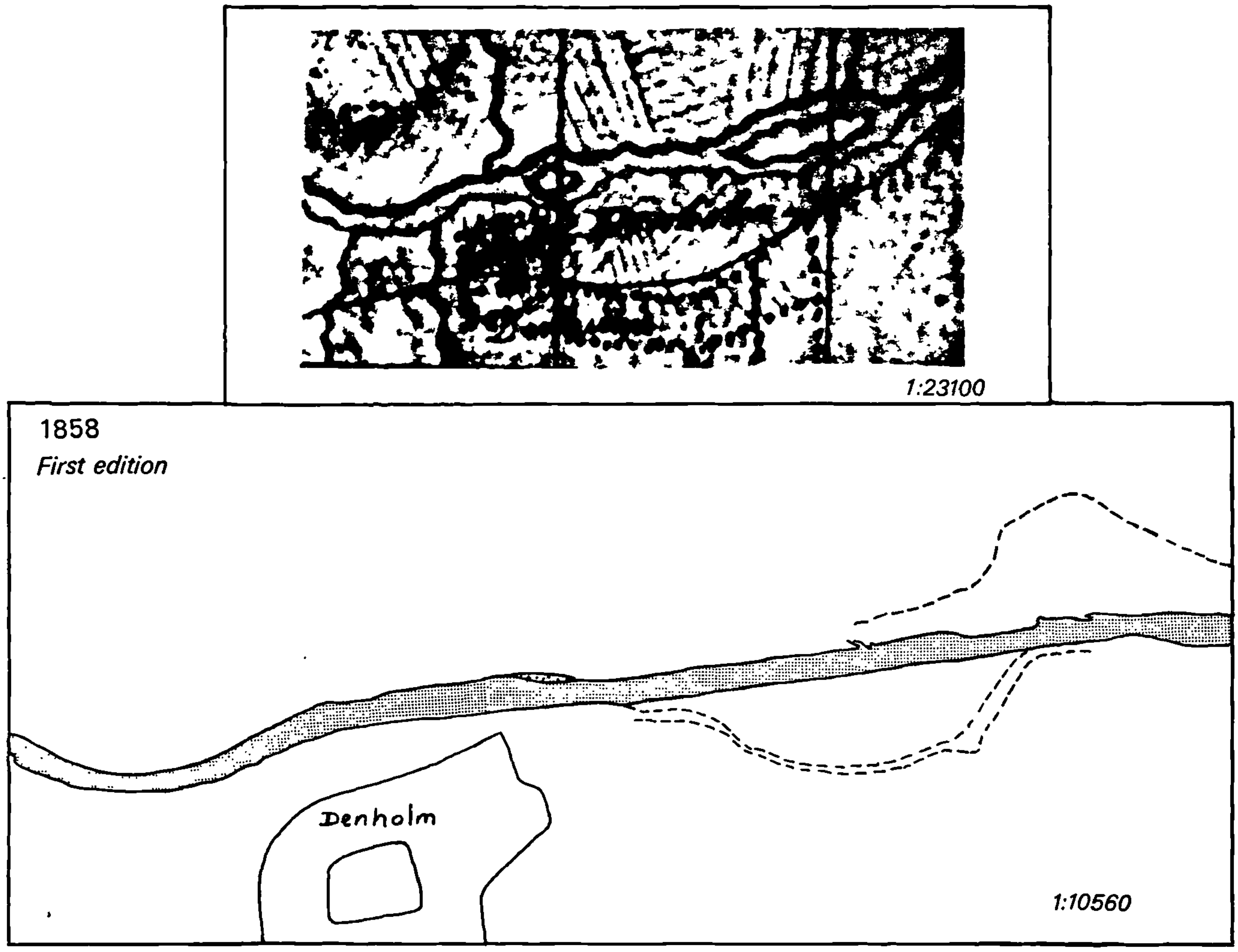


Figure 4.8.2.(iii)

Former islands on the Teviot near Denholm (Roy's map, 1750)



Similarly, the mainstream Leader had undergone a considerable reduction in sinuosity between 1750-1857 when Roy's map was compared with vegetational evidence in 1857, especially in its middle to upper reaches. For example, below the Boonreigh/ Leader confluence, the channel changed from "highly irregular with islands" to "straight/ sinuous" without any bars (Figure 4.8.2.(iv)). Similar trends occurred on the Bowmont and Whiteadder, for example near East Blanerne (Sample 31) where more sinuous palaeofeatures were verified by Roy's map. Clearly large areas of the floodplain must either have been reworked over the previous 100 years (pre-1850) or artificially constrained. The existence of Roman antiquities on the mainstream Tweed floodplain (O.S. 1:10650 First Edition) suggested that major planform change and floodplain reworking had not taken place over a much longer timespan. Minor changes from the old county boundary (which followed the middle Tweed at that time) occurred locally.

Finally, in terms of map-based information concerning "liability to flooding", a few areas were noted as particularly susceptible to frequent inundation, with no areas noted in the third edition, 2 on the second and 5 on the first. Reaches prone to flooding around 1870 were the Bowmont Water near Kirk Yetholm and on the lower Whiteadder haughs such as the Whiteadder near Prestonhall (Samples 72 and 30). Whether this can be interpreted as a reduction in frequency of flooding at sensitive sites, over the map study period, must be assessed later (Chapter 5). As stated in earlier sections, one has of course to be careful of changing cartographic conventions.

4.8.3 Temporal variation in channel pattern

4.8.3.1 Modes of channel adjustment

Categories of channel pattern change studied in the Tweed area were generally less dramatic than in the other two study areas (Table 4.8.3.(i)) and occurred locally where the strength of controls appeared to be relaxed. Such a control would be the localised removal of bedrock constraints on the lower Whiteadder, allowing the more extensive development of haughs. Alternatively, it may relate to locally increased sediment supply. Examples of the different types of planform adjustment are provided from the digitised map comparisons (Figure 4.8.3.(i)).

Channel changes, involving an increase in active area (Table 4.8.3.(i), Section A) were present but in fewer numbers in comparison with Speyside and Deeside. In contrast with the other two study areas, there were no examples of major channel metamorphosis involving an increase in channel active area, although there were several incidences of localised extra-channel avulsion eg. the lower Leader (Sample 10) and Whiteadder (Sample 24; Figure 4.6.3.(i)).

In general, reduction of active area took place at a more localised and smaller scale than in Speyside (Table 4.8.3.(i), Section B). However, an exception to that rule occurred on the Bowmont where at Sample 72, a definite change in channel character occurred between the first and second editions (1858-1899), associated with a major reduction in active area (Figure 4.8.3.(i)). There were several other examples on

Table 4.8.3.(i)

Categories of channel planform change within the Tweed study area

(A) Changes associated with an increase in width of active area

1: Complex extra-channel avulsion

2: Localised extra-channel avulsion

3: Local channel widening

(B) Changes associated with a reduction in width of active area

1: Channel metamorphosis

2: Reduction from complex to single channel

3: Reduction from split to single channel

4: Localised reduction from a split to single channel.

5: Breached partially confined upper meander bends

6: Cutoff through irregular meanders (a) neck (b) chute

(C) Changes associated with an equilibrium state

1: Slow lateral shift across active area

Figure 4.8.3.(i)

Examples of categories of channel pattern change sampled within the Tweed study area

Sample 24

Sample 10

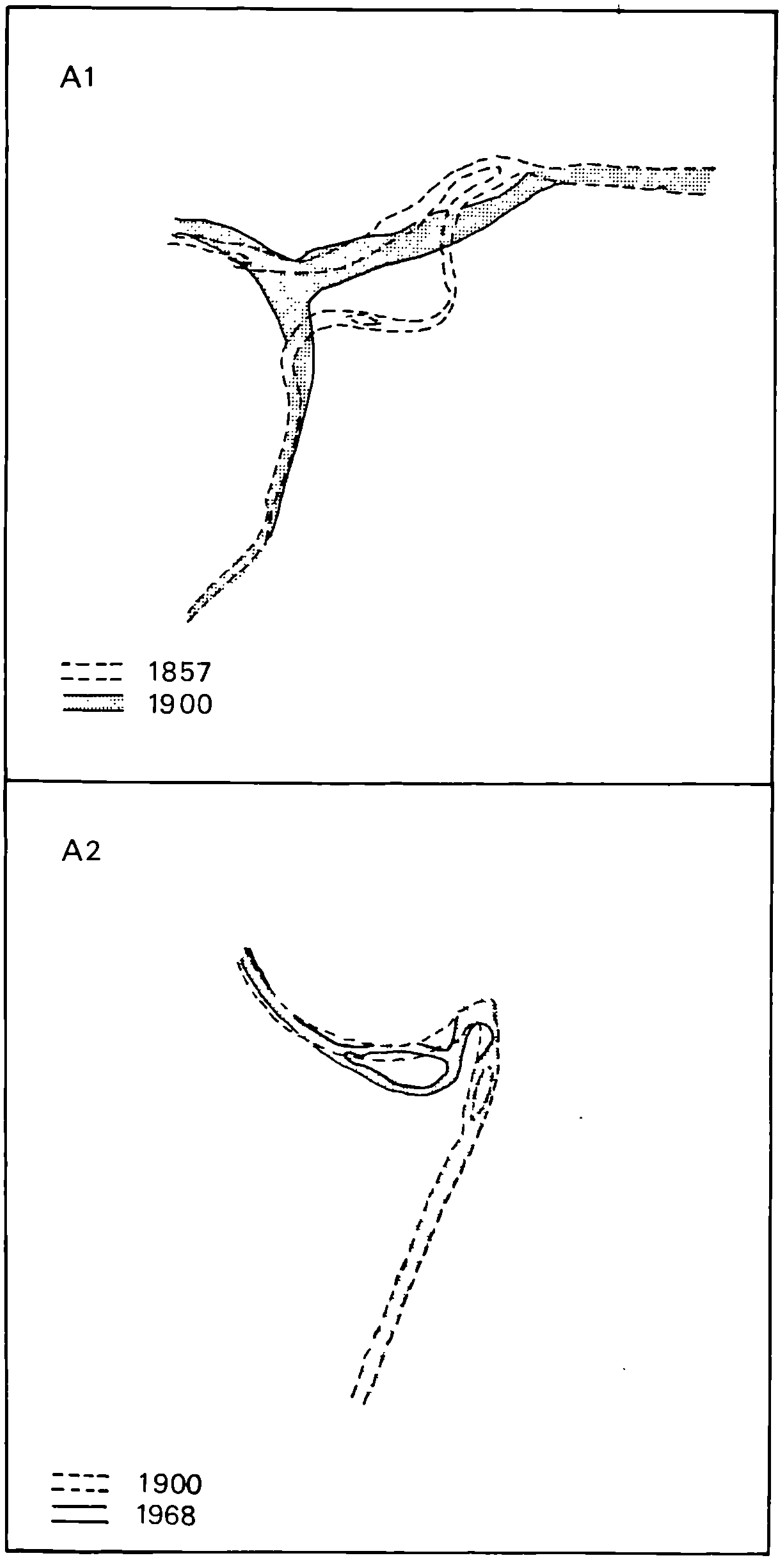
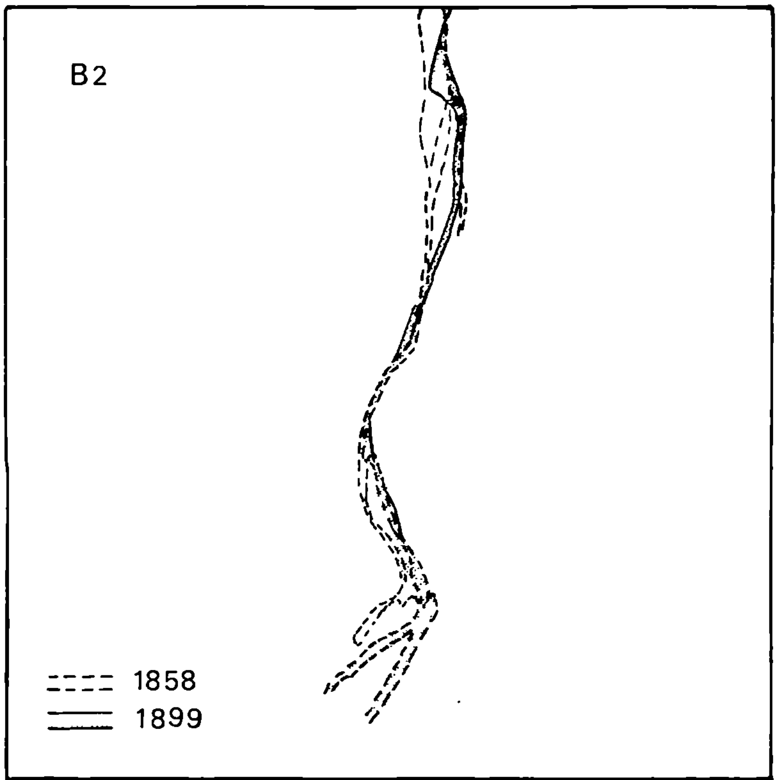


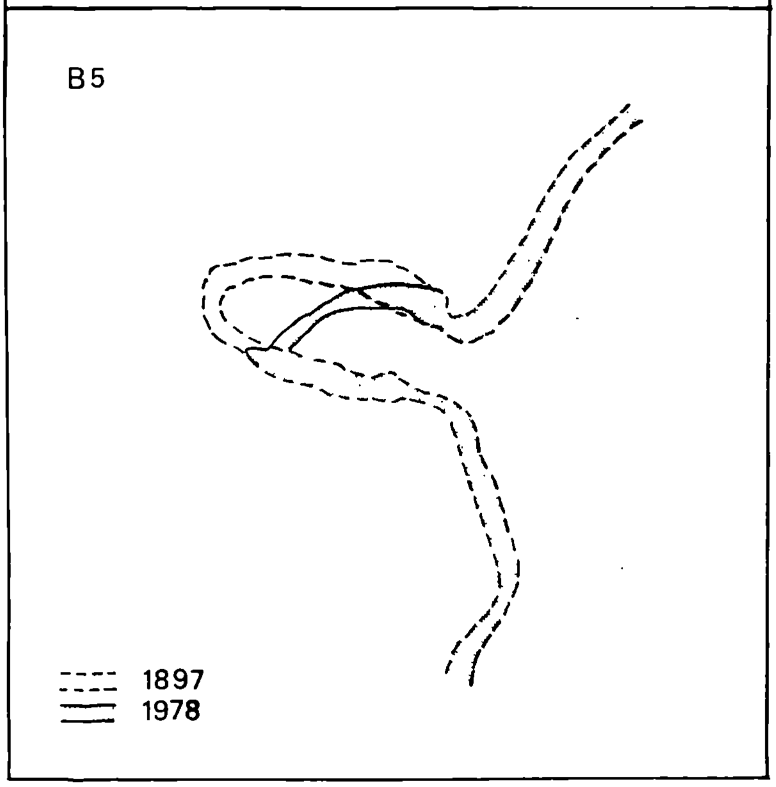
Figure 4.8.3.(i) cont.

Examples of categories of channel pattern change sampled within the Tweed study area

Sample 74



Sample 41



the Bowmont of reductions from a complex or split channel to a single channel (eg. Sample 69). This also occurred locally on the other tributaries such as Sample 40 on the Whiteadder or Sample 59 on the Teviot. Confined meanders were occasionally breached but not necessarily at the neck, as occurred for example, between 1897-1978 on the Whiteadder (Sample 41; Figure 4.8.3.(i)). Clearly, a large magnitude event must have been required to erode through such a large quantity of floodplain material. Unconfined irregular meanders (eg. Sample 42 on the Teviot) were also occasionally cut through by chute cutoffs. Whether such large local reductions in sinuosity were natural or man-induced, as a response to excessive reworking of agricultural land, will be evaluated later.

A large number of samples fell within Section (C) (Table 4.8.3.(i)), where there was no local confinement, for example related to an absence of bed outcrops. This pattern of response involved bank erosion and opposite bank restabilisation leading to a gradual shift across the floodplain, frequently at different rates at different cross-sections.

4.8.3.2 Rates of change as indicated by sinuosity index (CHASIN)

As seen in Table 4.8.3.(ii), median and IQR values were similar for both intermap periods and when the Wilcoxon test was carried out on the total sample, there was no significant difference between map-dates. However, study of Figure 4.8.1.(i) showed a shifting from 1.20-1.29 to 1.10-1.19 between the first and third editions. Certain sites must have

Table 4.8.3.(ii)

Rates of change in sinuosity index within the Tweed study area

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
CHASIN1-2	-0.03	0.00	0.02	0.05
CHASIN2-3	-0.04	0.00	0.02	0.06
AVSIN1-2	-0.0007	0.00	0.0006	0.0013
AVSIN2-3	-0.0005	0.00	0.0003	0.0008
Leader				
CHASIN1-2	-0.02	0.002	0.04	0.06
CHASIN2-3	-0.05	0.008	0.04	0.09
AVSIN1-2	-0.0005	0.00	0.001	0.0015
AVSIN2-3	-0.0007	0.0001	0.0005	0.0012
Whiteadder				
CHASIN1-2	-0.02	0.01	0.01	0.03
CHASIN2-3	-0.06	-0.01	0.01	0.07
AVSIN1-2	-0.0005	0.00	0.0003	0.0008
AVSIN2-3	-0.0007	0.00	0.0001	0.0008

Table 4.8.3.(ii) cont.

Parameter	25th Percentile	Median	75th Percentile	IQR
Teviot				
CHASIN1-2	-0.04	-0.01	0.01	0.05
CHASIN2-3	-0.02	0.00	0.02	0.04
AVSIN1-2	-0.0008	0.00	0.0003	0.0011
AVSIN2-3	-0.0003	0.00	0.0003	0.0006
Bowmont				
CHASIN1-2	-0.13	-0.06	0.01	0.14
CHASIN2-3	-0.04	0.10	0.06	0.10
AVSIN1-2	-0.0034	-0.001	0.0002	0.0036
AVSIN2-3	-0.0006	0.00	0.001	0.0016
Tweed				
CHASIN1-2	-0.03	0.0	0.01	0.04
CHASIN2-3	-0.03	0.0	0.02	0.02
AVSIN1-2	-0.0008	0.0	0.0002	0.0010
AVSIN2-3	-0.0004	0.0	0.0003	0.0007

decreased in sinuosity since the first edition, and whether this was related to natural stabilisation or channel training must be evaluated.

The basic statistical parameters for both total sinuosity change (CHASIN) and average annual sinuosity change (AVSIN), are shown in Table 4.8.3.(ii). The IQR rates for total sinuosity change were larger between the first and second editions (0.0013 yr^{-1}) than between the second and third editions (0.0008 yr^{-1}). This was typical of all catchments except the Whiteadder. The total changes for the individual sample points were plotted spatially, within the catchments, in Figures 4.8.3.(ii) to (vi). Data were also plotted against time, sub-divided by catchment (Figure 4.8.3.(vii) to (xi)). As seen in Figure 4.8.3.(x), the Bowmont underwent the largest reduction in sinuosity between the first and second editions, while there was a further increase at some sites with the third edition. For example, Sample 72 underwent a reduction from 1.20-1.06 between 1858 and 1896 and an increase from 1.06-1.24 by 1977. Whether this initial reduction was caused by channelisation or stabilisation due reduction in flood frequency between the second and third editions, will be assessed in Chapters 5, 6 and 9.

The other catchments also underwent changes in sinuosity though the trends were more variable. In aggregate, individual catchments showed only weak trends but in contrast, selected reaches displayed major changes in sinuosity. For example, the Whiteadder had several reaches (Figures 4.8.3.(iii) and (viii) that increased in sinuosity between the first and second edition and then reduced between the second and third editions eg. Samples 23 and 40. Again, study of the flood history must be undertaken to assess whether pre-1900 magnitudes and frequencies may

Figure 4.8.3.(ii)

Sampled changes in sinuosity index values falling within the Leader catchment for the two inter map periods

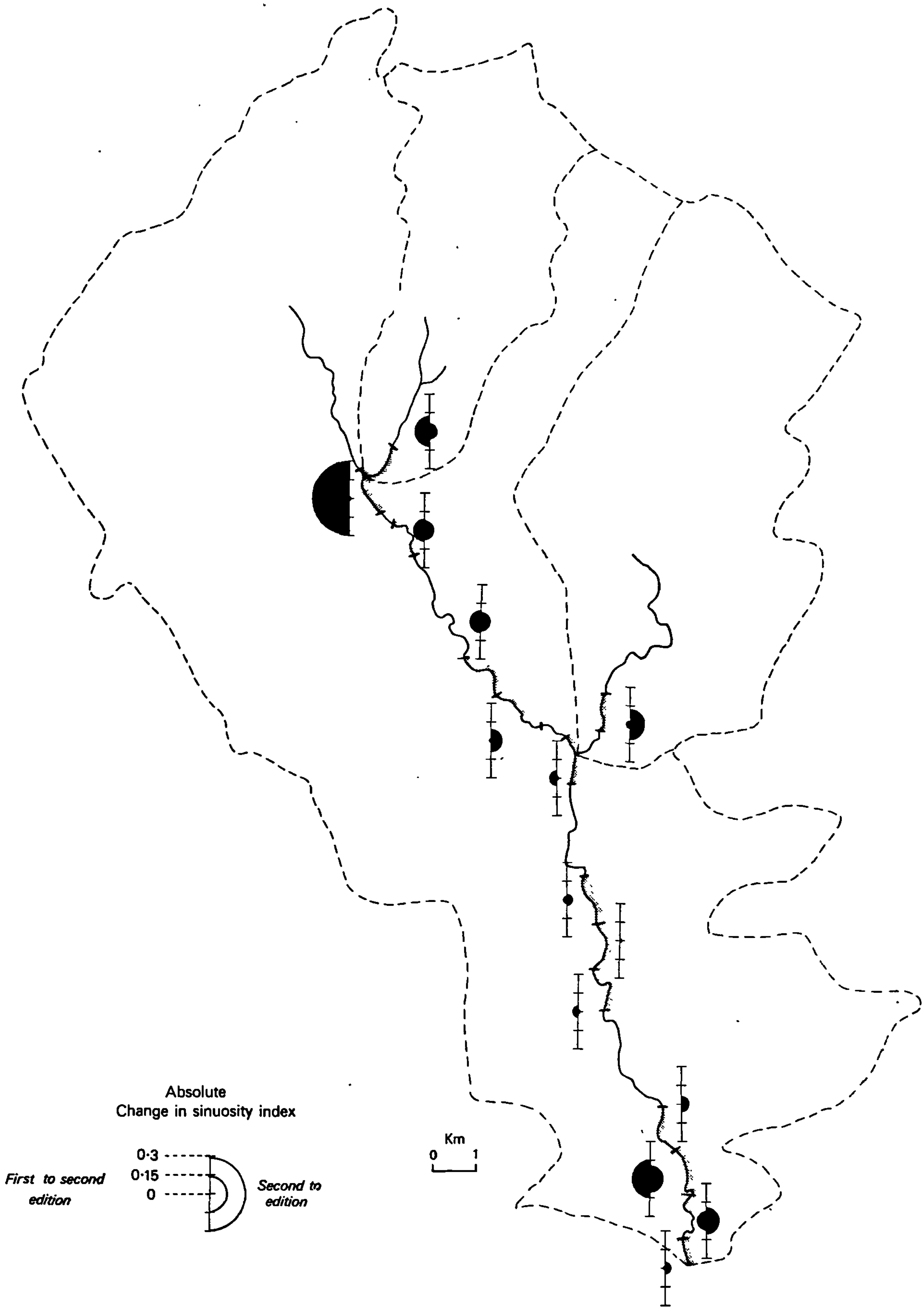
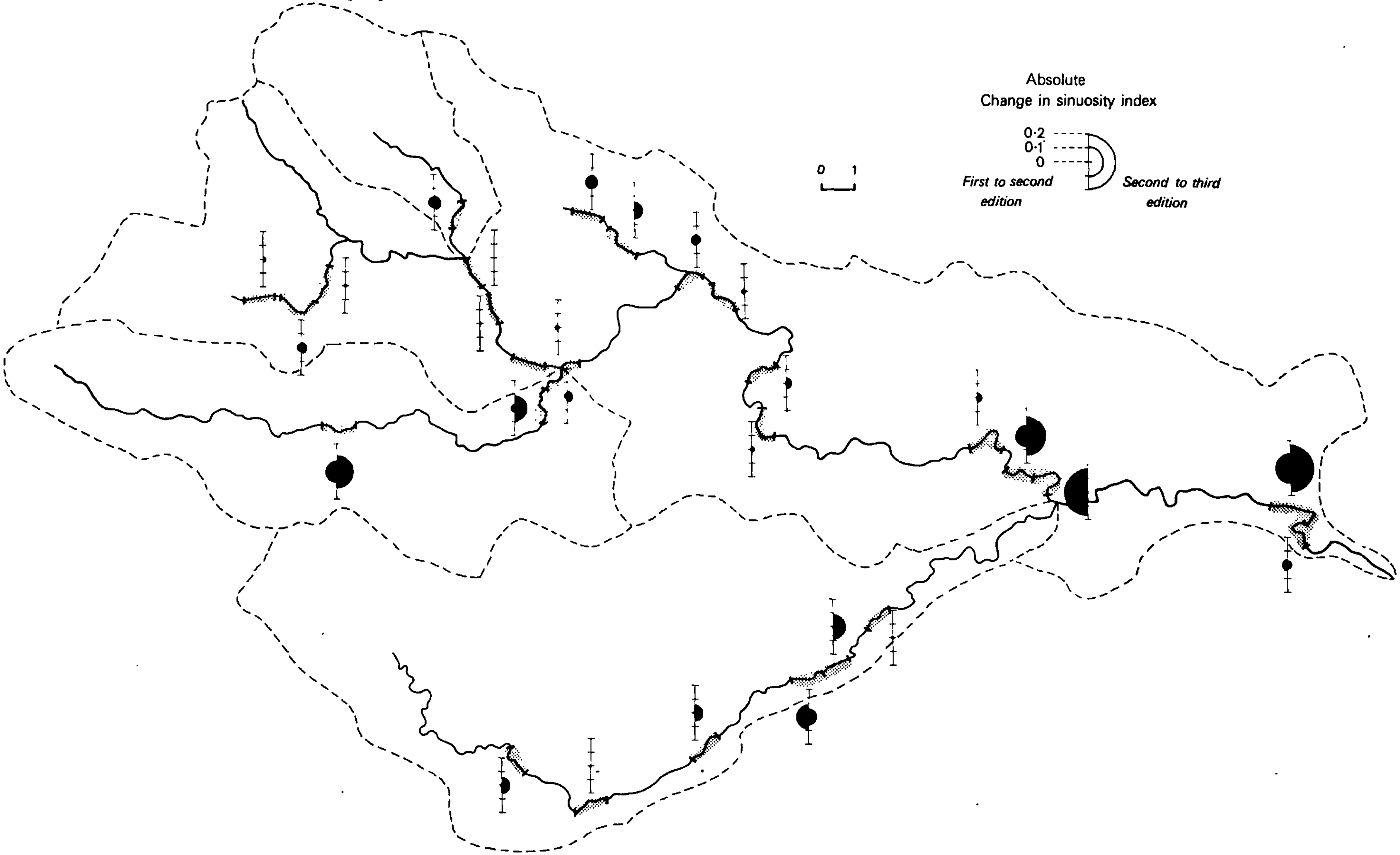


Figure 4.8.3.(iii) **Sampled changes in sinuosity index falling within the Whiteadder catchment during the two intermap periods**



Sampled change in sinuosity index values within the Teviot catchment during the two intermap periods

Figure 4.8.3.(iv)

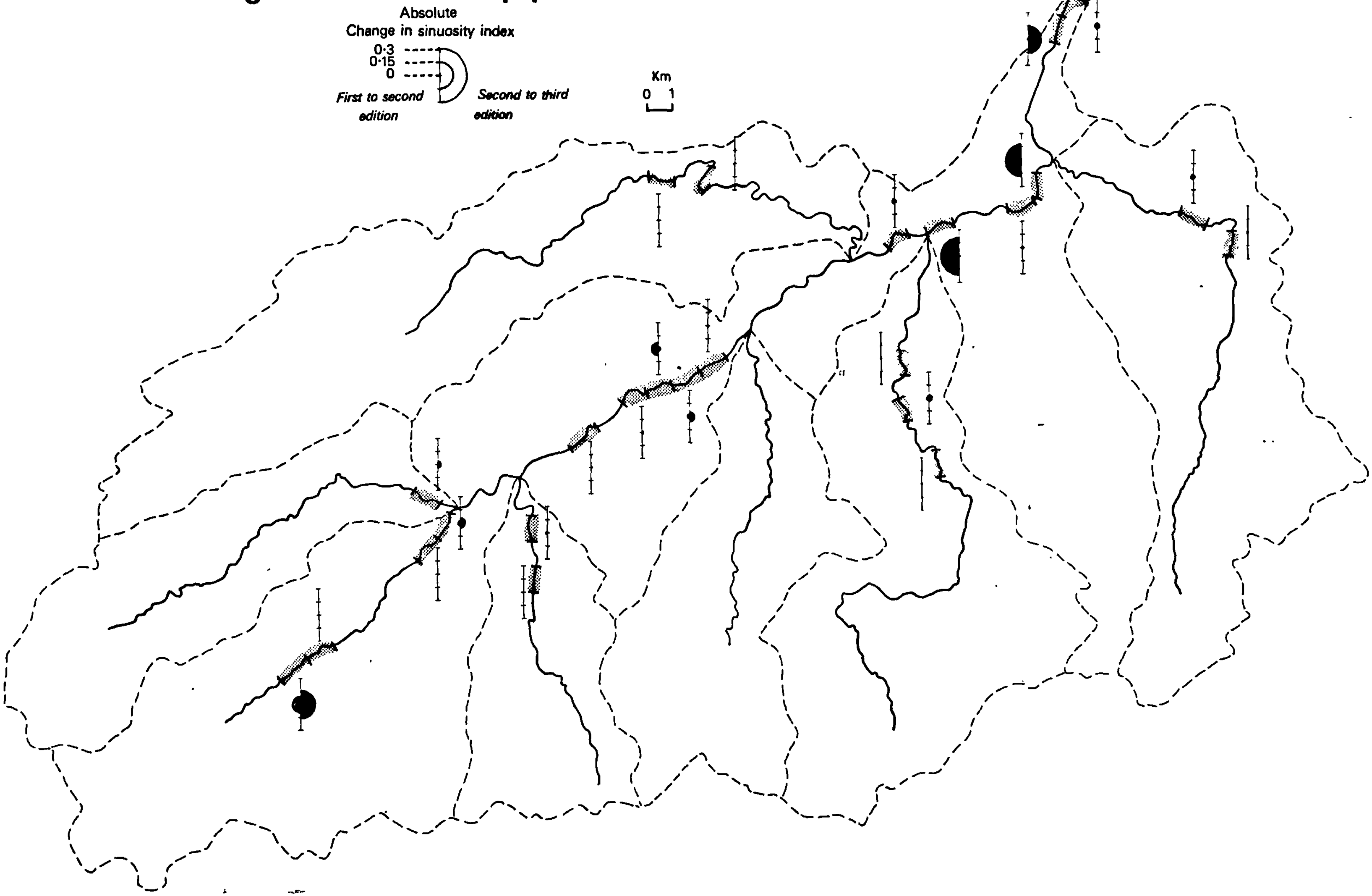


Figure 4.8.3.(v)

Sampled change in sinuosity index falling within the Bowmont catchment during the two intermap periods

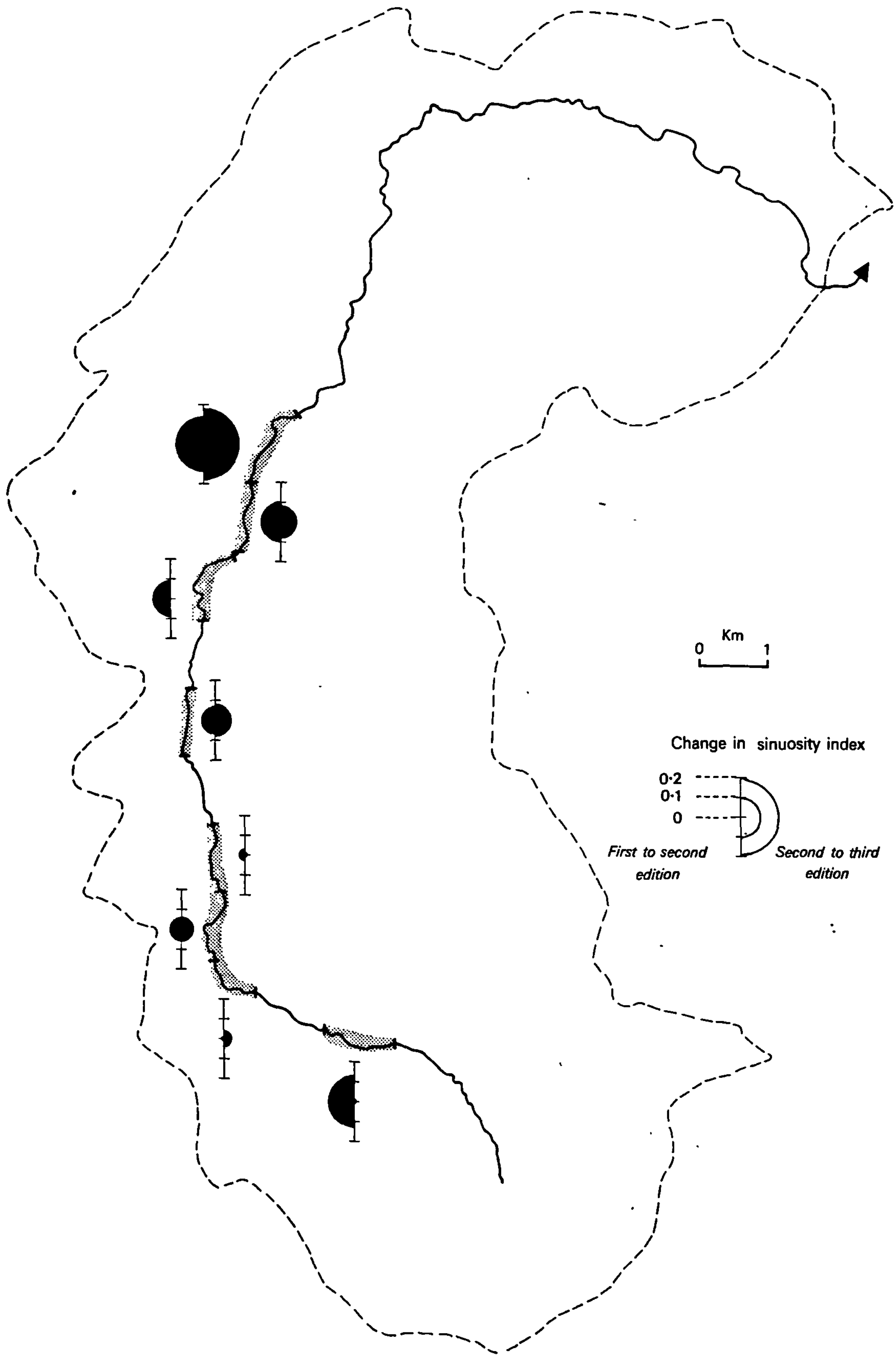


Figure 4.8.3.(xii)

Sampled change in sinuosity index falling on the mainstream Tweed during the two intermap periods

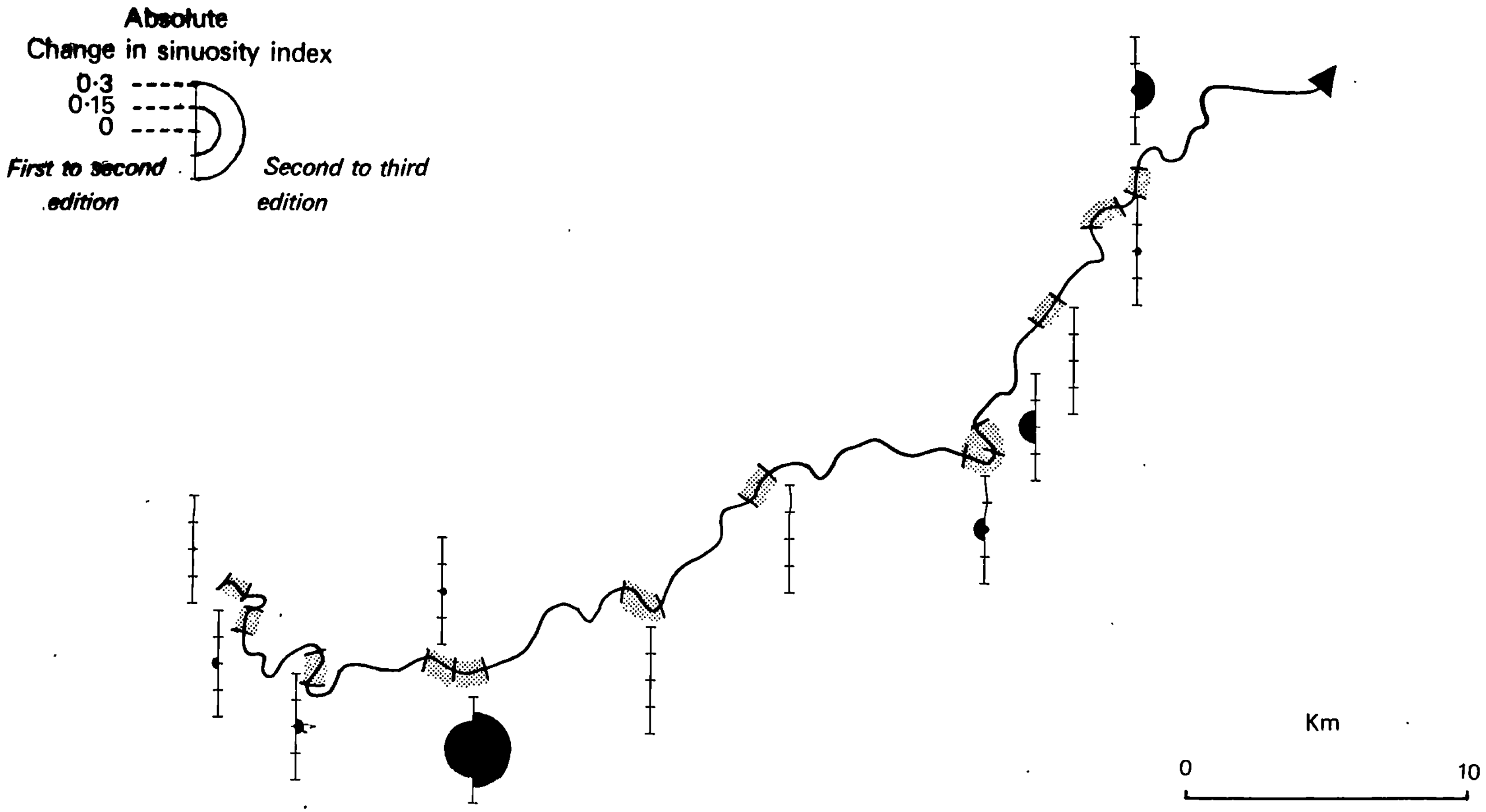


Figure 4.8.3.(vii)

LEADER:SINUOSITY INDEX CHANGE

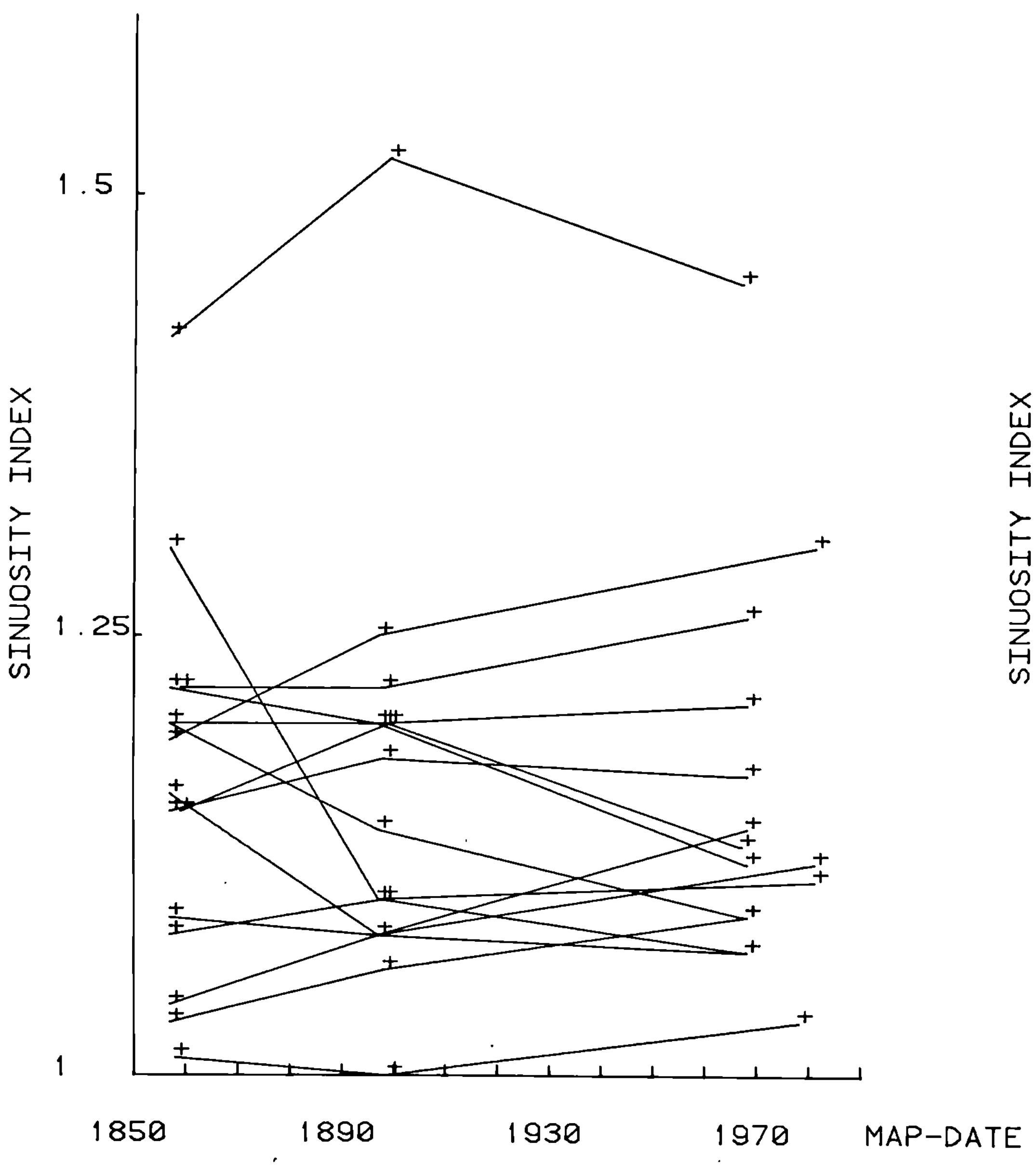


Figure 4.8.3.(viii)

WHITEADDER:SINUOSITY INDEX CHANGE

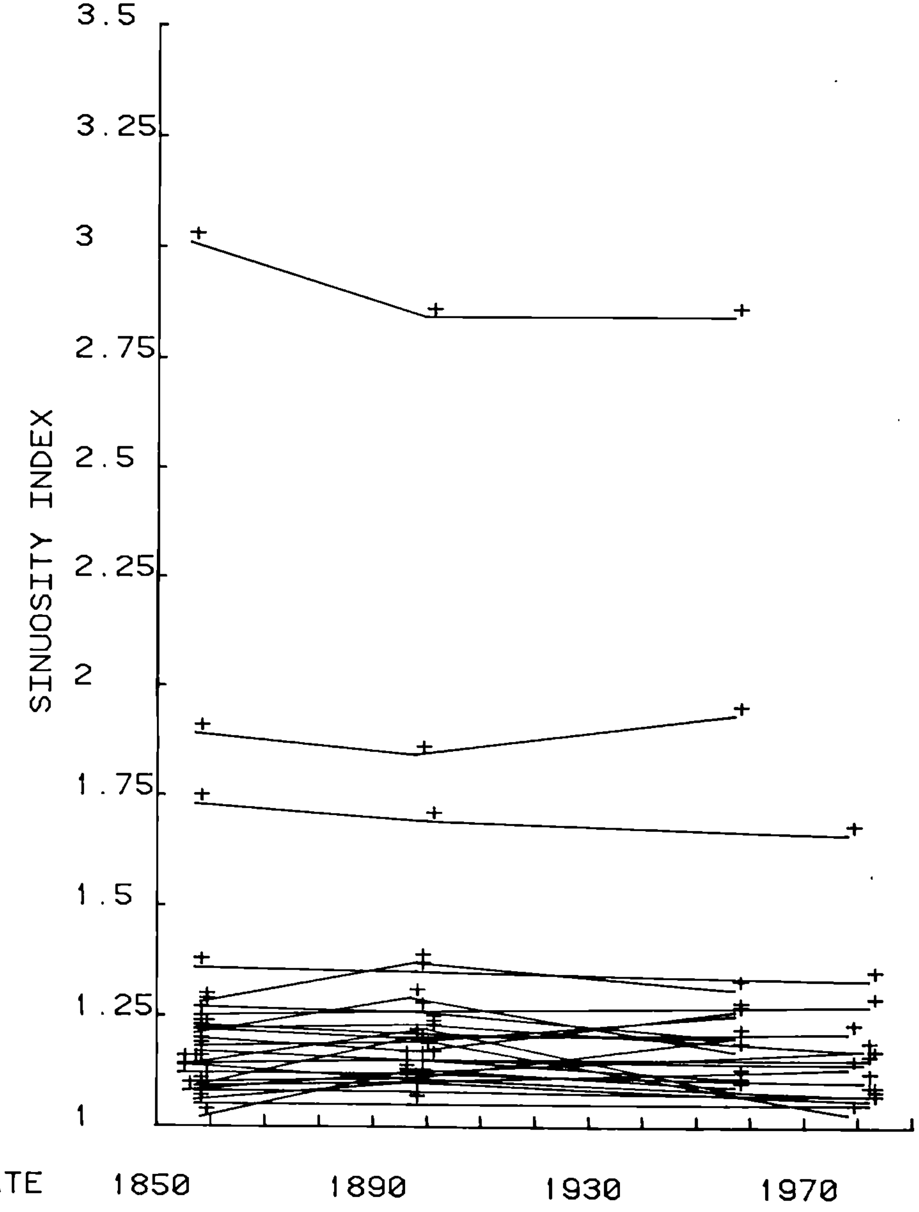


Figure 4.8.3.(ix)

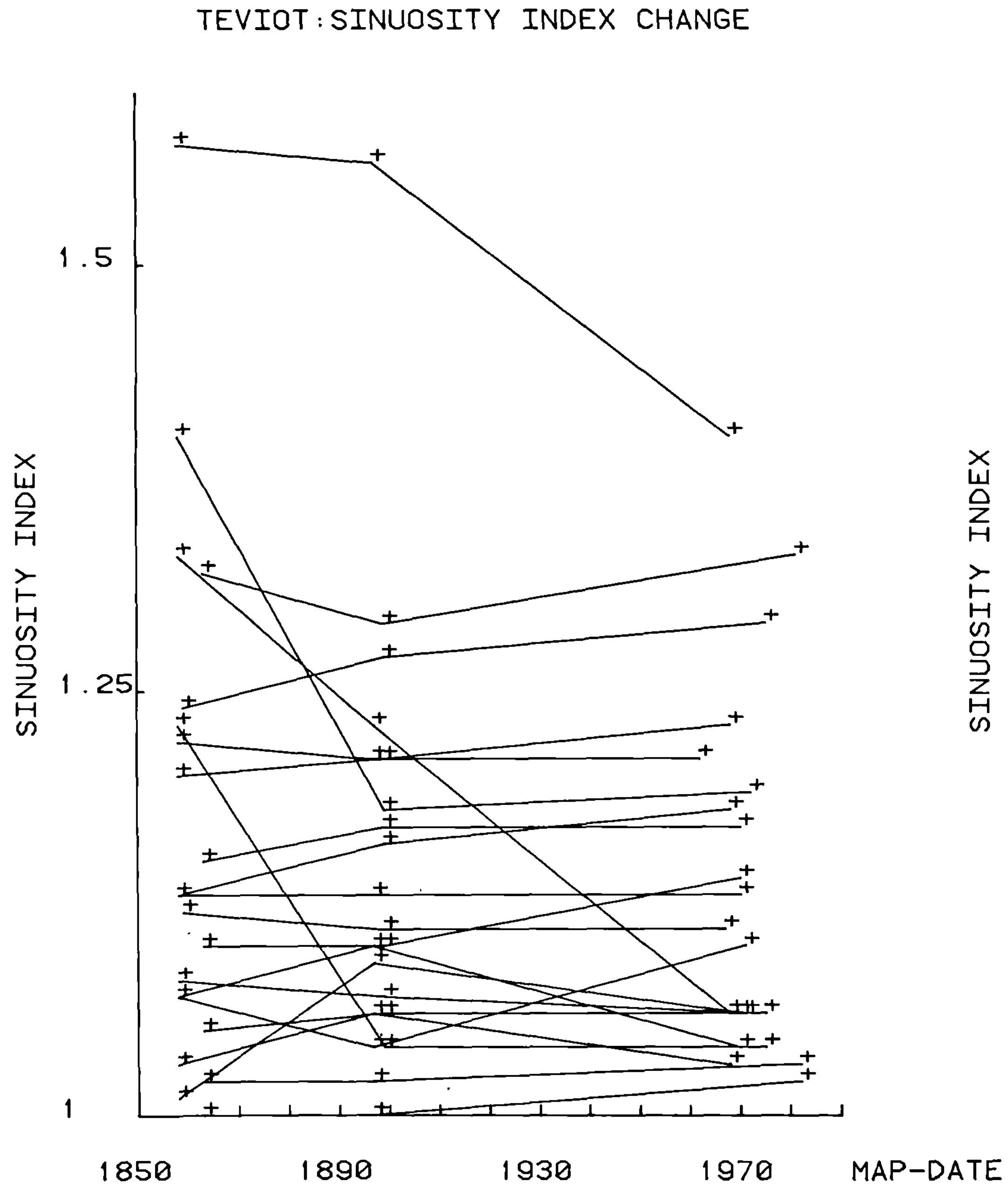


Figure 4.8.3.(x)

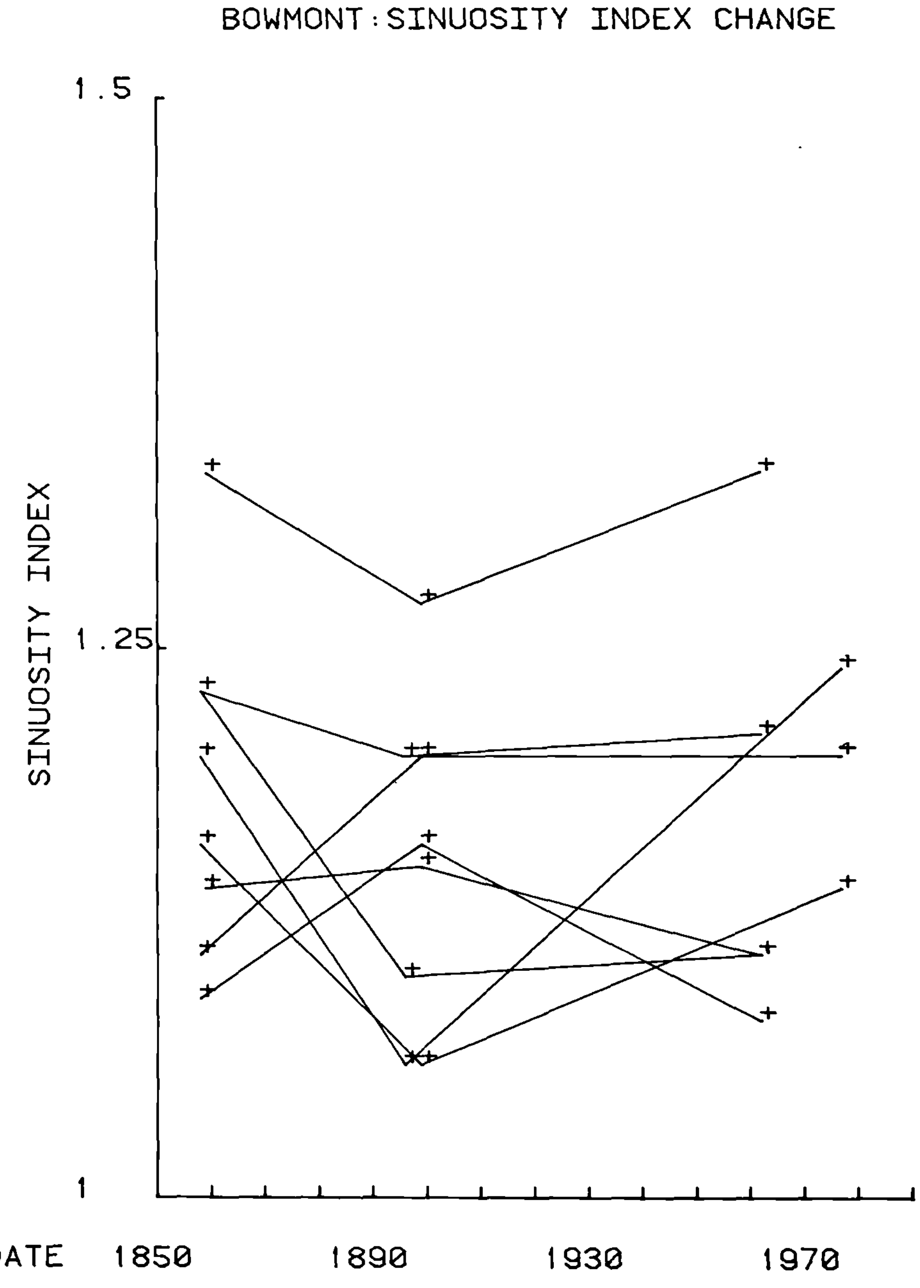
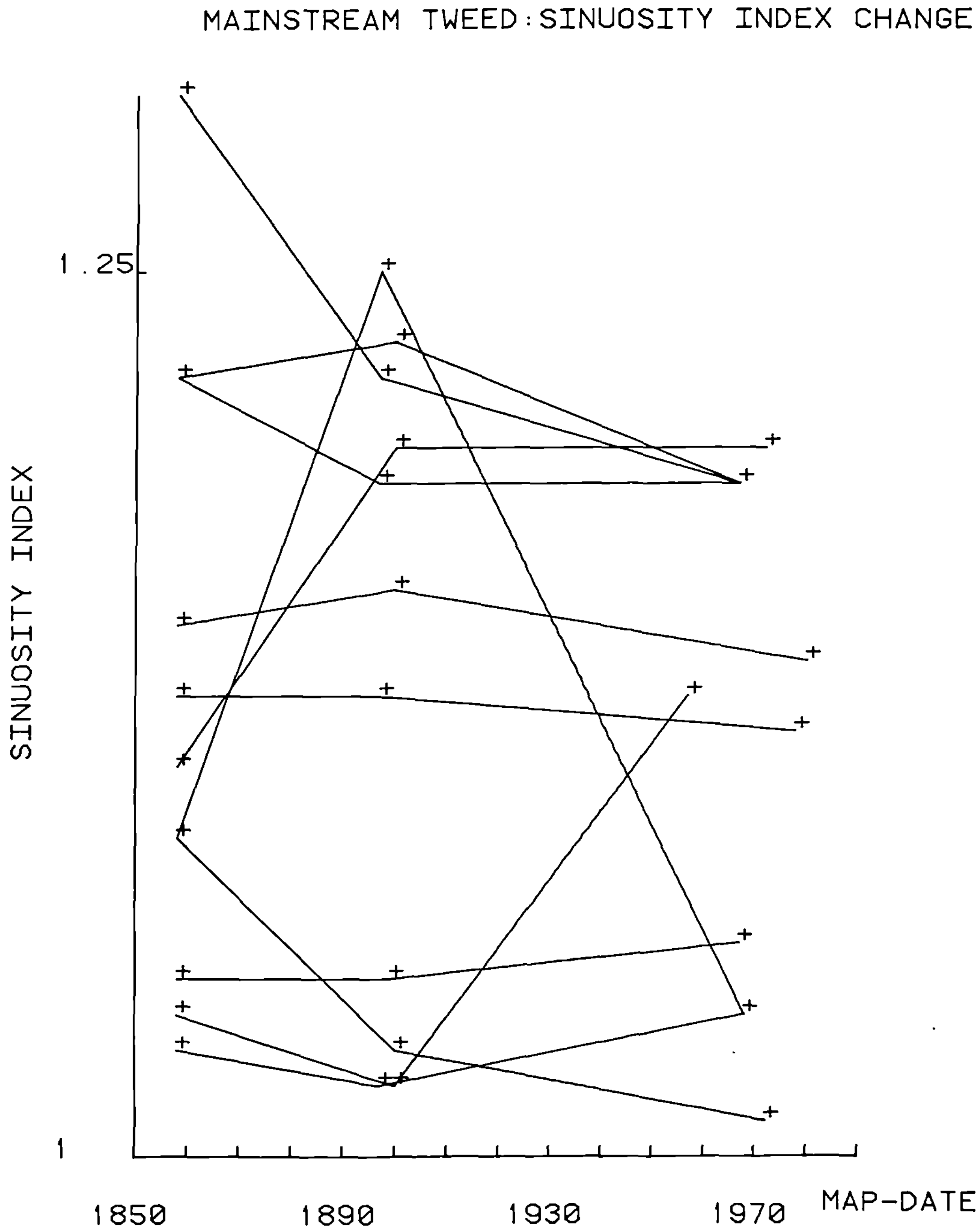


Figure 4.8.3.(xi)



have locally disrupted the system. It is noteworthy that a large number of sites were highly stable and underwent little or no change. Different reaches, depending on their former relationship to an equilibrium condition, must react in different ways to a similar stress to the system (with the assumption that reaches within individual catchments will have had similar flood frequencies). For example, there was a reduction from 1.40-1.18 on middle Teviot between 1858-1899 (Sample 58; see Section 7.4.6; Tweed study reach 7) and from 1.33-1.23-1.06 from 1858-1897-1968 at Sample 42 (Tweed study reach 6; Section 7.4.7) on the upper Teviot. Both examples represented cutting through of irregular meander bends. In contrast to these two reaches on the Teviot, the majority of the remaining samples were comparatively stable. It was notable that on the Teviot all major changes in sinuosity (>0.10) were reductions. Similar examples can be highlighted in the other catchments.

The importance of initial sinuosity was then assessed. Although when AVSIN2-3 was correlated with SIN2 for the total sample, no significant correlation was obtained when the data were subdivided by catchment, the Leader (r=-0.47) and Teviot (r=-0.53) did have significant negative correlations (Table 4.8.3.(iii)). Both these catchments had high reductions in sinuosity, occurring on more sinuous reaches within their upper to middle catchments (Figure 4.8.3.(vii) and (ix)), though the pattern was not strong for the whole of the study area. However, when AVSIN1-2 (total sample) was correlated with SIN 1, a highly significant correlation of -0.38 was attained. Thus, high reductions in sinuosity between 1858 and 1900 were associated with high initial sinuosity in the 1850s for the total sample.

Table 4.8.3.(iii)

The correlation (r) between subsequent sinuosity change and initial sinuosity index

Where the absolute change provided a better correlation, it is also tabulated.

		SIN 1	SIN 2
Total sample			
	AVSIN1-2	-0.38	-----
	AVSIN2-3	-----	-0.10
Leader			
	AVSIN2-3	-----	-0.53
	AVSIN1-2	-0.14	-----
	ABAVSIN2-3	-----	0.48
	ABAVSIN1-2	0.54	-----
Whiteadder			
	AVSIN2-3	-----	-----
	AVSIN1-2	-0.73	-----

Table 4.8.3.(iii) cont.

		SIN 1	SIN 2
Teviot			
	AVSIN2-3	-----	-0.53
	AVSIN1-2	0.51	-----
	ABAVSIN2-3	-----	0.59
	ABAVSIN1-2	0.38	-----
Bowmont			
	AVSIN2-3	-----	-0.34
	AVSIN1-2	-0.60	-----
Tweed			
	AVSIN2-3	-----	-0.70
	AVSIN1-2	-0.32	-----

4.8.3.3 Rates of change as indicated by braiding index (BRAID)

Median and IQR values for BI were similar at all three map-dates for the total sample and when the Wilcoxon test was carried out on the total data set, there was no significant difference between pairs of BI values at different map-dates. However, when the data was subdivided between tributaries, the Bowmont catchment was found to have a highly significant difference between BI2 and BI3 (Z= -2.52). Obviously, a major geomorphic change had taken place between 1900 and 1977.

To assess the spatial distribution of change within the total sample, absolute change in BI values were mapped within the drainage basin (Figures 4.8.3.(xii) to (xvi)) and each sample was also plotted against time (Figures 4.8.3.(xvii) to (xxi)). The summary statistics for change are provided in Table 4.8.3.(iv). Beginning with the spatial distribution of most extensive change, it was again clear that major changes had taken place on the Bowmont water, for example between 1858 and 1962, when there was a 0.42 reduction in BI at Sample 69 (Figure 4.8.3.(xx)). Although before 1900, there was one reach which dramatically increased its BI, the median trend was a reduction. After 1900 however, there was a considerable reduction in BI at all Bowmont sample sites.

Other catchments, such as the Leader, had more localised changes in BI; major reductions occurred along some reaches between the first and second edition maps at eg. on the upper Leader (1.31-1.00 from 1857-1897; Sample 1) and on samples from the two tributary streams (Cleekhimin Burn, Sample 2; Boonreigh Water, Sample 6). All three

Table 4.8.3.(iv)

Rates of change in braiding index within the Tweed study area

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
BRAID1-2	-0.05	0.00	0.00	0.05
BRAID2-3	-0.02	0.00	0.01	0.03
AVBRAID1-2	-0.0006	0.00	0.00	0.0006
AVBRAID2-3	-0.0005	0.00	0.0003	0.0008
Leader				
BRAID1-2	-0.11	0.00	0.05	0.16
BRAID2-3	-0.05	0.00	0.05	0.10
AVBRAID1-2	-0.0027	0.00	0.0013	0.0040
AVBRAID2-3	-0.0007	0.00	0.0006	0.0013
Whiteadder				
BRAID1-2	0.00	0.00	0.01	0.01
BRAID2-3	-0.01	0.00	0.02	0.03
AVBRAID1-2	0.00	0.00	0.0001	0.0001
AVBRAID2-3	-0.0002	0.00	0.0003	0.0005

Table 4.8.3.(iv) cont.

Parameter	25th Percentile	Median	75th Percentile	IQR
Teviot				
BRAID1-2	-0.045	0.00	0.00	-0.045
BRAID2-3	0.00	0.00	0.00	0.00
AVBRAID1-2	-0.0012	0.00	0.00	-0.0012
AVBRAID2-3	0.00	0.00	0.00	0.00
Bowmont				
BRAID1-2	-0.42	-0.07	0.04	0.46
BRAID2-3	-0.35	-0.14	-0.07	0.27
AVBRAID1-2	-0.0110	-0.0017	0.0011	0.0012
AVBRAID2-3	-0.0054	-0.0018	-0.0010	0.0044
Tweed				
BRAID1-2	0.0	0.0	0.26	0.26
BRAID2-3	0.07	0.0	0.10	0.03
AVBRAID1-2	0.0	0.0	0.007	0.007
AVBRAID2-3	-0.001	0.00	0.001	0.002

Figure 4.8.3.(xii)

Sampled changes in braiding index within the Leader catchment for the three map dates

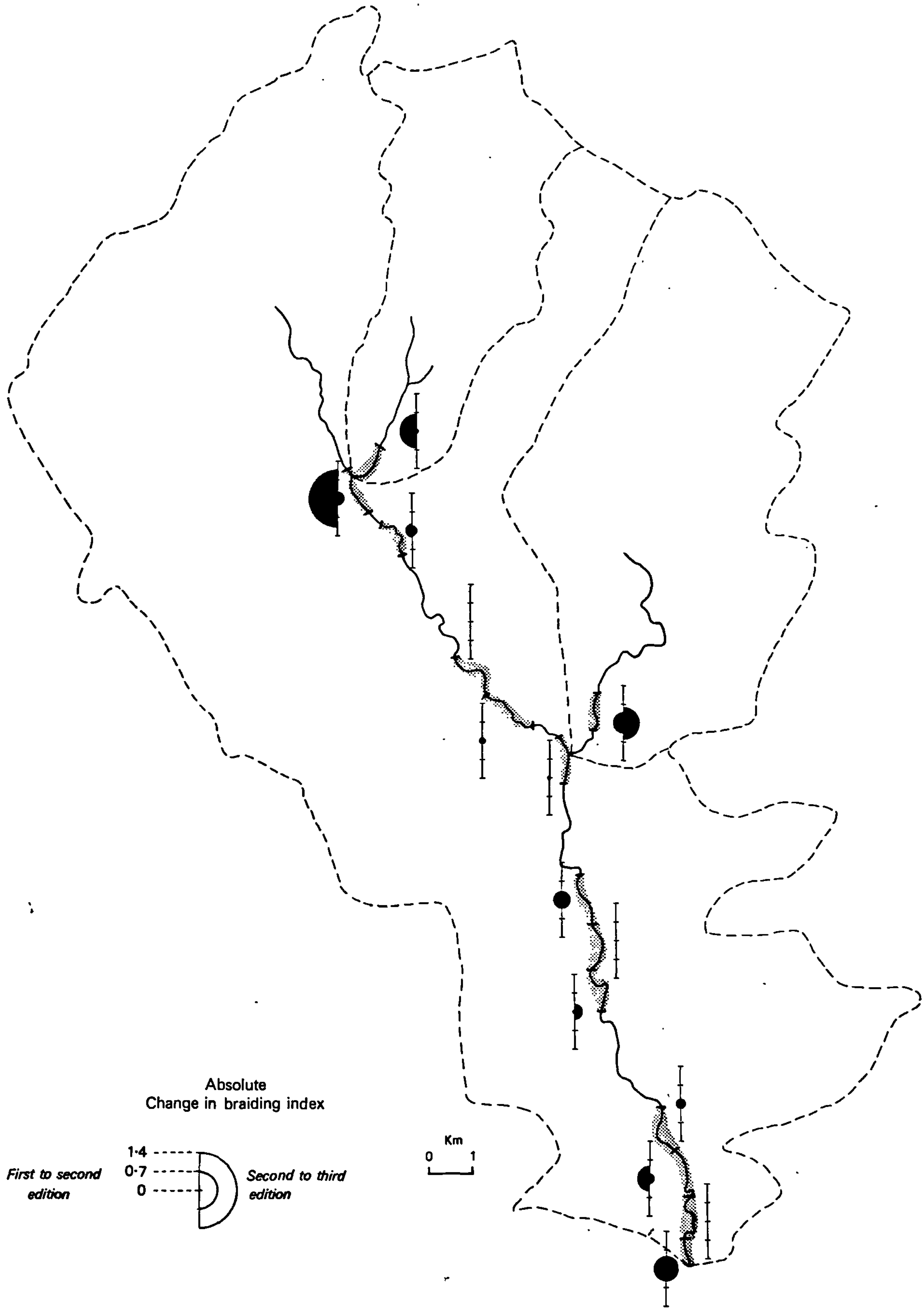
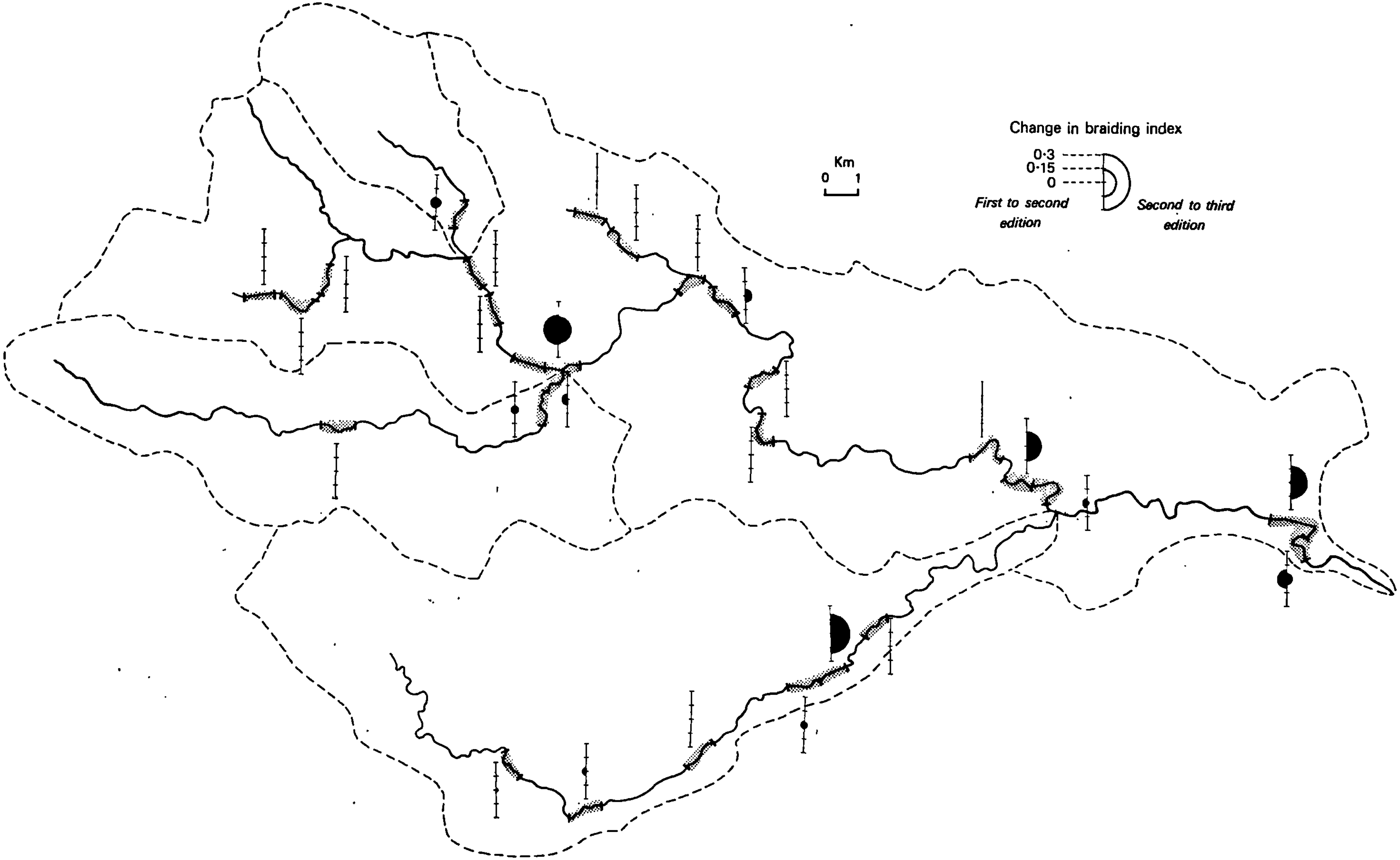


Figure 4.8.3.(xiii) Sampled changes in braiding index falling within the Whiteadder catchment during the two inter map periods



Sampled changes in braiding index values within the Teviot catchment during the two intermap periods

Figure 4.8.3.(xiv)

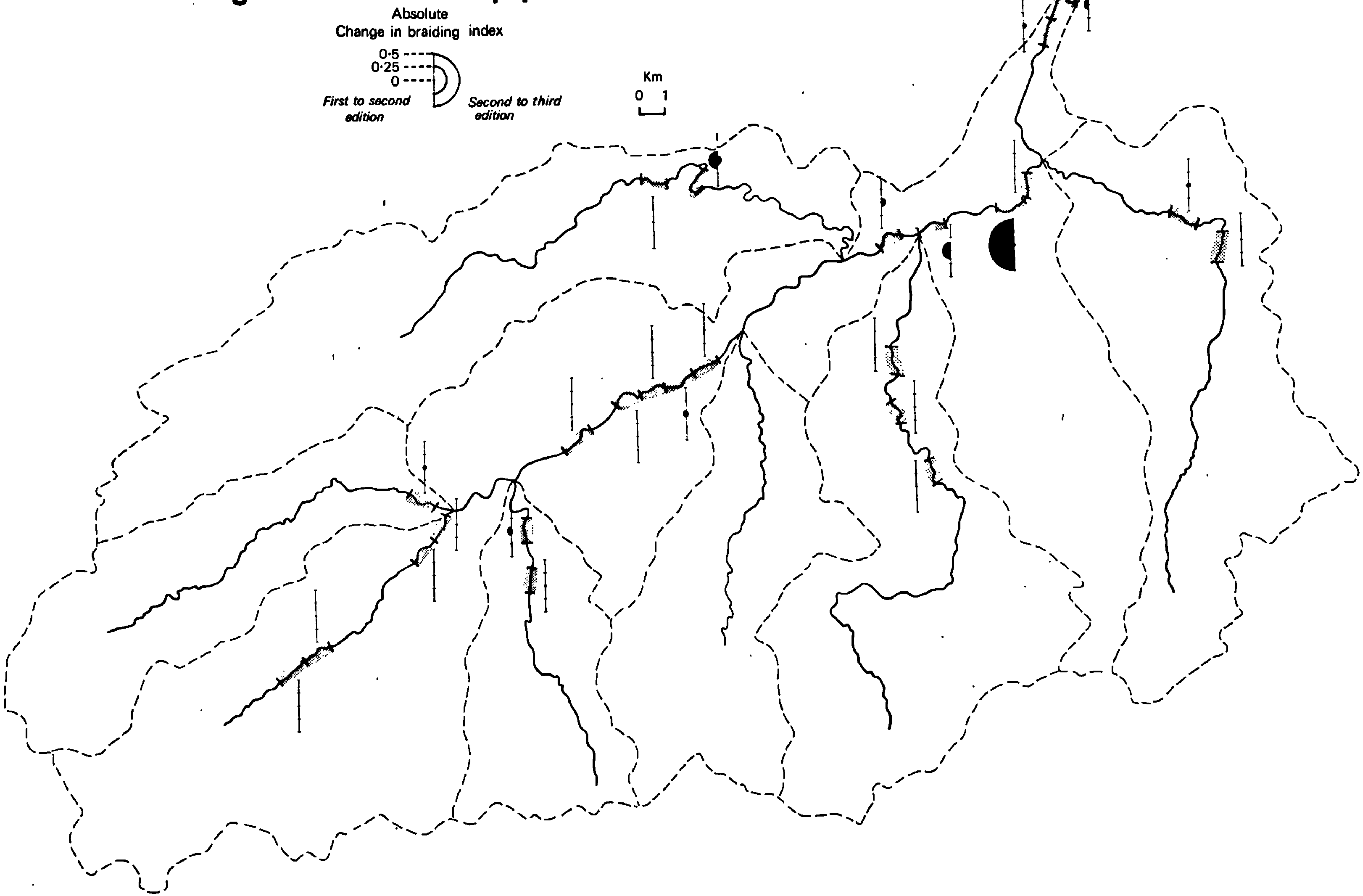


Figure 4.8.3.(xv)

Sampled changes in braiding index falling within the upper Bowmont Water catchment during the two intermap periods

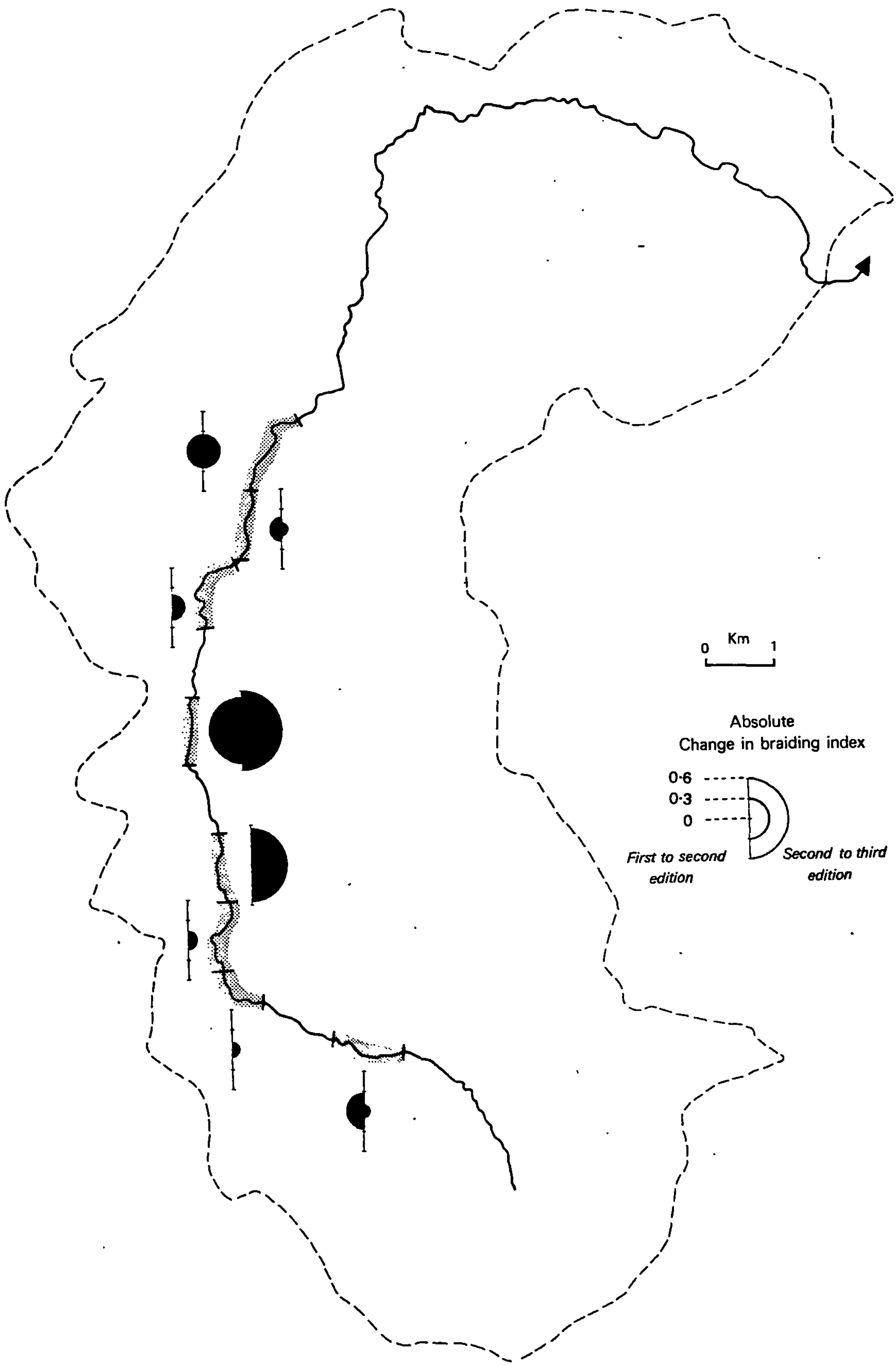


Figure 4.8.3.(xvi)

Sampled change in braiding index on the mainstream Tweed during the two inter map periods

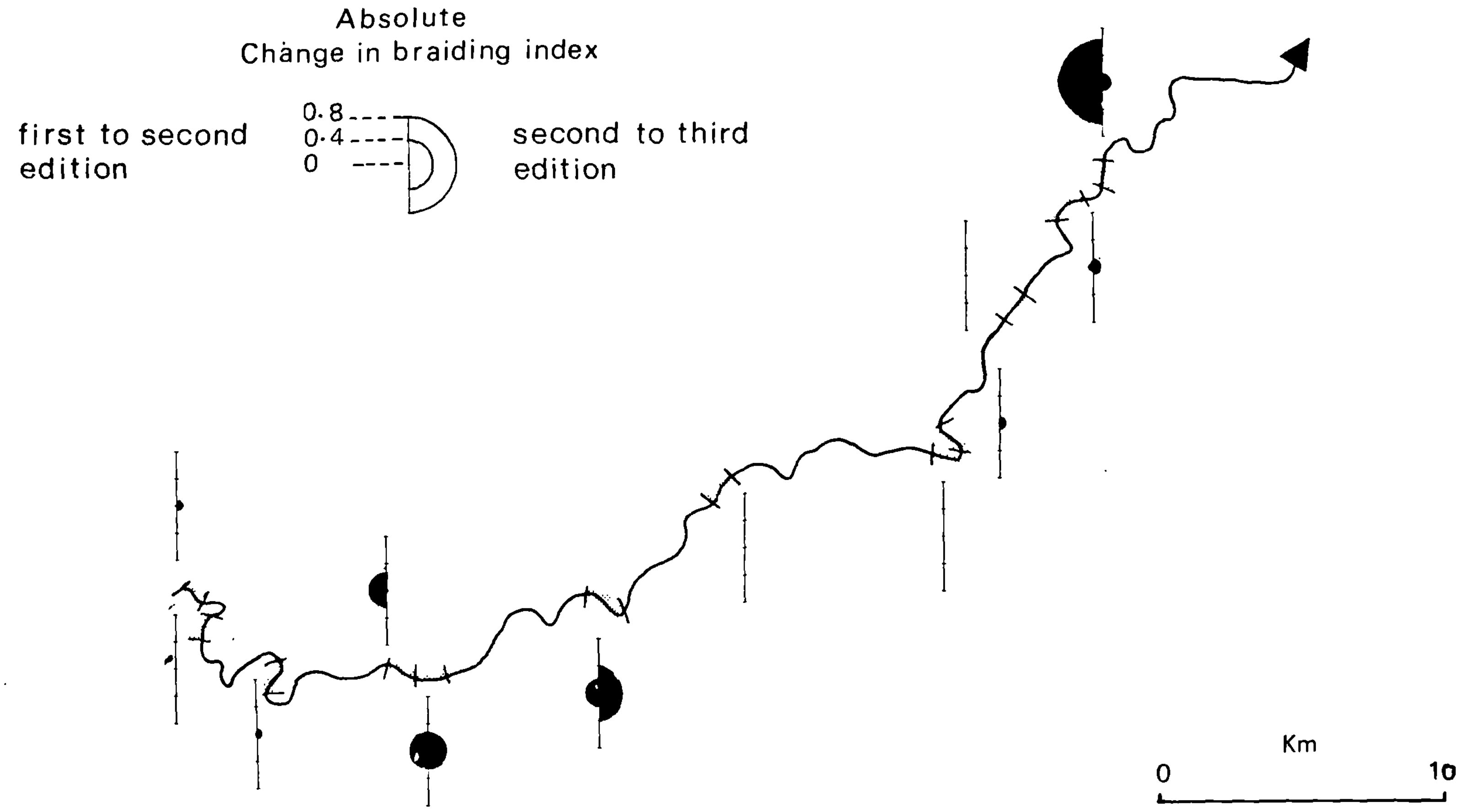


Figure 4.8.3.(xvii)

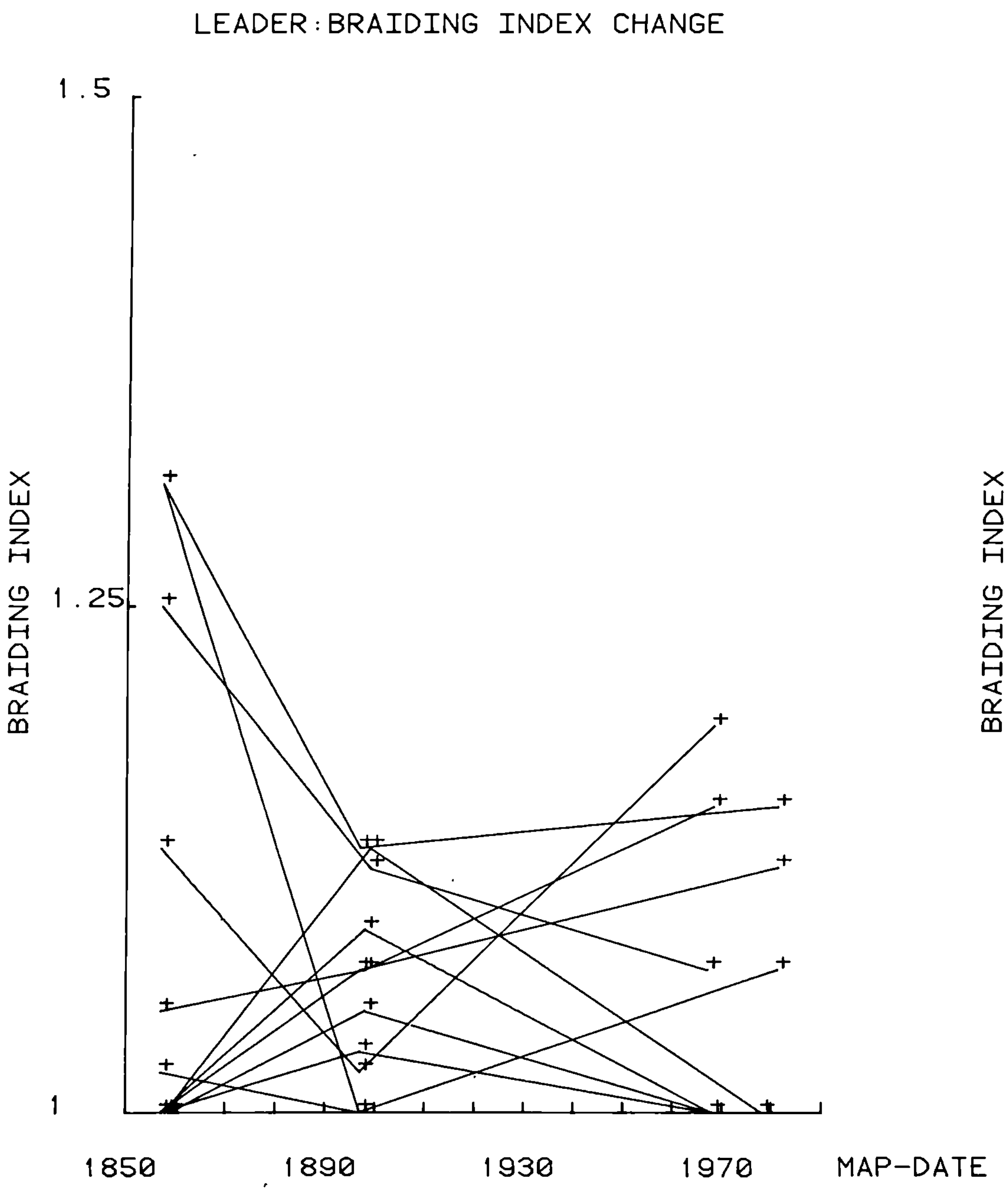


Figure 4.8.3.(xviii)

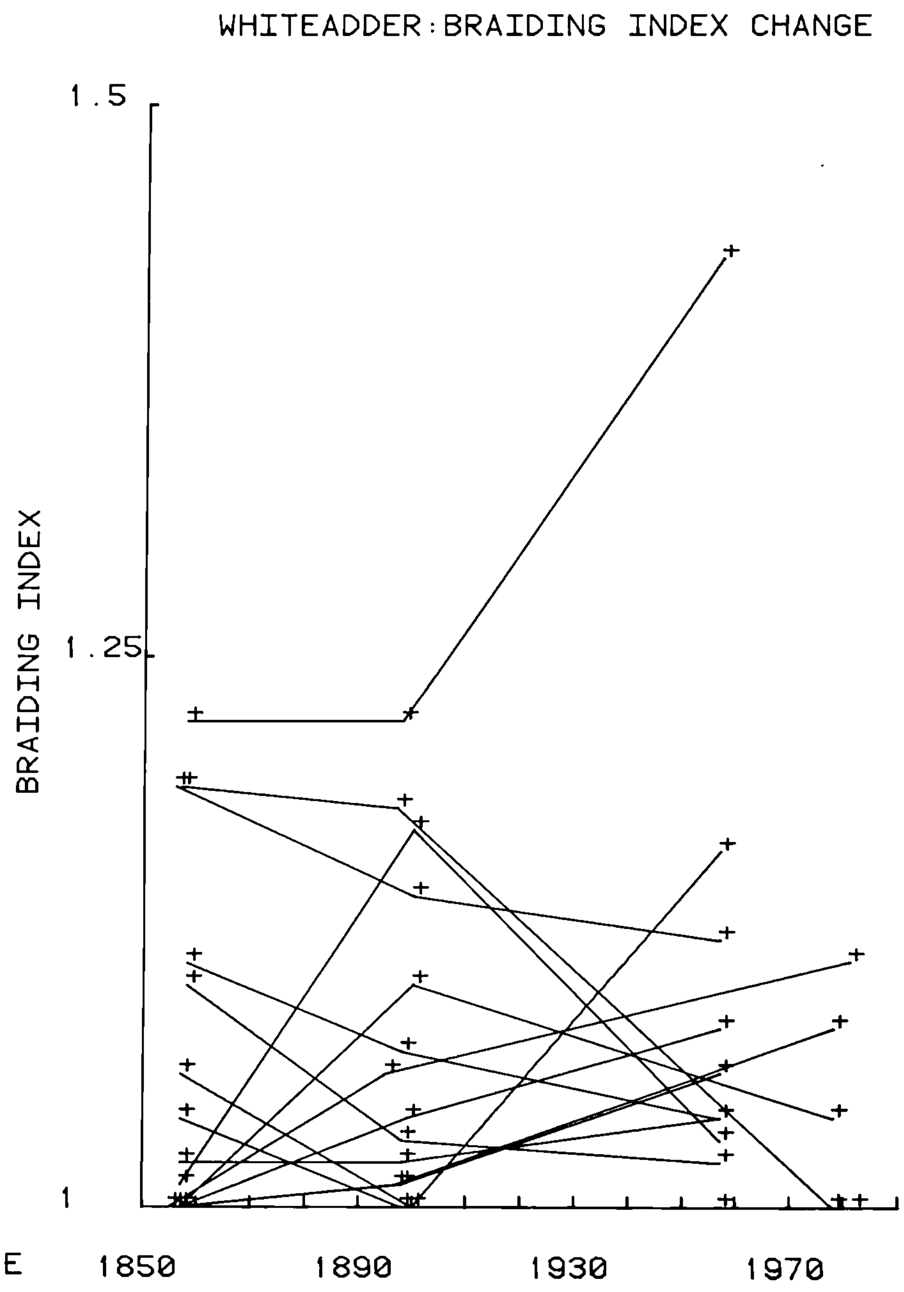


Figure 4.8.3.(xix)

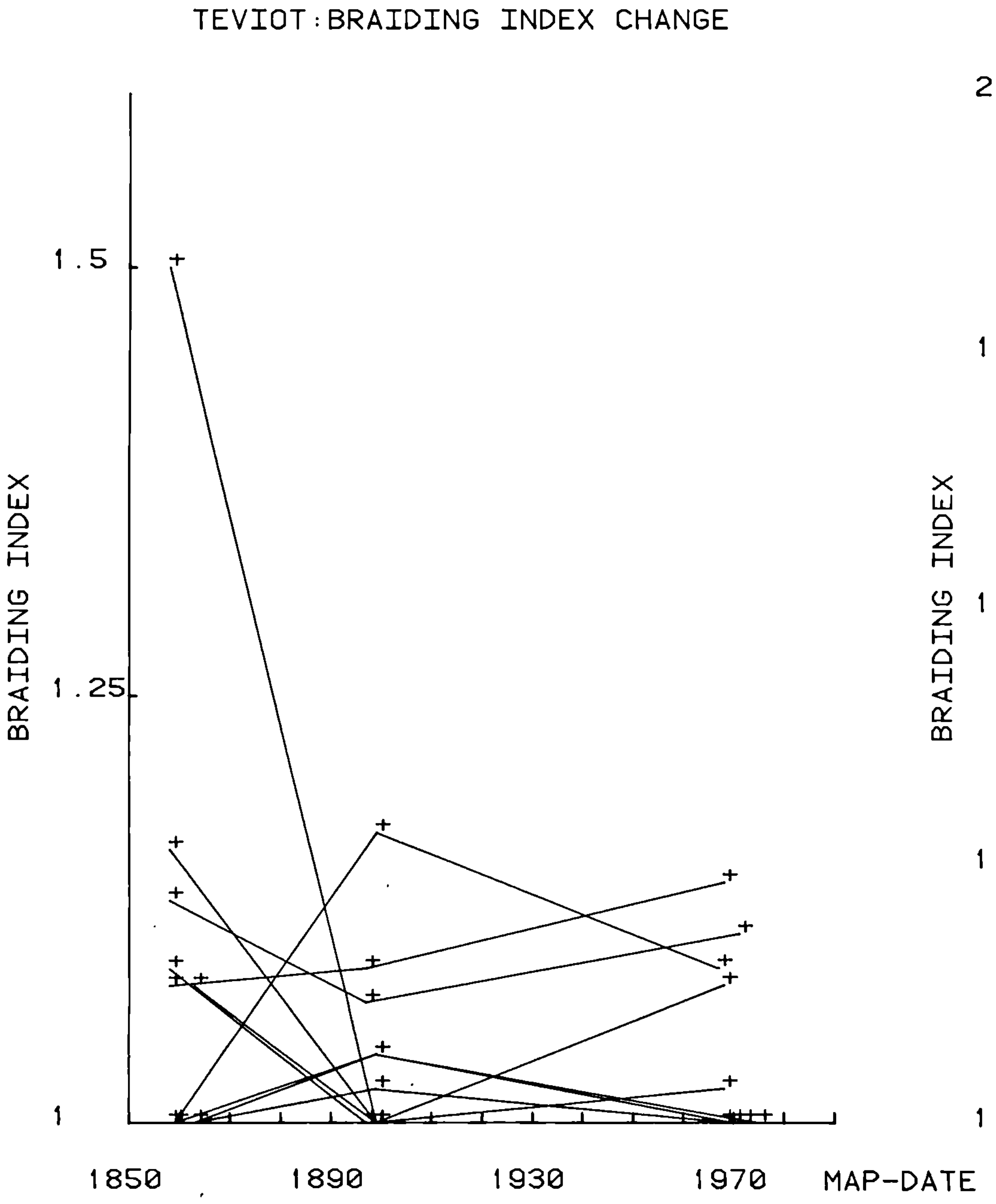


Figure 4.8.3.(xx)

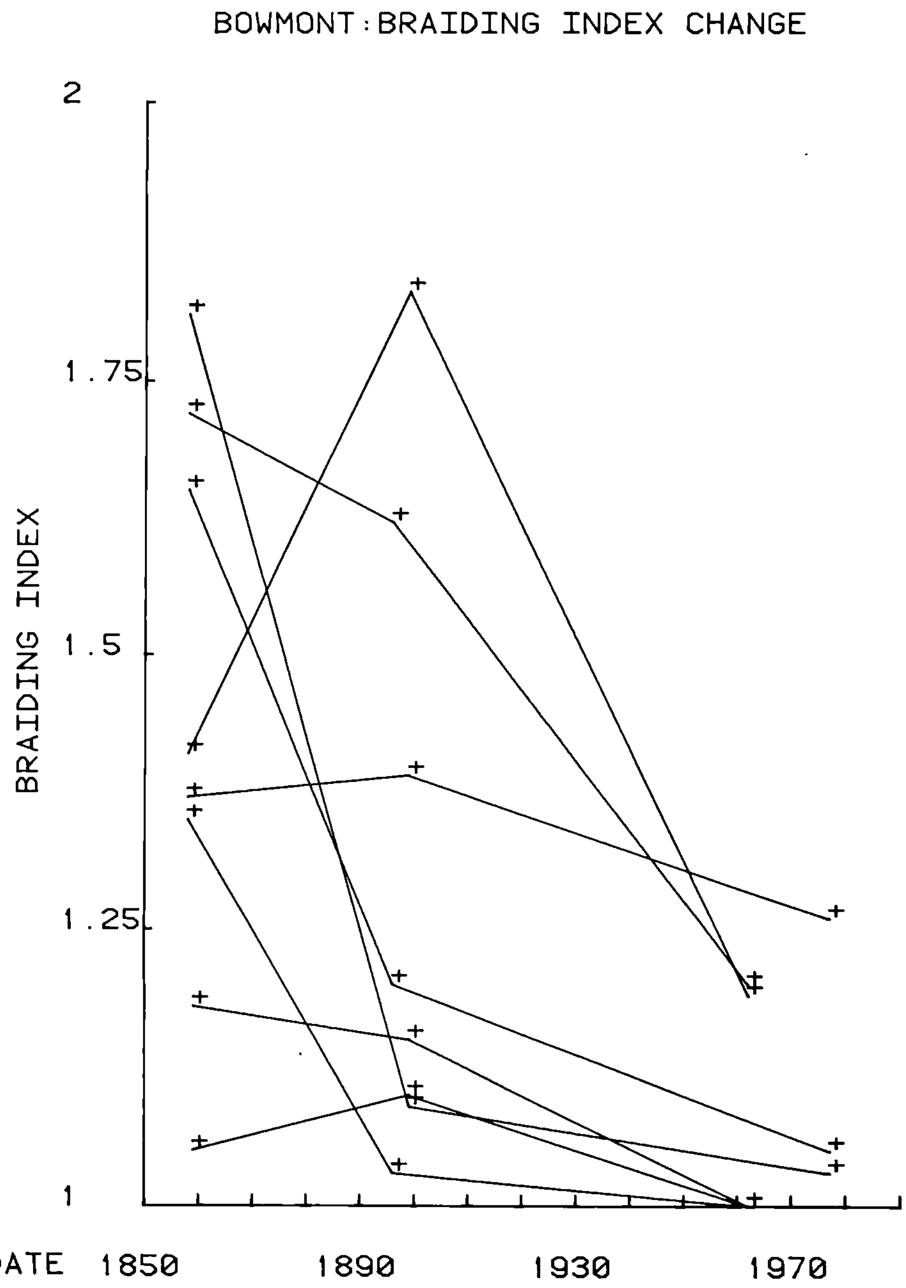
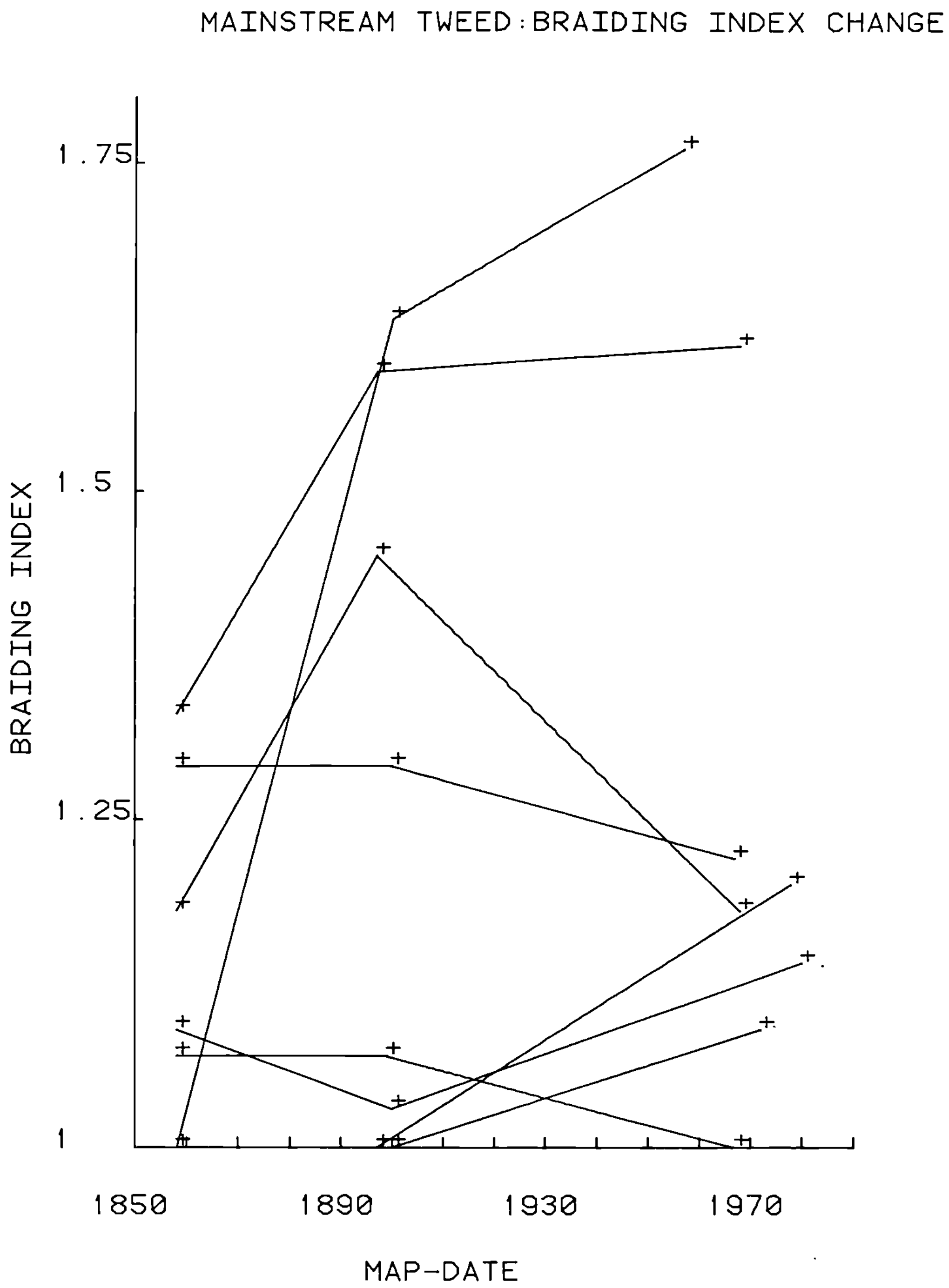


Figure 4.8.3.(xxi)



occurred on reaches with the highest initial BI values following which, considerable subsequent stabilisation had taken place, with the system either naturally undergoing relaxation or being artificially constrained. More generally on the Leader, the pattern was a slight increase in BI, over 1858-1900.

The Teviot catchment had very few split or braided sites and there was only one major change from 1858-1899 from 1.50-1.00 on the lower Teviot (Sample 59; Figure 4.8.3.(xiv) and (xix)), which involved a major planform reduction from a split to a single channel. Of the other Teviot samples, all those that had BI values in excess of 1.00 at the first edition had reduced by 1900 and this should be compared with the much greater bar frequency indicated on Roy's map.

On the Whiteadder also, changes in BI occurred between 1850 and 1900 but the pattern was more variable, with some reaches accumulating more sediment within the channel whilst others decreased. The IQR was however higher between 1900 and 1970, with a greater range of activity. The mainstream Tweed at several of the sampled reaches underwent increased BI over the three map-periods. For example, Sample 85 (Figure 4.8.3.(xvi) and (xxi)) increased from 1.00-1.63 between 1858-1900 and further to 1.76 by 1957. The periodicity of events of sufficient magnitude and frequency to transport these vast amounts of sediment must be considered. It was difficult to assess the importance of inputs from the tributaries but there was not the large build up of bars downstream of confluence sites cf. River Feshie (Spey study area). The importance of artificial structures must however be considered in relation to the location of sediment accumulations.

It was clear that different reaches were more susceptible to change and that a decrease in BI upstream may lead to an increase in BI at the next sediment store further downstream as erosion is triggered and lower sediment stores were reworked. Generally, however the Tweed sample was more stable in comparison with the Deeside and Speyside study areas, with fewer major disruptions of channel planform. There were four possible explanations, namely quick recovery from disruption that eluded the map-base, resistance to higher flows through more stable banks, the absence of disruptive flows over the period of study or man's intervention in the natural behaviour of the channel.

The importance of the initial braiding index in relation to subsequent change was then assessed. When the absolute rates of braiding index change (ABBRAID) were correlated with initial BI, high correlations were obtained (Table 4.8.3.(v)). It was interesting to note on the Whiteadder that although ABBRAID2-3 had a high correlation with BI2 (r=0.72), there was no significant correlation (0.08) between ABBRAID1-2 with BI1 ie. in this catchment between 1850-1900, locations of higher BI1 and subsequent changes were not necessarily coincident. This also measured the degree to which braiding was a stable channel form over the inter-map periods. Clearly, braiding planforms are either transient features or artificial change has taken place.

Incorporating the direction of the change, when AVBRAID2-3 was correlated with BI2, a highly significant correlation of -0.41 was found ie. high initial braiding index values were associated with high negative change in BI, especially between 1858-1900. Again, whether

Table 4.8.3.(v)

The correlation (r) between subsequent change and initial braiding index

Whiteadder	BI1	BI2
ABBRAID1-2	0.08	------
ABBRAID2-3	------	0.72

Leader	BI1	BI2
ABBRAID1-2	0.82	------
ABBRAID2-3	------	0.38

Bowmont	BI1	BI2
ABBRAID1-2	0.66	------
ABBRAID2-3	------	0.94

this reduction was natural or man induced will be assessed later. When AVBRAID1-2 was correlated with BI1, an even higher correlation of -0.54 was gained. Thus, not only high presence of change but also the direction of change were correlated with high initial BI within the Tweed sample.

4.8.3.4 Rates of change associated with lateral shift

Of the selected sites, 60 gave reliable values of lateral shift; again more upland sites had to be rejected due to a lack of accurate fixed points. The basic summary statistics are found in Table 4.8.3.(vi). The median rate of maximum annual lateral shift was slightly higher for pre-1900 than post-1900. Certain sites had exceptionally high rates of movement; for example, the Whiteadder/ Dye confluence (Sample 23, Tweed study reach 5; see Section 7.4.5) between 1857-1900, displayed a total maximum shift of 248 m and a total average of 37 m, associated with periodic avulsion across a low-level fan. Whether shifts over such large distances can be attributed to individual flood events or whether gradual migration occurred, will be assessed later. Large shifts associated with switches from split to single channels also occurred eg. Leader (Sample 1; max. 171m; av. 72m) from 1857-1897. Other large shifts, associated with irregular meander avulsions, occurred at different locations in the middle of all catchments eg. the Teviot at Falnash burn (Sample 42; Tweed study reach 6; see Section 7.4.6).

Table 4.8.3.(vi)

Parameter	25th Percentile	Median	75th Percentile	IQR
Total sample				
LSMAX1-2	18	29	71	53
LSMAX2-3	10	17	38	28
AVLSMAX1-2	0.20	0.35	0.50	0.30
AVLSMAX2-3	0.20	0.24	0.55	0.35
LSAV1-2	3	6	10	13
LSAV2-3	4	7	12	8
AVLSAV1-2	0.10	0.15	0.30	0.20
AVLSAV2-3	0.10	0.10	0.20	0.10

ALL VALUES IN METRES

Although for the total sample, higher rates of shift occurred between 1857 and 1900, Bowmont Water showed the opposite trend, with higher average shifts between the second and third editions. Whether this was due to differing magnitude and frequency of flood events affecting the Cheviot massif or artificial channel training will be assessed later. In contrast, many reaches remained highly stable over both time-intervals eg. those associated with more sinuous channels with low MAXWID related to bed-rock control and terrace confinement. Others remained stable despite the lack of apparent natural confinement, which either suggested an equilibrium form that was not easily disrupted or some artificial constraints.

4.8.4.1 Importance of quantitative parameters

The importance of contributing area was low in relation to both sinuosity and braiding index change within the total sample. However, when individual catchments were studied (Table 4.8.4.(i)), in the Leader sample there was a highly significant negative correlation (r=-0.48) between ABBRAID1-2 and area, ie. smaller catchments had higher rates of change of BI. This has implications in terms of the stream powers involved; smaller catchment areas must be able to generate stream powers sufficient to cause major channel adjustment. In contrast, on the Whiteadder at ABBRAID2-3, there was a significant positive correlation (r=0.36) ie. ABBRAID increased as area increased. Thus, differing catchments with differing types degrees of controls may display differing trends. For example, within the Leader catchment, annual average lateral shift 1-2 not surprisingly showed a similar relationship, with a highly significant correlation of r=-0.83 with area. Thus, rates of lateral shift had a strong negative relationship with area on the Leader, relating to degree of bedrock control, which increased downstream. With height, there was a high negative correlation (r=-0.45) between ABBRAID2-3 and height on the Whiteadder. Low rates of change were thus associated with the greater heights.

The only catchment where MAXWID was an important control (cf. the Spey and Dee study areas) was on the Bowmont (ABBRAID1-2; r=0.50). Obviously the channel was making more use of its available floodplain area with its glacially deepened trough between map editions 1 and 2, there being no correlation between ABBRAID2-3 and MAXWID. In other catchments eg. the Teviot, the maximum floodplain width was not utilised often due to terrace confinement, and position within the catchment could be more important eg. the Leader.

4.8.4.2 Importance of qualitative parameters

Local and upstream sediment availability both showed interesting changes over the three map-dates. When cross-tabulation was performed on both sediment classifications at different map-dates, there were deviations from the dominant axis, especially between 1 and 2/3 editions. In the 1870s, there were more sites within the total sample with large amounts of channel-side sediment. This was slightly reduced by 1900, however since 1900 there has been a major stabilisation of channel banks. A possible interpretation was a pre-1850 disruption of channel planform, followed by a subsequent return to a more stable form. Change was very marked on the Bowmont Water where there was a considerable stabilisation from "large amounts of channel side sediment" to only "localised sediment".

To assess the association of rates of activity index change with the typology classifications, a "breakdown" analysis was carried out. When ABBRAID2-3 and ABAVSIN2-3 were broken down by LOCSEDMT2, there were no significant differences between categories. However, when AVLSMAX 2-3 and AVLSAV 2-3 were broken down by LOCSEDMT2, there was a highly significant difference between categories. Highest maximum rates of change (1.7 m yr^{-1} and 1.2 m yr^{-1} respectively) were associated with those reaches with large amounts of unstabilised sediment, which could easily be reworked. One has to be aware however of the chicken and egg problem, ie. is the area reworked because of the large amount of unstabilised sediment there or vice versa.

When ABBRAID1-2 and ABAVSIN1-2 were broken down by LOCSEDMT1, there was a highly significant difference between categories, with the largest rates of change in BI (0.006 yr^{-1}) and sinuosity (0.002 yr^{-1}) taking place in reaches with large amounts of gravel availability and unstable banks (Table 4.8.4.(ii)). Reworking of "in situ" sediment stores must occur. Thus, high activity areas were associated with large amounts of sediment availability and regular reworking of the floodplain, whereas this was mainly not the case with changes between the second and third editions.

Table 4.8.4.(i)

Selected results of the "breakdown" analysis within the Tweed study area

(a) AVLSMAX2-3 broken down by LOCSEDMT2

LOCSEDMT2	Mean	SD	
0	0.69	0.58	
1	0.72	0.59	Sig = 0.05
2	1.18	0.45	

(b) AVLSMAX2-3 broken down by PLNVEG2

PLNVEG2	Mean	SD	
1	0.37	0.46	
2	0.75	1.01	Sig = 0.02
3	0.54	0.50	
4	0.97	0.61	
5	0.91	0.55	
6	1.40	0.20	

Table 4.8.4.(i) cont.

(c) AVLSMAX1-2 broken down by PLNVEG1

PLNVEG1	Mean	SD	
1	0.21	0.41	
2	0.29	0.03	Sig = 0.008
3	0.96	0.98	
4	1.61	1.15	
5	1.93	1.28	
6	2.25	2.12	

(d) ABAVSIN1-2 broken down by LOCSEDMT1

LOCSEDMT1	Mean	SD	
0	0.0008	0.0010	
1	0.0011	0.0014	Sig = 0.02
2	0.0018	0.0015	

Table 4.8.4.(i) cont.

(e) ABBRAID1-2 broken down by LOCSEDMT1

LOCSEDMT1	Mean	SD	
0	0.002	0.0052	
1	0.001	0.0018	Sig = 0.008
2	0.0061	0.0075	

(f) AVLSMAX2-3 broken down by LOCSEDMT2

LOCSEDMT2	Mean	SD	
0	0.24	0.34	
1	0.42	0.52	Sig = 0.001
2	1.39	1.47	

When flood vegetation was studied, there was a shift in modal class, from category (3) "trees and rough-grassland" to category (5) "agricultural" by the third edition. Bank vegetation was however fairly similar over the three map-dates with channels frequently tree-lined, despite proximity to agricultural land. When AVLSMAX2-3 and AVLSAV2-3 were broken down by floodplain vegetation 2, there was a highly significant difference between category means. Lowest rates of shift (0.01 m yr^{-1}) were associated with dense tree-cover, while rough-grassland had the highest rates (2.7 m yr^{-1}). Similarly, when AVLSMAX 1-2 was broken down by floodplain vegetation 1, a significant difference was found between means. Highest mean rates of shift were associated with heather and rough-grassland categories. The complex circular control of vegetation must again be appreciated, but unlike the other study areas, trees (especially tree-lined rather than localised occurrence) did appear to restrict lateral activity. When bar vegetation was assessed as an index of increasing or decreasing stability of bar features, there was a slight increase in well stabilised bars by the third edition and a similar reduction in those under rough-grassland. It appeared that in this environment, if bars were allowed to stabilise, they became a more permanent feature of the fluvial landscape.

When channel pattern was cross-tabulated between map-dates, there was a slight shift from categories (5) and (6) ("irregular" and "irregularly meandering") to category (3) ("sinuous"). However, in terms of the strength of the association, there was more similarity between first and second than second and third editions. Within the

ISLANDS classification, there were 14 samples greater than or equal to category 3 ("frequent bars") at the first edition, this was reduced to 9 and 6 respectively in the second and first editions. When the importance of channel pattern category was considered in relation to rates of change, the following results were found. There was no significant difference when ABAVSIN2-3 was broken down by CHANPATT2 ie. changes in sinuosity occurred independent of category. However, when ABAVSIN1-2 was broken down by CHANPATT1, a significant difference was found with highest rates of change in categories 3 and 5-8 ie. "sinuous" and "irregular" to "regular meandering" channels. Other activity indices broken down by channel pattern classification did not show significantly different rates of change.

4.9 Summary

In summary, there were distinct differences between the three study areas in terms of the dominant range of channel types, and the relative importance of controls as indicated by the typology. The spatial and temporal distribution of the associated rates of change also showed considerable contrast.

In terms of spatial variation as indicated by the typology, several points were highlighted. Position and height of the reach within a catchment were not important, especially within the Dee study area, due to the dominance of glacial controls. Highest sediment availability, from both local and upstream sources, occurred within Speyside. In contrast, within the Tweed sample the majority of reaches had much lower local sediment supply, especially by 1970. The importance of vegetation on the banks varied. Whereas tree-lined banks on the Tweed tributaries were mainly associated with undivided channels, within the Spey sample, they were not necessarily an indicator of stability. They were thus associated with a range of planform types, from stable sinuous to highly active wandering channels. Larger bar features appeared to have had more chance to stabilise within the Spey catchments than within the other two study areas. However vegetation was found to be one of the least important features within the typology.

In terms of channel description, in all three areas sinuous channels represented a dominant channel pattern. Furthermore, these were much more frequent (cf. other planforms) within the Tweed study area. Straight channels were rare in all three study areas. A larger percentage of wandering channels occurred within the Spey and Dee samples, suggesting that the controls required for such planforms are not common within the Tweed study area. The Spey study area was characterised by fewer reaches within the irregular meandering to tortuous meandering categories of channel pattern.

The "islands" classification perhaps most usefully brought out the contrasts between areas; the Spey had the highest frequencies within the split to reticulate categories while the Tweed sample had a very low percentage. This indicated large differences in the amount of sediment stored with a channel. All three study areas had evidence of channel shift pre-1850. However, the lower values within the Dee sample were probably due to the total reworking of floodplain areas and palaeochannels. Within the Dee study area, even after avulsion, abandoned parts of the floodplain continued to be within reach of further reworking by subsequent floods cf. the Spey and Tweed study areas.

In terms of change in channel pattern, pre-1900, the most dominant feature within the Dee study area was the statistically significant difference in BI between 1869 and 1900. Clearly at least one geomorphically active stress had been applied to the system causing an expansion of channel planform, and which involved both reworking available sediment and accessing new reserves. There had been transport of sediment through the system between lower competence storage zones. There was however considerable spatial variation in channel response between catchments, depending on slope, location in relation to local baselevels, availability of sediment and width of floodplain area.

Pre-1900 within the Spey study area, although overall BI values were higher than for upper Deeside, results differed between tributaries. For the total sample, the median and IQR for the braiding index were higher pre-1900 (both BI1 and BI2) and this does compare with Lewin and Weir's (1977) findings of increased braiding on the lower Spey, pre-1900. However, this was not consistent within the tributaries. For example, the Nethy and Feshie had lower rates of change pre-1900, maintaining a greater tendency to braid, in comparison to a large scale decrease and increase in BI respectively, post-1900. In each case, the pre-1900 magnitude and frequency of discharge events must be analysed both regionally and specifically within each study catchment. Any landuse effects that could have caused increased sediment mobilisation with the catchment and thus sediment inputs to the channel at that time must also be assessed.

The Tweed sample also showed differences between catchments, which were highly dependent on degree of confinement by both terraces and bedrock. Overall values of change were however small in all activity indices. Both the Leader and the Bowmont had the highest rates of change in BI. Again, comparison between both regional and catchment specific flood histories is required. The general reduction in both sinuosity and braiding index suggests that the role of artificial stabilisation must be considered as a possible reason for discrepancies between catchments.

Within the Dee study area post-1900, there was an apparent relaxation and stabilisation of the system in all catchments apart from the Clunie, which underwent some disruption. In comparison with the 19th century disruption of planforms, the 20th century has seen a large number of channels regaining a more stable form. It is therefore necessary to assess how the flood history was different before 1900, whether any major disparities existed between catchments and whether any major landuse changes had occurred pre-1900.

Within Speyside, in terms of post-1900 planform change, the Nethy catchment stood out with its significant increases in sinuosity. Whether this was due to the differing magnitude/ frequency of discharges over that period or whether it was a response to disruption of earlier artificial stabilisation must be questioned. In terms of post-1900 BI change there was a general trend towards reduction of BI, especially within the Avon and Tromie catchments, where splits and anabranches were either cut through or abandoned. The Spey in contrast had increased its BI suggesting a gradual stabilisation and increase in the size of bar forms.

Within the Tweed study area, one of the major changes since 1900 occurred in the braiding index on the Bowmont where the IQR was over 3 times its pre-1900 value. This involved both expanding and reworking of the channel planform. Much slower rates of change were associated with the other catchments, frequently contracting the area of floodplain used. Thus, the stream powers required to cause dramatic planform alteration occur infrequently or else artificial constraints have been imposed.

Assessing the application of the typology, there are some aspects that could be improved. For example, some indication of the scale of channel patterns would be useful, to allow comparison of small amplitude, irregular meanders on the Derry with the much larger irregular features on the Whiteadder. The spatial discontinuity of controls within upper Deeside sometimes caused difficulty in assigning a 1 km reach to one category, and the dominant form had therefore to be used. The major disadvantage with such a typology is the necessity to keep the criterion for measuring variables constant through time. This is difficult as there is an natural tendency to modify and improve your assessment of features as the sample progresses. This was however minimal, as the initial classifications fitted the map-based detail well.

Assessing the activity indices, arguments can occur over their representativeness and sensitivity to different types of planform change but the application was through the testing. Sinuosity was not a sensitive indicator of common modes of channel adjustment within the Spey and Dee study areas, however it did reveal important changes within the Tweed sample. In contrast, braiding index did reflect frequent modes of adjustment, highlighting differences in planform response between reaches.

CHAPTER 5

The magnitude and frequency of extreme rainfall and runoff

5.1 Introduction

To place the rates of channel planform change analysed in Chapter 4 in perspective, it is now necessary to turn to analysis of the available flood record for each study area. Concentrating initially on flood frequency, there may be persistence in the occurrence of flood events of a certain magnitude associated with known or unknown rates of geomorphic activity. The magnitude of discharge events of a certain recurrence interval may also vary through time, depending on the period of study. It is important to assess whether flood events of varying magnitudes and durations have undergone periods of increased or decreased frequency in the past 250 years, or whether they have been maintained relatively constant through time (Chapter 1: Question 5). Seasonality of flood events may be important in terms of the magnitude of the flood flow with respect to storm type. Absolute rainfall amount must also be compared with likely soil moisture level (Archer, 1981) or winter rainfall acceptance potential. Seasonality may have implications as to the types of geomorphic work done and in an applied sense, it may also affect the extent of economic damage eg. to floodplain crops.

It is necessary, too, to look for possible climatic fluctuations and associated variations in the allogenic input of rainfall ie. changes in the frequency of events likely to cause flooding. These are controlled by the mechanism of variations in the prevailing wind circulation through:

> "The proximity of the paths of frequent cyclonic activity and frontal passages to the place of interest and the prevalence or otherwise of onshore and/or upslope winds" (Lamb, 1982, p211).

In England, it has been found that in known periods with the highest prevalence of westerly winds in middle latitudes, rainfall was enhanced except in physiographically sheltered areas. Whether or not this applies also to the magnitude of random rainfall events has not been documented.

The frequency of snow falls must also briefly be considered here, though only in a more limited and qualitative way. Frequency and duration of snow cover is also influenced by changes in the frequency of westerly winds and of course changes in temperature, especially where there is a marginal event. The abundant historical information on snow-fall events (eg. British Rainfall) would provide an interesting comparative study, but had to be limited due to time restrictions. There is however very little information concerning snowmelt flooding within Scotland in published hydrological studies. The known British, and where available Scottish, records of climatic fluctuations must therefore be assessed. To quote Ladurie (1971):

> "Climate is a function of time. It varies; it is subject to fluctuation; it has a history." (Ladurie, 1971 p8)

This analysis concentrates on three main characteristics of rainfall events which affect the allogenic contribution to flood frequency, namely their magnitude, frequency and duration. The recurrence interval (RI) of a rainfall event may not be equal to the return period of the runoff response (see Larson and Reich, 1973). It is necessary to assess the subsequent implications of this discrepancy to flood frequency analysis and the explanation for varying rates of channel planform change.

It is first important to establish the known hydrometeorological categories of flood events (eg. Colman (1953), Ward (1978), Werritty and Acreman (1985)), which are documented as typical of the Scottish upland environment. Although meteorological factors cause floods, physiographic factors can intensify them (Ward, 1978). There are several different types of hydrometeorologic situation which, with favourable catchment characteristics, may increase the likelihood not only of a moderate increase in stage, but of the generation of a large magnitude flood event. If a large quantity of precipitation falls for 24 hours or longer either in a frontal or orographically reinforced storm (Lambor, 1956), the saturation level of the catchment may be exceeded. Under such conditions, saturation overland flow may take place, accelerating the rate of runoff to the stream channel.

This flood situation may be exacerbated by such temporally variable catchment characteristics as antecedent soil moisture levels, ie. whether there is a soil moisture deficit, saturated soil state or perhaps a frozen ground surface at the onset of more intense precipitation. Other circumstances which can exacerbate flooding, eg. hydraulic discontinuities, are discussed by Werritty and Acreman (1985). Also latent catchment characteristics, which are more constant in steady time, such as indurated soil horizons or a predominance of bedrock near the surface, will affect surface runoff. Shorter duration, more localised, convective thunderstorms can also produce high intensity rainfall. This in turn can cause high rates of saturation overland flow, or even Hortonian overland flow, to occur as the ground becomes incapable of allowing infiltration of the increased amount of rainfall. Such a situation will of course be heightened in its flood risk by similar catchment features as the longer duration rainfall events.

Another complicated category of flood event is caused by the accumulation of snow, perhaps within drifts in the catchment, over periods in winter. With a slight increase in temperature, a fall of warm rain or a combination of the two, there may be a sudden increase in discharge, far in excess of the diurnal snowmelt fluctuations in flow that are frequently recorded in upland Scottish catchments between February to April (eg. within the Feshie catchment; Werritty and Ferguson (1983)). There may thus be a temperature dependent time-lag between the fall of snow and its influence on river flows. These three subsets are obviously artificially delineated as most flood-producing events are highly complex, but it is important to assess how important each of these hydrometeorologic types is in the Scottish upland context, plus other flood-risk situations if they become identifiable.

Each of these genetically distinct, flood-risk categories must be envisaged as operating on different but interlocking recurrence interval scales ie. a 20 year convective storm event could take place in the same year as a 20 year RI snowmelt event. It is therefore important when studying discharge frequency to understand the origins of the flood situation, as the slope of the discharge frequency curve may vary in steepness eg. with summer convective versus summer frontal storms. No study is currently known in the U.K. which breaks down a discharge peak over threshold (POT) series into genetic categories when assessing possible return periods. It has however been attempted in the western U.S.A. and Italy for snowmelt floods and floods associated with frontal storms.

5.2 Data sources and methodology

Discharge and rainfall records, of varying lengths and numbers of stations, exist for each of the study areas and the information they provided will be discussed below. However, if this study was limited to these quantitative sources, much valuable magnitude/ frequency information would be omitted. In the first instance, much useful information over the longer timespan of more than 150 years was extracted from a variety of other sources. The types of details which could be collated include information about antecedent conditions, relative heights of flood peaks down a river, and qualitative and quantitative information on rates of hydrograph rise and recession.

In terms of rainfall analysis, the intensity, frequency and duration of storm events were frequently reported. Potter (1978) documents similar sources in order to extend the flood record in England and Wales. There are several published meteorological records and reports of reliable accuracy and these provided information over a large timespan, although the type of information published becomes more quantitative, and in some ways less useful, after 1950. The actual volumes studied were British Rainfall (1860-1968), Symon's Meteorological Magazine (later the Meteorological Magazine) (1866-1976) and The Journal of the Scottish Meteorological Society (1865-1910). These provided both quantitative values of rainfalls for a variety of durations, eye-witness accounts of storm and associated flood events, and discussions on various meteorological phenomena. Information was however highly selective in both the locations and the events covered, and varied in the detail given.

Other valuable sources of information were from newspaper reports at both a national and local level. The Scotsman, for example, was first published in 1817 and many local papers extend back in date beyond this, although names change and runs of editions may vary in length eg. The Kelso Mail. Information from such sources is of course of lesser reliability as the accuracy of details given was dependent on both the objectivity of the reporter and the accuracy of his data source, which cannot now be checked. However, details were frequently confirmed through other sources, for example stage heights and areas of inundation. The potential of newspaper information as a data source to reveal the impact of natural events was demonstrated by Gregory and Williams (1981).

Contemporary and near contemporary literature also gave useful information and these included travel guides eg. Burton's (1864) account of his mid 19th century exploits in the Cairngorm mountains or Hume Brown's (1891) collection of early accounts of travellers in Scotland, local histories eg. Maxwell (1809), and local administrative reports eg. the Report of the Commission for Highland Roads and Bridges (1802-1856). Very rarely, historic documents on specific floods or a series of flood events have been written eg. Lauder's (1830) accounts of the 4th August, 1829 floods in Inverness-shire, Moray and Aberdeenshire. Alternatively, papers and accounts have been submitted both to local journals eg. The Deeside Field and also to specific academic/ technical journals such as The Journal of the Institution of Civil Engineers. A variety of information was extracted from estate records but this was highly time consuming and studies were limited by availability and ease of access. Such a source has potentially the information to extend the known flood record back in detail to pre-1800.

The three study areas naturally varied in the availability of these sources, and information was collected for both regional flood events and events specifically important to individual basins within each study area. The Scottish Development Department, in an internal report (Speyside Drainage Report, 1958), collated some information on historic floods for the entire Spey basin but much more detail was collected, specific to the Spey study area. The collated flood histories are presented in detail in the Appendices 1.1-1.12 and in each case, the date, the source and a summary of the important points are included. The most important events will be discussed in later case studies (see Sections 5.4.8, 5.5.8 and 5.6.8). Where only the presence or absence of a flood event at a certain date is known, these dates will be tabulated in figures within the text. It must be remembered that the frequency of such reports depends on an individual's perception of flood events ie. the concept of a perception stage above which floods are noticed (Gerard and Karpuk, 1979). Much information referred to the effect of the flood (eg. if damage was done) rather than the event itself. Flood reports became less frequent, the further back in time researched. Presumably then, we were only gaining information about the most extreme flood events as more moderate events will have been forgotten and unrecorded as they were superseded both in stage and memory.

This data base did allow study of the seasonality of extreme rainfall events in the past, and where relative stage information was available, ranking was carried out and thus an approximate assessment of RI within the known record was possible. Incorporation of such longer term data within the gauged record allows a more realistic flood recurrence interval to be calculated (Dalrymple, 1960). Against this qualitative background, the more quantitative information provided by discharge and raingauge records will be evaluated.

Discharge records are of course a valuable data source of recent flood frequency but their length tends to be limited to under 40 years. Analysis of available discharge records will now be discussed. The extreme values in the series were studied in terms of their magnitude, recurrence interval and associated rates of catchment runoff. For the purposes of this study, "extreme" magnitude refers to an event in excess of 50 years RI while a "moderate" event may have a RI of 10-50 years. Acreman (Ph.D. in preparation) is acknowledged for updating up to the end of 1982 the POT series for Scotland, published in the FSR (NERC, 1975). The POT flood series was identified on the basis of a threshold that allowed approximately 5 peaks per year. Acreman is also responsible for the estimation of return periods for stations in this data set, using probability weighted moments and the Generalised Extreme Value Distribution (Acreman, Ph.D. in preparation). It must however be remembered that within such analysis, it is an important assumption that the annual maximum series represents a random sample from a single unknown probability distribution. Frequency analysis has thus the problems of gaining as long a data-set so as to fit the correct distribution, as well as assuming homogeneity over these longer periods. For example, a discharge record commencing in 1970 may be highly atypical of 1900-1982, and thus cannot be accurately extrapolated beyond that time-period since it will give unrealistic estimates of high return period floods.

The maximum runoff rates found from these discharge records were compared with the Scottish maxima recorded for Scotland in the I.C.E.'s (1933; 1960) Floods in relation to reservoir practice. Occasional flood levels, eg. marks on bridges indicating the maximum flood height reached at that cross-section, allowed the magnitude of historic flood flows to be ranked. It must be remembered in any subsequent interpretation that each flood event will reach different heights at different locations upstream and downstream, and although a spot ranking is available for a particular cross-section, along other reaches of the river the relative magnitude of events may be different. Also, the hydraulic geometry of the cross-section may have been altered through time.

After considering the discharge record, the spatial pattern of longer term raingauge information was assessed. Unfortunately, such gauges tend to be located at lower altitudes and so unavoidable extrapolation maybe a gross underestimate of precipitation on the neighbouring mountain masses. The altitudinal modification of airflows, with the uplift of fronts over mountain masses, causes increased mixing of air, and thus the likelihood of heavy intense rainfall (Taylor, 1976). Similarly, with convective events when an unstable air mass rises due to contact with a mountain slope, this may trigger instability and intense convective thunderstorms. Thus, although mountain masses do not generate their own separate climatic regions with any regularity in upland Scotland, they may increase the intensity of different types of synoptic rainfall events (see Smithson, 1969a).

There was little available information concerning precipitation gradients, although Gloyne (1968) gave an increase in precipitation of 2.81 mm m^{-1} increase in altitude for west slope gradients and 0.88 mm m^{-1} increase in altitude for east slope gradients, generally for the Scottish Highlands. Obviously, aspect, exposure and physiographic factors must be important. Ballantyne (1983) found strong curvilinear gradients on the east-facing slopes of individual massifs in Wester Ross, with a 2-fold increase in rainfall between 18 and 690 metres on An Teallach. He concludes for his particular site that long term average precipitation may be considerably greater than the values depicted on standard period rainfall maps. This also has implications for the concept of maximum probable precipitation in upland areas.

In terms of the rainfall analysis carried out, the longest records of available data from daily rainfall gauges (pre-1920) were collected from the official yearly data sheets (Meteorological Office, Edinburgh), this data taking several forms. The annual total rainfall gave a general comparison between years and runs of years as being typically "wet" or "dry". It also gave an indication of the probable water balance of the catchment and the likelihood of long-term ground saturation. Broken records and records where the gauge had been moved were homogenised by the method of average annual ratios (Jones, 1980, 1981, 1983a; Craddock, 1975, 1976). This involved calculating the ratios of annual catches between segments of the broken record and another nearby continuous record. With the subsequent conversion factor, the shorter segments of the broken record were adjusted to a similar ratio as that of the longer record. Thus, a composite record was constructed and the ratio would be expected to become more accurate with increased length of segments.

After recognising the wettest years, a 5 year running mean and running standard deviation were put through the annual total data to assess how much fluctuation there was around a mean value. This also showed if any long term changes in precipitation amounts had occurred. A similar approach has frequently been adopted eg. Salter (1921), Lamb (1968). Such fluctuations were then compared with previously documented trends (eg. Lamb, 1982), outlined in Section 5.3. The

Kolmogorov-Smirnov test and the runs test were used to investigate whether the total data set came from the same parent population (Davis, 1973). It has been suggested that there may be periodicities that influence rainfall amount (Brunt, 1927) eg. related to sunspot cycles. Cycles of roughly 11 years (9-14) and 22 years have been found (Lockwood, 1979; Wilcox, 1976; Schneider and Mass, 1975; Meadows, 1975) but the current view is very much anti such simple cycles. Generally, more complex superimposed combinations of shorter and longer lengths have been identified eg. 4, 80-90 years (Lamb, 1972, Tabony, 1979).

To see if any repeats, periodicities or cyclicities occurred in the annual total data, autocorrelation analysis (AC) was carried out at up to 50 years lag, using the Box-Jenkins test for analysis of time-series (Box and Jenkins, 1976; Fuller, 1976). The number of years for such lags was limited by the length of the data set. The Box-Jenkins analysis was facilitated by the use of the SPSS computer package (Nie et al., 1981). If we compare the sequence of annual totals with itself at successive positions, then it is possible to locate the maximum correspondance and to calculate the degree of similarity between corresponding segments (Davis, 1973).

Significant auto-correlation was taken as a value in excess of 2 standard errors. The assumption of the latter is that the time-series constitutes white noise ie. is totally random. Partial auto-correlation was also carried out ie. autocorrelation with the influence of other trends held constant. It must be remembered that several cycles may be coincident. Fourier/ spectral analysis (Bath, 1974) was also used on the data, allowing comparison with the

Box-Jenkins results. The power spectra represents the plot of power (square of the amplitude) against frequency. Such spectral analysis has been carried out by Tabony (1979) for long-term rainfall records in England and Wales.

Having assessed any periodicities in the annual total data, peak over threshold data, extracted for varying durations, were collected at each of the chosen stations for each year of record. These could then be compared with extreme 24 hour values for Scotland as presented in Table 5.2.(i) and with the probable maximum precipitation (PMP) as estimated by the FSR (NERC, 1975), from maps of estimated maximum 2 and 24 hour rainfall. The upper limits of PMP, in the British context, are difficult to define eg. with Bonacina's report of a one day rainfall greatly exceeding 400 mm (Reynolds, 1979). As Lamb (1978) points out, with a small number of raingauges, it is by no means unlikely that sampling of the greatest downputs of rain are an underestimate.

For the 24 hour data, every rainfall greater than or equal to 1 in (25.4 mm) was recorded. For 48 and 72 hour events, the threshold was increased to 1.5 in (38.1 mm). These thresholds were chosen because on average there were 2-5 POT per year, although of course this varied between study areas. A threshold in imperial units was chosen for ease of data collection (adopted also by Perry and Howells, 1982; Lawler, 1985). To see if the total number of peaks over threshold (POT) had any strong relationship with the annual total, the two were correlated at each station. A continually wet year may not necessarily be one with a large number of POT. The temporal pattern of these events was then assessed in the first instance, by plotting frequency against year of

Table 5.2.(i)

Extreme 24 hour rainfall values for Scotland

Station	Rainfall (mm)	Date
Sloy Main Adit	238.4	17/01/1974
Kinlochquoich	208.3	11/10/1916
Dalness, Glen Etive *	199.1	17/12/1966
Croy, Dalcross Cas.	179.3	25/09/1915
Blackwater Dam, South	171.7	17/12/1966

* documented also by Reynolds (1967)

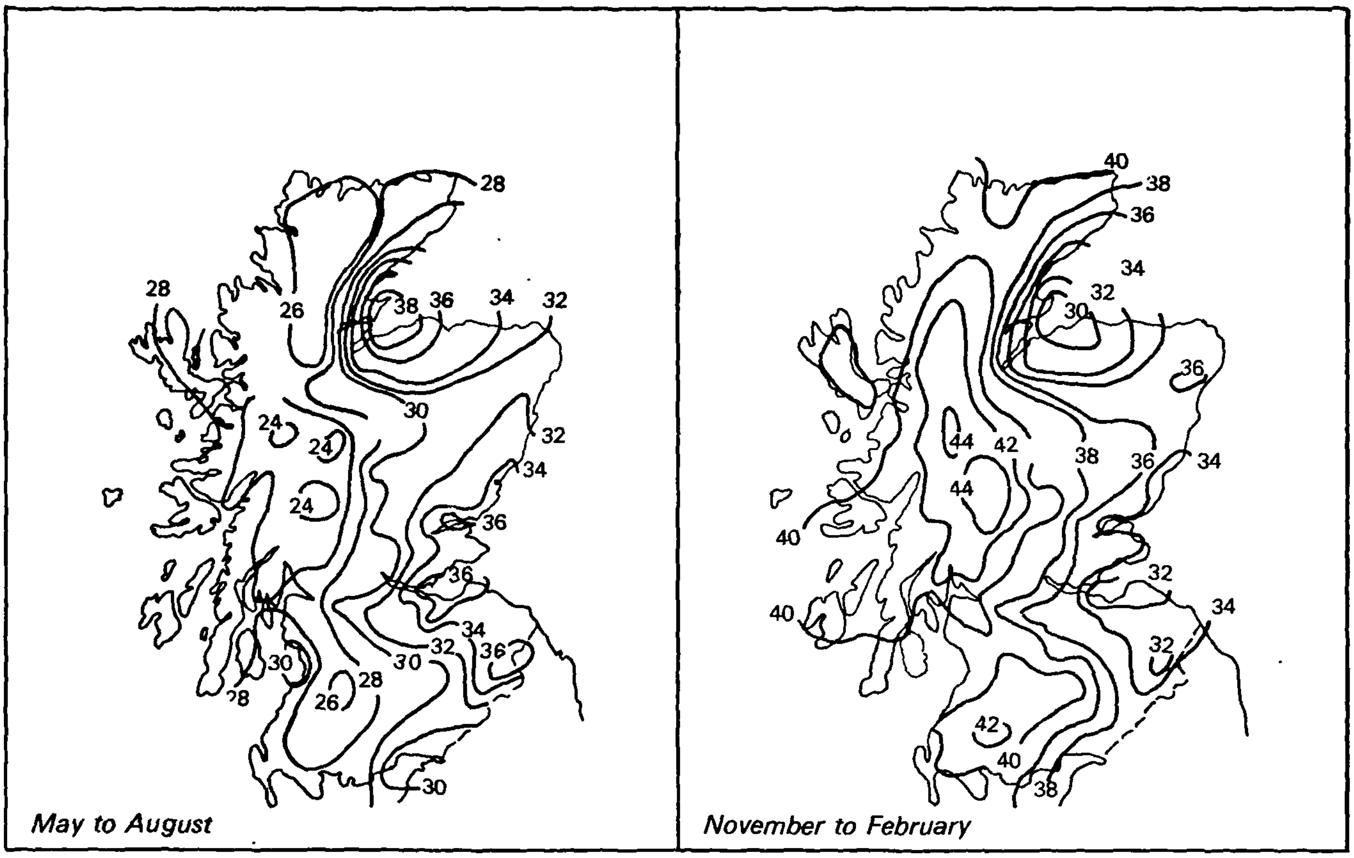
(Data sources: British Rainfall (1860-1968), Bilham (1935)) record for each station. The seasonality of these POT were also studied, both for the total sample and for subdivisions by decade through cross-tabulation. These were compared with the Meteorological Office maps of seasonal rainfall across Scotland as a percentage of the average annual total (Bleasdale, 1961-1969; see Figure 5.2.(i)). The seasonality of POT during exceptionally wet decades was also assessed.

It was also necessary to estimate the recurrence intervals of varying magnitudes of rainfall at these different durations. However, to do this certain assumptions have to be made about the homogeneity of the data set. This was difficult as it is not really known how robust Gumbel type distributions are in these circumstances. Inconsistency and non-homogeneity may introduce dependence into independent time-series (Yevjevich and Jeng, 1967). Certain periods may be identified as having a higher or lower recurrence interval of events of different magnitudes because of serial correlation and persistence. A Gumbel type Extreme Value distribution 1 (FSR, NERC 1975) was initially fitted to the annual maximum series, for each duration. The distribution function for the EV1 distribution is given in Equation 5.1 and Figure 5.2.(ii) illustrates the varying values of k within the EV family of distributions.

$$F_{(x_1)} = \exp(-e^{-(x_1 - u)/\alpha}) \qquad (5.1)$$

Figure 5.2.(i)

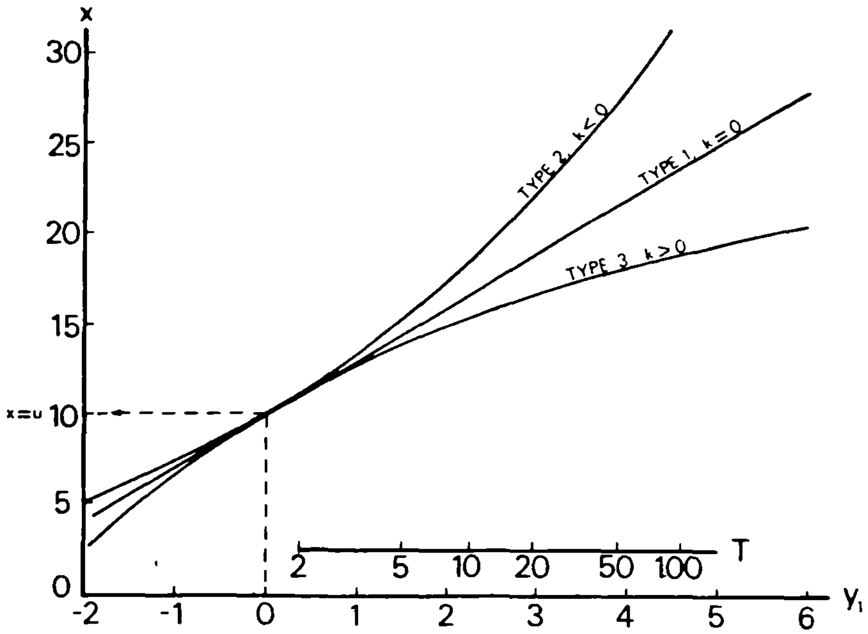
Equipercentile maps



Average rainfall for four months 1915-1950 as a percentage of the annual rainfall average

Bleasdale (1961-1965) pp 257 and 259

Figure 5.2.(ii)



The three types of extreme value variate shown as functions of the type 1 reduced variate by the relation:

$$x = u + \alpha (1 - e^{-ky_1}) / k$$

(Source: FSR, NERC 1975 p41 Figure 1:10)

As this was found not to fit the most extreme values of every data set, the General Extreme Value distribution (GEV) was also fitted (Jenkinson, 1975), using the method of maximum likelihood estimate (Kite, 1977). The distribution function is given in Equation 5.2 and this attempts to fit a curve (K>0 or K<0) rather than a straight line (K=0) to the data set. This involves estimating three parameters (u , α and k) rather than two (u and α).

$$F_{(x)} = e^{-[1-k(x-u)/\alpha]\ 1/k} \qquad (5.2)$$

The maximum likelihood estimate estimates the probability of "n" individual maximum events actually being observed as "n" annual peaks as a maximum (Kite, 1977). Both these distributions (EV1 and GEV) were fitted to the total population for each station and also on overlapping or running sub-sets of the data. This allowed assessment of how far the magnitude of rainfall for a set RI has varied through time. It also showed the danger of extrapolating from small data sets. The 48 hour values were compared with those 2 day falls of 5 year RI of Jackson (1977).

To assess the possible magnitude and frequency of even longer duration rainfall events, ranging in length from 4 to 31 days, accumulated rainfall amounts were analysed. Rainfall values were extracted from the raw data set in each study region and rainfall amount was plotted against $\log_{10}$ time. These envelope curves were ultimately to be used for geomorphic purposes rather than final definitive meteorological statements. They were however compared with a prepared envelope curve for the whole of Scotland and associated estimates of PMP (FSR, NERC 1975). This allowed the estimation of an envelope curve for the study area and an upper limit of recorded rainfall within the number of years on record. This was compared with the maximum probable precipitation estimates of the FSR (NERC 1975), which are based on rather large regions, and both the notable falls during intervals of a few days in Great Britain documented by Hawke (1942) and Glasspoole's (1929-1930) study of areas covered by intense, widespread falls of rain. For example, at the upper extreme within Scotland, 16.72" (424.7 mm) of rain was measured at Portree, Isle of Skye during 5-8th Dec, 1863. Such rainfall events are interesting because with high antecedent moisture conditions, there will be sufficient time for the catchment soils to reach saturation level. Long periods of consistently heavy rainfall, with bursts of higher intensity, would be expected to be associated with major flood events.

Finally, all the previously discussed rainfall analyses refer to a minimum of a 24 hour rainfall event, however at the other extreme the importance of short duration, localised rainfall events, especially in smaller tributary catchments, must not be ignored. There is much evidence for the existence of highly intense localised rainfall, especially in areas of physiographic intensification (eg. Austin Miller's (1952) study on the cause and effect of a Welsh cloud burst where the fall possibly exceeded 3 inches (76.2 mm) in one hour or Grant's (1953) report at Eskdalemuir where in June, 1953, 80 mm fell in 30 mins.) Some documented reports of historic "waterspout" occurrence in

Scotland are tabulated in Table 5.2.(ii). Other qualitative eye-witness accounts of the geomorphic impact of "water spout" events are provided by Meaden (1984) in "Point deluges in the Pennines." However, the meteorologic cause of such highly intense rainfall is not always clear but there is growing evidence of their geomorphic significance, especially in smaller catchments where high runoff rates can be maintained. For example, the Hermitage Water storm of July, 1983 caused slope erosion, peat failures and channel adjustments within headwater catchments (pers. comm. M. C. Acreman and A. Werritty).

Less extreme short duration events are frequently associated with convective thunderstorm events. However, in Jackson's (1977) study of the 50 highest 2 hour falls of rain in the British Isles, only three were Scottish (eg. 26th June, 1953 in Morayshire; 90 mm in 2 hours) but there is a paucity of gauges in the uplands. Reynolds (1978) also showed the geographical locations of 2 hour storms with a rainfall greater than 100 mm occurring this century, all being south of a line joining Anglesey to Newcastle. When the 25 largest falls in 3 and 5 hours over 1900-1976 were studied (Jackson, 1979), no values were recorded for Scotland.

Despite this, in all three study areas there was historic and more recent evidence of the geomorphic effectiveness of short duration events, even if their magnitude was not comparable with the upper extremes of the range. Unfortunately, because of the very nature of such storms, they will escape being recorded by raingauges; or at the very least, a neighbouring gauge is unlikely to record the peak intensity of the storm event due to the extreme nature of the

Table 5.2.(ii)

Reported historic occurrences of "waterspout" phenomena within Scotland

Date	Location	Rainfall	Source
?/5/1682	Oxford	"Point deluge and tornado" approx. 2 feet [0.6 m] water fell in 15 mins.	Meaden (1979)
30/8/1879	Hill above Drum Park	Approx. 4 ins (127 mm) in 4 hours.	BR1879
5/12/1863	Portree	Succession of waterspouts 12.48 ins [317.0 mm] in a day.	JSMS (1866)
2/8/1883	Ochils between Dollar and Alva	Within a 1/4 hour, street was in flood, some parts being fully 3 ft [0.9 m] under water. Flood came so suddenly that people who had crossed stream only 3 mins before almost had to swim back to their houses.	MM1883
17/8/1897	Moffat district River Annan	Abnormal downpour, which lasted about 20 mins. river Annan rose rapidly.	MM1897
12/8/1884	Blane valley, Stirling	No estimate of rain but Finglen Burn rose 12 ft [3.7 m] in 30 mins.	BR1884
2/7/1893	Cheviots	As the result of a "waterspout", 30-40 acres [12-16 ha] of upper layer peat ploughed to a depth of about 5 ft [1.5 m] and piled in enormous masses. If this damage was really due solely to rain- must have been exceptionally heavy, falling practically as a solid mass.	B & G

MM Meteorological Magazine

BR British Rainfall

JSMS Journal of the Scottish Meteorological Society

B & G Brooks and Glasspoole (1928)

precipitation gradient. For each study area, rainfall amount and duration were collated from a variety of sources, including British Rainfall and some official gauged records from the Meteorological Office. Data from neighbouring basins were included to gain a more regional sample. These were plotted on a graph of quantity against $\log_{10}$ time to produce a local envelope curve. This was compared both on an inter-area level with the Scottish maxima for different durations, and also with the maximum possible precipitation estimates of the FSR (NERC, 1975).

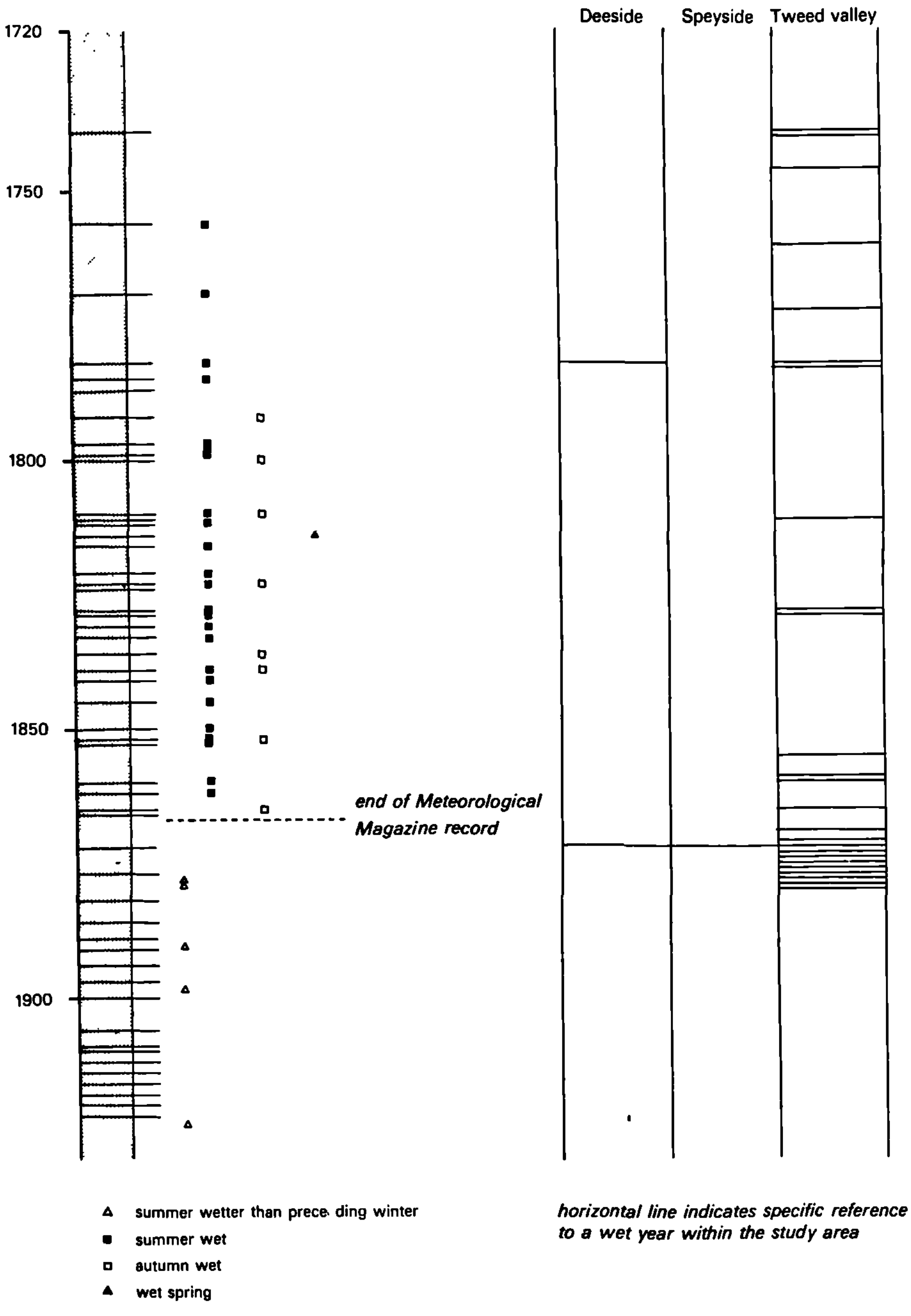
Finally, the characteristics of certain major storm events, associated with major floods, were studied in more detail for their hydrometeorologic conditions. This allowed similarities and dissimilarities to be highlighted and estimates to be made of the likely recurrence interval of a recurrent storm pattern rather than a point rainfall value. A series of rainfall maps were constructed, using the SURFACE computer package (Sampson, 1978). The study catchments were superimposed to assess magnitude/duration within the study area, both for the period of most intense rainfall but also, where data allowed, for antecedent conditions. Unfortunately, with pre-1940 rainfall events, the distribution of gauges becomes more sparse and this had to be remembered when studying storm profiles so that unnecessary extrapolation could be avoided. As certain regional storm events affected both the Spey and Dee study areas, both areas are shown on the same diagram. The synoptic situation and other exacerbating factors were researched from British Rainfall and newspaper information. It should be noted that the studied flood events were not necessarily the largest on record because events with a principally snowmelt contribution were not analysed in this way.

5.3 Climatic fluctuations over the last 500 years

It is first necessary to ascertain whether climatic change or fluctuation has taken place in Britain over the last 250 years and how that period of climate compares with a longer 500 year period. A review of the literature on longer timescale Holocene climatic change is discussed in Chapter 3. Initially, climatic change must be defined, here implying a continual drift in mean climatic index as opposed to a random fluctuation. Such climatic conditions vary over both very short and much longer timespans (ie. seconds to thousands of years) and clearly definition of change is highly timescale dependent. Thus, the 250 year time-base of this study must therefore be stressed. Salter (1921) also makes the important point that it is extremely rare for total rainfall of any year to be above or below average in all parts of the country at the same time. It is also important to appreciate the resolution of the changes we are dealing with. The total rate of climatic change at the onset of the Little Ice Age (mean change 0.01^{o}C yr^{-1} from 1250-1400 AD) was no more than the possible difference between mean values for successive winters (1^{o}C). Figure 5.3.(i) brings together information on climatic fluctuation from a variety of sources (see key) for the British context. Symons' (Meteorol. Mag., 1867) also gives an inventory of cold and wet years from 1739-1866 (see Figure 5.3.(ii)).

Figure 5.3.(ii)

Known history of wet years for the British Isles



Sources: Salter (1921) Brooke and Glasspoole (1927) Thomson (1901) British Rainfall
Symons Meteorological magazine (1867)

Unfortunately, there is a scarcity of information on the upland Scottish environment. In the Scottish context, the following climatic periods have been differentiated for the purpose of this study. Within the Sub-Atlantic period, two important climatic fluctuations have occurred, between 1000-1200 AD and 1430-1850 AD. The latter period, referred to as the "Little Ice Age", is very important in terms of the climate during the study period, and therefore the following discussion is divided into Little Ice Age and post-Little Ice Age periods.

Between 1430-1850, there was a world climatic cooling which has been related to the combined affect of a rapid decline in sunspot activity from 1645-1715 (Schneider and Mass, 1975), coupled with several cases of explosive volcanic activity (Lamb, 1970). The first decreased total solar, radiational output while the second caused an intensified dust veil in the upper atmosphere. There is a great controversy about the importance and status of this Little Ice Age in upland Scotland (Sugden, 1977; Sissons, 1979). This period has been dated in a general way from 1430-1850 AD, but the Little Ice Age reached its most extreme between 1745-50 and 1809-1818 (Manley, 1949). It is possible that 1692-1701 and 1880 should also be included. This has been substantiated through both periglacial evidence (Matthews, 1942) and lichenometric evidence.

There is some indication that this fluctuation saw a colder Cairngorm climate with more permanent snow-beds. Sugden (1977) however suggested that inactive glacier ice existed in corries sheltered from the sun as recently as the late 17th century. This has now been disputed (Sissons, 1979) but it seems that glaciation in upland Scotland only represents a marginal deterioration, as Hawke and Champion (1954) demonstrated in their study of the altitudinal proximity of the mountains to glaciation, based on the frequency of melting of permanent snow patches. It seems more likely that there was an increase in both the size and number of permanent snow patches. Many articles cite the descriptions of permanent snow by early travellers eg. Taylor 1618 and Pennant 1772 (Hume, 1895). Lamb (1968) quotes from Taylor who visited Deeside about 1610 and wrote of the Cairngorms:

> "There I saw Mount Ben-Awne [Ben Avon] with a furr'd mist upon his snowie head instead of a nightcap. For you must understand that, the oldest man alive never saw but the snow was on the tops of the divers of these hills, both in summer, as well as in winter." (Brown, 1891 p120)

The most important factor in conclusion is that the most intense climatic periods only lasted 1-2 decades and thus, glacier ice is unlikely to have formed. The main characteristic of the Little Ice Age was that atmospheric circulation was more meridional than present, with long, cold winters and short, wet summers (Lockwood, 1977; Lamb, 1977). Thus, the late 17th century was the nadir of the Little Ice Age in Scotland. Continual severe weather conditions are reported in the Borders for over 15 years (1684-1699) with increased incidence of storms and flooding (Whyte, 1980). Lamb (1977) also reports cold, wet summers during the late 1600s within Speyside (from the Statistical Account of Duthil and Rothiemurchus). Bonacina (1927) reported the most snowy winters in the British Isles (1876-1925) as shown in Table 5.3.(i).

Table 5.3.(i)

Worst winters for snow according to Bonacina (1927)

Whole of British Isles:	1875-76; 1877-78; 1880-1881; 1885-1886; 1891-1892
Affecting Scotland :	1916-1917; 1918-1919

Therefore, some of the worse winters in Scotland actually occurred post-Little Ice Age.

There is evidence in the literature for all three study areas that periods between 1700-1800 were very wet, with poor summers and harsh winters (see Figure 5.3.(ii)). For example, the diary of the wife of the Laird of Kemnay gave in detail the weather from 1758-1795 in Aberdeenshire (Paul, 1881). Certain years were typified by wet and storminess eg. 1766:

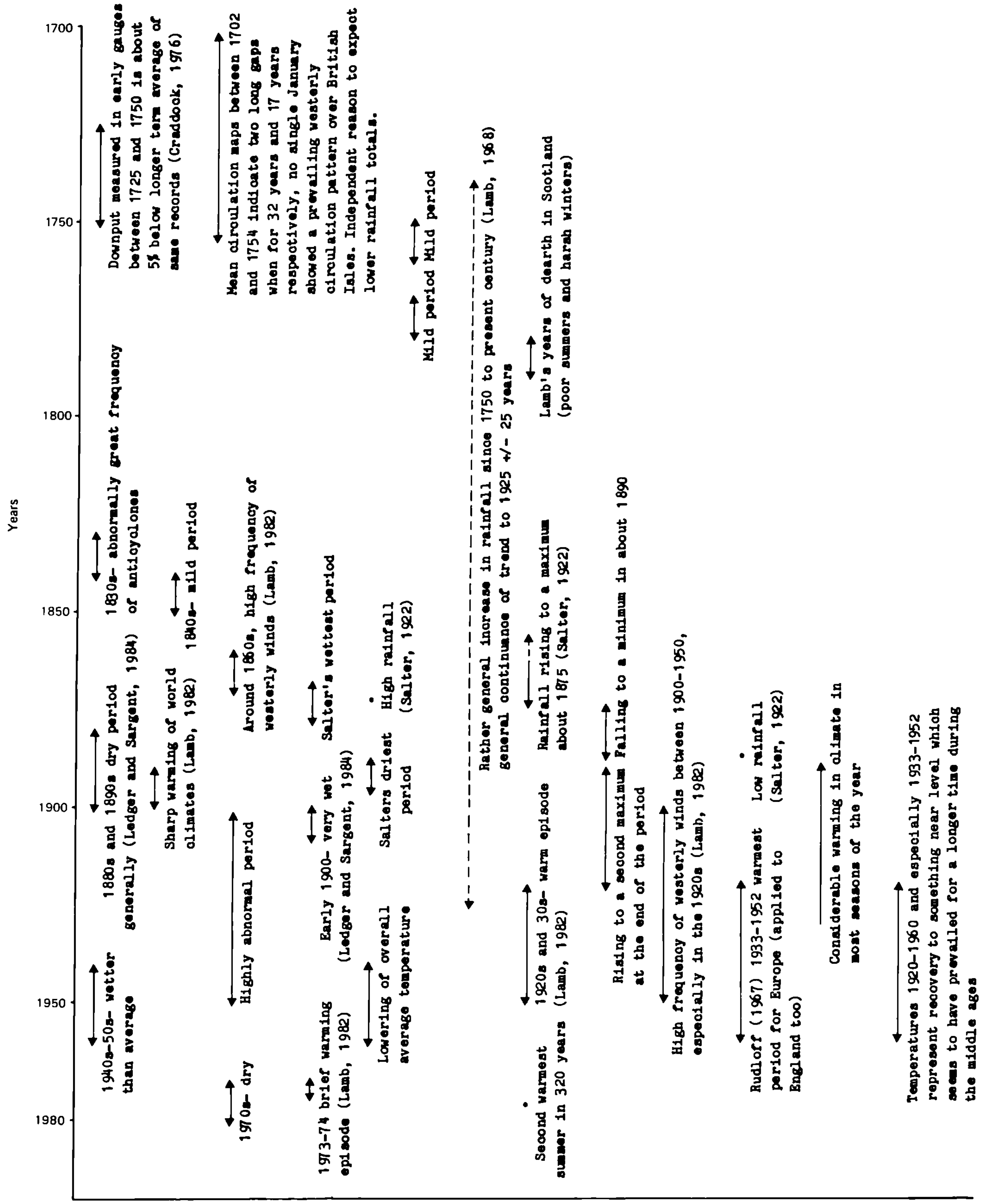
> "This year, 80 days rain, 35 days snow on the ground and 17 days wind." (Paul, 1881 p113)

The worst year was 1782 with frost setting in as early as August. This was succeeded in Aberdeenshire by two further bad years, namely 1783 and 1784. Thus, the 1780s were years of dearth in Scotland (Lamb, 1977). This is again substantiated by Symon's (1867) table of pre-1867 "remarkably cold and wet years" for the British Isles.

A second much shorter fluctuation occurred post-Little Ice Age, between 1890 and around 1938, when world air temperatures rose 0.5 ^{o}C (see Figure 5.3.(i)). The average temperature for January in central England was approximately 2.5 ^{o}C higher than prevailing temperature levels in the 1780s.

Figure 5.3.(i)

Documented climatic fluctuations within the 18th to 20th centuries



(Sources: Lamb 1971, 1977, 1982; Ledger and Sargent, 1984; Salter, 1922)

5.4 Dee study area: Magnitude and frequency of extreme runoff

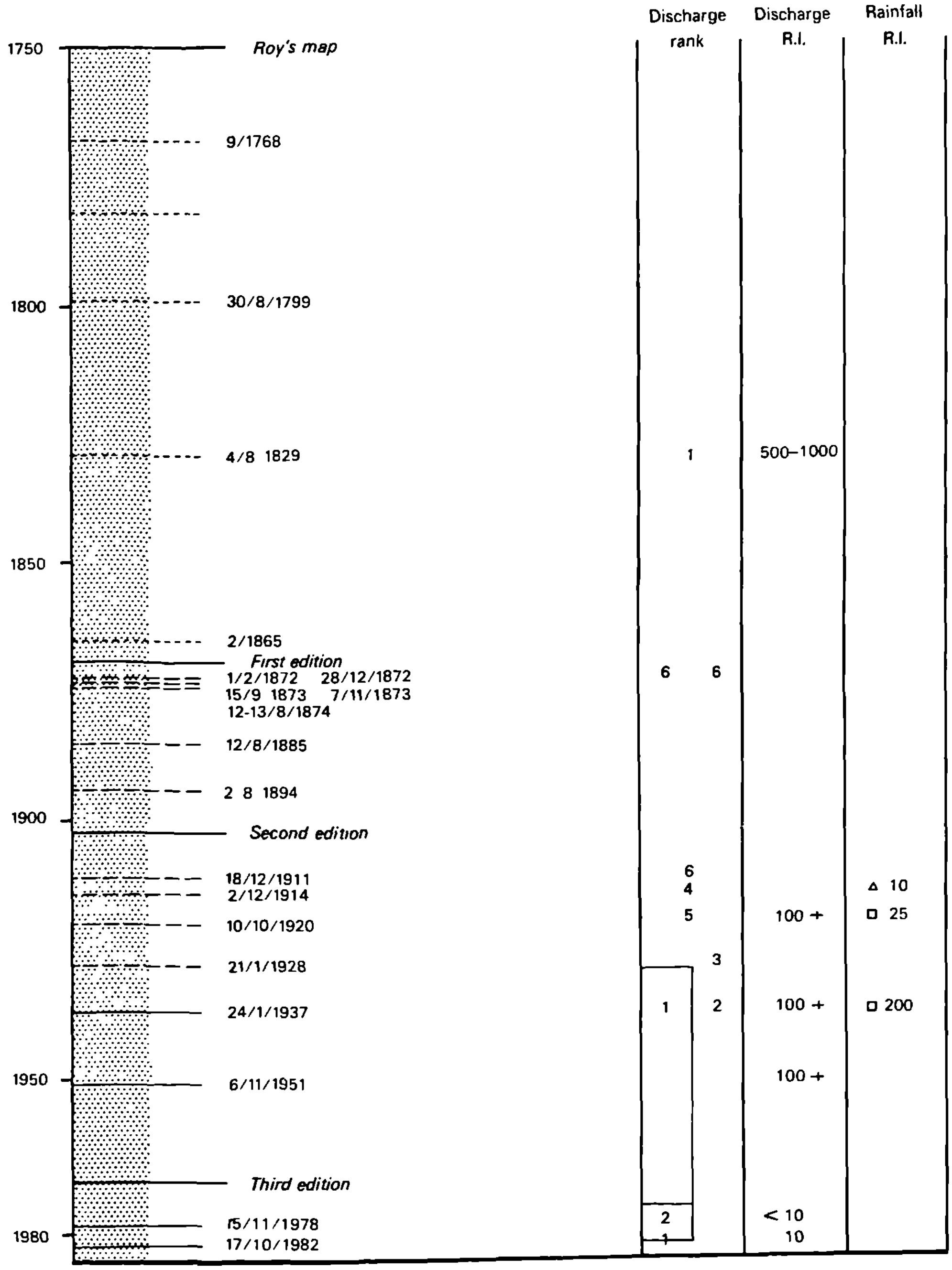
Reconstruction of the magnitude and frequency of extreme rainfall and runoff within the Dee study area will now be presented.

5.4.1 Documented flood history on the River Dee

The collated flood histories for upper Deeside are shown in Figures 5.4.1.(i) to (iv), with an attempt to place a recurrence interval on historic flood events; further details are found in Appendix 1.1 and 1.2. The earliest record of flooding on the Dee was on 2nd Feb, 1642, by Old Spalding (cited Lauder, 1830). Since that time however, several major flood events have been documented in the literature eg. 4th Aug, 1829 (Lauder, 1830) and 31st Jan, 1868 (Nairne, 1895). There are maximum flood height markings at several locations down the Dee but the flood plaques at Balmoral Castle Gardens are the only known ones within the study area. From these flood plaques, the ranking of flood events from 1872 to 1937 is given for that cross-section. Obviously each flood event will produce relatively different discharges at different locations downstream, depending on major areas of runoff contribution. Figure 5.4.1.(v) shows the ranking of flood events downstream from a variety of data sources, including Bremner (1922). However, the Balmoral, Danzig and Ballater ranking will of course be more indicative of stage in the upper catchment. If the Balmoral record is studied without knowledge of the pre-1872 events, it might be interpreted that flood magnitude has increased from 1872 to 1937. However, the 4th Aug, 1829 flood in upper Deeside, especially within the north to south facing

Figure 5.4.1.(i)

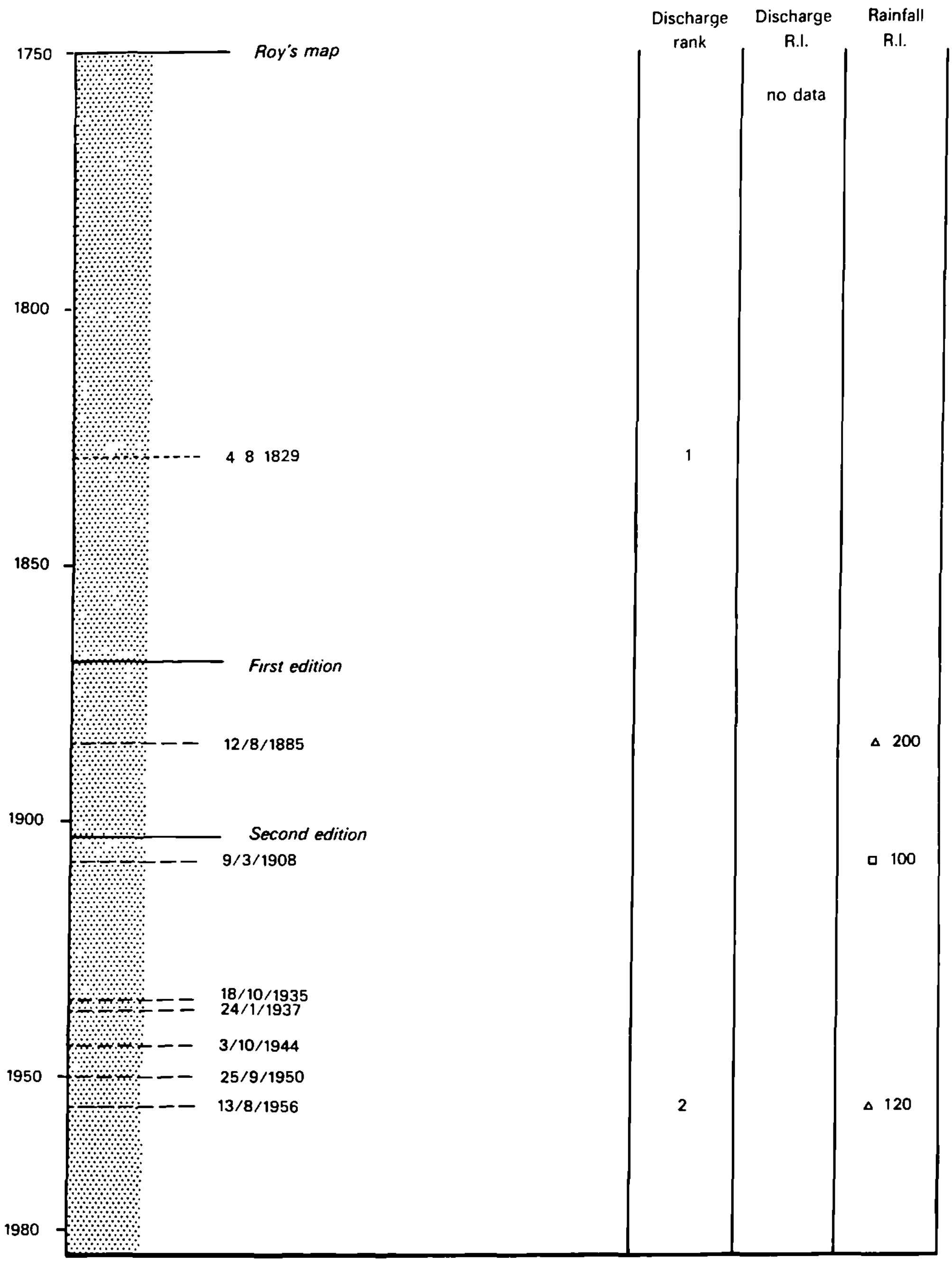
The flood history on the mainstream Dee above Crathie



Rainfall gauge: Braemar
Discharge gauge: Pollhollick

Figure 5.4.1.(ii)

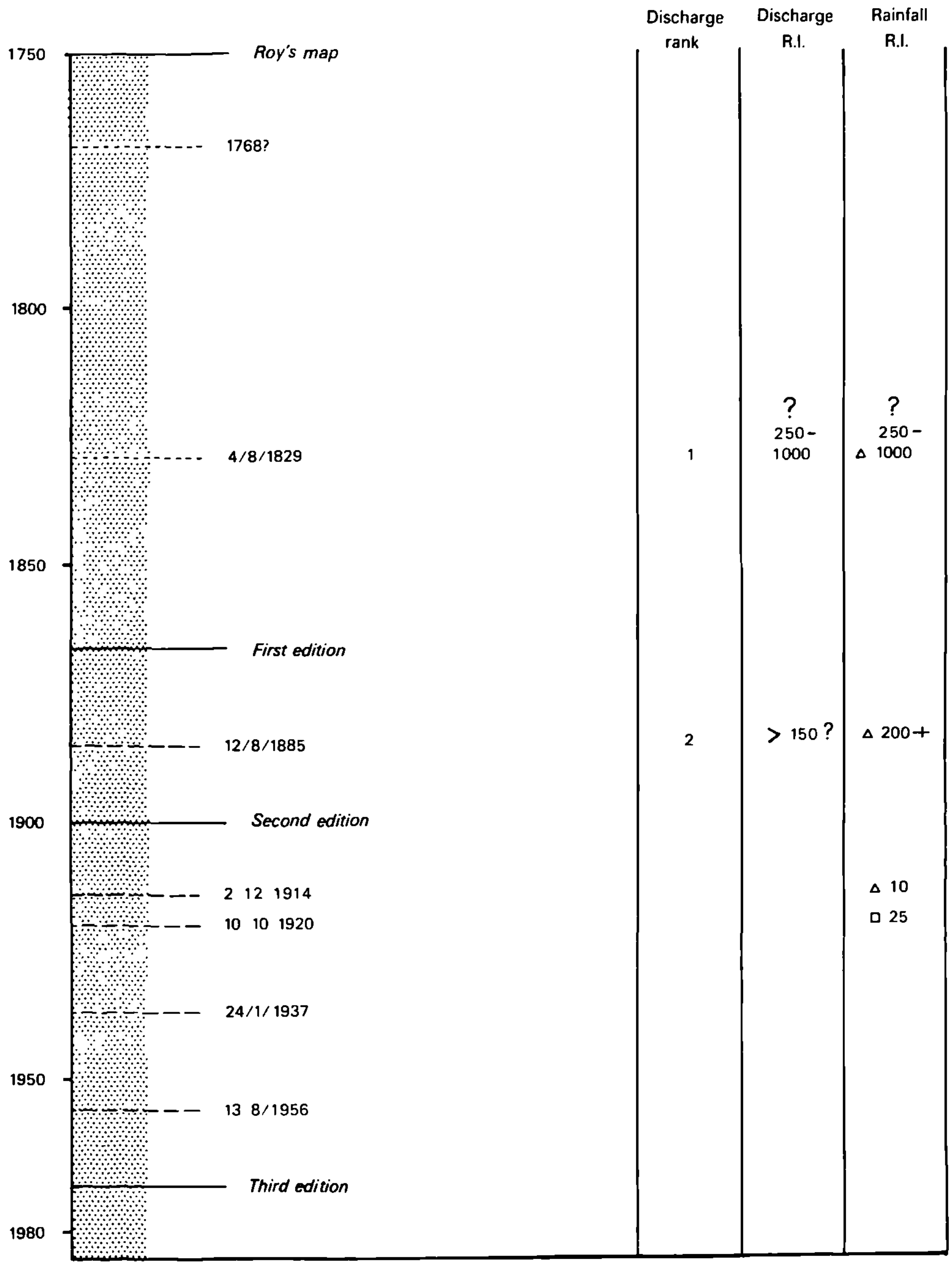
The flood history within the Lui Water catchment



Rainfall gauge: Derry Lodge

Figure 5.4.1.(iii)

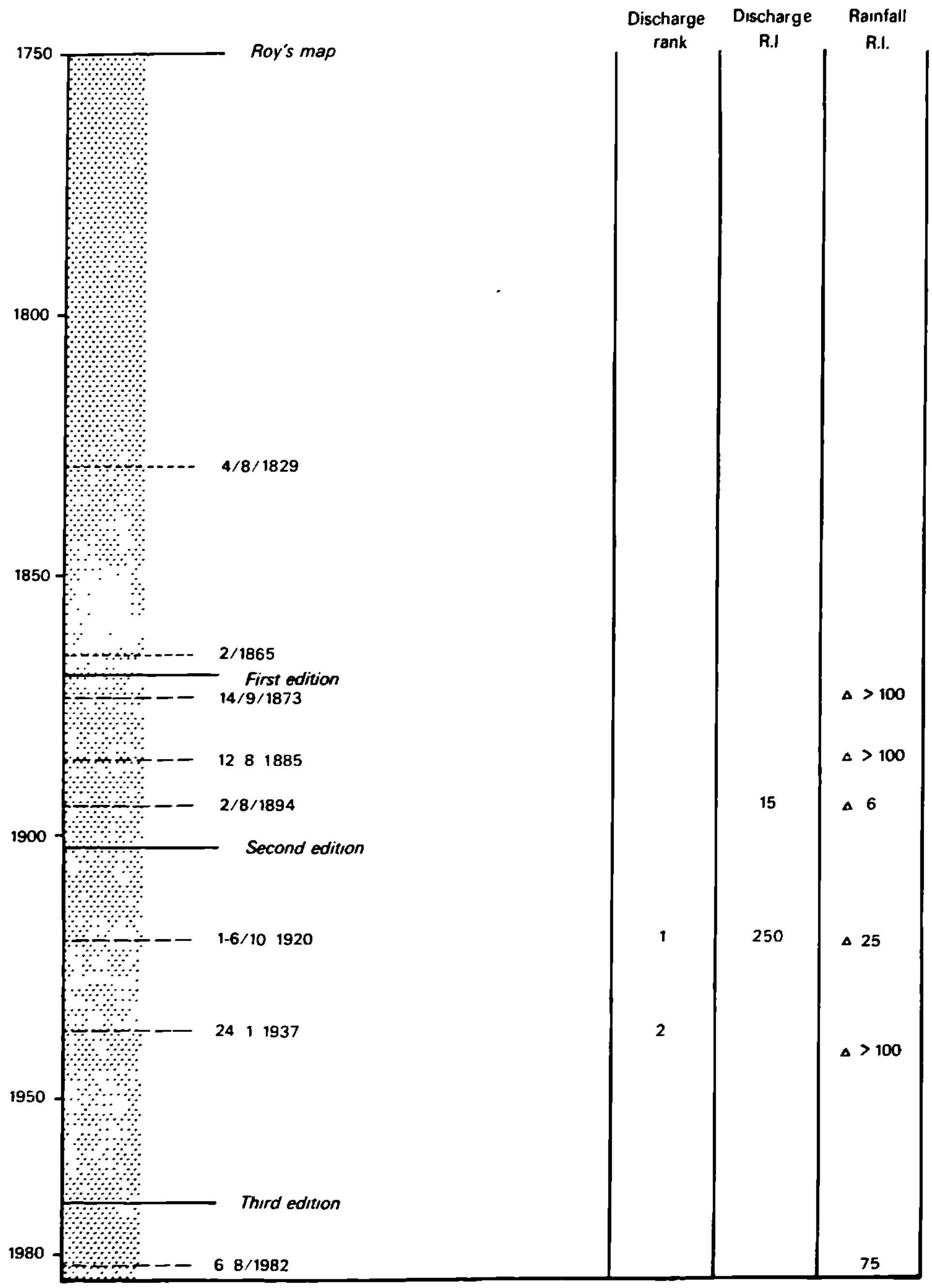
The flood history within the Quoich catchment



Rainfall gauge: Braemar

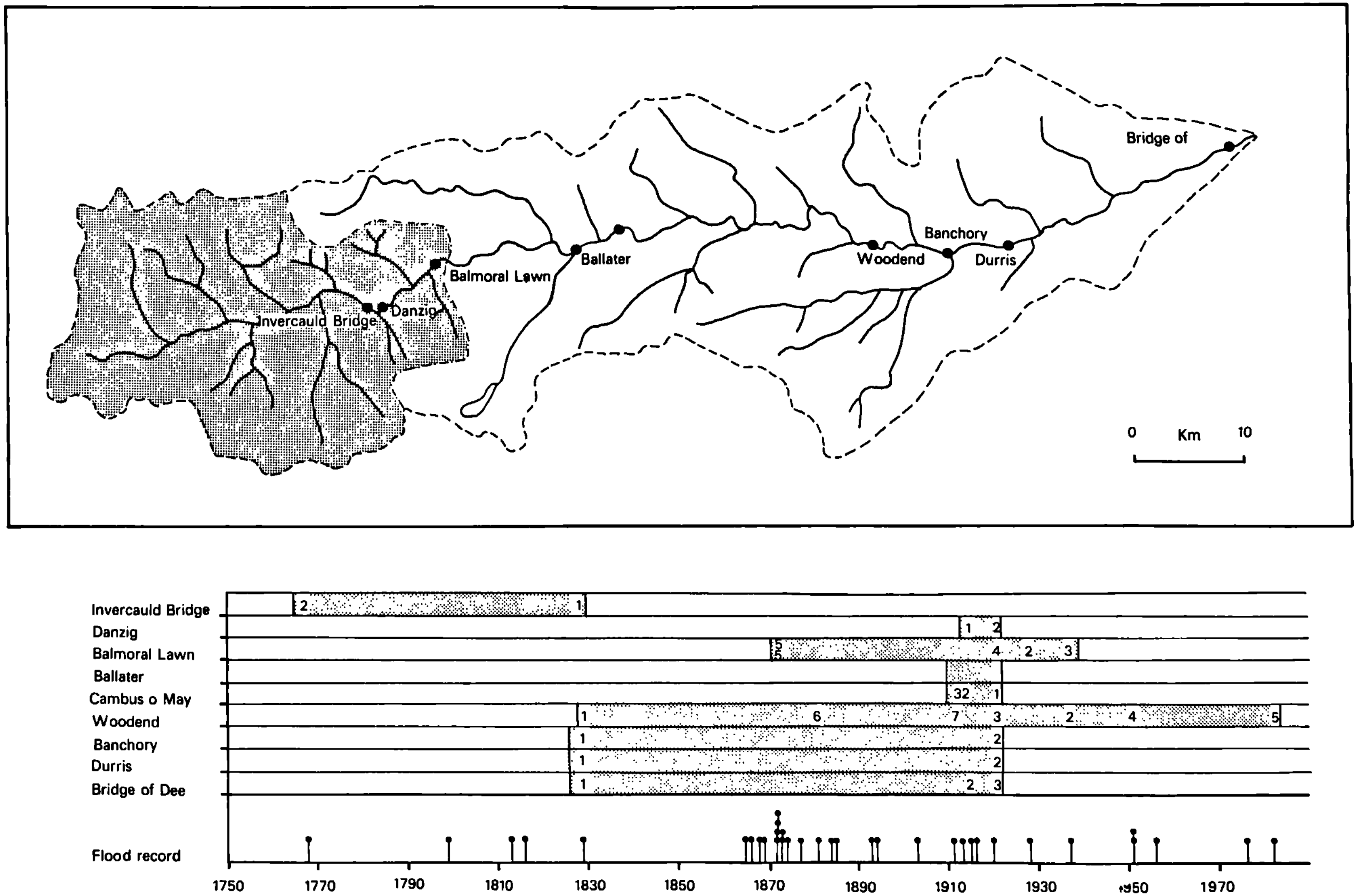
Figure 5.4.1.(iv)

The flood history within the Clunie catchment



The ranking of major discharge events within the Dee catchment

Figure 5.4.1.(v)



tributaries, was the most major event on record. In contrast, Roberts (1919) believed the floods were larger on the Dee, pre-1900:

> "There are marks cut on bridges and benches in the river, indicating the heights of phenomenal floods in 1829, 1877 and 1881, the heights of these floods are greater in every case (the greatest being 1829) than those in more recent years [written 1919] including the phenomenal flood of 1913."
> (Roberts, 1919 p6)

Of course, the major 1920 and 1937 events had yet to occur.

The seasonality of historic flood events (where month is recorded), over different time-periods, is shown in Figure 5.4.1.(vi) and the hydrometeorologic characteristics, where available, are tabulated in Table 5.4.1.(i). Typically, flood events associated with longer duration frontal storms occurred between December and February. However, August was the most frequent flood month when both the most extreme pre-1850 flood events occurred. Again, between 1850-1899, the importance of August as a flood month was highlighted. Higher extreme rainfall amounts must have been required to generate such floods as in summer SMD and potential ET rates are at their greatest. However, one has to be careful of the different roles of convective and frontal storms. Convective storms were more likely to affect individual catchments rather than cause a major event on the mainstream Dee. Lesser flood events, still of high magnitude, occurred within September to November, some associated with the autumn equinox. In contrast, post 1900 there seemed to be a considerable shift in the seasonality of major

Table 5.4.1.(i)

Hydrometeorological characteristics of historical flood events on the upper Dee

A Regional storm

B Convectional storm

C Intense rainfall

D Thunder and lightening

E Wind

F Wind direction

G Snow melt contribution

H Ice break

I Temperature rise

J Ground frost

K Pre-flood ground saturation

Event	A	B	C	D	E	F	G	H	I	J	K
02/02/1642				*	*	*					
00/09/1768	*										
04/08/1829	*		*	*	*		(*)				
00/02/1865							*	*			
00/03/1866							*				
31/01/1868	*										
00/10/1869	*				*						*
01/02/1872	*										
26/02/1872	*										*
25/10/1872	*										
28/12/1872	*										*
07/11/1873	*						*				
11/08/1884		*									
12/08/1885	*										
08/06/1893		*									
02/08/1894		*		*							
18/12/1911	*										
07/05/1913	*				*	E					
14/12/1914	*										
28/10/1915	*										
10/10/1920	*				*	SE					*
24/12/1937	*				*	NE					*
25/06/1951	*										*
06/11/1951	*				*						
28/07/1956	*										
13/08/1956		*									

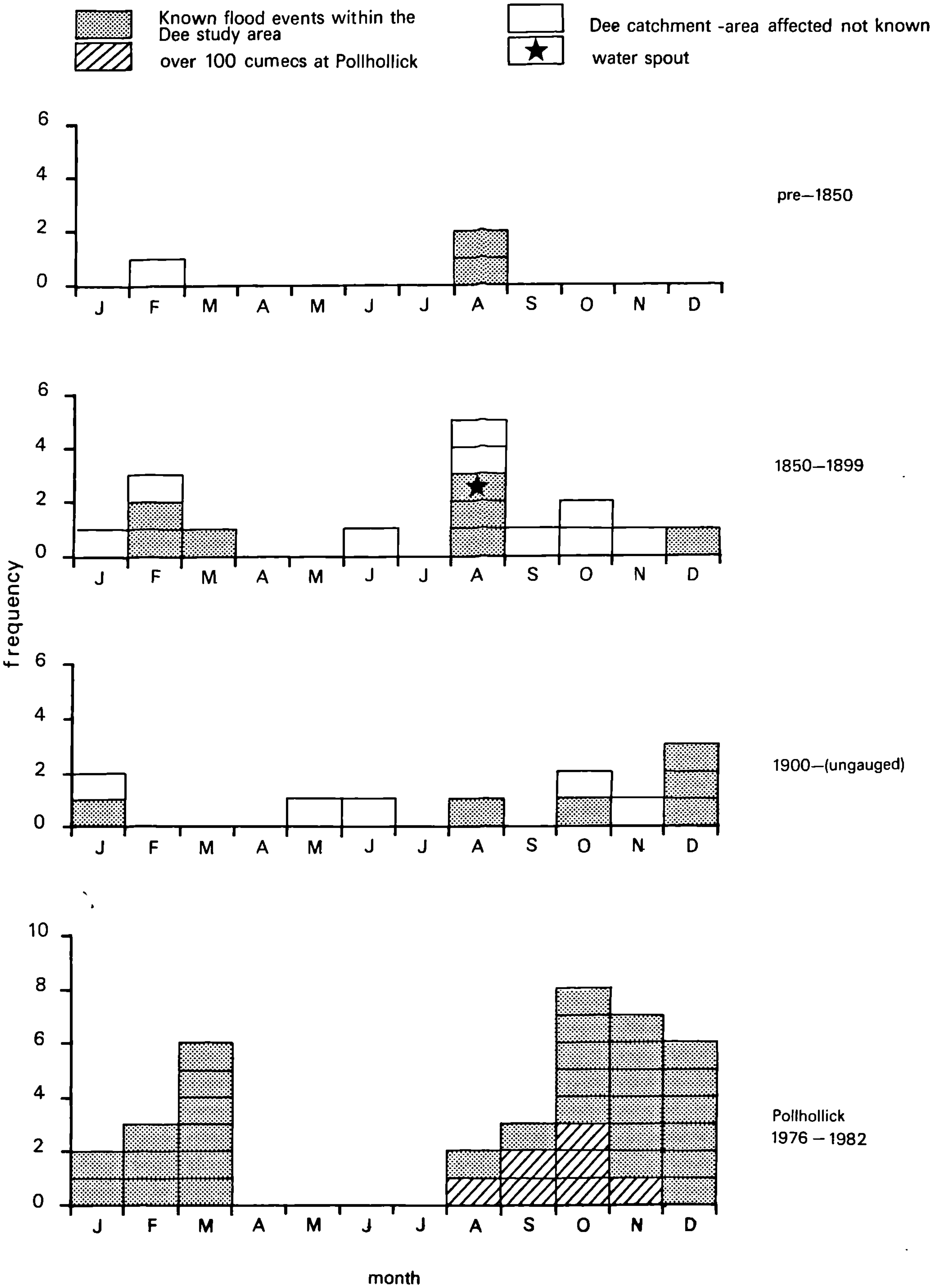
flood events (Figure 5.4.1.(vi)), with the majority falling between October and January, associated with longer duration frontal storms eg. 24th Dec, 1937. Looking at frequencies of flood events in qualitative terms, between 1850-1899 although the flood events were not the largest on record, there was a higher documented frequency of floods than during the subsequent period 1900-1976 (see Figure 5.4.1.(i)). In terms of synoptic origins, the majority of flood events were associated with longer duration storms of more than 24 hours. The only events with a major recorded snowmelt contribution were all 1850-1899 (Table 5.4.1.(i)).

5.4.2 Dee study area: Analysis of discharge records

The availability of discharge records on the River Dee was outlined in Section 3.2.5. The record at Woodend initiated by Captain W.N. McClean, the pioneer in the development of Scottish river flow measurement (Werritty, 1983), is of unusual and valuable length in the Scottish context (1929-1982). Unfortunately, there is no gauge within the Dee catchment above Crathie but the Polhollick gauge, though short in record, is located just above Ballater (690 km^2). Fortunately, no major tributaries join before that point. As well as continuous gauging, there have been single measurements and calculations of individual, historic flood flows, which provide a useful inter-event comparison of magnitudes and possible recurrence intervals.

Figure 5.4.1.(vi)

The seasonality of flooding on upper Deeside



The first step in analysing the discharge record was to extend the Polhollick record by correlating the peak over threshold data for the overlapping 1975 to 1982 period, with the corresponding discharges on the much longer Woodend record, of twice the basin area (1370 km^2). Of course this is approximate, as catchment characteristics are not necessarily constant. Flood pulses may be delayed in working their way downstream and also storms of different synoptic origins may cause varying rates of contribution from different parts of the catchment during different flood events. However, of the 38 peaks over a threshold of 96 m^3 s^{-1} at Polhollick (1975-1982), only 5 did not register a peak over threshold of 195 m^3 s^{-1} at Woodend. These presumably must have been localised, convective storm events in the upper part of the catchment. The correlation coefficient was encouragingly high at 0.86 and the regression equation was Equation 5.1

$$POT_p = 66.09 + (POT_w * 0.50) \qquad (5.1)$$

It seems that an event at Woodend in excess of approximately 400 m^3 s^{-1} is very likely to have a peak over threshold recorded at Polhollick. The actual POT at Polhollick were also correlated with the estimated POT for that period at Polhollick calculated by areal reduction and again the correlation coefficient was high (r=0.81). The difference is hardly significant, but upstream of Crathie with changing slope and density of the drainage network, such areal reduction may be more inaccurate.

The major flood events were then established from the gauged discharge data. From the Woodend record, two events stood out well above the rest of the series, namely 1st Jan, 1937 and 6 Nov, 1951, ranked first and second respectively. Using Equation 5.1, an approximation of the discharges at Polhollick was calculated, along with catchment runoff rates (see Tables 5.4.2.(i) and (ii)). The highest POT on the actual Polhollick record were on the 17th Oct, 1982 with 397.0 m^3 s^{-1} (0.58 m^3 s^{-1} km^{-2}) and 15th Nov, 1978 with 392.7 m^3 s^{-1} (0.57 m^3 s^{-1} km^{-2}) (see Appendix 2.1). There were also estimates of historical floods from the known stage at Cairnton (Woodend), from a variety of sources (Table 5.4.2.(i)). In certain cases, estimates of discharge were calculated from the areal runoff contributions provided by Roberts (1919). The most extreme event on record in discharge terms was the 4th Aug, 1829 flood with a stage of 5.93 m (FSR, NERC 1975), which gave a discharge of 1900 m^3 s^{-1}. Reducing these discharges for Polhollick, again using the regression Equation 5.1, produced the estimated results of Table 5.4.2.(ii).

It was then necessary to ascertain the expected RI of floods of this magnitude. When comparing these historical flood discharges with Acreman's estimates of floods of different RIs, derived from fitting the EV1 distribution to the limited discharge record as seen in Table 5.4.2.(iii), the outlier qualities of the extreme 1829 event were apparent. The discharge of 1900 m^3 s^{-1} at Woodend was two times the estimated 100 year event. The 4th Oct, 1920, 1st Jan, 1937 and 6th Nov, 1951 events all exceeded the 100 year RI of discharge (949.6 m^3 s^{-1}), ie. in just over 30 years, even when the regional growth curve was

Table 5.4.2.(i)

Stages of historic flooding at Cairnton

Flood	Stage (m)	Discharge (m^3 s^{-1})	Information Source
4/8/1829	5.93	1900	FSR (NERC, 1975)
1881	***	481 (approx)	Roberts (1919)
9/5/1913	***	317	Roberts (1919)
4/10/1920	4.2	1133	McClean records

Table 5.4.2.(ii)

Point estimates of discharge for major floods at Cairnton

Year	Woodend discharge (m^3 s^{-1})	Specific runoff (m^3 s^{-1} km^2)	Polhollick discharge (m^3 s^{-1})	Specific runoff (m^3 s^{-1} km^2)
4/8/1829	1900	1.39	1012	1.47
1881	481	0.35	306	0.44
9/5/1913	317	0.23	224	0.32
4/10/1920	1133	0.83	630	0.91
1/12/1937	1137	0.83	632	0.92
6/11/1951	1020	0.74	576	0.83
13/10/1982	683	0.50	408	0.59

Table 5.4.2.(iii)

Deeside: Magnitudes of discharge for different recurrence intervals, fitting the Extreme Value Distribution 1 (EV1) to the annual maximum series

Station	Q(2)	Q(5)	Q(10)	Q(20)	Q(50)	Q(100)
Polhollick	282.9	370.5	420.0	466.7	519.2	554.2
Woodend	407.6	555.0	654.7	745.8	862.9	949.6
Park Bridge	538.4	572.4	796.4	866.8	931.4	974.5
Polhollick+	262.5	350.0	423.0	528.0	618.4	723.4

Q(i) discharge with a recurrence interval of i years

All values are in $m^3 s^{-1}$

+ using regional curve ordinates

(Source: Acreman, Ph.D. Thesis in preparation) used. This makes one suspicious of the FSR (NERC, 1975) growth curves. In fact at Woodend, the 1920 and 1937 events were approximately equal in magnitude at that point. This can be compared with the earlier ranking at different locations on the Dee, provided from a variety of sources, especially with the series of flood plaques in Balmoral Castle Gardens and the ranking of flood events from 1872 to 1937 (Figure 5.4.1.(v)). The 1829 event clearly has not been exceeded over 1768-1984 and as such it must have a RI at the very minimum of 220 years. Discharge estimates from other sources (Thorne, Ph.D. in preparation; Acreman, Ph.D. in preparation) suggest the RI may be as high as 500-1000 years ie. a catastrophic event. A summary table for the results of the discharge analysis is given in Figure 5.4.1.(i).

5.4.3 Dee study area: Analysis of rainfall records

In the Dee study area and its lower catchment, there are several daily rainfall gauges, which extend their record back at least 70 years and a few beyond this date. The four data sets studied in detail are summarised in Table 5.4.3.(i) and their locations are shown in Figure 3.2.4.(ii). The Derry Lodge record provides a useful comparison due to its higher altitude (427 m). Unfortunately, the Derry Lodge rainfall record is broken by a shift in location of the gauge in 1957.

If the Derry Lodge and Braemar annual totals were correlated for the two separate segments of record, a highly significant correlation coefficient of 0.61 was gained as opposed to a similar value of 0.62 for the second segment. If the accumulated totals of both the Derry Lodge

Table 5.4.3.(i)

Studied long rainfall records in upper Deeside

(short records used in brackets)

Station	N.G.R.	Height (m)	Length of Record
Braemar	3152 7914	339	01/01/1857-31/12/1982
Balmoral	3260 7964	283	01/01/1904-31/12/1982
Ballater	3376 7964	193	01/06/1908-01/05/1952
Ballater (I.F.)	3380 7965	193	01/06/1951-31/12/1982
Derry Lodge (old)	3036 7932	427	01/01/1903-30/09/1957
Derry Lodge (new)	3037 7932	427	01/10/1957-31/08/1976
[Danzig Shiel	3202 7904	320	01/01/1971-30/9/1983]

(Length of record refers to monthly totals. There are occasional gaps in the daily data set, for example the Braemar daily record has 1906-1912 missing).

Table 5.4.3.(ii)

Ratioed annual rainfall data: Derry Lodge versus Braemar

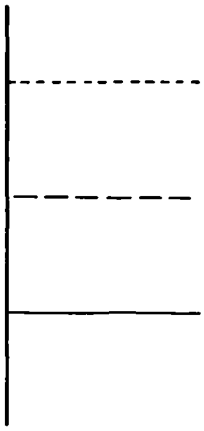
	Average annual ratio
Record 1 (1903-1957):	0.916
Record 2 (1957-1976):	0.856

Therefore the conversion factor to extend the longer record up to 1976 is: 0.856/0.916 = 0.935

and Braemar records were plotted as a double-mass plot, there was no obvious break in the scattergram. The annual ratios were calculated as seen in Table 5.4.3.(ii). A composite record for Derry Lodge was thus constructed. With the altitudinal difference however, it may be that the discrepancy associated with raingauge shift will be less than the fluctuations in difference between Derry Lodge and Braemar. Fortunately, the correlation coefficients were significantly high. A similar problem occurred with the Ballater record and it was homogenised in the same way.

It was first necessary to study annual total rainfall to assess the range within which fluctuation took place. In terms of the maximum and minimum range of values of total annual rainfall at each of the stations, the wettest year on record was 1872 on the Braemar record with 1504.9 mm; the second was 1320.1 mm in 1982 and the third, 1916 with 1214.0 mm (Figure 5.4.3.(i)). Although the Braemar record is the only record with a value for 1872, the ranking for the other years seemed otherwise confirmed by the Balmoral and Ballater (composite) records (Figures 5.4.3.(ii) and (iii)). 1872 was supposed to be the wettest year in the Aberdeen district since 1782 (British Rainfall, 1872). This high rainfall period (1872-1875) corresponds to Salter's (1921) wettest period (Figure 5.3.(ii)). For the higher altitude Derry Lodge station, with a higher average rainfall, the wettest years were slightly different, namely 1903 (1332.5 mm), 1923 (1271.8 mm), and 1946 (1235.5 mm), not following the consistency of lower altitudes (Figure 5.4.3.(iv)).

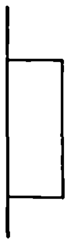
Key for flood history diagrams



contemporary source

rainfall record

discharge record



rank within the gauged record

△ *24 hour rainfall*
□ *48 hour rainfall*
■ *72 hour rainfall*

All rainfall and discharge recurrence intervals in years

Figure 5.4.3.(i)

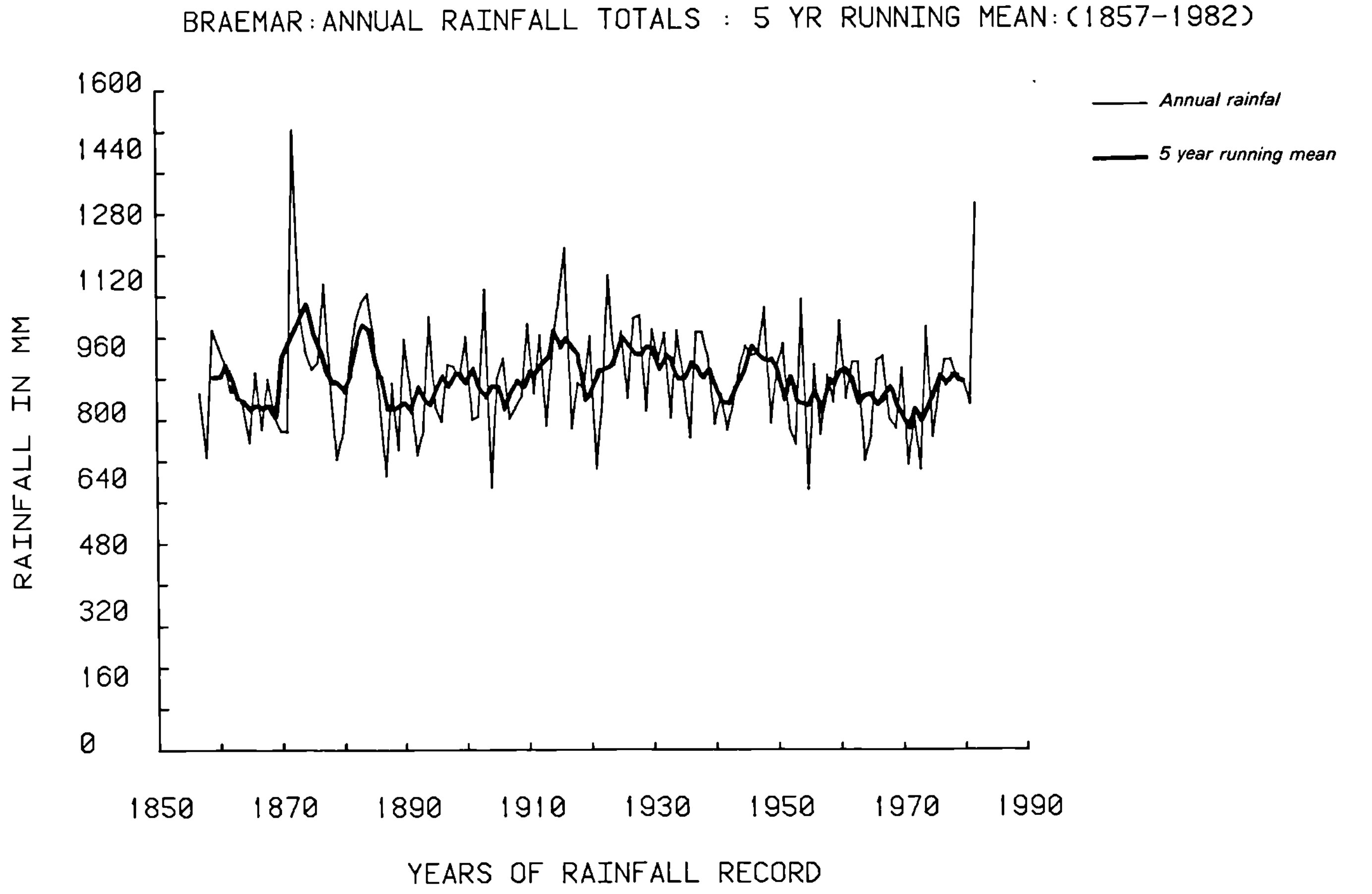


Figure 5.4.3.(ii)

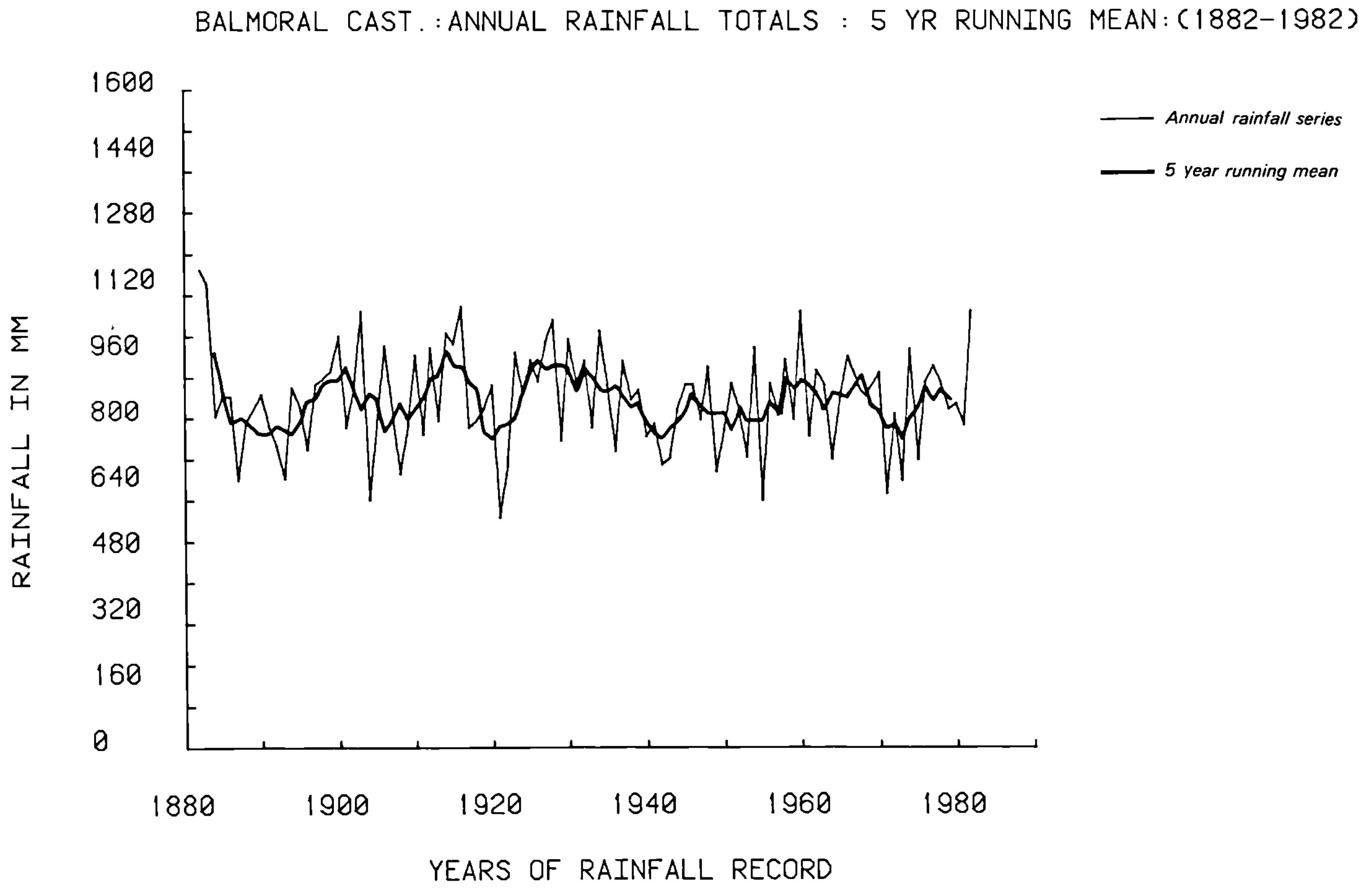


Figure 5.4.3.(iii)

BALLATER:ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN:(1910-1982)

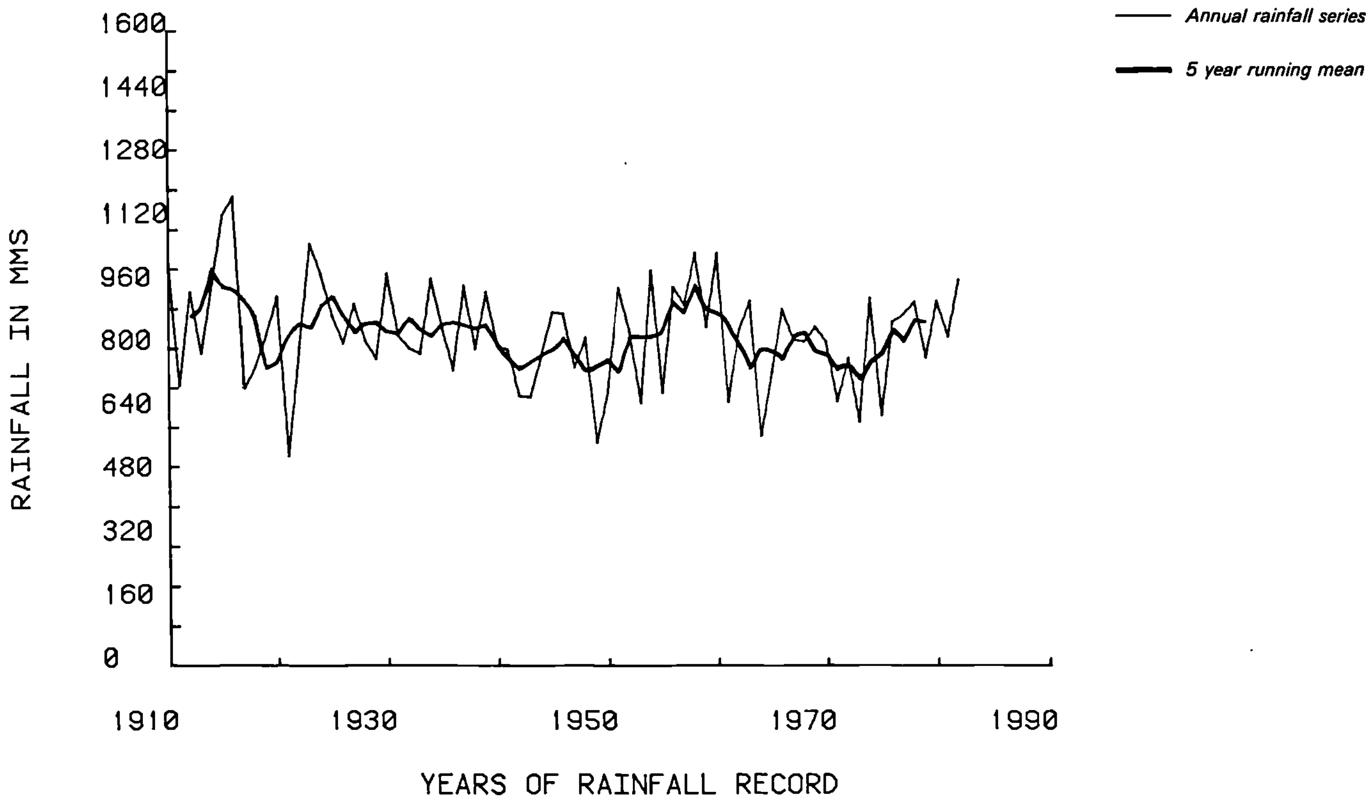
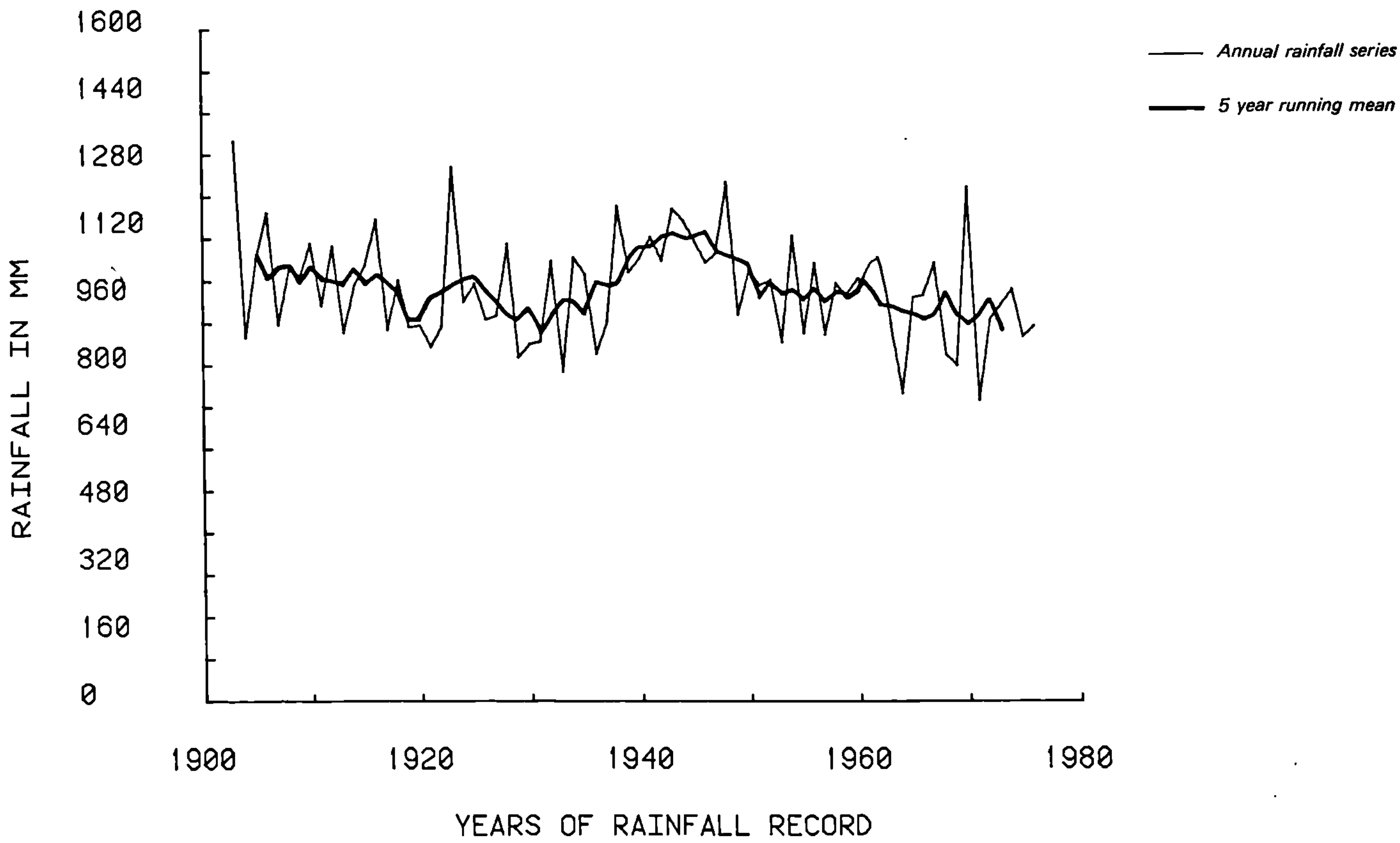


Figure 5.4.3.(iv)

DERRY LODGE:ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN:(1903-1976)



To see if any oscillations had occurred, the Braemar annual total series was plotted with a 5-year running mean, showing peaks at mean years of 1872-75, 1883-1884 and 1914. On the Derry record, the longest run of "above average" years (13) was 1938-1950, as can be seen in Figure 5.4.3.(iv), each with an annual total in excess of 1000 mm except 1949 (991.6 mm). When the Braemar and Balmoral annual total data were tested by the Kolmogorov-Smirnov and runs test, the Z statistics showed that the data set was not significantly different from a normal distribution.

When Box-Jenkins time-series analysis was carried out on these annual total series to assess any periodicities, highly significant autocorrelations (AC) were found at a number of years lag. Initially, each station will be discussed individually and then it will be seen if some common pattern occurs as would be expected with regional rainfall. The Braemar record had significant AC at 12 (0.13) and 22 (0.20) years lag and this pattern was confirmed by the spectral density analysis. A highly negative autocorrelation was also found at 27 years lag (-0.21). In terms of confirmation from other records, although the Balmoral record had a positive AC at 22 years lag, it did not exceed 2 S.E. The most significant AC was at 14 years lag with a value of 0.16. Again however, there was a high negative AC at 27 years lag (-0.27) and this AC stood out too when partial auto-correlation was carried out. At Derry Lodge, the annual total series showed high positive AC at 4, 35, 38 and 42 years but no significant periodicity. This 42 year periodicity was substantiated by the existence of a large peak in the spectral frequency analysis at just over 40 years and another at around

3-4 years. Therefore, in terms of common periodicity (ie. not just the peculiarities of one record), there was considerable variation and no predominant cycle but the 4 and 22 years lag consistently gave high positive, if not significant, AC. The high negative lag at 27 years also seemed consistent, possibly suggesting a high positive autocorrelation at 54 years lag.

5.4.4 Dee study area: Analysis of the rainfall POT series

The temporal distribution of 24 hour POT against magnitude for each station, are seen in Figures 5.4.4.(i) to (iii)). If initially years were assessed purely according to their frequency of occurrence of POT, certain years stood out. The maximum number of POT > 1 inch (25.4 mm) in 24 hours was 12 in 1982 on the Braemar record (see Figure 5.4.4.(i)), while at Danzig Shiel gauge, 14 POT were recorded in 1982. Considerable spatial variation was apparent however because at Balmoral for the same year there were only 5. It should be noted that 1872, the wettest year on the Braemar record, had only 5 POT ie. suggesting more continuous rainfall throughout the year. If 24 hour POT were grouped by decade (Figure 5.4.4.(vii)), 1870-1879 had the highest frequency with 34 cf. 1970-1979 with 19. In terms of higher magnitude 24 hour events (Figure 5.4.4.(iv)), 1873 was an exceptional year with 4 POT that were in excess of 1.5 in (38.1 mm), and 3 in excess of 2 in (50.8 mm).

Figure 5.4.4.(i) BRAEMAR:YEARS WITH 24 HR POT 1857-1982

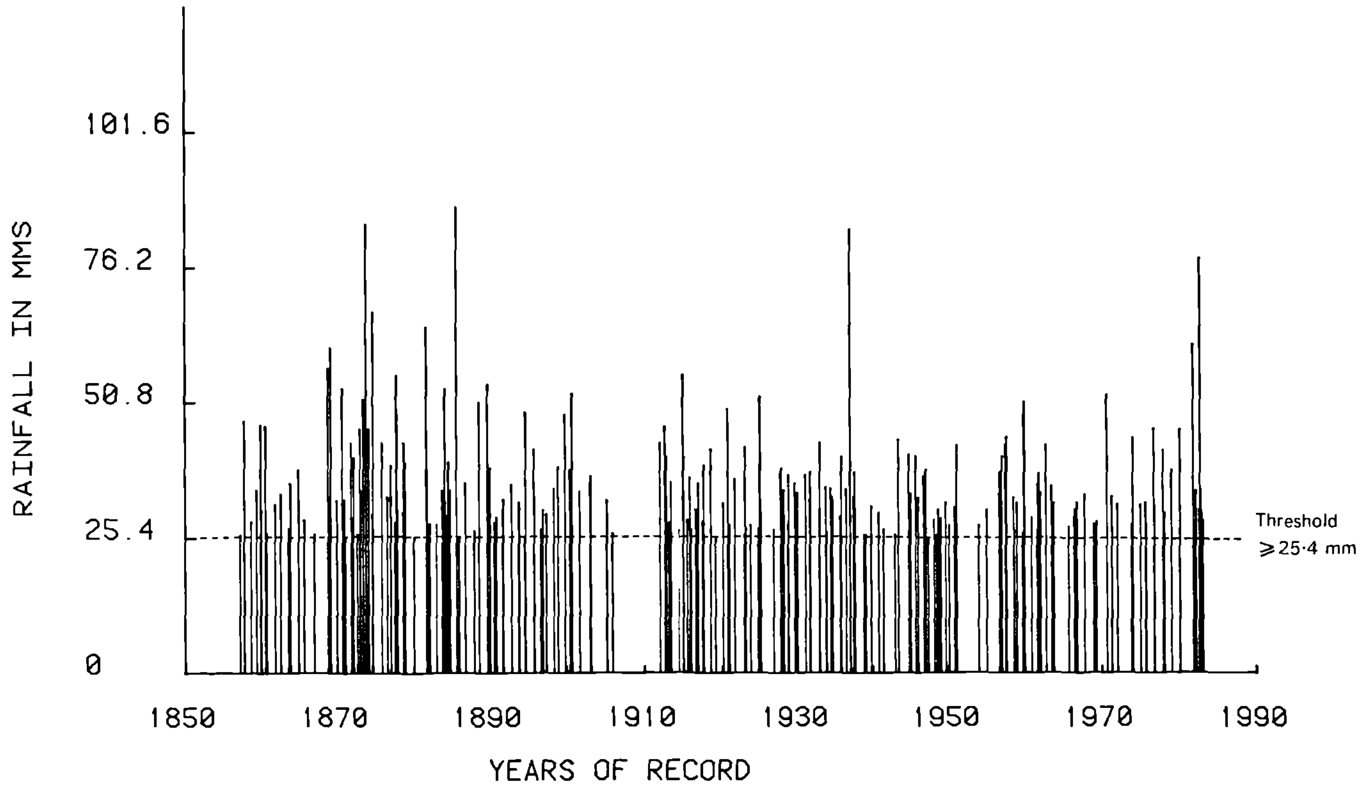


Figure 5.4.4.(ii) BALMORAL CASTLE:YEARS WITH 24 HR POT 1906-1982

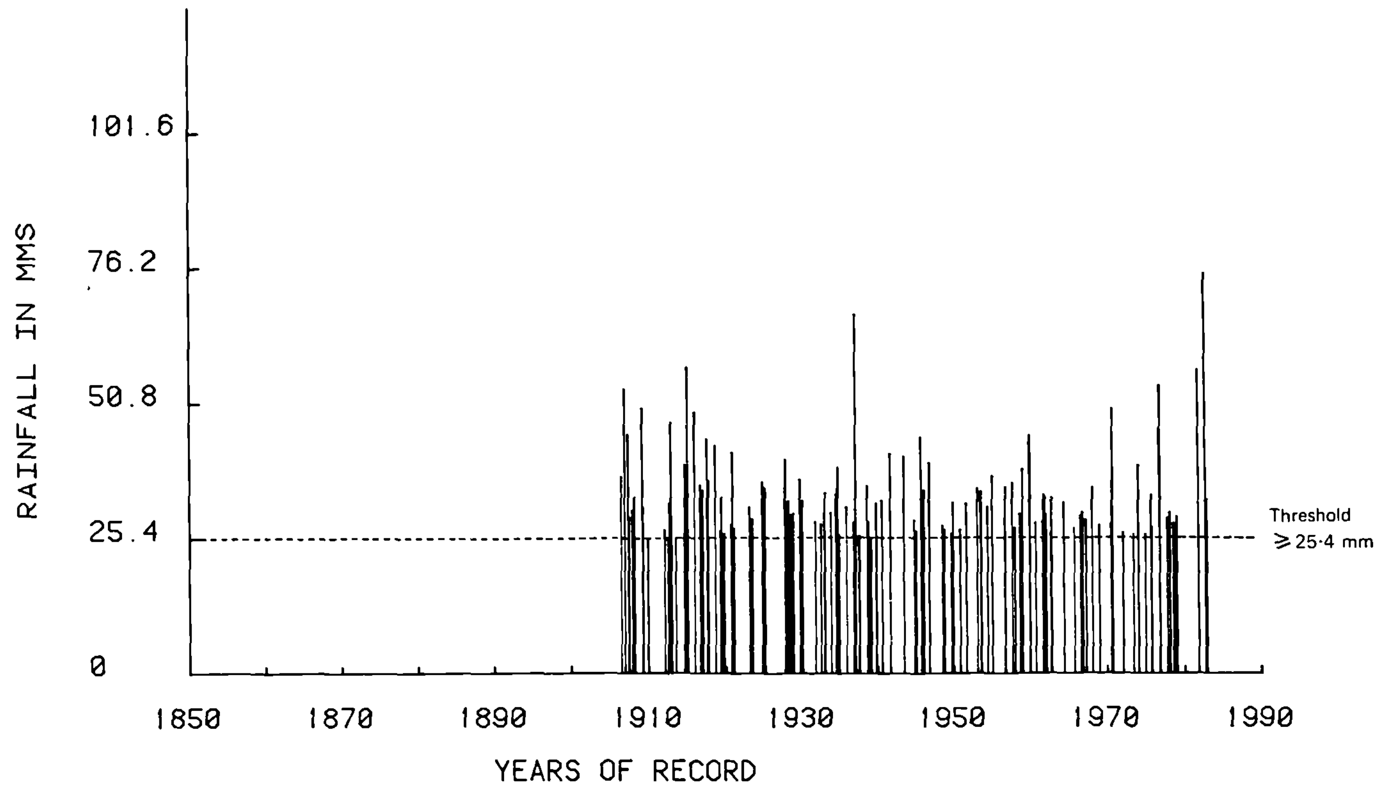


Figure 5.4.4.(iv)

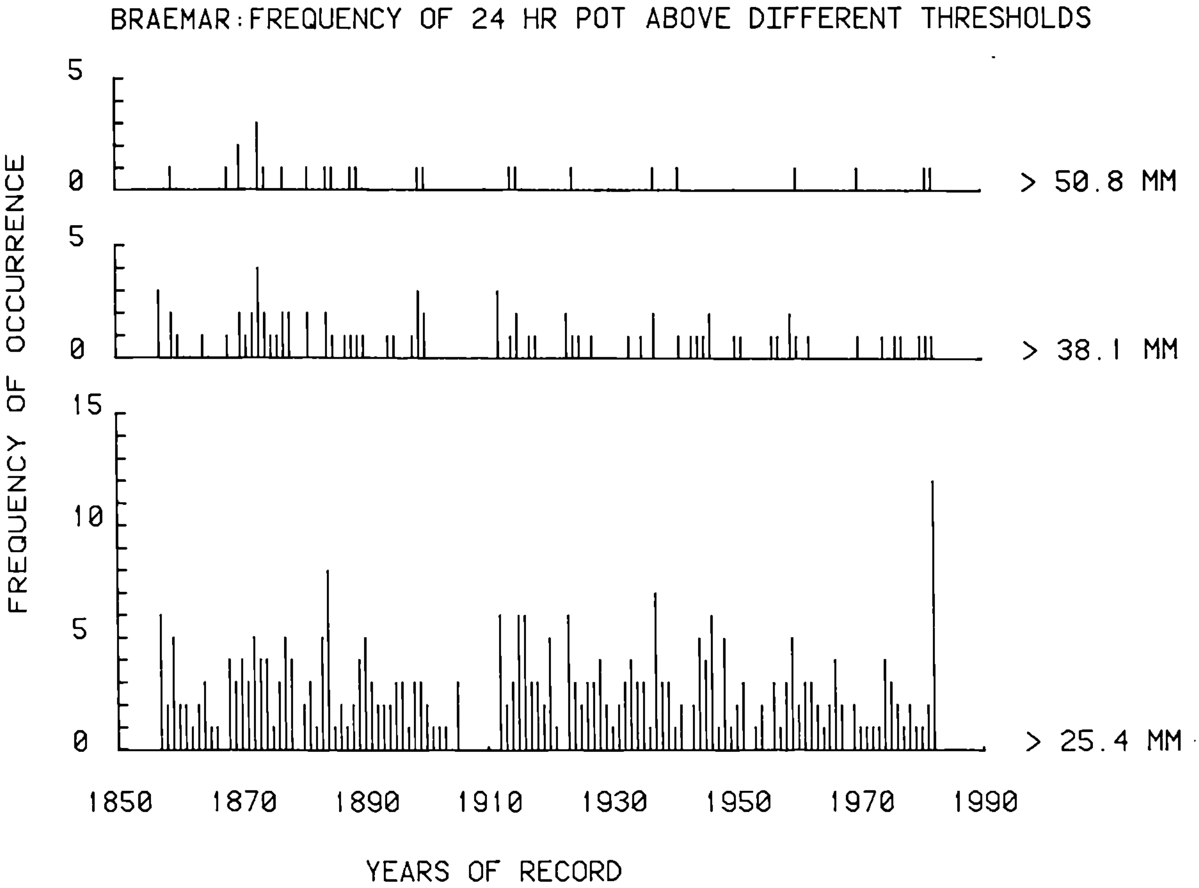


Figure 5.4.4.(v)

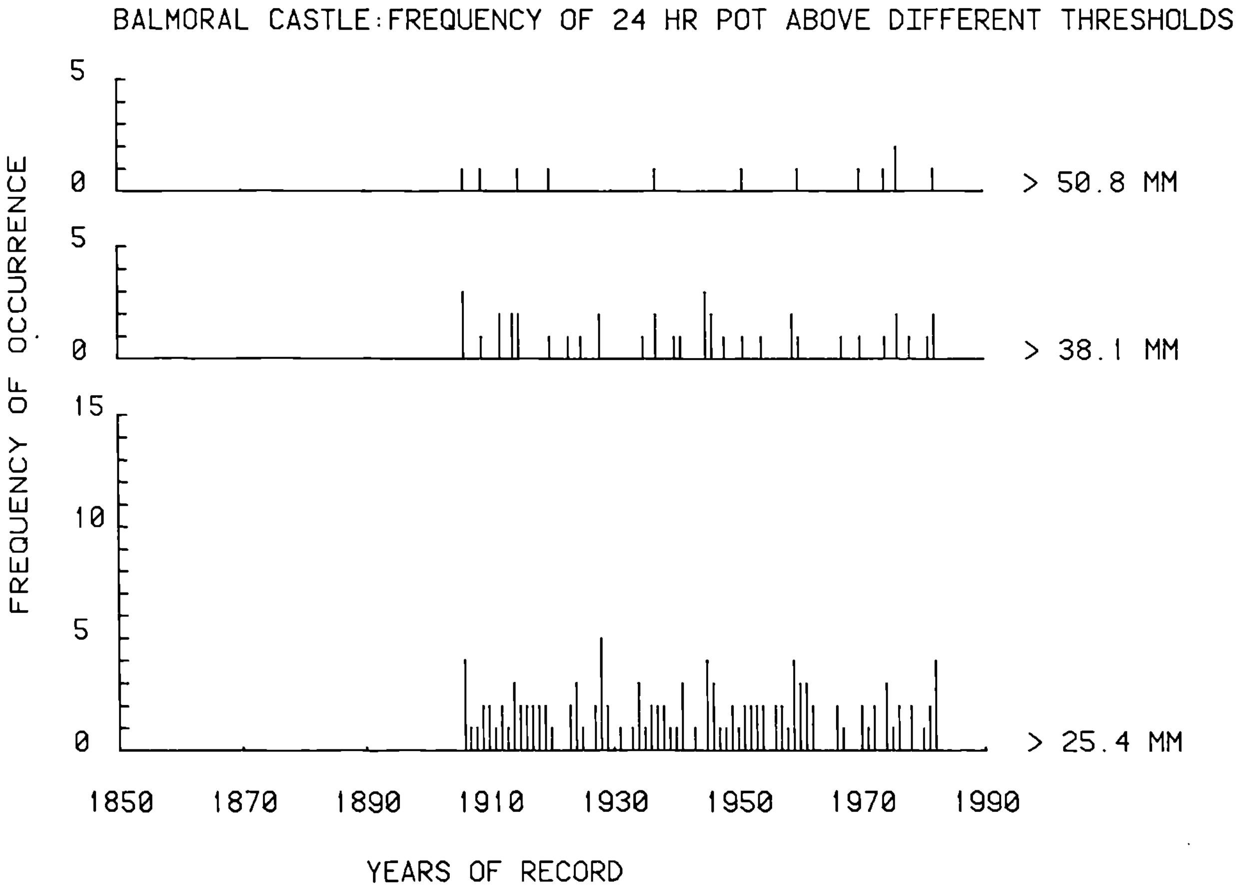


Figure 5.4.4.(vi)

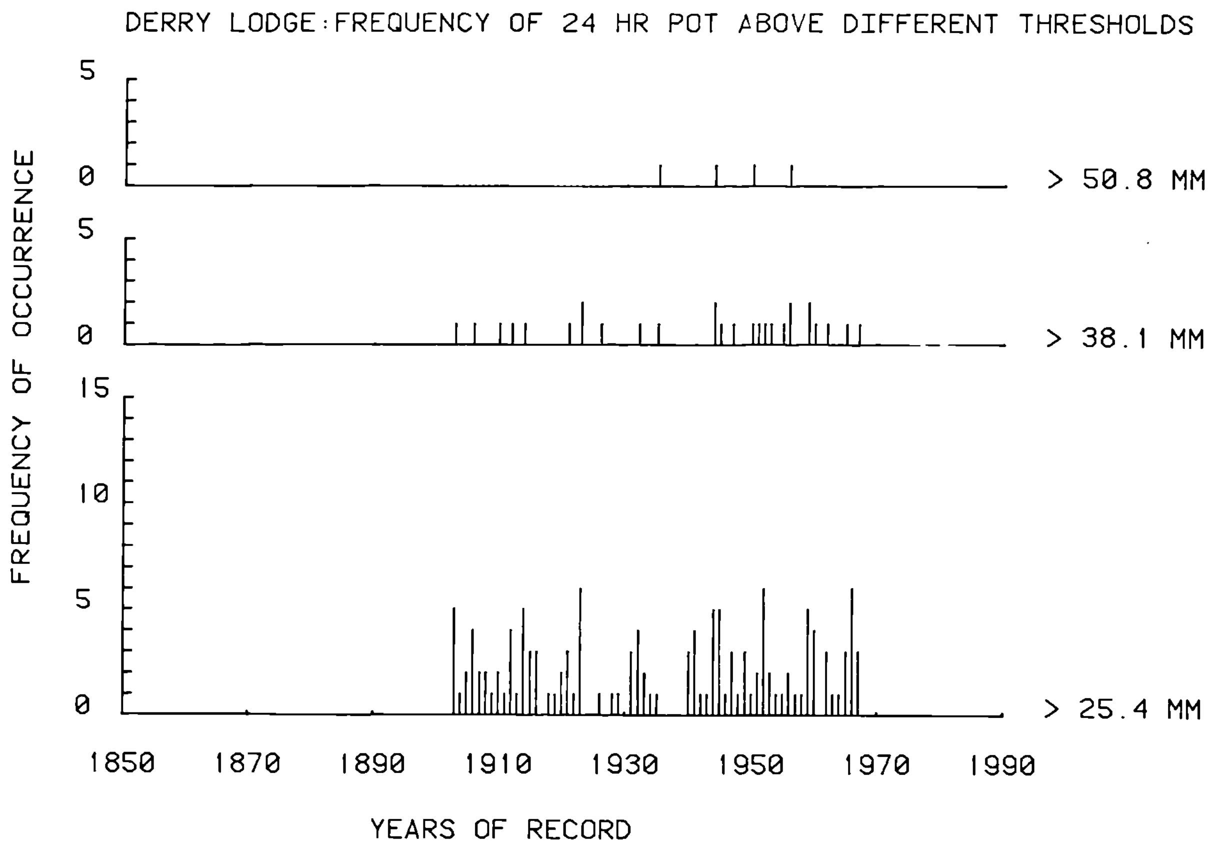
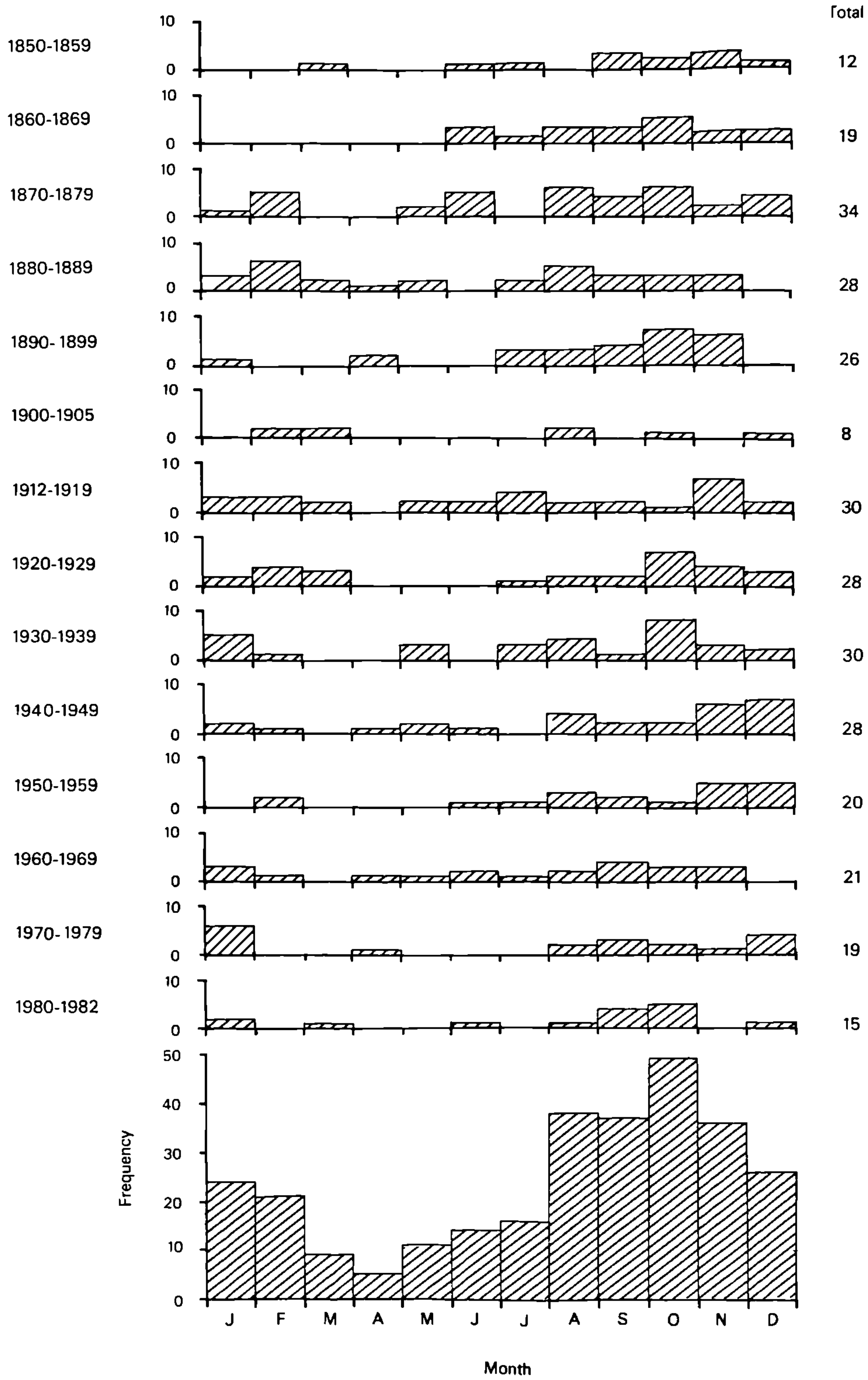


Figure 5.4.4.(vii)

The seasonality of 24 hour POT at Braemar (subdivided by decade)



The increased frequency of rainfall POT of the 1870s period is obvious when the temporal distribution of POT in excess of 2 in (50.8 mm) was studied (Figure 5.4.4.(iv)). Only 4 years had one fall in the range of 3-3.5 in (76.2-88.4 mm) in 24 hours- 1873,1885,1937 and 1982. The wettest period in terms of 24 hour POT frequency was from 1868 to 1878 when only one year (1875) had less than 3 POT in a year. From the Balmoral record, the years with the highest frequency of 24 hour POT were 1914 (6), 1928 (7) and 1937 (6); all years associated with major flood events (Figure 5.4.1.(i)). The differences in frequency of occurrence even with a slight altitudinal change of 56 m and a distance of 9.5 km must be noted; the decline in frequency of 24 hour POT can be seen in the comparison of Figures 5.4.4.(iv) (Braemar) and 5.4.4.(v) (Balmoral). Derry Lodge, in contrast to the expected higher frequency of POT, seemed to be fairly similar in frequency of events of the same magnitude as in the Braemar record (compare Figure 5.4.4.(iv) with Figure 5.4.4.(vi)).

To assess the relationship between the number of POT and actual annual total rainfall, thus enabling predictions to be made for individual stations where daily data was missing, the two series were correlated. For the Braemar station, a positive correlation of 0.56 was found for 24 hour POT > 1 in (25.4 mm) but for 24 hour POT > 1.5 in (38.1 mm) and > 2.0 in (50.8 mm), the correlation reduced considerably to 0.32 and 0.23 respectively.

When the temporal distribution of 48 hour rainfall POT was broken down in relation to the exceeding of different thresholds at each station, the results are plotted in Figures 5.4.4.(viii) to 5.4.4.(ix). The maximum number of 48 hour POT at Braemar in excess of 1.5 in (38.1 mm) was 9 in 1982, but other years stood out as having above average frequency. Of the 4 years on record with two 48 hour peaks in excess of 2.50 in (63.5 mm), all 4 were pre-1900 (Figure 5.4.4.(viii)). Thus clustering of 48 hour events in excess of 50.8 mm was even more pronounced than with 24 hour events (Figure 5.4.4.(ix)). Within the Balmoral record, the only year with two POT in excess of 2.50 in (63.5 mm) was 1976 (Figures 5.4.4.(x) and (xi)), thus confirming the pattern suggested by the Braemar record.

5.4.5 Dee study area: Seasonality of POT

In terms of the seasonality of events of different durations, both the 24 and 48 hour series showed a similar distribution. In terms of pure numbers, the greatest number of 24 hour events occurred from August to January. However, from Figures 5.4.5.(i) and (ii), the concentration of high magnitude events from mid-July to mid-August can clearly be seen and must be explained either by summer convective or frontal storms. These will only generate high magnitude discharge events if absolute rainfall amount greatly exceeds soil moisture deficit, or after antecedent rainfall temporally reduces this deficit. In terms of frequency, in the Braemar record 42% of the 24 hour POT > 1 in (25.4 mm) fell between September to November. This is very interesting because of

Figure 5.4.4.(viii) BRAEMAR: FREQUENCY OF 48 HR POT ABOVE DIFFERENT THRESHOLDS

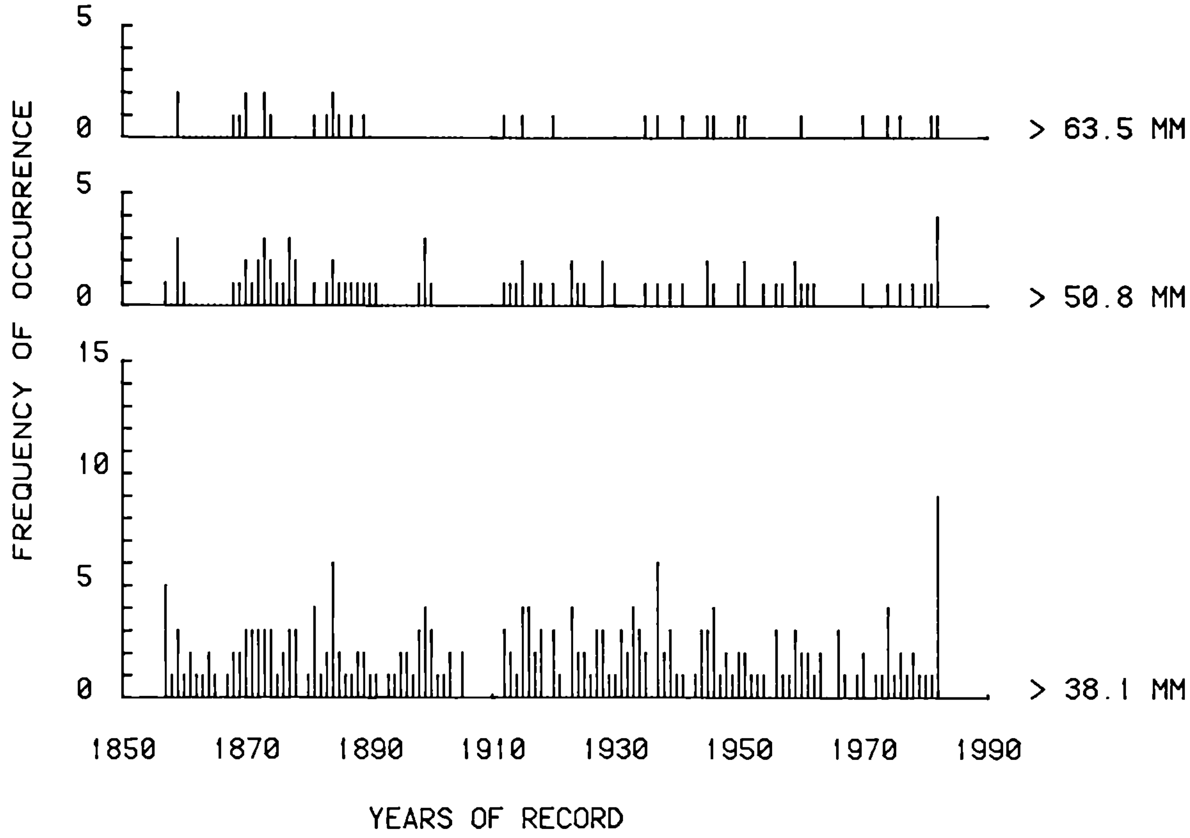


Figure 5.4.4.(x) BALMORAL CASTLE: FREQUENCY OF 48 HR POT ABOVE DIFFERENT THRESHOLDS

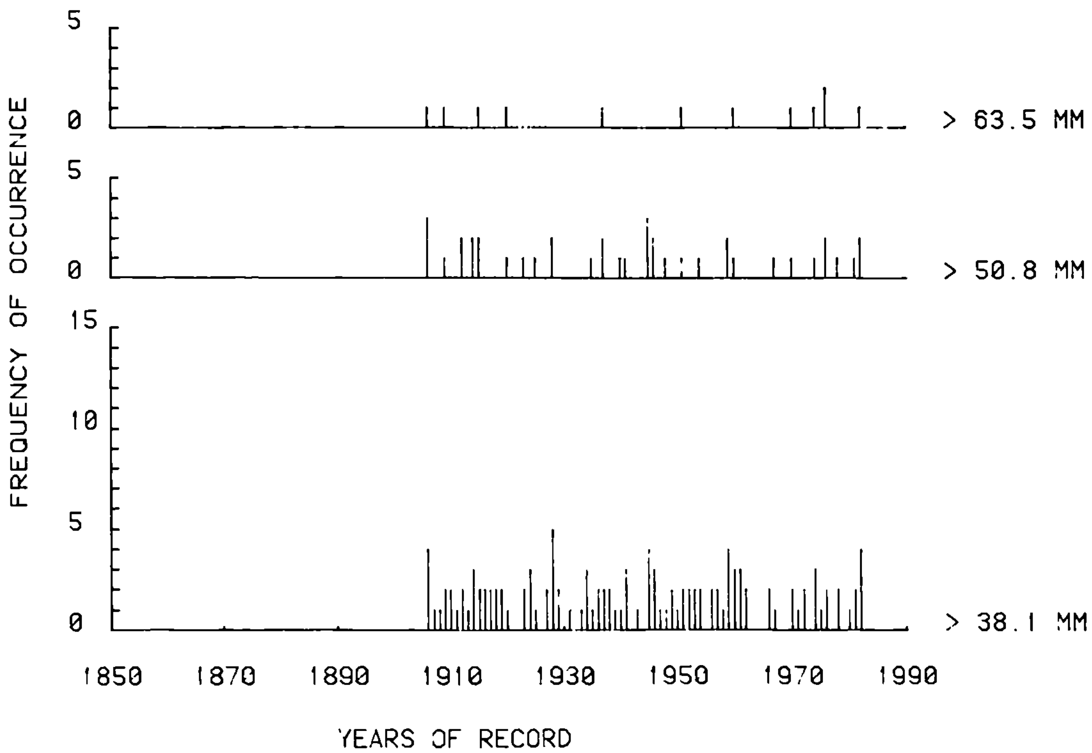


Figure 5.4.5.(i)

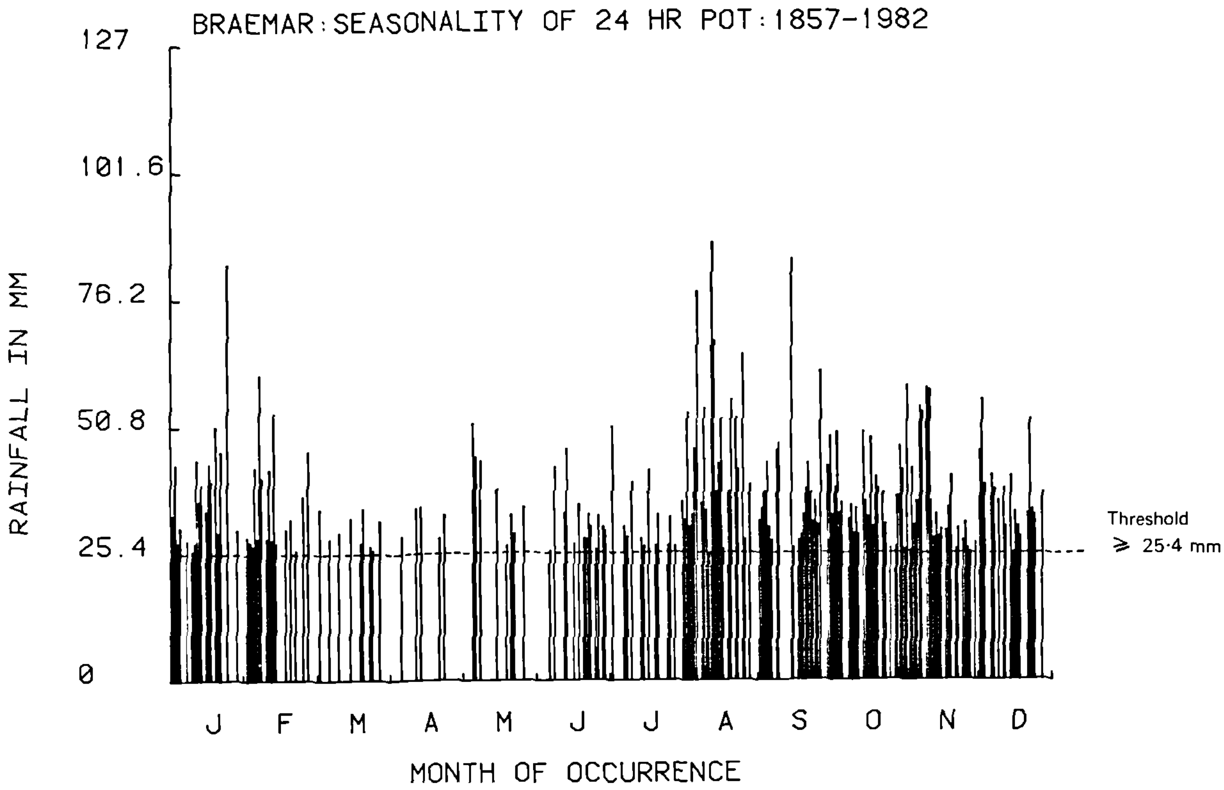


Figure 5.4.5.(ii)

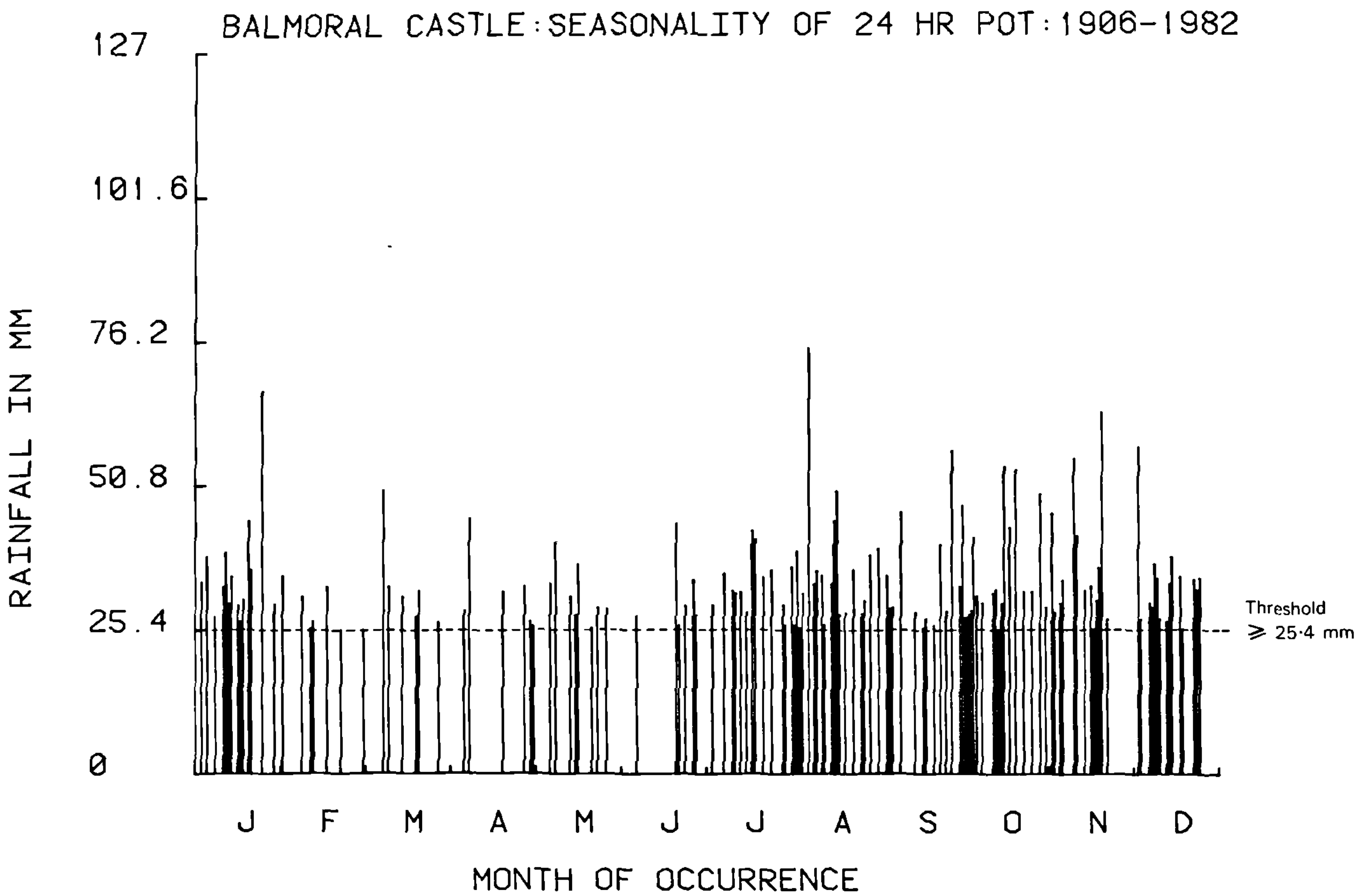


Figure 5.4.5.(iii)

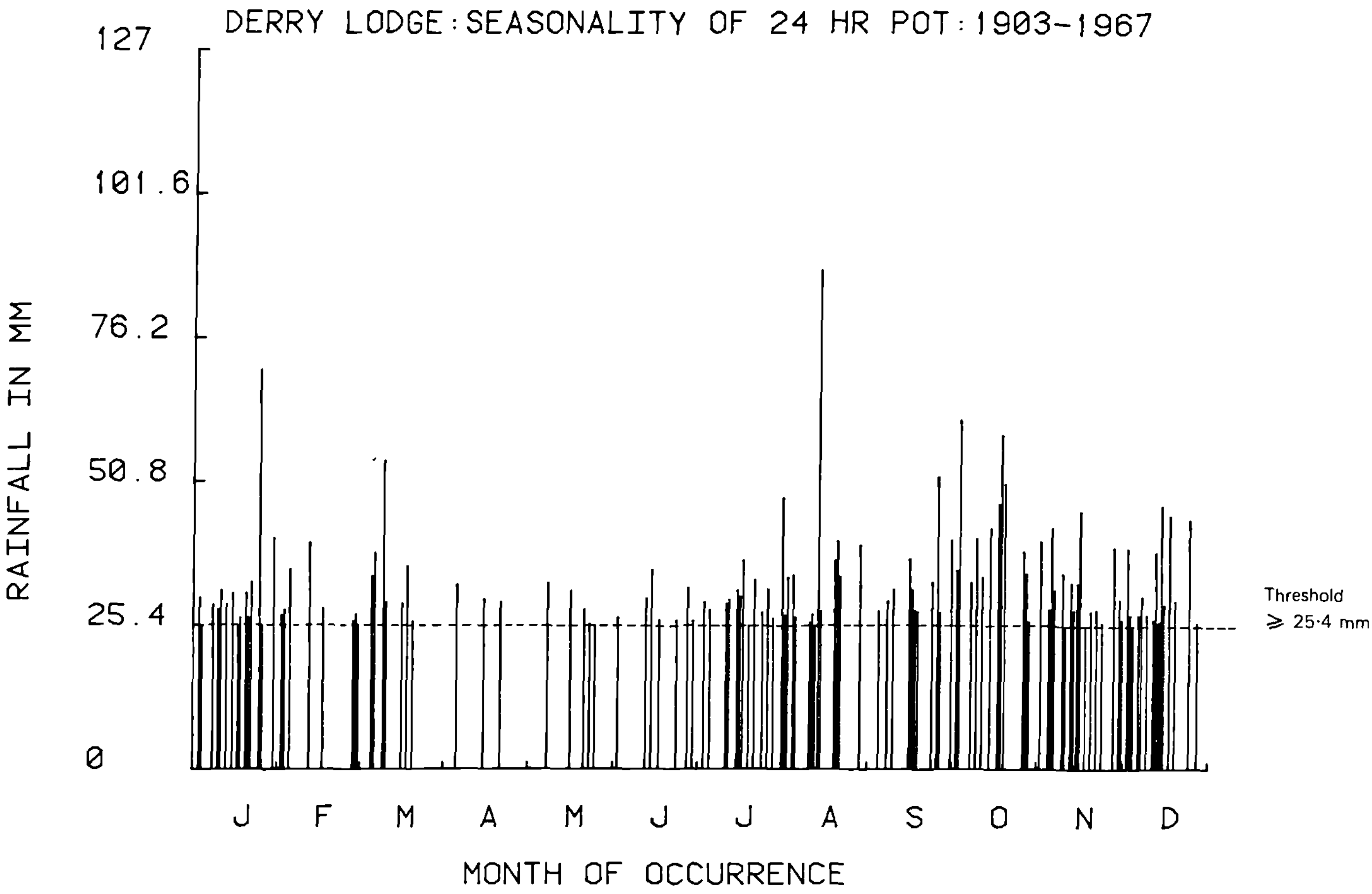


Figure 5.4.4.(ix)

BRAEMAR: YEARS WITH 48 HOUR POT 1857-1982

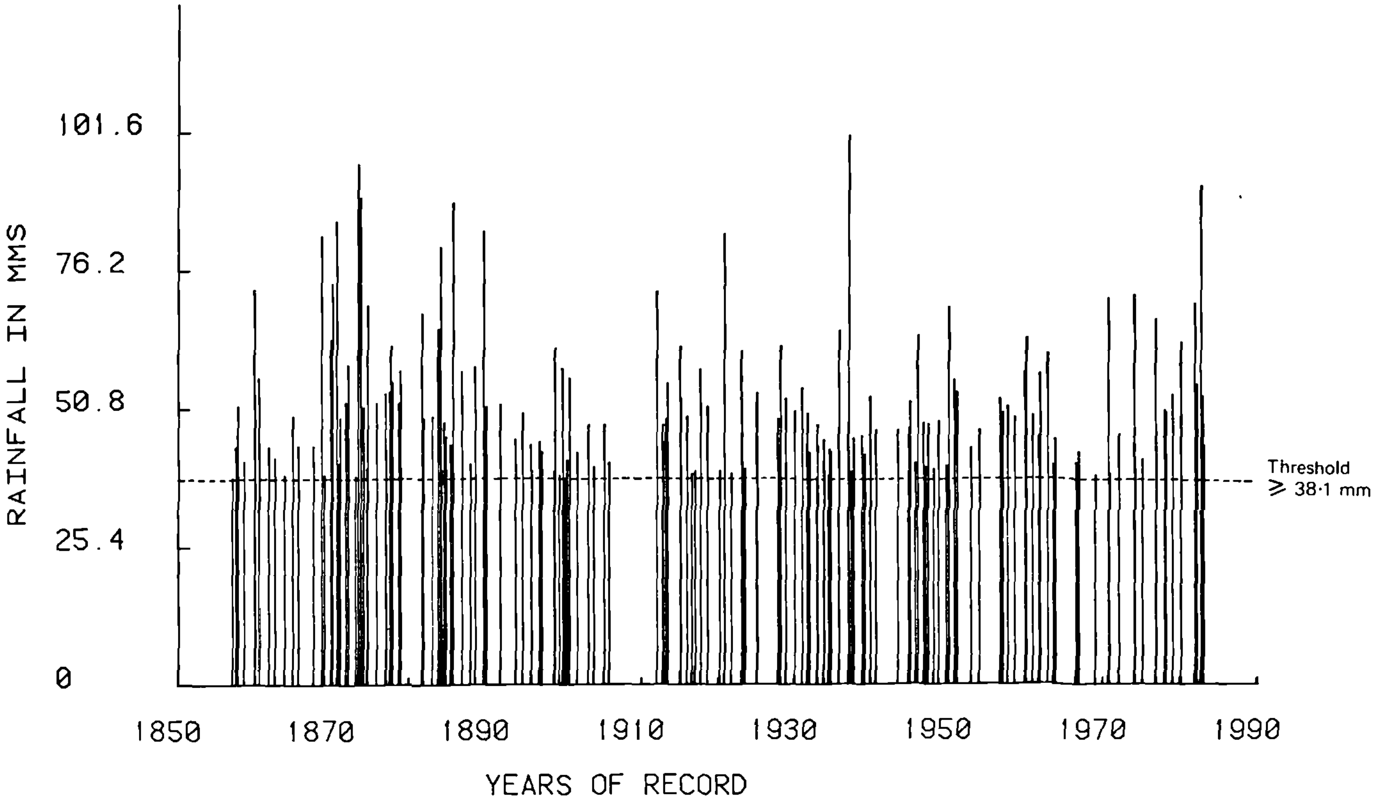
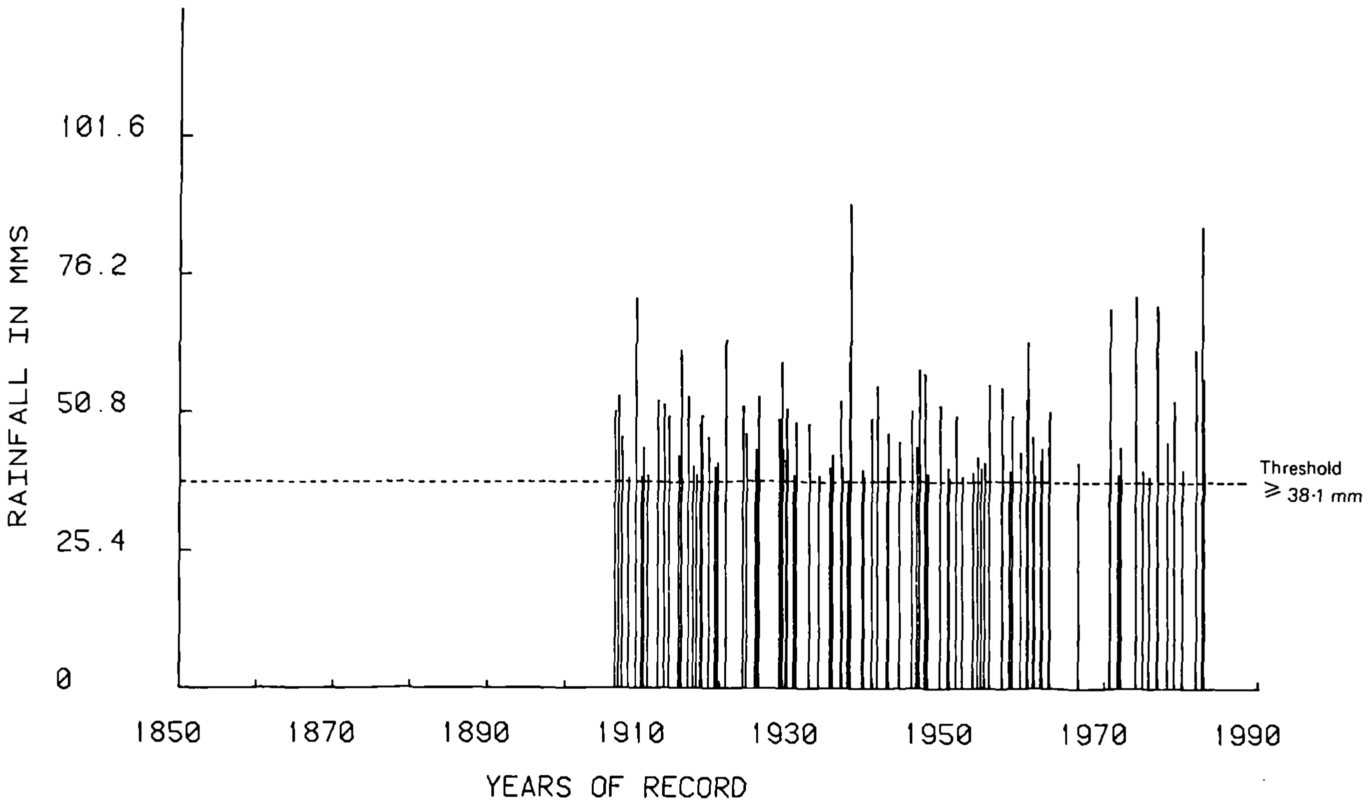
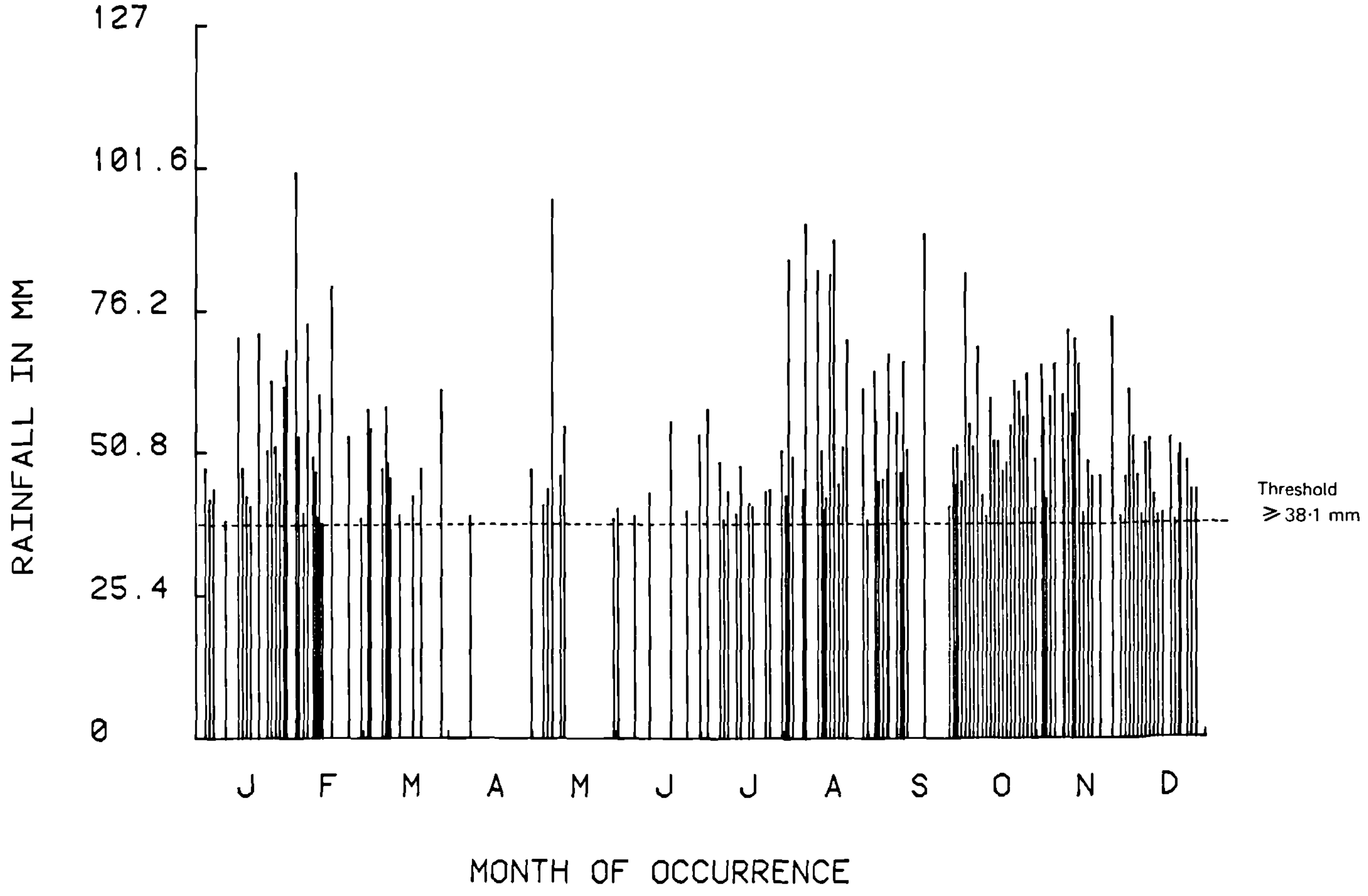


Figure 5.4.4.(xi)

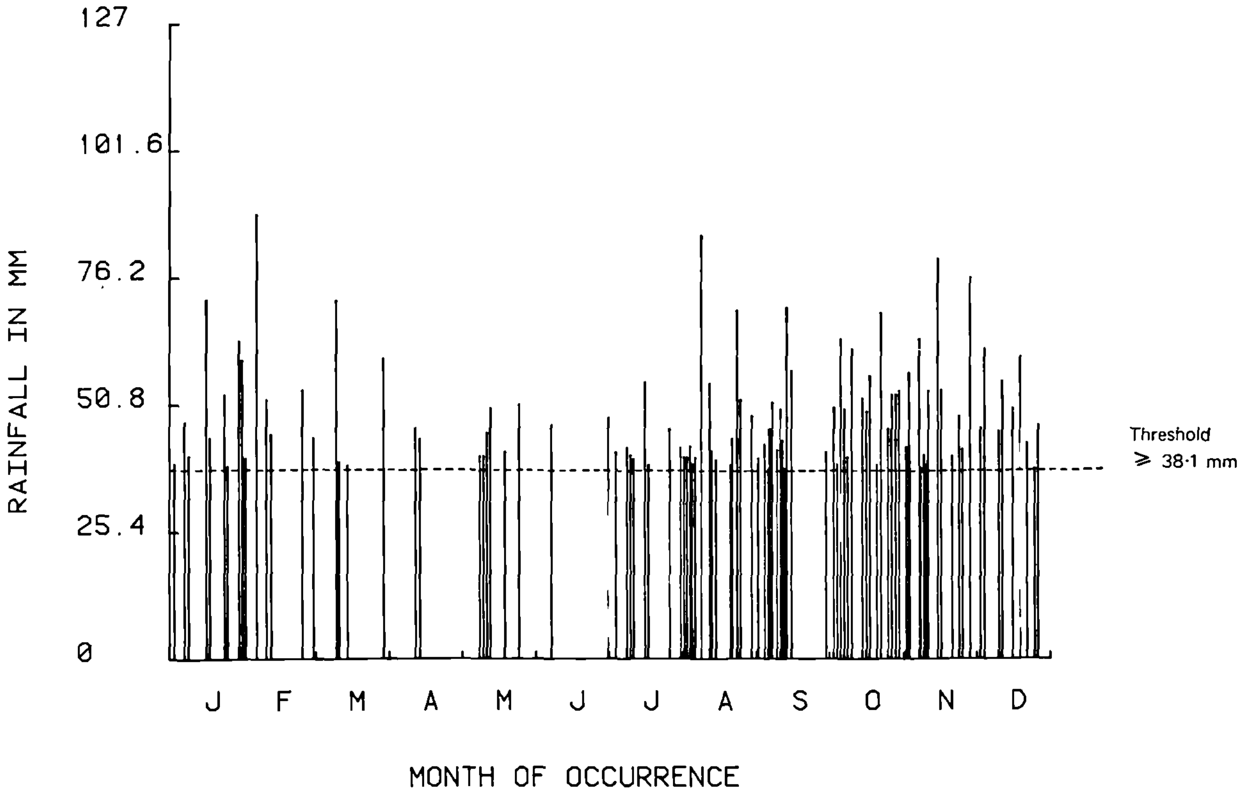
BALMORAL CASTLE: YEARS WITH 48 HOUR POT 1906-1982



BRAEMAR: SEASONALITY OF 48 HR POT: 1857-1982 **Figure 5.4.5.(iv)**



BALMORAL CASTLE: SEASONALITY OF 48 HR POT: 1906-1982 **Figure 5.4.5.(v)**



its relation to soil moisture deficits returning to zero in late autumn. This peak was also seen from the 48 hour data set (Figure 5.4.5.(iv)) but the relative number of large events occurring August to October increased. There was a notable gap in the POT seasonality of occurrence in March and April. A similar pattern was found on the Balmoral record (see Figure 5.4.5.(v)).

5.4.6 Dee study area: Rainfall annual maximum series

The results of the extreme value analysis of the annual maximum series are now presented, to establish the recurrence interval of annual maximum rainfall events of different magnitudes and durations. It is known from the FSR (NERC, 1975) that the maximum probable precipitation for a 24 hour duration is 250-300 mm. Fitting the EV1 distribution to the data, the following RI for rainfall of a certain value as an annual maximum were found (Figure 5.4.6.(i)). When the general extreme value distribution (GEV) was also fitted, using the method of maximum likelihood, the RI(50) and RI(100) are indicated below in brackets (Table 5.4.6.(i)). This was to assess whether the GEV distribution fitted the more extreme values better. The importance of depletion effects, comparing the magnitude of 24 hour cf. 48 and 72 hour events, can clearly be seen.

Using the maximum likelihood method for fitting the GEV distribution for the Braemar record did not significantly alter the the results eg. annual maxima with 100 year RI was 81.6 mm using GEV and 80.7 mm using EV1 (see Table 5.4.6.(i)). From the results of the 24

Table 5.4.6.(i)

Magnitudes of rainfall annual maxima (AM) for different recurrence intervals, fitting the EV1 distribution (and GEV distribution) for the upper Deeside Stations

24 hour	AM(2)	AM(10)	AM(25)	AM(50)	AM(100)
Braemar	39.9	56.9	66.5	73.7	80.7
				(73.9)	(81.6)
Balmoral	35.3	49.6	57.6	63.6	69.6
				(****)	(****)
Derry Lo.	37.0	53.6	63.1	70.1	77.1
				(72.6)	(82.5)
Ballater	41.8	58.2	67.5	74.4	81.2
(ratioed)				(76.0)	(84.1)
48 hour	AM(2)	AM(10)	AM(25)	AM(50)	AM(100)
Braemar	53.1	72.3	83.2	91.3	99.3
				(90.0)	(97.0)
Balmoral	47.8	65.6	75.7	83.1	90.5
				(82.8)	(89.9)
Derry Lo.	50.4	71.8	83.9	93.0	101.9
				(96.6)	(111.2)

Table 5.4.6.(i) cont.

72 hour	AM(2)	AM(10)	AM(25)	AM(50)	AM(100)
Braemar	61.5	82.5	94.5	103.3	112.1
				(101.2)	(108.5)
Balmoral	56.8	75.7	86.4	94.4	102.3
				(91.3)	(97.0)

(72 hour annual maximum could not be accurately calculated for Derry Lodge or Ballater due to several composite (rather than daily) rainfall totals measured in excess of 48 hours).

AM(i) annual maximum rainfall with a recurrence interval of i years

All values in mm

**** data unavailable

hour annual maxima data from the Braemar record, the annual maxima which have a RI in excess of 100 years were 14th Sept, 1873 with 84.1 mm, 12th Aug, 1885 with 87.4 mm and 24th Jan, 1937 with 83.3 mm. The 1885 fall (see Figure 5.4.6.(i) and Figure 5.4.1.(i)) was the highest on record, with an estimated RI of 200 years. The highest value from the Balmoral record was 6th Aug, 1982 with 74.7 mm and this had an estimated RI of 150-200 years (Figure 5.4.6.(ii)). The second largest fall was 24 Jan, 1937 with 67.3 mm and a RI around 75 years. There have thus been two major 24 hour falls post 1900.

As might be expected, Derry Lodge showed a slightly different pattern. The highest annual maximum was on the 13th Aug, 1956 with 3.46 in (87.9 mm) and a RI of over 100 years. On the 9th March, 1908, another high rainfall was recorded with 3.34 in (84.8 mm). The fall of 24th Jan, 1937 was 2.77 in (70.4 mm), giving a recurrence interval of 36 years (see Figures 5.4.6.(iii) and 5.4.1.(ii)). None of these values approach the estimated probable maximum precipitation for 24 hours duration of 225-300 mm (FSR, NERC 1975).

When the recurrence interval of annual maxima of 24 hour rainfall was associated with the incidence of known floods (Figure 5.4.1.(i)), it was notable that although 1872 was the wettest year on record, the RI of its annual maximum was not high (45.7 mm with a RI of 5 years). Rainfall was more evenly distributed throughout the year. One of the larger rainfall events (Jan, 1937) on the Braemar (3rd) and Balmoral (2nd) records produced the largest flood event recorded, but in 1937 there were exceptionally three other POT in that month. The hydrometeorological aspects of this event will be discussed later in

Figure 5.4.6.(i) DEE CATCHMENT AT BRAEMAR 1857-1982
24 hour annual maxima

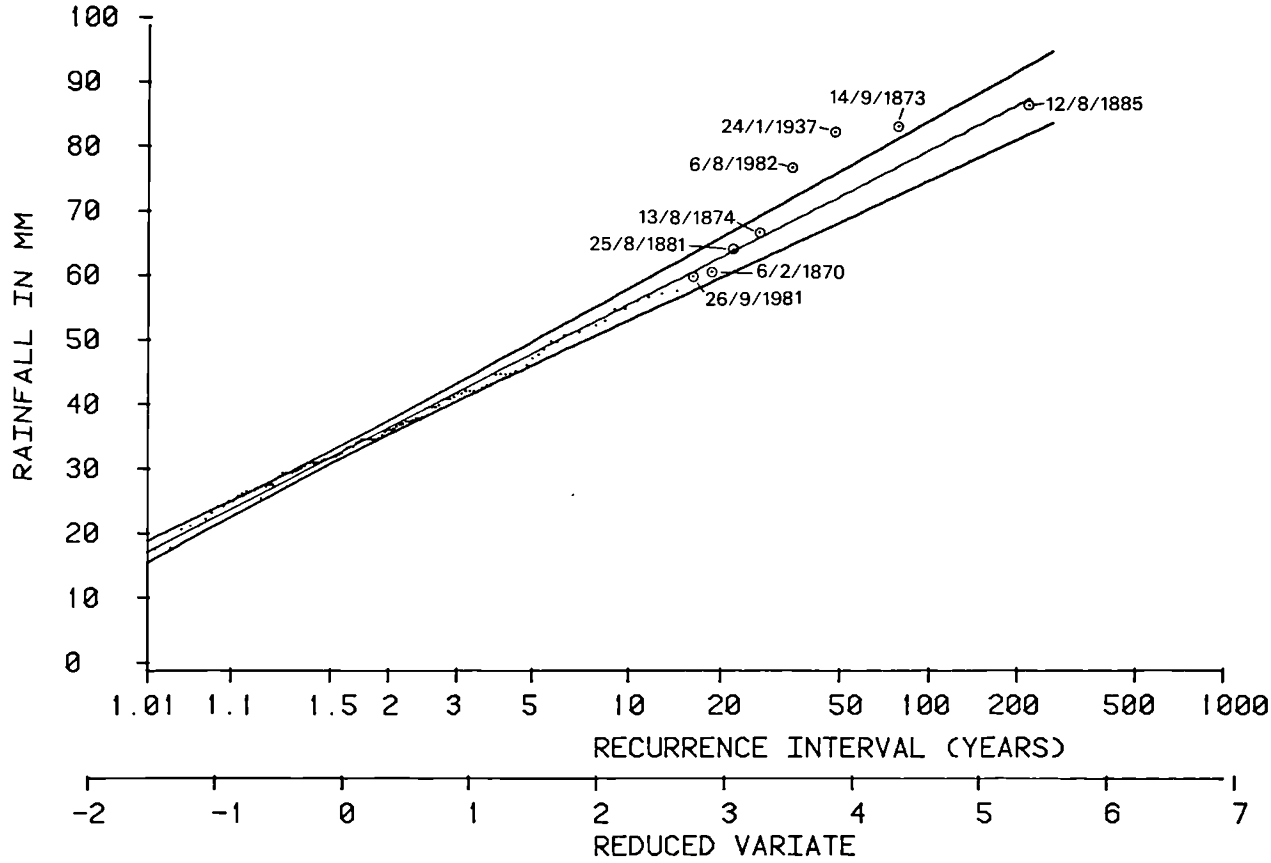


Figure 5.4.6.(ii) DEE CATCHMENT AT BALMORAL CASTLE 1904-1982
24 hour annual maxima

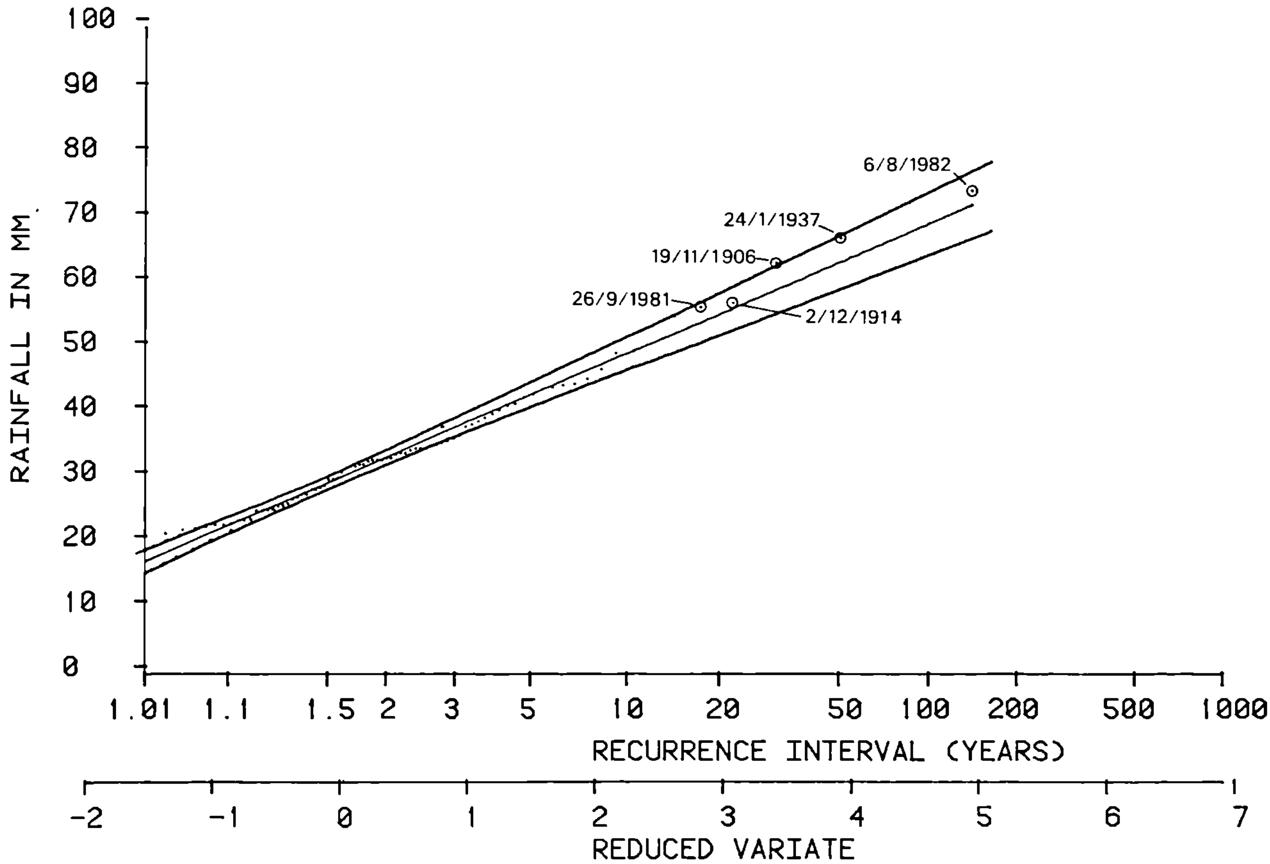
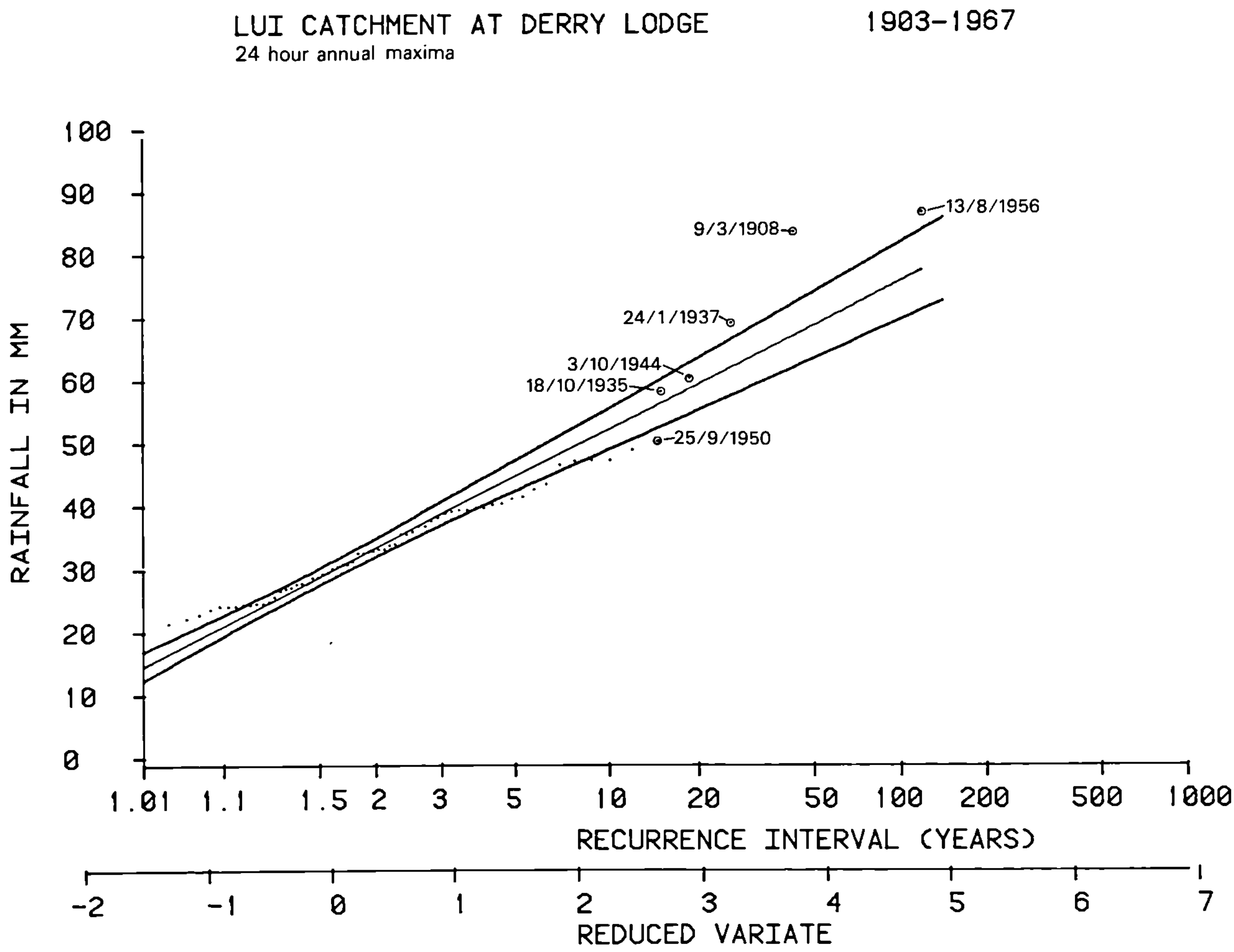


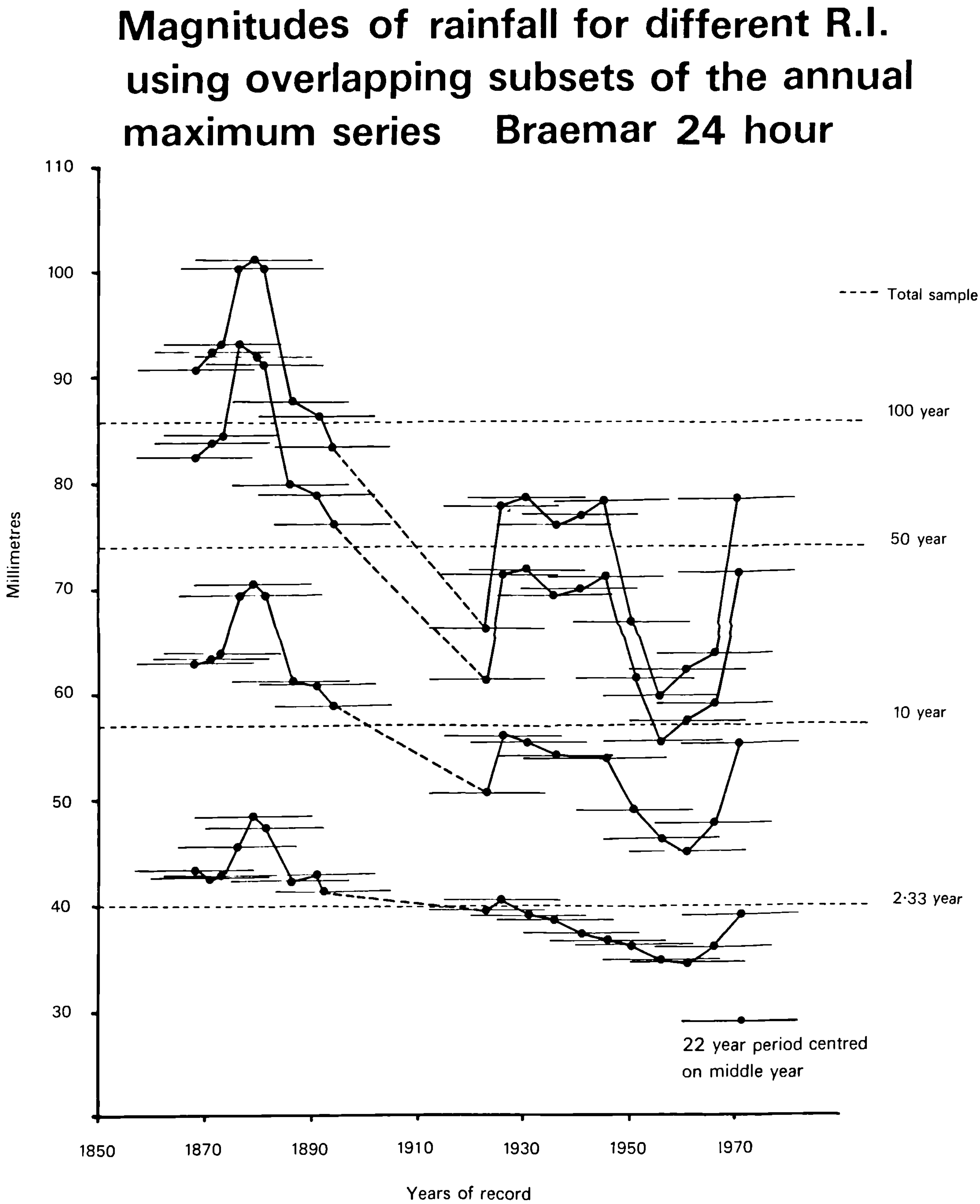
Figure 5.4.6.(iii)



Section 5.4.8.

If distribution analysis was carried out on subsets of the data, divided into 22 year periods with a 5 year overlap, this allowed some idea of the changing RI of rainfall within different sub-sets of the record. The time period was based on the predominant periodicity of the annual total data set. When these values were plotted year of record against rainfall magnitude, the frequency of annual maxima of different magnitudes was shown to fluctuate in different periods. As seen in Figure 5.4.6.(iv), the variation in magnitude of annual maxima within each 22 year period for a set RI could be large. For example, in the Braemar record when comparing 100 year RI annual maxima, the difference between 1865-1887 and 1950-1972 was 1.58 in (40.2 mm). The patterns of change in magnitude of annual maxima for a set RI are shown in Figure 5.4.6.(iv). Even for lower RI (2-10 years), there could be considerable variation. At Braemar between 1868-1890, there was a peak of 48.3 mm for rainfall with a 2 year RI, cf. 34.0 mm between 1950-1972. This discrepancy in annual maximum value for a known RI, naturally increased with an increase in RI. There was a considerable peak pre-1900 (1865-1887), while the period 1950-1975 showed a marked trough in Figure 5.4.6.(iv). The comparative increase between 1920-1940 was however much smaller than the earlier pre-1900 peak. A similar pattern, for the years of overlap, was found on the Balmoral record. When this pattern was compared with general climatic patterns for the U.K., as outlined in Table 5.3.(i), there was a basic similarity.

Figure 5.4.6.(iv)



The highest 48 hour annual maximum at Braemar occurred on 5-6th May, 1873 with 95.8 mm, closely followed by 24-25th Jan, 1937 with 95.5 mm and 5-6th Aug, 1982 with 91.5 mm (Figure 5.4.6.(v)). The 1873 event however had no recorded incidence of flooding and thus must either have been highly localised or was associated with a very high soil moisture deficit. It is notable that the highest values of 48 hour POT were not much larger than the highest values of 24 hour POT. This suggests depletion effects with an exhaustion of water content in the air mass as it reaches the eastern side of the country. As can be seen in a comparison with the 24 hour annual maxima series, there was only a difference of 8.1 mm between the two. The 1873 and 1937 values had a RI of approximately 75 years while the 1982 rainfall had a RI of approximately 50 years (Figure 5.4.1.(i)). From Figure 5.4.6.(viii), the fluctuation in 48 hour annual maximum series for set RIs can be seen and like the 24 hour annual maxima, there was a peak between 1865-1890. However, the secondary peak around 1915-1940 was much larger than the corresponding 24 hour peak. This pattern was confirmed by the Balmoral record for the overlapping period of years.

Extending the storm duration even further on the Braemar record, the highest 72 hour annual maxima was again 22-24 Jan, 1937 with a rainfall of 110.5 mm (4.35 in) and a RI of just under 100 years. Other high 72 hour rainfall events can be identified on Figure 5.4.6.(ix). The high 72 hour RI of the 1937 event was confirmed by the Balmoral record with a 72 hour rainfall of 96.5 mm and a RI of over 50 years (Figure 5.4.6.(x)). The 1937 event was thus exceptional in both its magnitude and duration. In terms of fluctuations within the 72 hour

Figure 5.4.6.(v) DEE CATCHMENT AT BRAEMAR 1857-1982
48 hour annual maxima

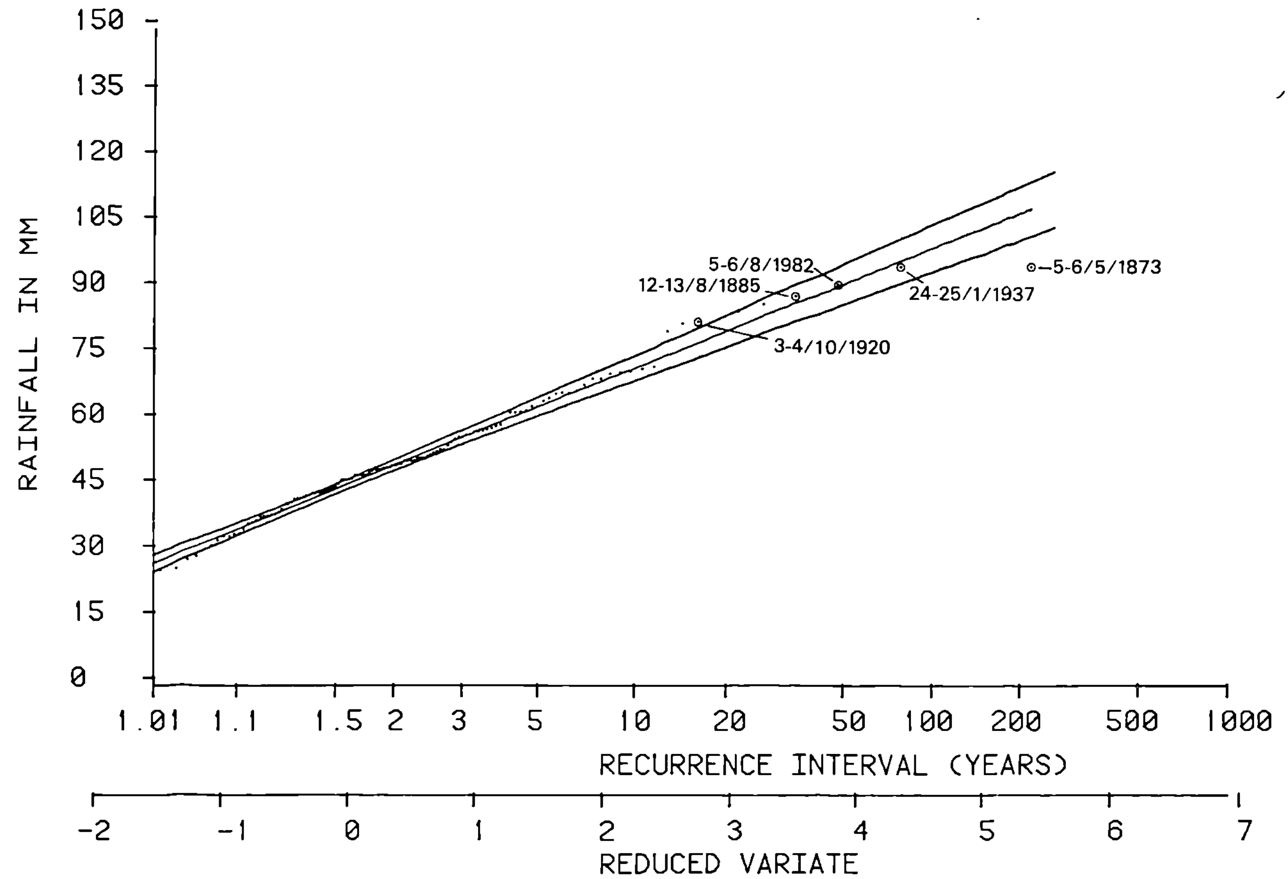


Figure 5.4.6.(vi) DEE CATCHMENT AT BALMORAL CAST. 1904-1982
48 hour annual maxima

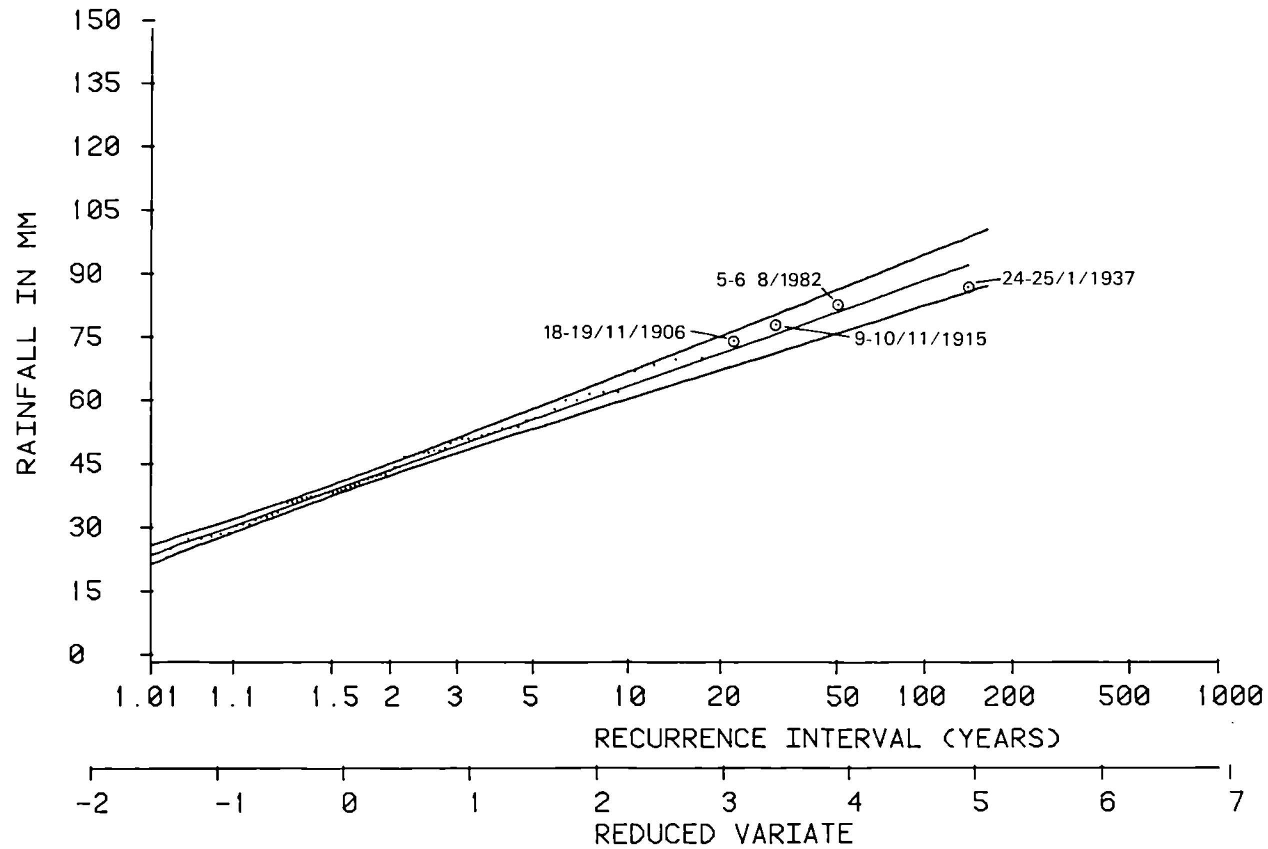


Figure 5.4.6.(vii)

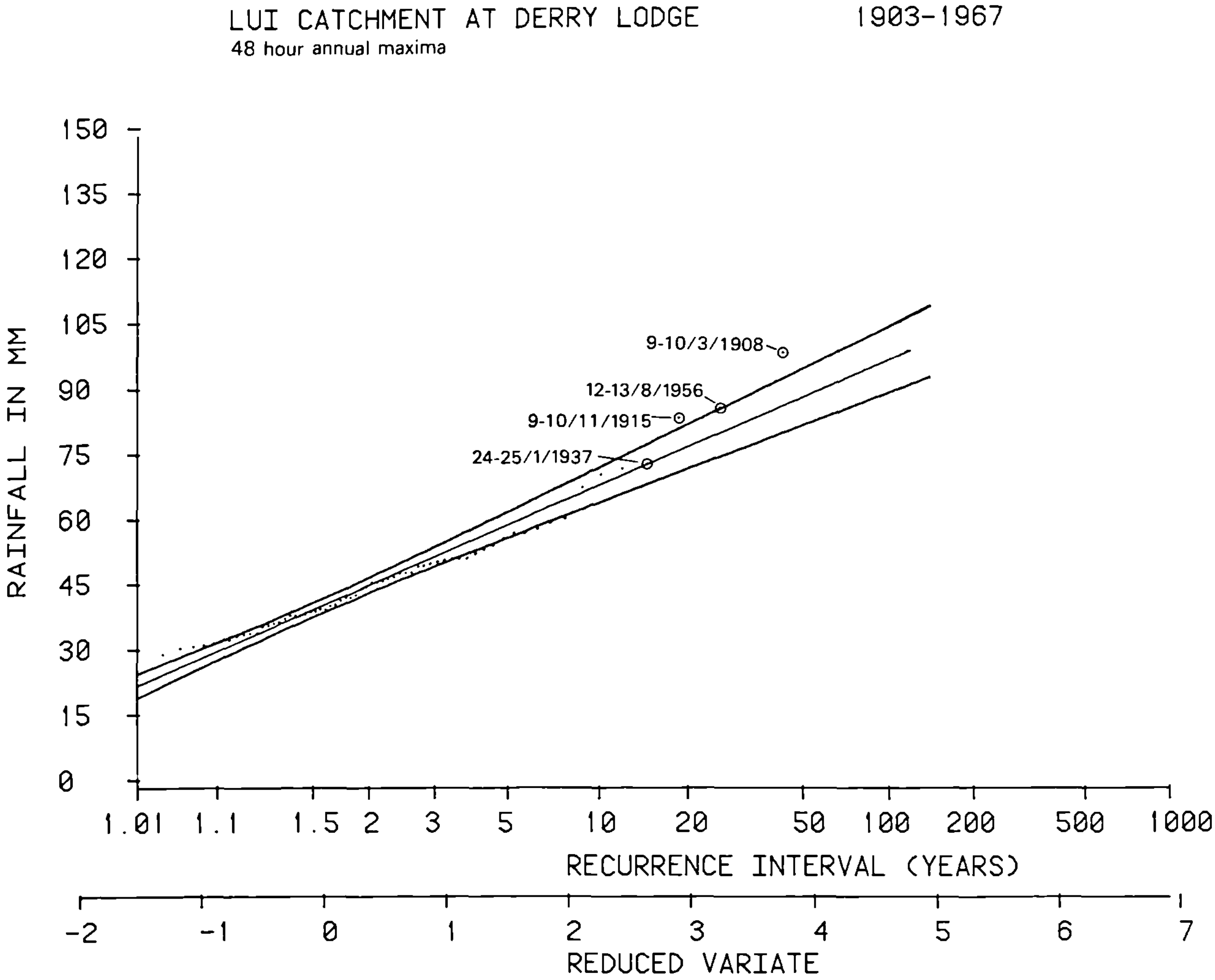


Figure 5.4.6.(viii)

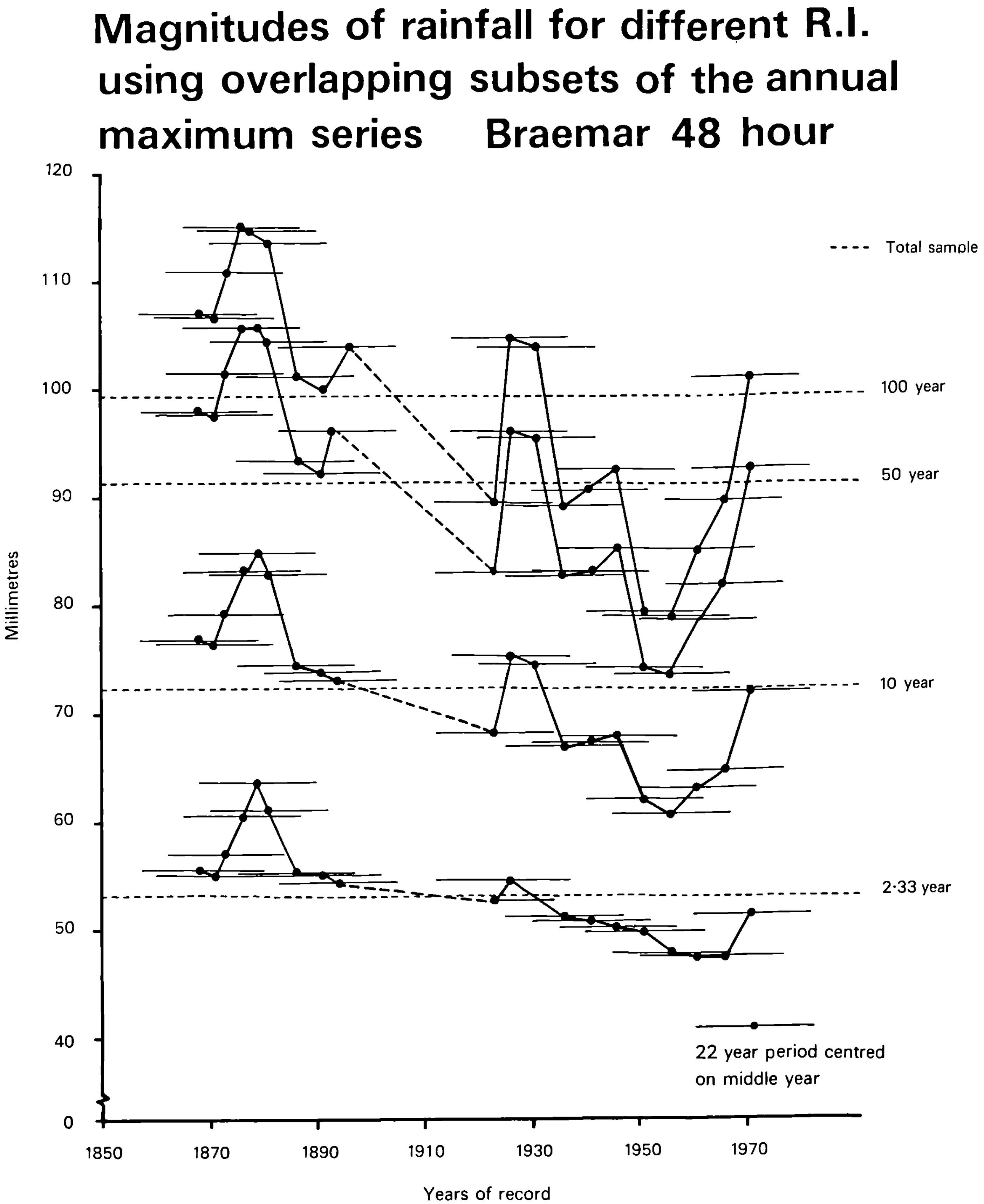


Figure 5.4.6.(ix)

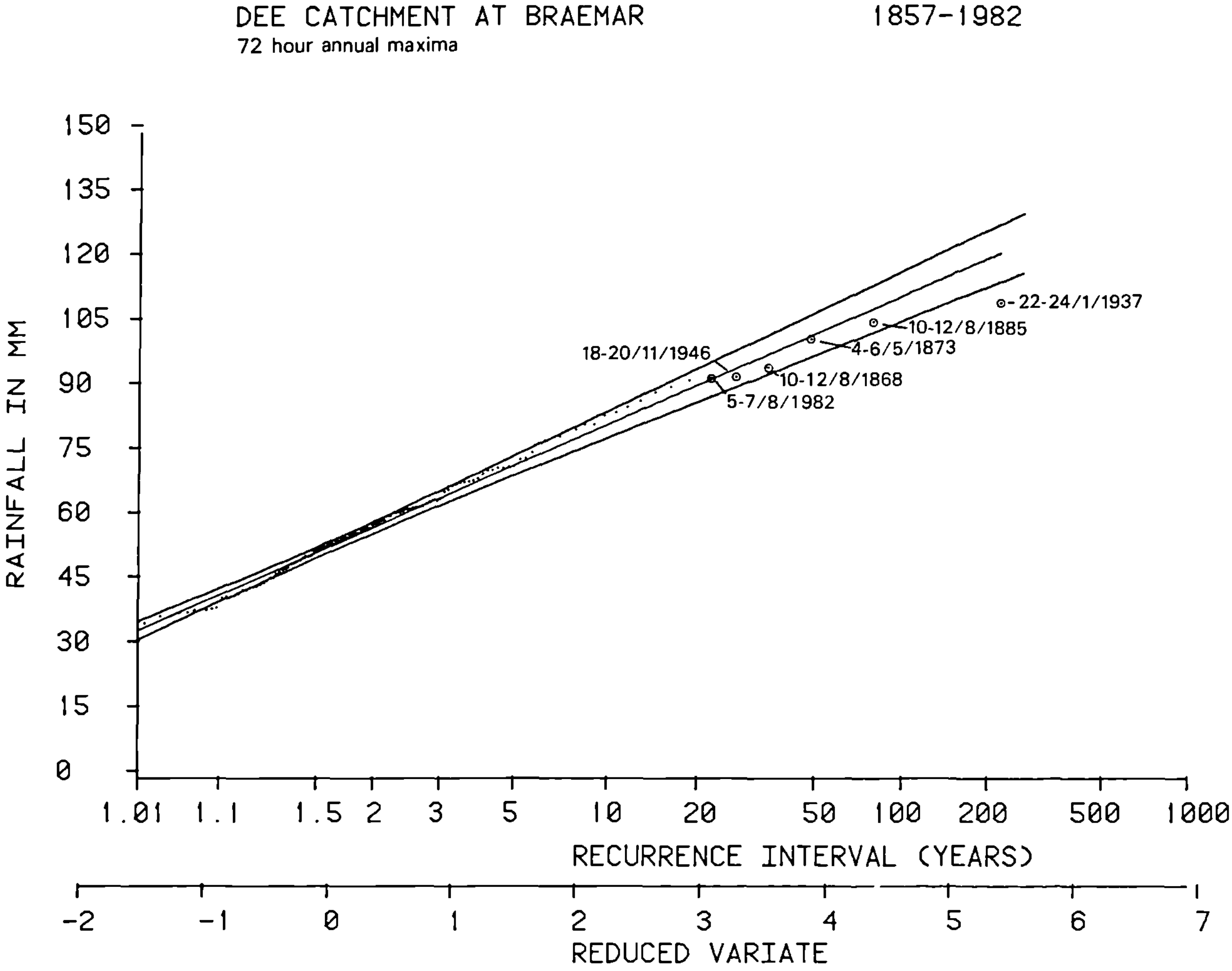
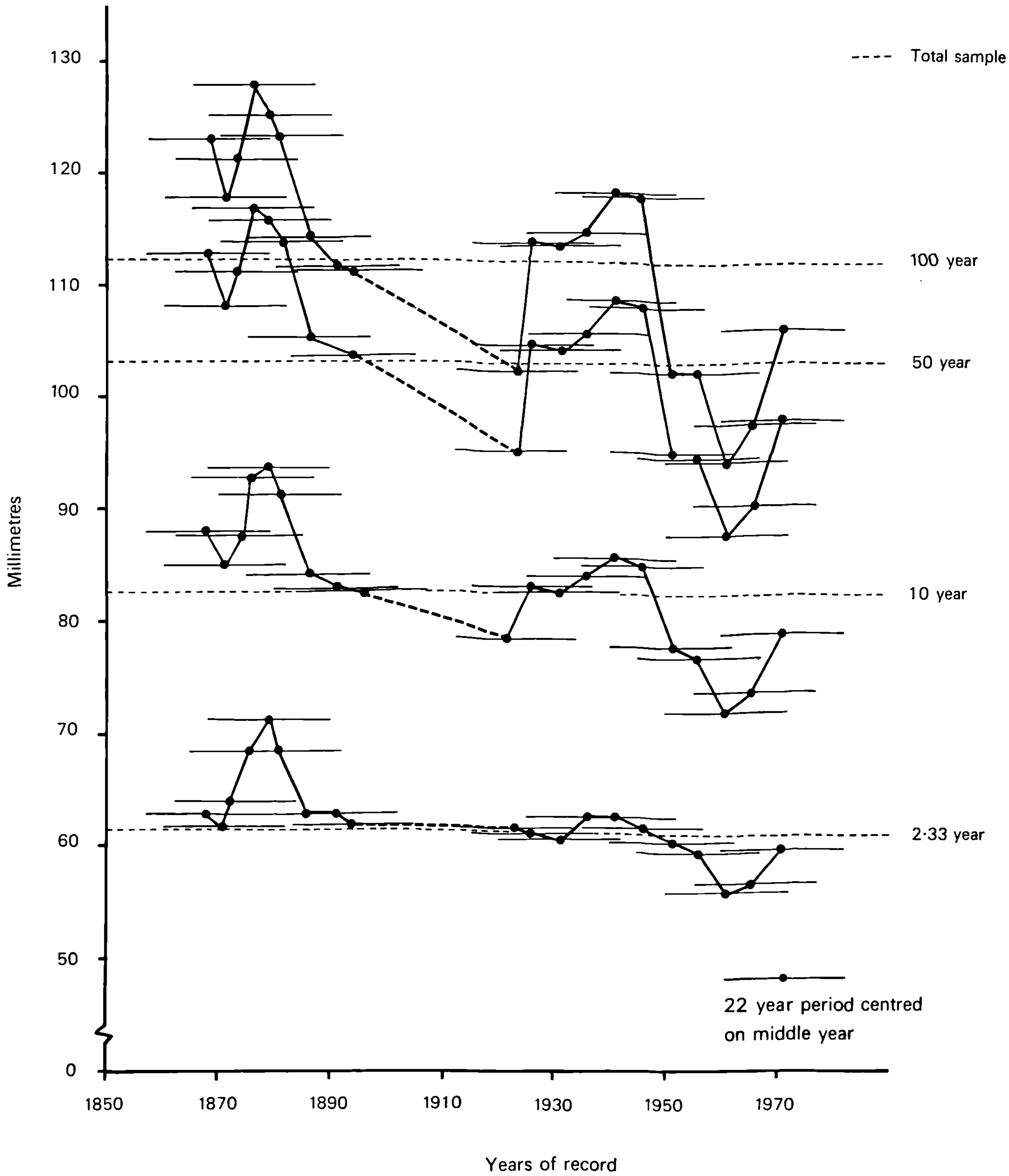


Figure 5.4.6.(xi)

Magnitudes of rainfall for different R.I. using overlapping subsets of the annual maximum series Braemar 72 hour



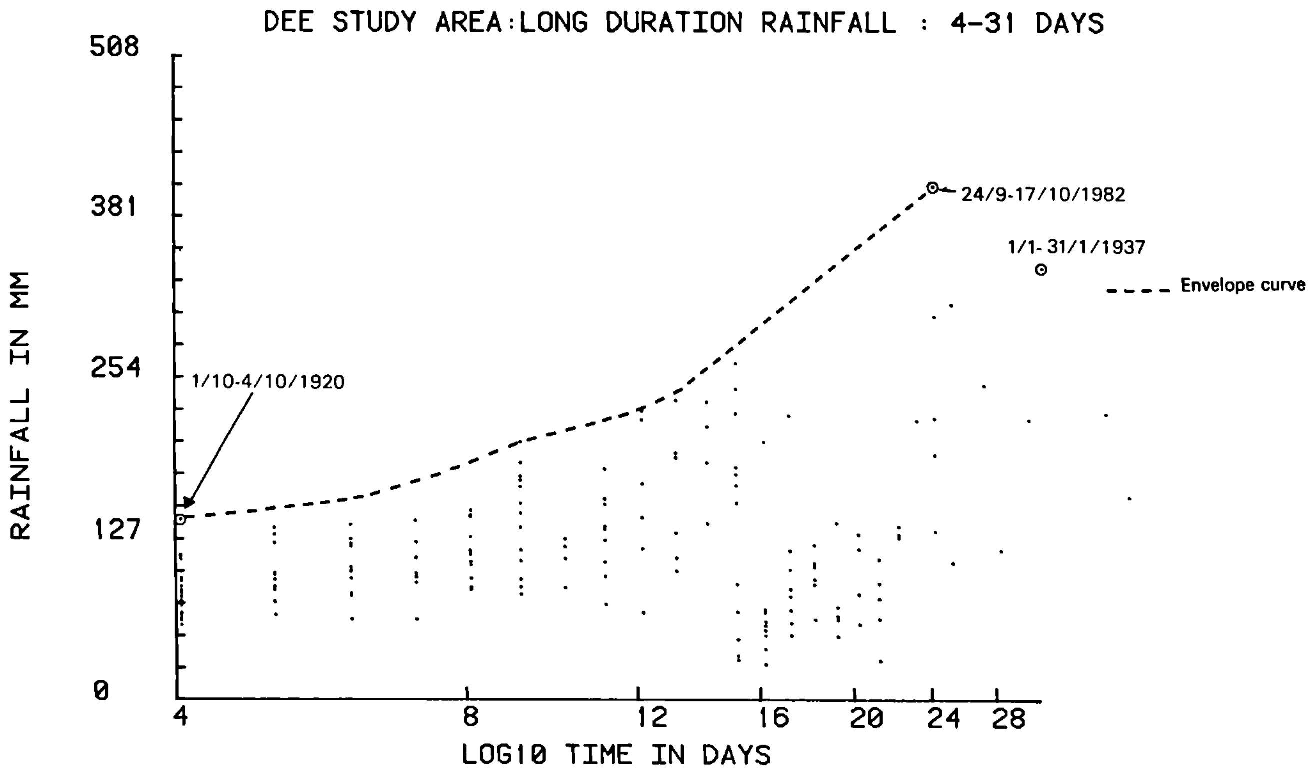
annual maximum series, the peaks and troughs followed a similar pattern to those of 48 hour duration (see Figure 5.4.6.(xi)).

5.4.7 Dee study area: Longer and shorter duration events

Long duration data sets of 4-31 days were extracted both from the raw data set and British Rainfall and plotted against $\log_{10}$ time (Figure 5.4.7.(i)). This incorporated both major 24-72 hour falls and high antecedent rainfall, associated with slow moving fronts and progressions of cells. Certain events stood out clearly from the data set. Most exceptional were the Jan, 1937 storms recorded at Balmoral and Braemar and the 1-17th Oct, 1982 storms at Braemar, which form the limits of the envelope curve. The 1937 event showed high values at all stations within the area eg. Braemar 1-31st January had 13.4 in (340.6 mm), while at Braemar between 25th Sept and 17th Oct, 1980, 11.9 in (302.3 mm) rainfall fell. There seemed to be a general pattern of levelling off after 4-7 days. If however the rainfall exceeded this duration, there was a likelihood that magnitude would increase non-proportionally with duration due to depletion effects.

As outlined in Section 5.1, there is considerable evidence to suggest that with an increase in altitude, the frequency of intense, localised, convective storms also increases. In terms of frequency in Deeside, there were several documented reports of cloudbursts or waterspout phenomena, producing geomorphic stresses both within the channel and slope systems.

Figure 5.4.7.(i)

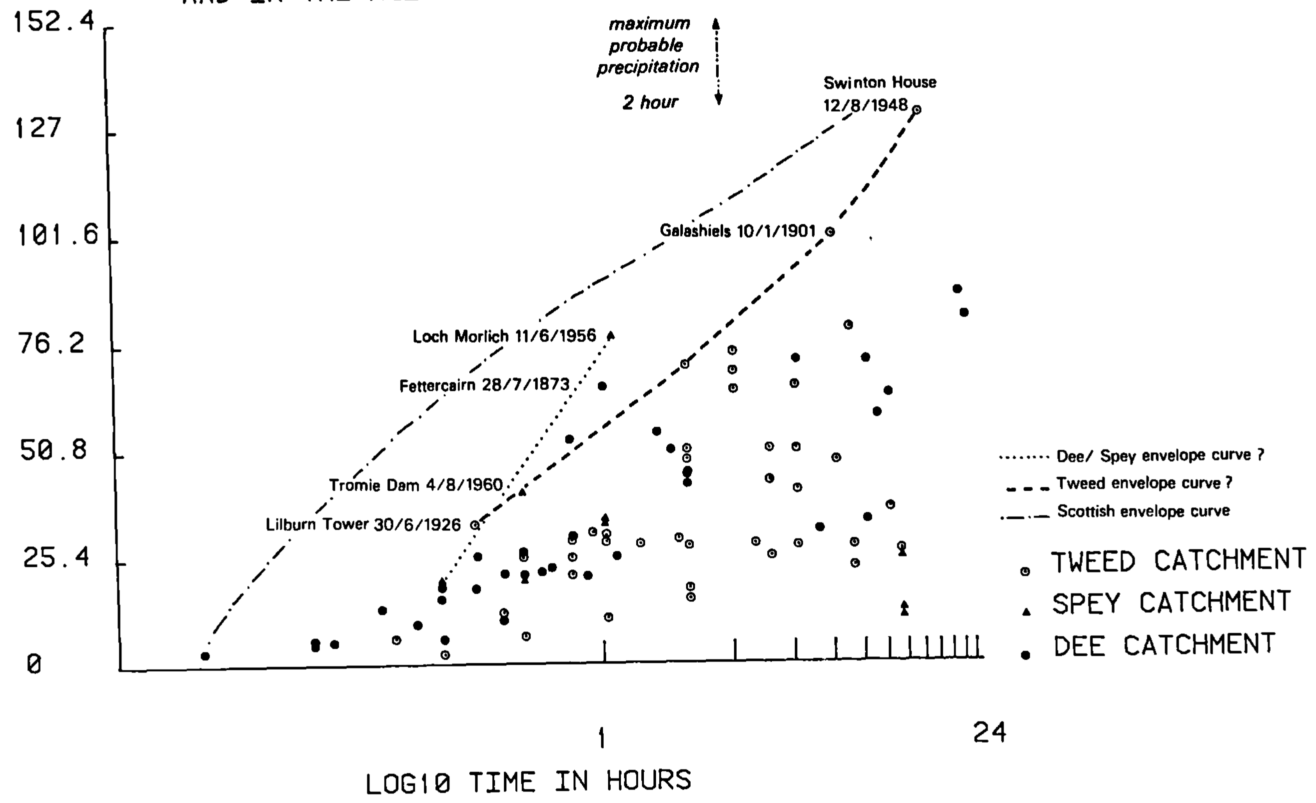

DEE STUDY AREA:LONG DURATION RAINFALL : 4-31 DAYS
RAINFALL IN MM
508
381
254
127
0
4
8
12
16
20
24
28
LOG10 TIME IN DAYS
1/10-4/10/1920
24/9-17/10/1982
1/1-31/1/1937
Envelope curve

The short duration rainfalls for Deeside, as collected from the Meteorological Magazine and British Rainfall, were plotted amount against $\log_{10}$ time in Figure 5.4.7.(ii) to gain an approximate envelope curve. The 2 hour probable maximum precipitation of 130-140 mm depending on altitude, calculated in FSR (NERC, 1975), was also plotted, as was the probable envelope curve for Scotland. As seen on Figure 5.4.7.(ii), the most "rare" event was 2.10 in (53.3 mm) in 45 minutes at Braemar on the 2nd Aug, 1870, giving an hourly rate of 2.8 in (71.1 mm). This event must have been very localised as at Ballater no rain was recorded. Another longer duration fall of 3.44 in (87.4 mm) in under 12 hours occurred in Braemar on 12th Aug, 1885, approaching in intensity the 1948 event on the Tweed but not nearly with the same areal extent. It is known from The Scotsman's report that the Dee was in high flood, with the Quoich and Milton Burn especially affected.

> "A rainfall of unprecedented severity, and to which even an approximate parallel has not been found within the memory of the present generation took place at Braemar and generally over upper Deeside..." (Scotsman, 8/1885)

This event, although regionally localised, was recorded on gauges downstream. Thus, at Midmar Blackstock, a rainfall of 2.52 in (64.0 mm) in approximately 11 hours occurred. It is important to note that antecedent rainfall was also reasonably high with 1.51 in (38.4 mm) falling in the preceding 3 days at Braemar.

Figure 5.4.7.(ii) SHORT DURATION RAINFALL WITHIN THE STUDY AREAS AND IN THE NEIGHBOURING REGIONS



Other reports of similar intense local rainfall events occur in the literature without an estimated rainfall value, and although a complete list has been included in Appendix 1.1, specific examples will be described here. Brief mentions of highly localised waterspout phenomena are noted in the Meteorological Magazine. On 11th Aug, 1884, the following took place:

> "In Braemar, some trees were struck and a short distance beyond the Linn of Dee on the north side of the river, the face of the hill was serrated and scores of tons of rocks etc. dislodged. At the close of the storm, rain fell for about 5 mins very like a waterspout". (Meteorol. Mag., 1884)

However, this event seems to have been most important in initiating slope failure and possibly affected channel sediment supply. It must have had some geomorphic effect on the neighbouring channels, although there was no recorded rise in discharge on the mainstream Dee.

Another example was the "Cloudburst on Upper Deeside", documented by Robertson (1935), which occurred on the 13th July, 1927 and affected the Aberarder district and the Fearder Burn.

> "The inn [Inver] was surrounded by water and flooded inside to a considerable depth. A few yards to the east of the inn, the usually modest, wimpling Fearder Burn was a raging, roaring torrent, which had broken bounds from its ordinary course and claimed, after a lapse of 107 years, its

original bed from which it had at that time been diverted......the strength of the torrent threatened to carry the men off their feet....a huge wall of water rushing down the hill, which lies to the north of the Inver inn.... chasm which the mighty flood was cutting out on its course, a chasm which on the subsidence of the water was found to be some 15 feet [4.6m] in depth" (Robertson, 1935 p9).

As can be seen in Table 3.2.1.(i), the Fearder Burn has a small catchment area of 27.6 km^2 but a high basin relief of 107 $m\ km^{-1}$. The river by 1972 had returned to its 1903 (second edition) course; in fact probably the flood channel became immediately disused again at lower stages. The old flood channel is now grassed over and obviously has been without flow and stabilised since this extreme storm event. The storm discharge must have exceeded the threshold stage required to reoccupy the 1820 channel (as verified on Roy's map). As can be seen from the 24 hour rainfall at the closest gauges, Balmoral and Braemar, low rainfall totals of 19.6 mm and 2.0 mm respectively were recorded. Another major localised convective storm event occurred in Aug, 1956 but this is discussed in detail by Baird and Lewis (1957) and this will be studied as a specific example in the aerial photograph case studies. (Section 7.2.6).

These events may be important geomorphically if they occur over small catchments and are associated with sudden increases in discharge and stream power. There is clearly a considerable range of areal extents associated with convective events, and thus effects may differ depending on the location of the cells of highest rainfall. Otherwise, the most important contribution is the mobilisation of sediment on the slopes. Although the frequency of occurrence of such events over the same catchment is low, the frequency of occurrence sampled within the whole region is much higher. Thus, with an increase in the sampling interval, time is substituted for space (ie. the ergodic hypothesis) in the frequency of occurrence (and documentation) of these rare events.

5.4.8 Dee study area: Storm case studies

The following major storm events were studied for Deeside and were chosen mainly because an estimated discharge was available. Profiles of 24 hour, 48 hour and antecedent rainfall were reconstructed where appropriate.

Flood event	Rank at Balmoral	Figure
(1) 2-14 Dec 1914	3	3.4.8.(ii)
(2) 1-6 Oct 1920	4	3.4.8.(iii)
(3) 12-25 Jan 1937	1	3.4.8.(iv)

(1) 2-14 Dec, 1914

The 1914 event seemed to be the most localised of the 3 storms studied:

"One remarkable feature of the flood of the 2nd Dec, 1914 is that, while it reached exceptionally high levels at and above Ballater, neither the Aberdeen Burgh Surveyor nor the harbour authorities have records of anything exceptional in the state of the water near its mouth at that date. The flood water was no doubt derived from excessive precipitation in the upper part of the Dee basin, but receiving little alteration from the lower hills to the E-ward, it failed to have any notable effect- or any worthy of record- on the lower reaches of the river." (Bremner, 1922 p31)

Two main periods of heavy rainfall intensity occurred, firstly as a 24 hour event on the 2nd and secondly as a 48 hour event on the 12-13th. The earlier 24 hour event had a RI of under 10 years at Braemar and rainfall in excess of 2 in (50.8 mm) at Balmoral and Abergeldie Castle Gardens. The later 48 hour event increased in magnitude and RI with movement downstream to Ballater (see Figure 5.4.8.(ii) and Table 5.4.8.(i)). Between and including these two events, there was nearly continuous rain over a period of 12 days, and thus soil moisture level would have been high. Rainfall totals for a 13 day period ranged from 6.55 in (166.4 mm; Derry Lodge) to 9.10 in (231.1 mm; Ballater) and areas of highest intensity are shown in Figure 5.4.8.(ii). It is really necessary, but very difficult, to estimate the RI of such high antecedent rainfall followed by a high RI rainfall event. Clearly it is well in excess of 100 years, despite the low RI of the individual POT.

Figure 5.4.8.(i)

Location of the study areas within the storm profiles

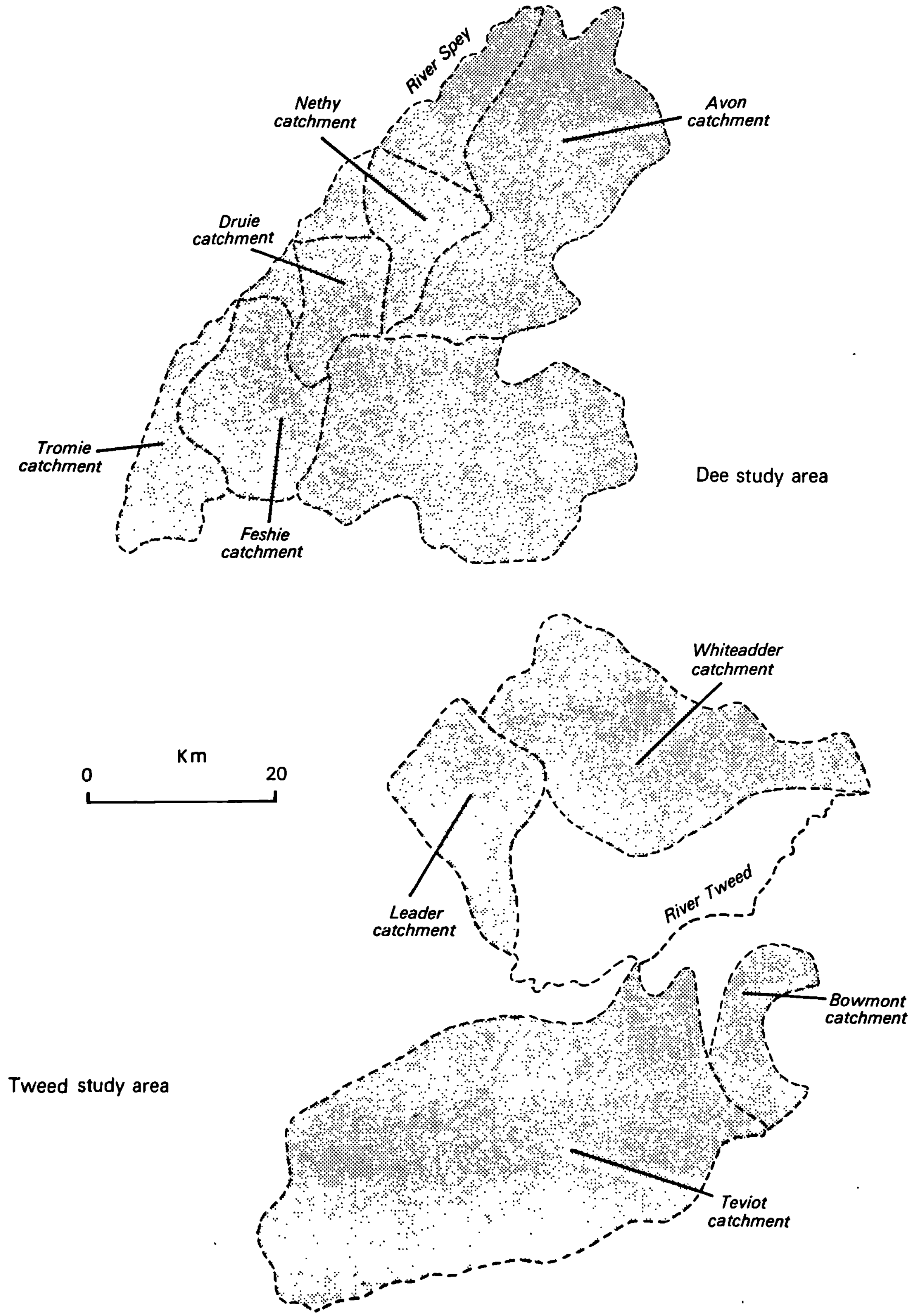


Figure 5.4.8.(ii)

Rainfall 2-14 December, 1914 Dee study area

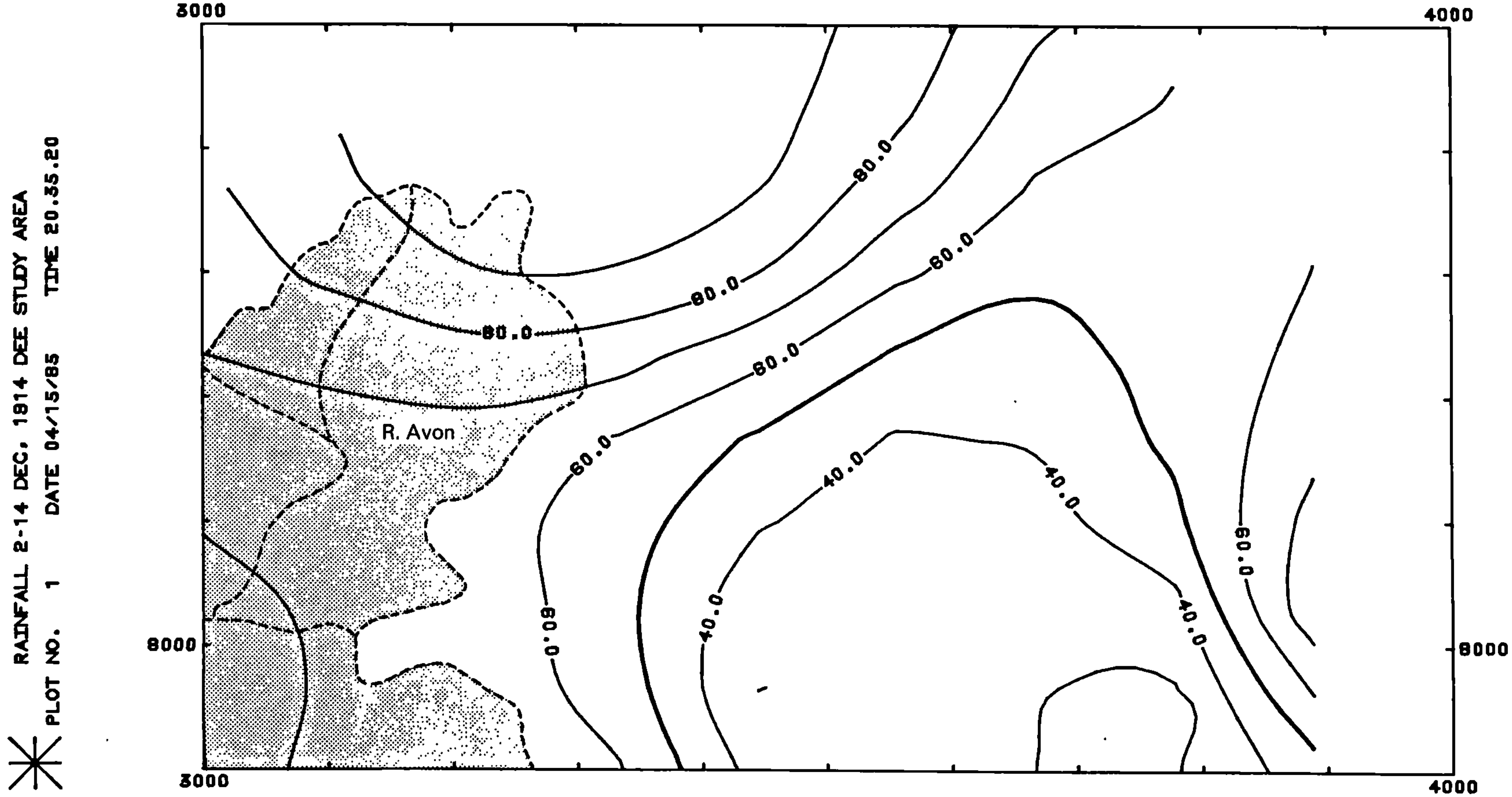


Table 5.4.8.(i)

Rainfall recurrence intervals for the 12/1914 event *

Station	24 hour (2/12/1914)	48 hour (2-3/12/1914)	72 hour (2-4/12/1914)
Braemar	55.9 mm (10 yr)	62.0 mm (6 yr)	77.7 mm (8 yr)
Balmoral	57.4 mm (25 yr)	61.9 mm (8 yr)	71.1 mm (8 yr)
Derry Lodge	25.4 mm (1 yr)	52.8 mm (3 yr)	+
Ballater	48.3 mm (5 yr)	+	+

\+ RI unknown but higher than for shorter durations at the same Station

* as an annual maximum

(2) 1-6 Oct, 1920

The synoptic situation associated with the 1920 flooding was as follows:

> "On 4th Oct, a south easterly gale brought heavy rainfall in the east of Scotland. There is evidence that the fall was unduly augmented by local pressure irregularities." (*British Rainfall*, 1920)

From the 1-4th, falls of 4 in (127 mm) or more occurred along the River Dee from Braemar to Drum. The most intense rainfall fell on the 3rd and 4th and at Braemar, this 48 hour event had a RI of 25 years (Table 5.4.8.(ii)). Figure 5.4.8.(iii) shows the spatial distribution of rainfall totals from the 1st-6th (6 days). The area of most intense rainfall in fact was to the E. and S.E. of the study area with 6.01-7.77 in (152.7-197.4 mm) recorded in those 6 days. The maximum value in the Dee study region was 5.80 in (147.3 mm) at Ballater with Derry lodge, Balmoral and Braemar recording 2.41, 3.99 and 4.32 in (61.2 mm, 101.3 mm, 109.7mm), respectively. These included at least one 24 hour POT in excess of 1 in (25.4 mm). Over this period, the north to south flowing tributaries, eg. Clunie Water, would be contributing the major portion of the flow as the precipitation gradient was shown to be high over the White Mounth to Glen Muick. The Derry Lodge fall suggested considerably less rainfall on the south-facing slopes.

Table 5.4.8.(ii)

Rainfall recurrence intervals for the 10/1920 event *

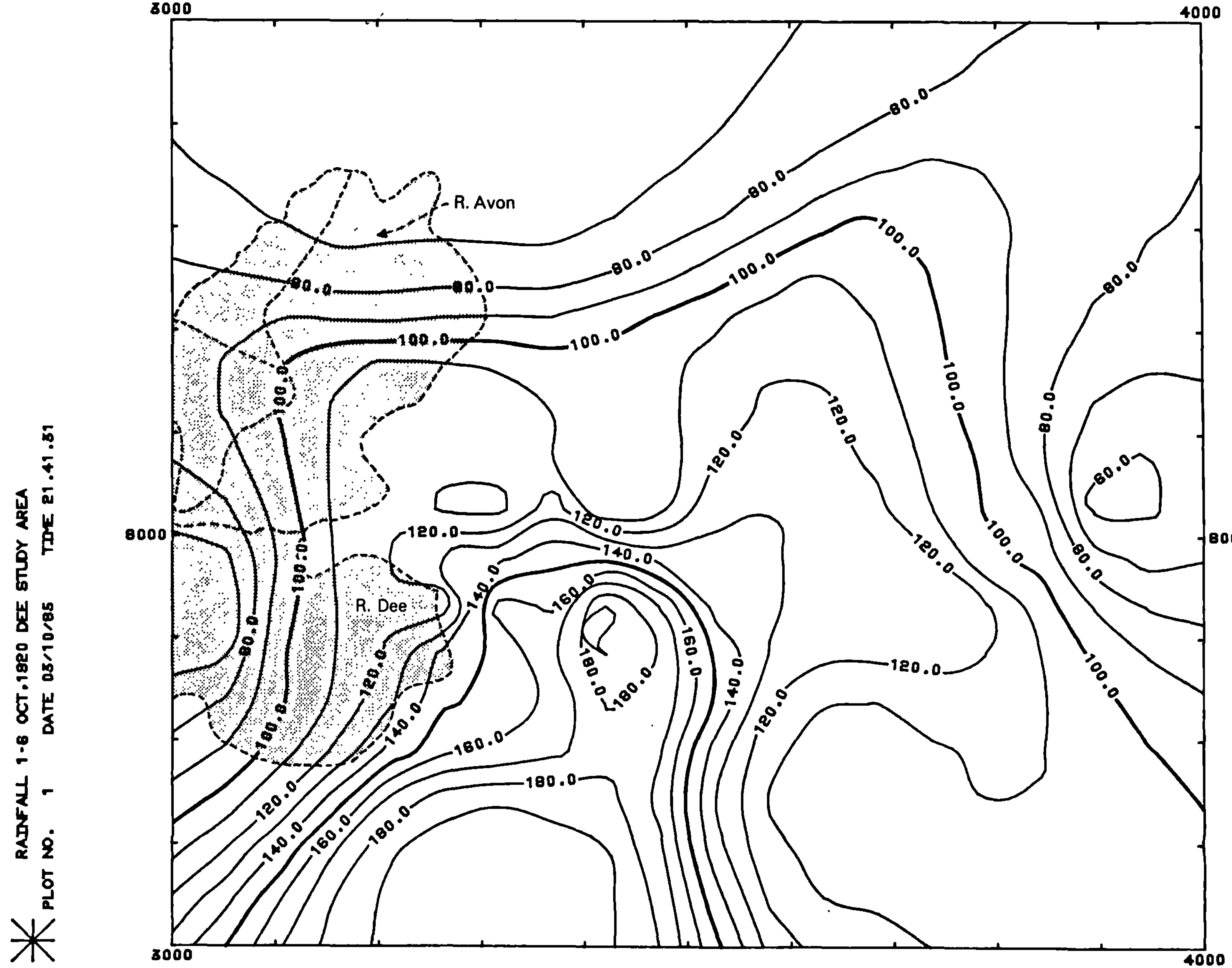
Station	24 hour (3/10/1920)	48 hour (3-4/10/1920)	72 hour (3-5/10/1920)
Braemar	49.5 mm (7 yr)	82.8 mm (25 yr)	90.9 mm (21 yr)
Balmoral	41.4 mm (6 yr)	64.0 mm (9 yr)	76.5 mm (26 yr)
Derry Lodge	26.4 mm (1 yr)	37.8 mm (1.5 yr)	-----
Ballater	63.5 mm + (19 yr)	-----	-----

* as an annual maximum

----- data unavailable

+ occurred on 4/10/1920

Figure 5.4.8.(iii) Rainfall 1-6th October, 1920 Dee study area



(3) 12-25th Jan, 1937

According to British Rainfall (1937):

"A very wet period over the whole country generally commenced on the 11th and continued until the 25th....On the 24th, a deep depression W. of Ireland moved slowly N.N.W and troughs of low pressure moved quickly N.E. wards across the British Isles, causing heavy rain and gales generally." (British Rainfall, 1937)

As detailed in Section 5.4.6, the RI of the 48 hour rainfall for the 24-25th was the highest on record at Braemar, Balmoral and Ballater with a RI of over 200 years at Braemar. Most exceptional about this event was its antecedent conditions, in terms of long duration rainfall (see Section 5.4.7). On the 12th-23rd (12 days) previous, 4.89 to 6.06 in (124.2-157.9 mm) of rain fell, and within this period Braemar and Balmoral recorded 4 daily values in excess of 1 in (25.4 mm) (Ballater and Derry Lodge recorded 3). However, this area was not the most extreme case. The station at Aitnach, in the neighbouring Muick catchment, recorded 4.49 in (114.0 mm) on the 24th and a 12 day antecedent rainfall of 9.48 in (240.8 mm).

The distributions of both 24 hour and antecedent rainfall in relation to the study area are shown in Figures 5.4.8.(iv) and 5.4.8.(v). In terms of the 24 hour fall, the study area was just removed from the centre of peak rainfall intensity over Glen Muick. Braemar was within the threshold of 3 in (76.2 mm) in 24 hours. Thus, the contribution from

Table 5.4.8.(iii)

Rainfall recurrence interval for the 1/1937 event *

Station	24 hour (24/1/1937)	48 hour (24-25/1/1937)	72 hour (22-24/1/1937)
Braemar	83.3 mm (100+ yr)	100.6 mm (200 yr)	110.5 mm (90 yr)
Balmoral	67.3 mm (80 mm)	88.9 mm (90 yr)	96.5 mm (150 yr)
Derry Lodge	70.4 mm (36 yr)	85.1 mm (28 yr)	-----
Ballater	72.4 mm (73 yr)	-----	-----

* as an annual maxima

------ data unavailable

Figure 5.4.8.(iv)

Antecedent rainfall 12-23 January, 1937 Dee study area

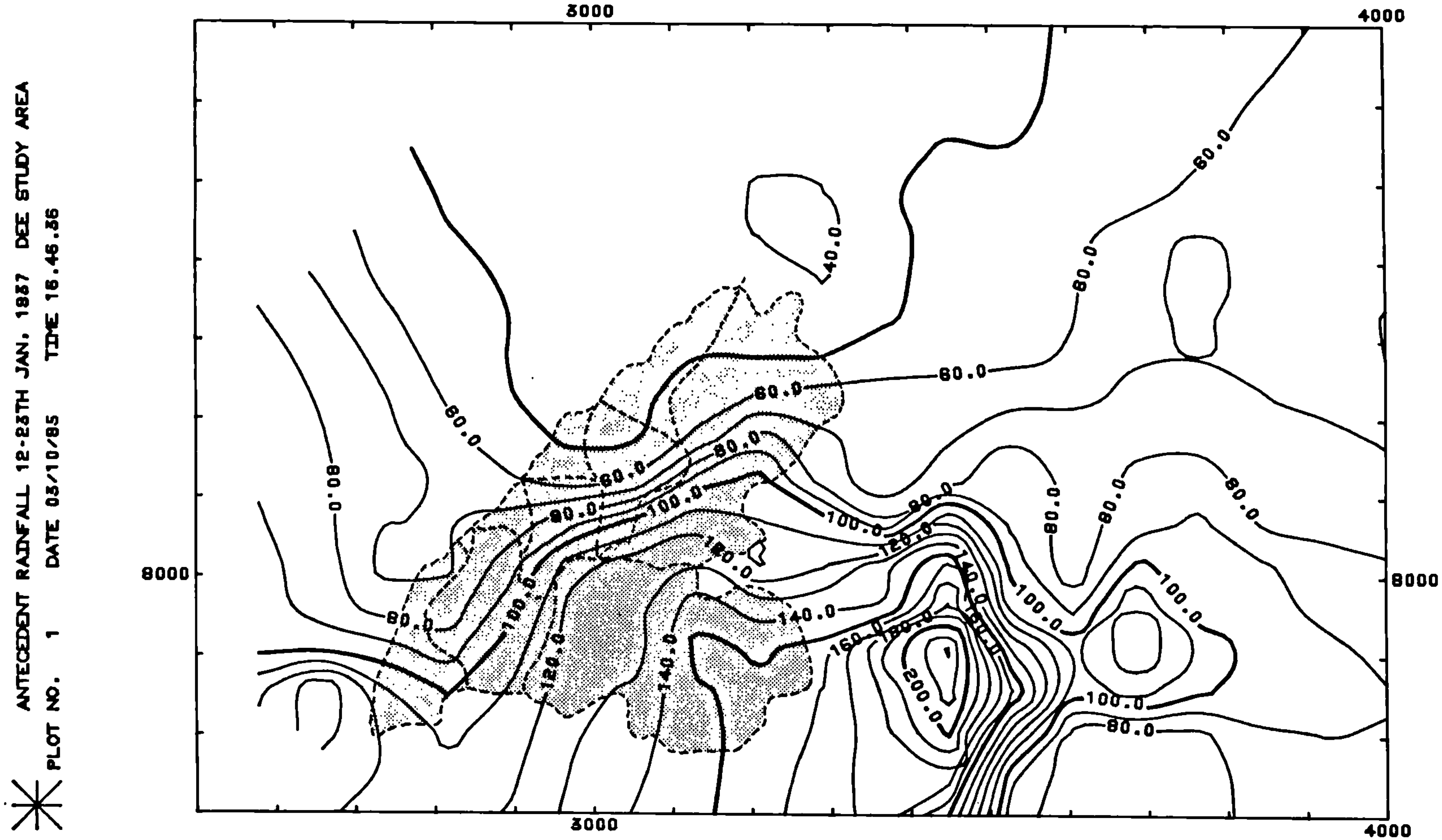
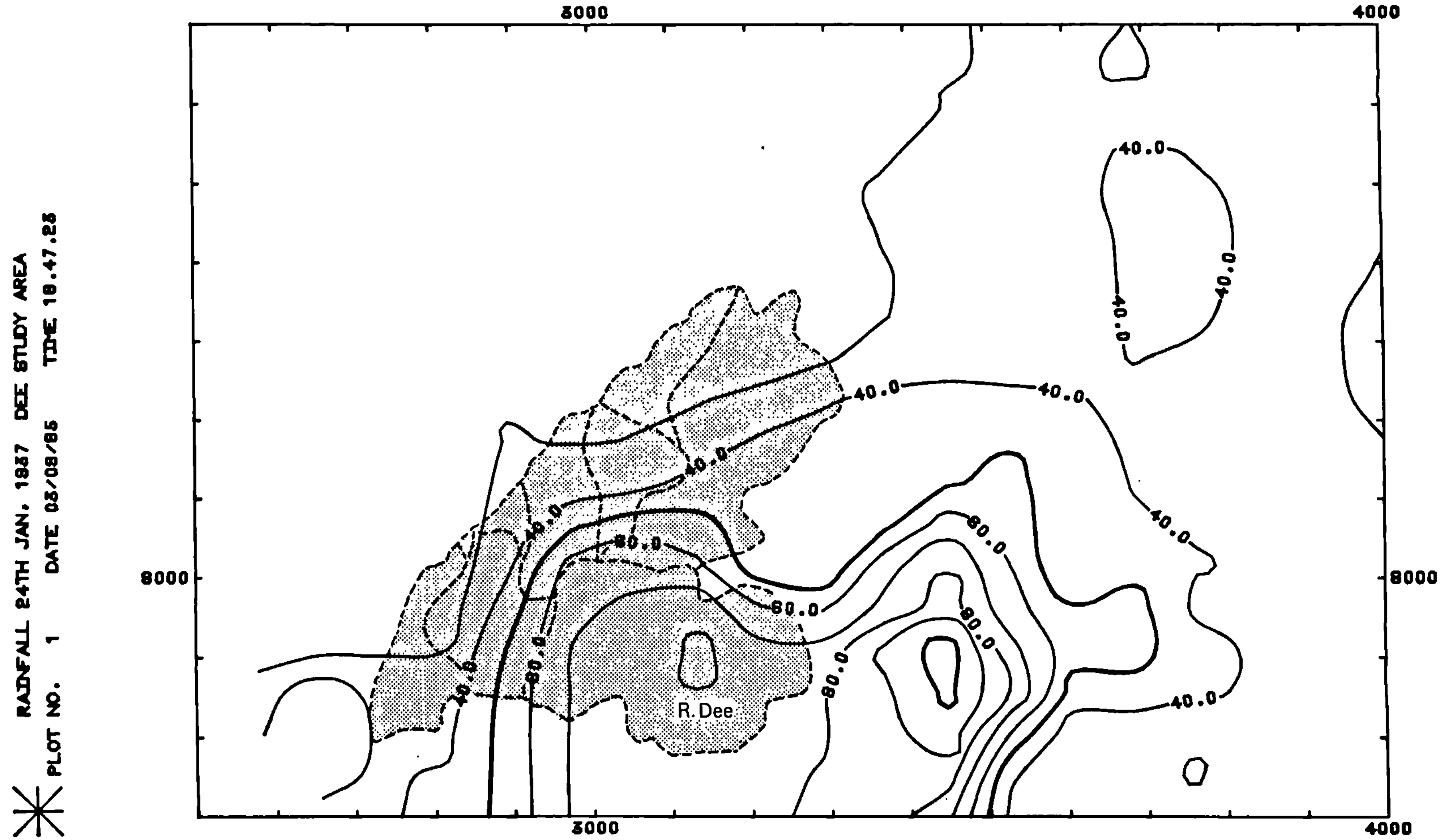


Figure 5.4.8.(v)

Rainfall 27th January, 1937   Dee study area



the Muick catchment to the highest Woodend stage must have been considerable. Had the 48 hour rainfall occurred without the antecedent conditions, it seems highly likely that a flood event would have occurred. However, the situation was exacerbated by the unusual antecedent conditions, which must have brought the catchment soils quickly to saturation level. From the Woodend record, there were three other discharge POT previous to the 24th, on the 12th, 18th and 22nd with specific runoff rates of 0.29, 0.15 and 0.37 $m^3 s^{-1} km^{-2}$, respectively. The 1920 event showed considerable spatial similarities to that of 1937 but antecedent conditions were not as extreme with the absence of POT in the series.

It is important to note that in none of these synoptic reconstructions did the cell of highest intensity rainfall centre on the upper Dee catchment. This has implications in any assessment of the maximum probable flood. It is also significant that in none of these events was snowmelt the major contributing factor. In terms of the causes of flood producing rainfall, major 48 hour events with RI of around 20 years, plus associated high antecedent conditions often involving an earlier rainfall POT, seem to precipitate major flood events. Major POT in themselves do not necessarily produce the most extreme runoff events. Lower RIs, as seen in the 1913 and 1914 events, can still produce major flood events far in excess of their 24/48 hour recurrence interval, provided they are coupled with high high antecedent rainfall conditions. Easterly winds seem to be the dominant synoptic pattern and it has been suggested that fronts move west, become stationary and then move east again, thus accounting for the double POT frequently found (Acreman, pers. comm.).

5.4.9 Summary

There have been several climatic fluctuations, which have affected the magnitude, frequency and duration of rainfall and runoff events within upper Deeside. Pre-1900, there was an increased frequency of moderate to extreme rainfall POT (both 24 and 48 hour duration) which was accompanied by a fluctuation in higher annual rainfall totals. These were followed by some of the highest annual maxima within the rainfall POT series (RI > 50 years), especially associated with summer frontal storms in August. In the 1870s and 1880s, this increased frequency of both moderate and extreme rainfall events independently confirmed the reconstructed discharge record. There was an increase in moderate to extreme flooding over this period (RI 20-50 years), though these floods when ranked were not the largest on record.

Since 1900, there have been minor climatic fluctuations with the trough of low rainfall POT of the 1970s being the most pronounced. This period was also associated with an increase in SMD and winter rainfall acceptance potential. Random high magnitude rainfall events, which generated high recurrence interval discharge events, were also more pronounced post-1900. In contrast to the discharge record pre-1900, these major floods were generated by winter cyclonic storms of much longer duration (48 hours +), with exceptionally high antecedent rainfall conditions. Localised high RI convective storms (100-200 year rainfall RI) were also recorded in some tributary catchments, eg. at the Derry Lodge gauge, since 1900.

5.5 Spey study area: Magnitude and frequency of extreme runoff

Reconstruction of the magnitude and frequency of extreme rainfall and runoff within the Spey study area will now be presented.

5.5.1 Documented flood history within the Spey study area

The historical flood information collected is found in Appendix 1.3, both for the mainstream Spey and where possible sub-divided for each tributary (Appendices 1.4-1.6). The earliest mention of flooding in Speyside, within the literature, occurred about 1630 and referred to the Spey at Badenoch:

> "Oftimes this river in tyme of speat or stormie weather will be also bigg as if it were a logh, and also as broad, and overflowes all the low corne lands of the country next to itself." MacFarlane's Collections (cited Speyside Drainage Report, 1952-1958 p6).

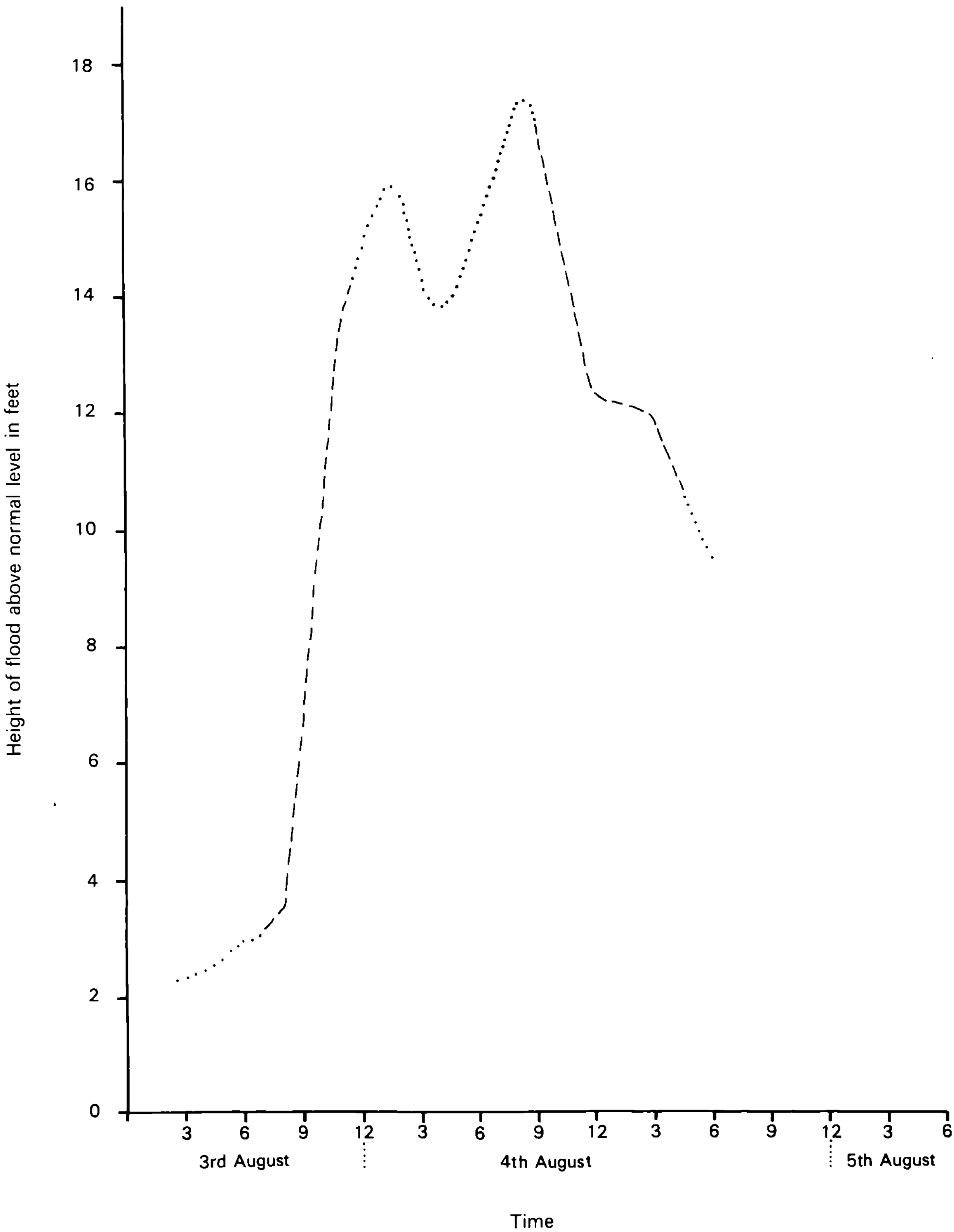
The earliest account of any specific major flood event was in Sept, 1768. This was regarded by Lauder (1830) as the worst flood prior to 4th Aug, 1829 and it was documented as being particularly destructive on the River Avon (Appendix 1.4). Major storms were also known to have taken place on Speyside in 1799 and 1812. The Aug, 1829 flood event was the largest event on record both in terms of size of area affected and general magnitude. The flood affected zone within the upper Spey valley had high relative discharge rises reported in Lauder (1830) and these indicated clearly that the floodplain was inundated for much of its breadth from Kingussie downstream. Information on both rainfall and discharge was limited, although Lauder cites the report of 3.75 in (95 mm) in 24 hours at Huntley, within the Findhorn valley. This value was a gross underestimate of the falls on higher ground, which must have been in the region of 150 mm. Estimate of Q_{max} at Boat o' Brig (Werritty, pers. comm.) was 1665 m^3 s^{-1} calculated from the reconstructed hydrograph, shown in Figure 5.5.1.(ii) (Thorne, Ph.D. in preparation) and the estimated RI was 500-1000 years. It is clear within the Spey study tributaries and especially the Avon, that rates of both rainfall and associated runoff were very high:

> "At Poll-du-ess [upper Avon], ...the river is bounded by perpendicular rocks on each side. There the bed of the stream is 44 ft [13.4 m] broad and the flood was 23 ft [7 m] above the usual level. Deep as the ravine was, the river overflowed the top of it. From correct measurements taken, the column of water that passed here, with intense velocity, appears to have been about 1200 square feet [111 m^2] in its transverse section." (Lauder, 1830 p184-5)

Thus, even if velocities were only 2 m s^{-1}, this would give specific runoff rates of over 2.5 m^3 s^{-1} km^{-2}; velocities were most certainly much higher.

Figure 5.5.1.(i)

The hydrograph of the August, 1829 flood event at Boat o' Brig in the Spey valley



Source: P. Thorne Ph D in preparation Information from Lauder (1830)

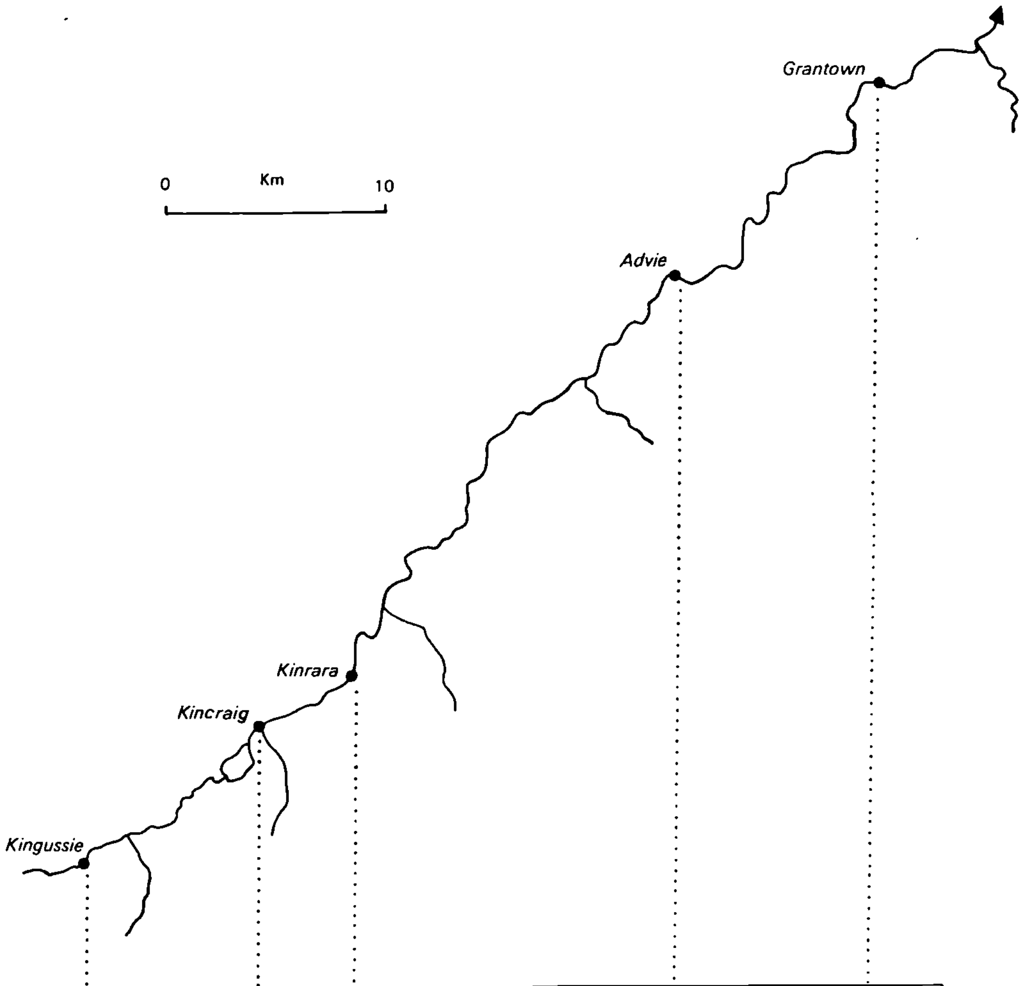
Since 1829 but pre-1900, several other major flood events have occurred eg. 24-26th Jan, 1849 and 30-31st Jan, 1868. An attempt was made to rank their relative magnitude at various cross-section locations downstream, as shown in Figure 5.5.1.(ii). Clearly there was a difference in ranking downstream, eg. Kingussie flood history compared with that of Gordon Castle, depending on areas of most intense rainfall contribution. It should be expected with such a large catchment that many events would seem to affect either the upper or lower river, especially if some of the major storms come off the sea. The most memorable floods within the whole Spey catchment affected both.

It is notable that several major events pre-1900 occurred with a spacing of approximately 20 years (in 1829, 1849, 1868 and 1892/4), eg. the 1869 flood event was generally only 2 ft (0.6 m) below the 1829 event. Quite frequently within the same year, one major flood occurred and was followed by another of similar magnitude. There were numerous examples of this eg. Aug, 1829 (Appendix 1.3), 30th Jan, 1892 and 8th Oct, 1892, Grant's account (see Appendix 1.3.1) of the 1st Jan, 1868 event, July, 1956 and July/ Aug, 1970. The collated historical information is tabulated in Figures 5.5.1.(iii) to 5.5.1.(x) and hydrometeorological characteristics given in Table 5.5.1.(i).

The seasonality of these major flood events pre-1900 was frequently October to January, associated with winter cyclonic or orographic storms. In contrast, the extreme 1829 flood was a summer frontal storm. Within the gauged tributary record however, the importance of August as a flood month must be noted. In terms of their

Figure 5.5.1.(ii)

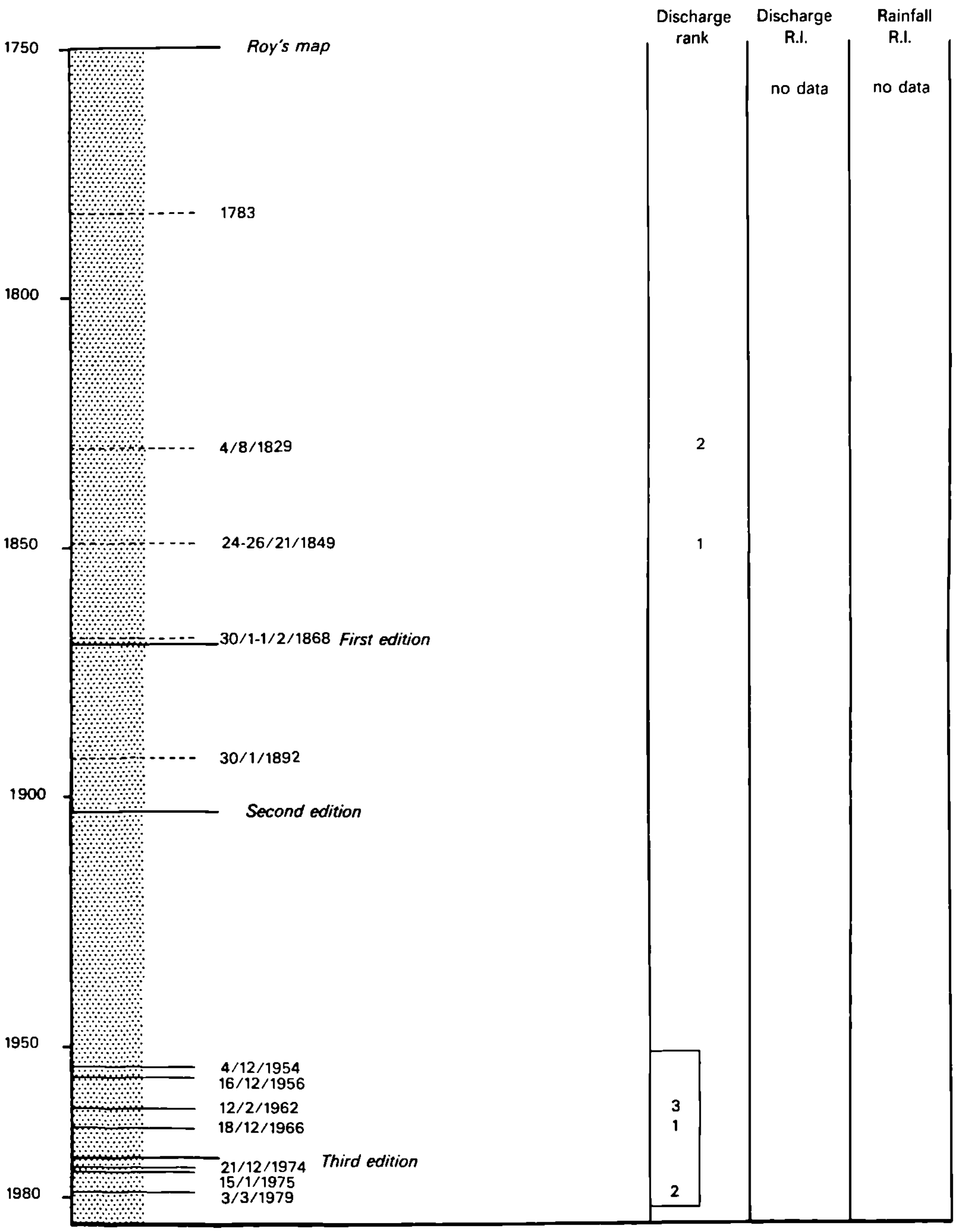
Ranking of the relative magnitude of pre-1900 flood events within the Spey study area



4/8/1829	2	2	2	1	1
24-26/1/1849		1	1		
30/1-1/2/1868				2	3
2/1 1892				2	2
5/2/1894	1				

Figure 5.5.1.(iv)

The flood history on the River Spey at Kinrara



Discharge gauge: Kinrara

Figure 5.5.1.(v)

The flood history on the Spey at Grantown

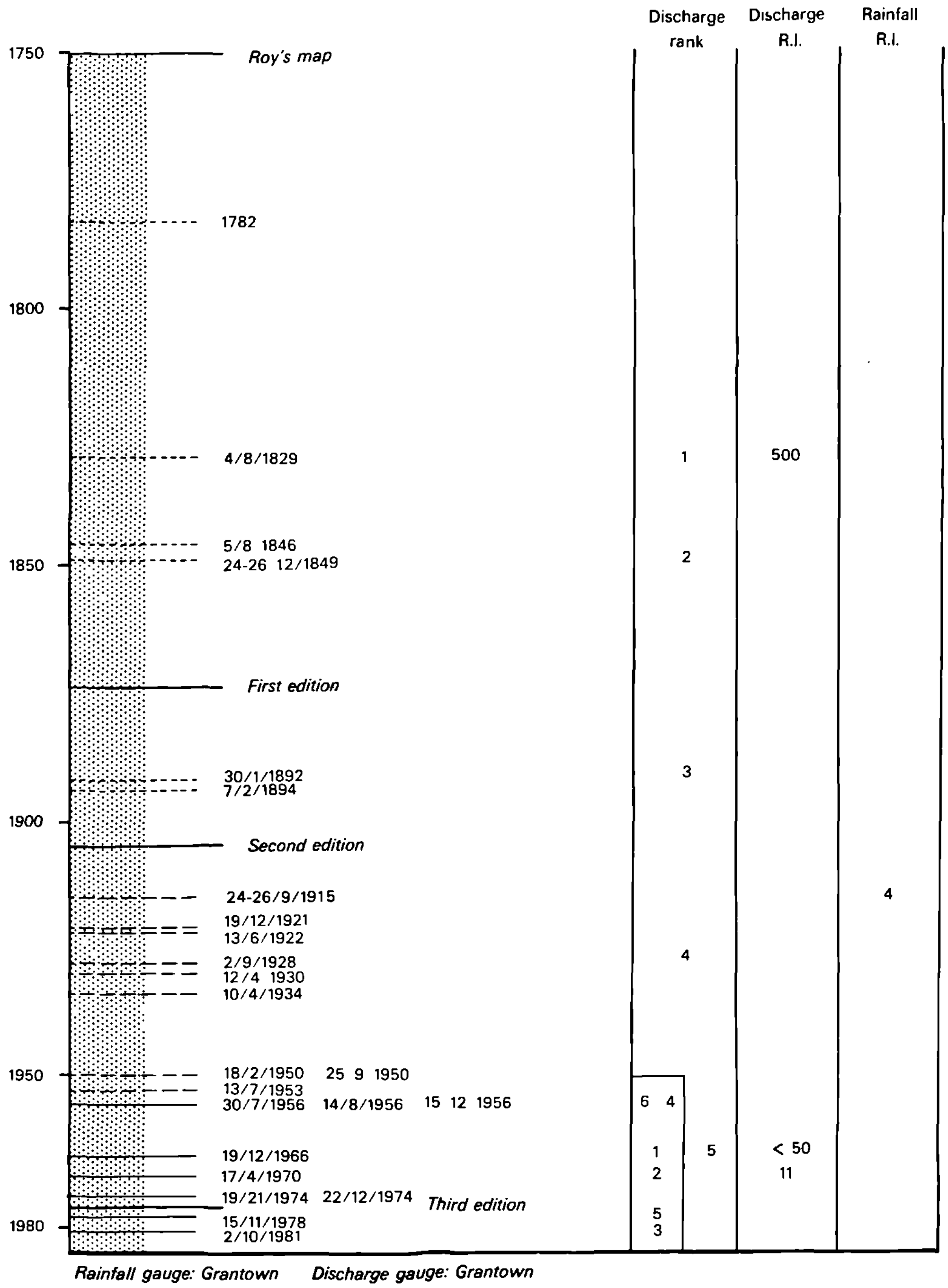
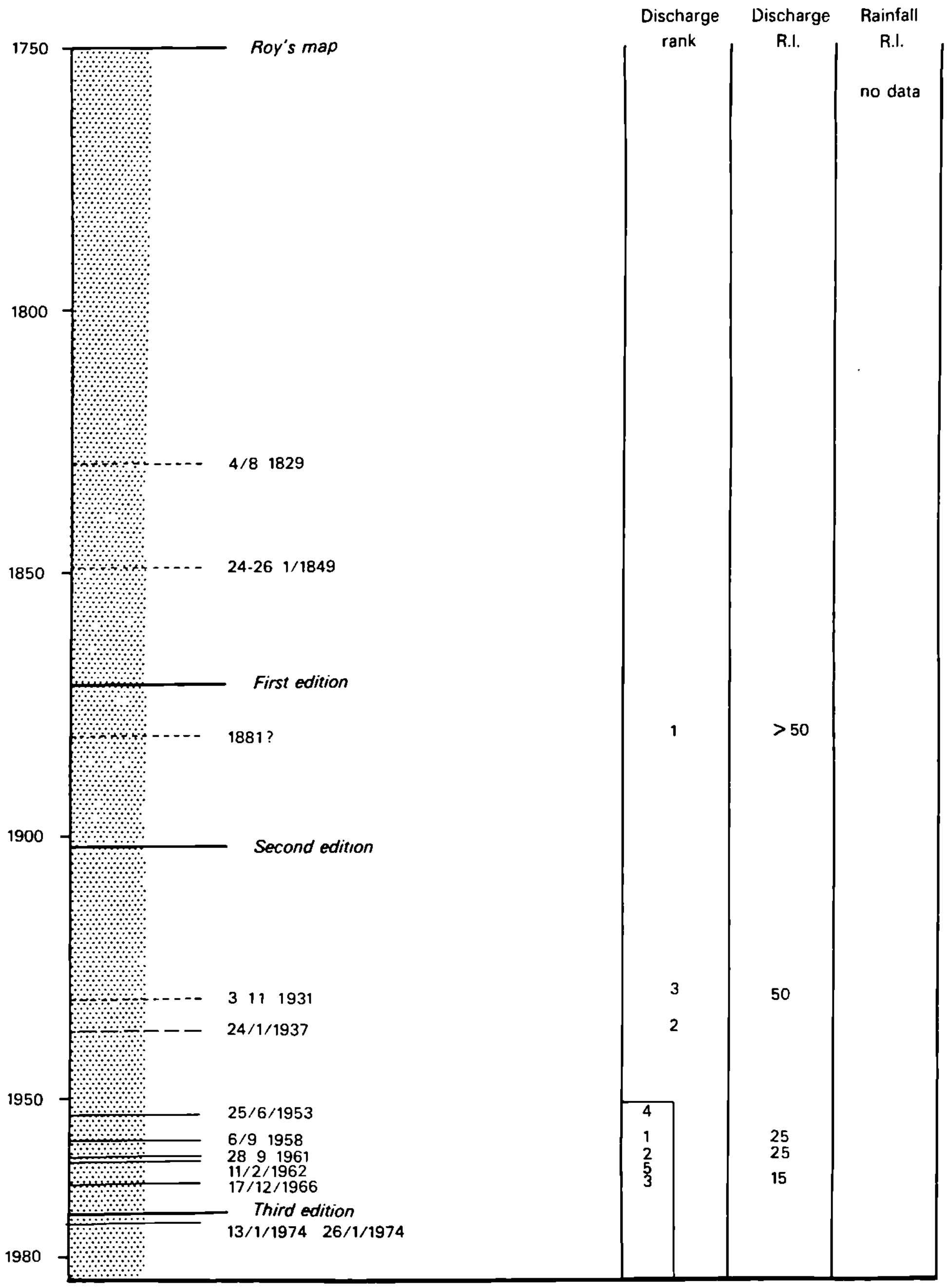


Figure 5.5.1.(vi)

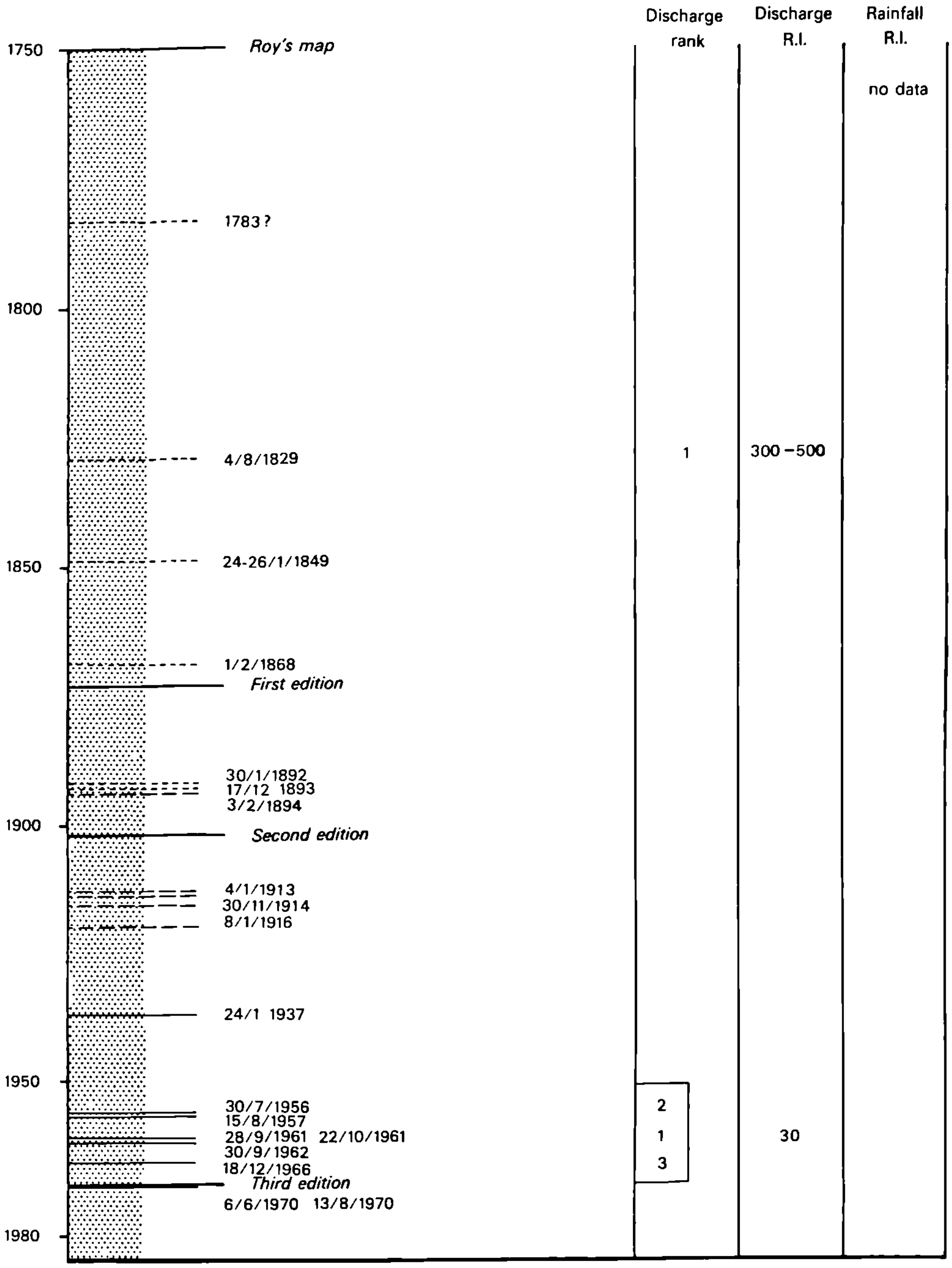
The flood history within the Tromie catchment



Discharge gauge- Tromie bridge

Figure 5.5.1.(vii)

The flood history within the Feshie catchment



Discharge gauge: Feshiebridge
St Andrews Univ. gauge

Figure 5.5.1.(viii)

The flood history within the Druie catchment

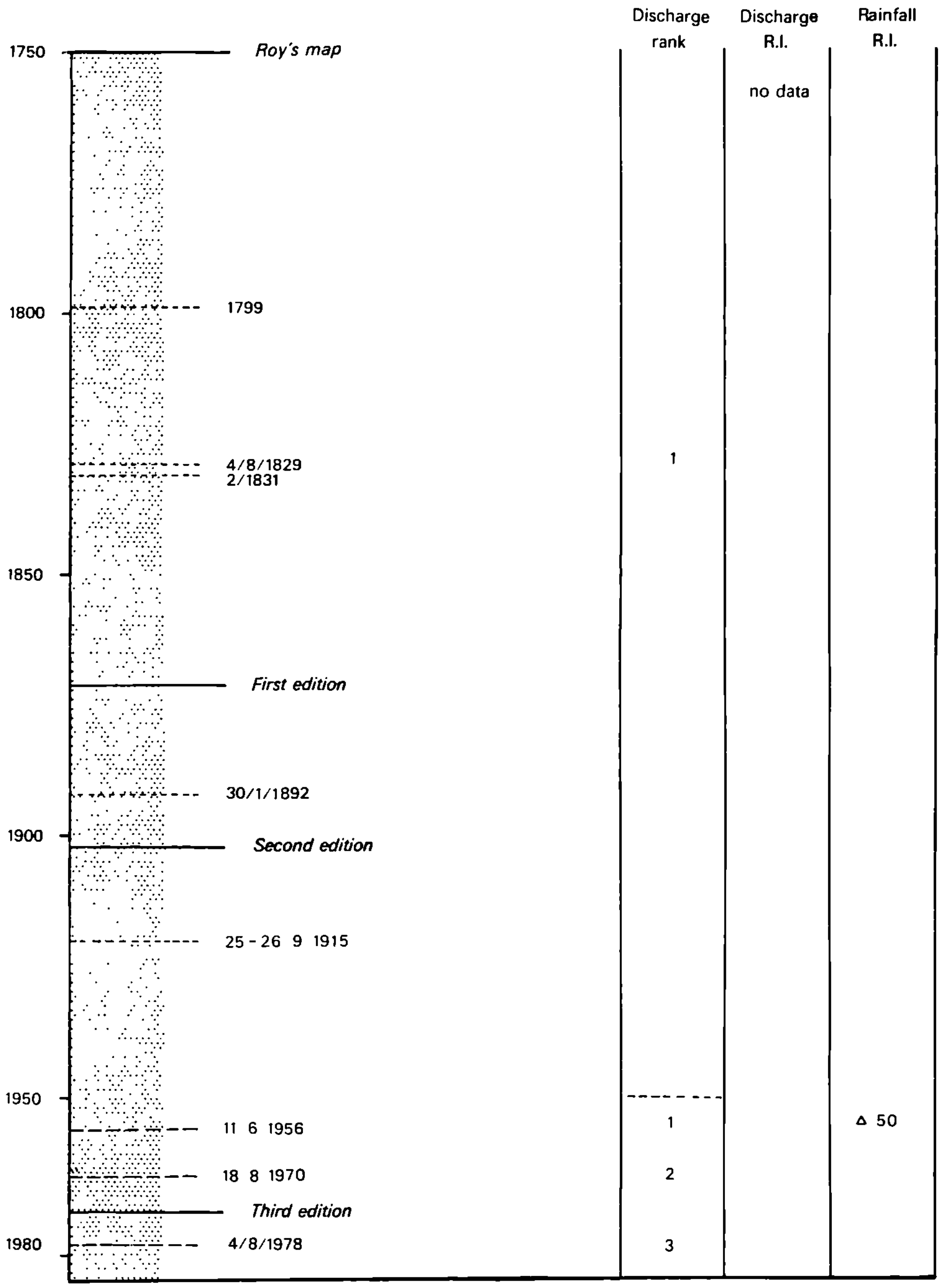
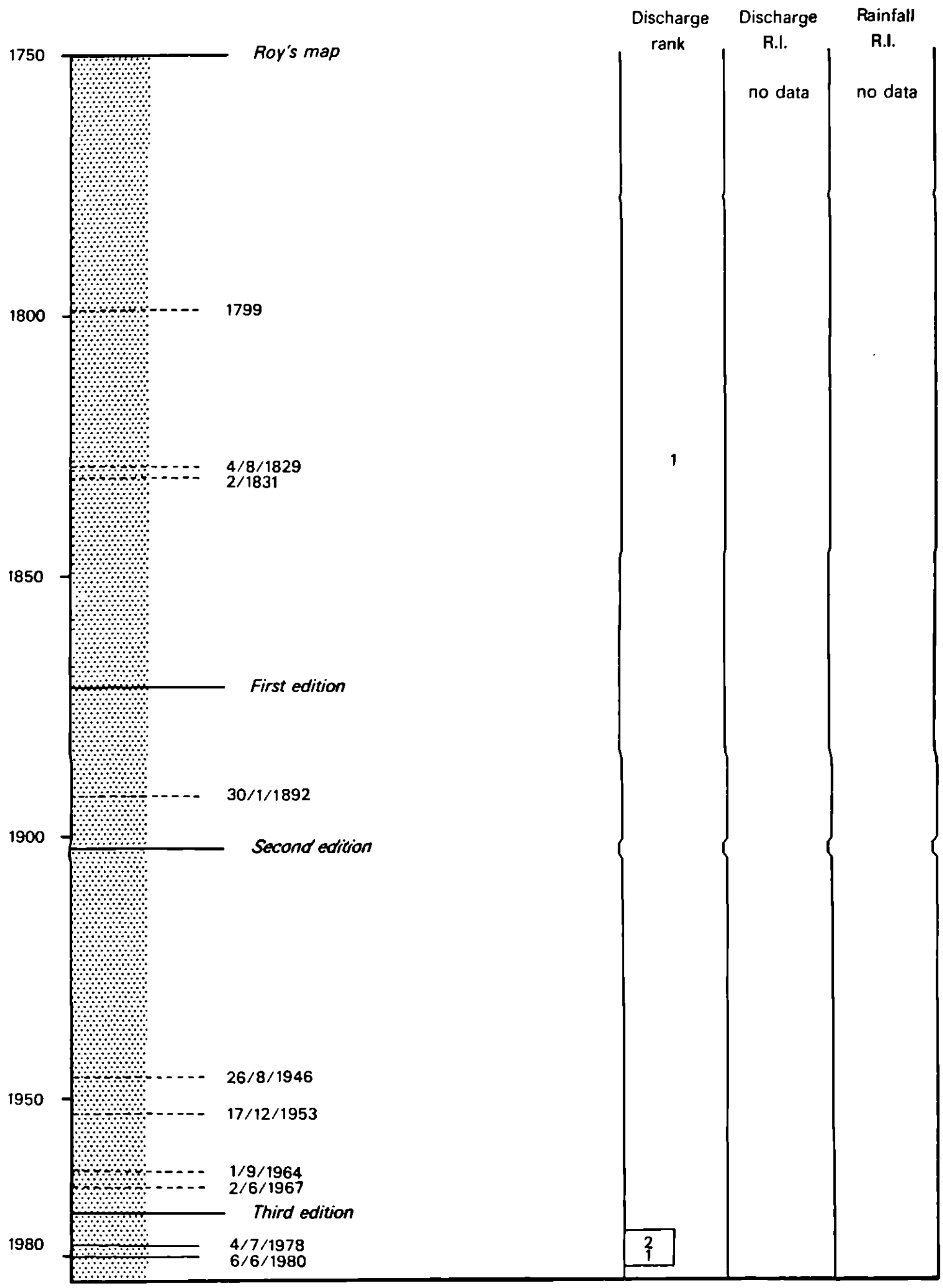


Figure 5.5.1.(ix)

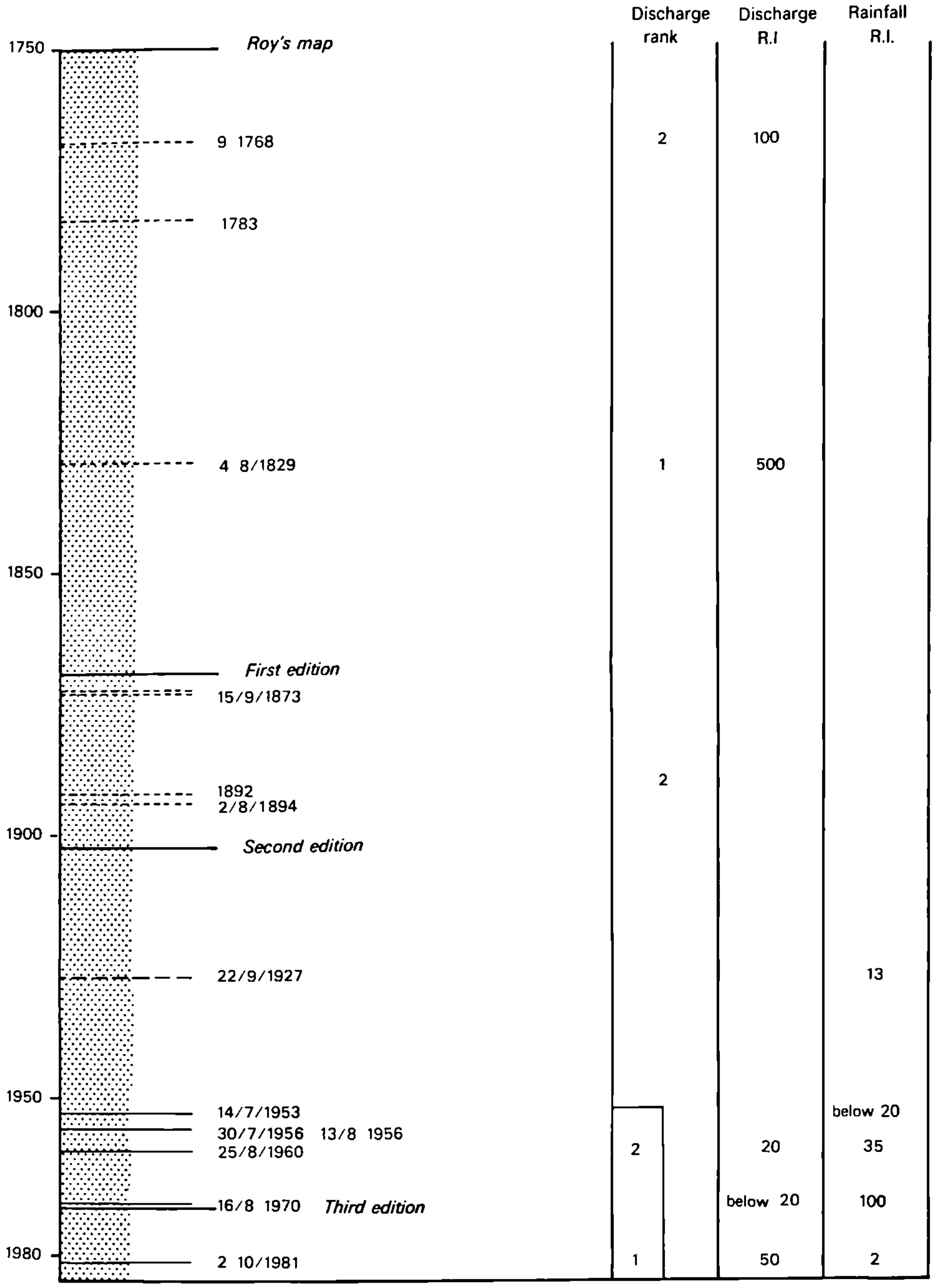
The flood history within the Nethy catchment



Discharge gauge Nethybridge

Figure 5.5.1.(x)

The flood history of the River Avon catchment



Rainfall gauge: Ballindalloch
Discharge gauge: Delnashaugh

Figure 5.5.2.(i)

The seasonality of flooding on the River Spey at Kingussie

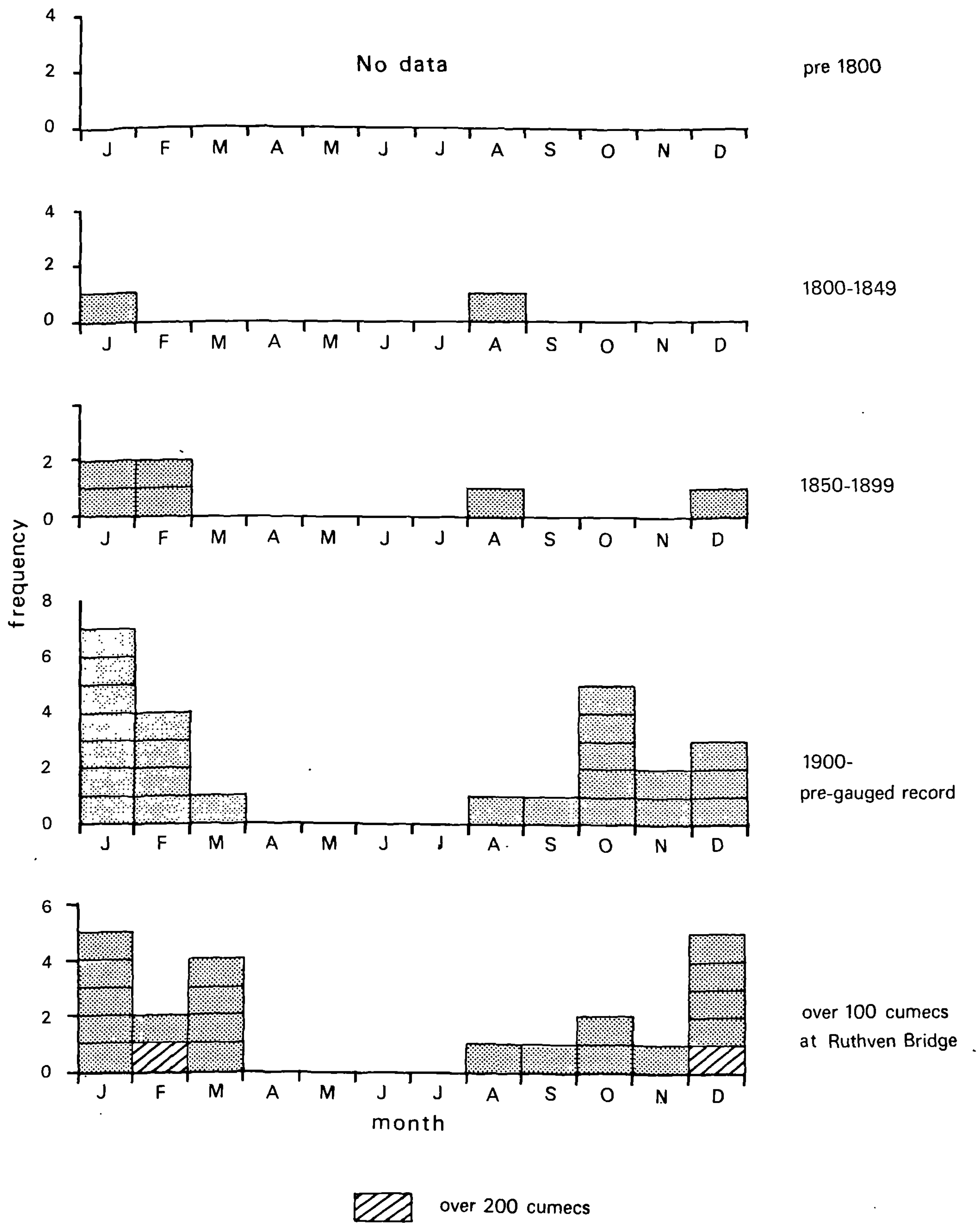


Figure 5.5.2.(ii)

The seasonality of flooding on the River Spey at Grantown

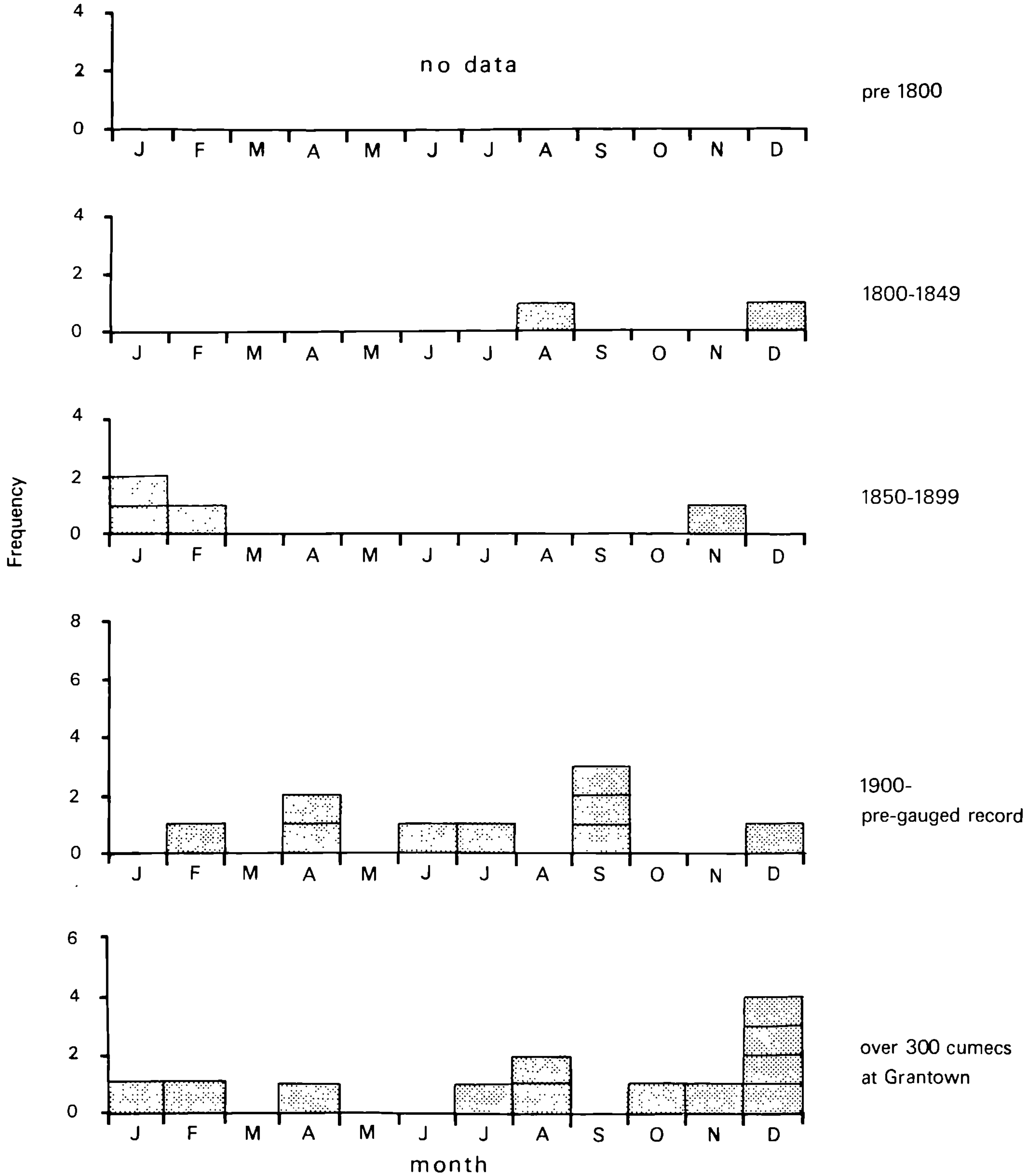


Figure 5.5.2.(iii)

The seasonality of flooding on the River Tromie

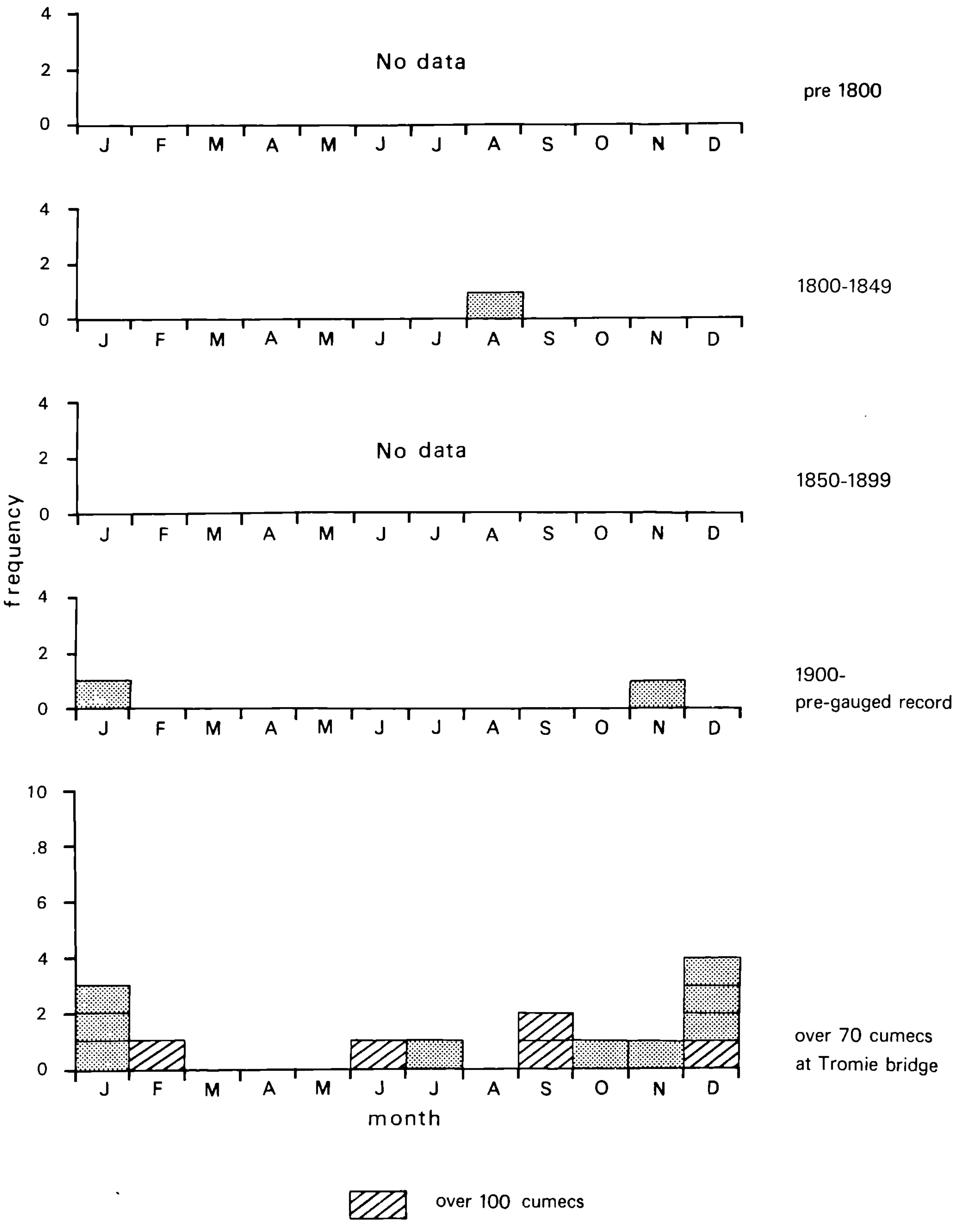


Figure 5.5.2.(iv)

The seasonality of flooding on the River Feshie

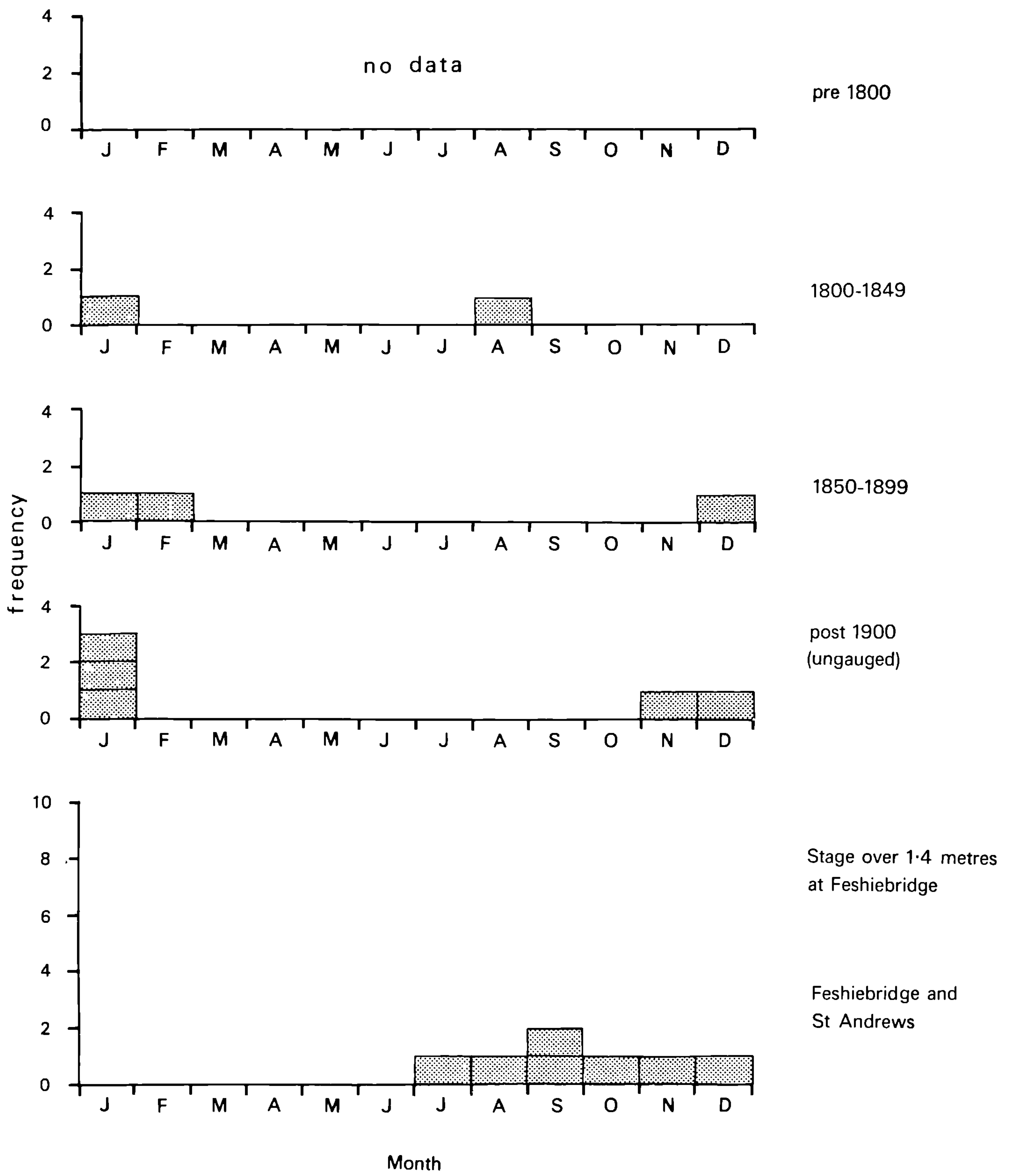


Figure 5.5.2.(v)

The seasonality of flooding on the River Druie

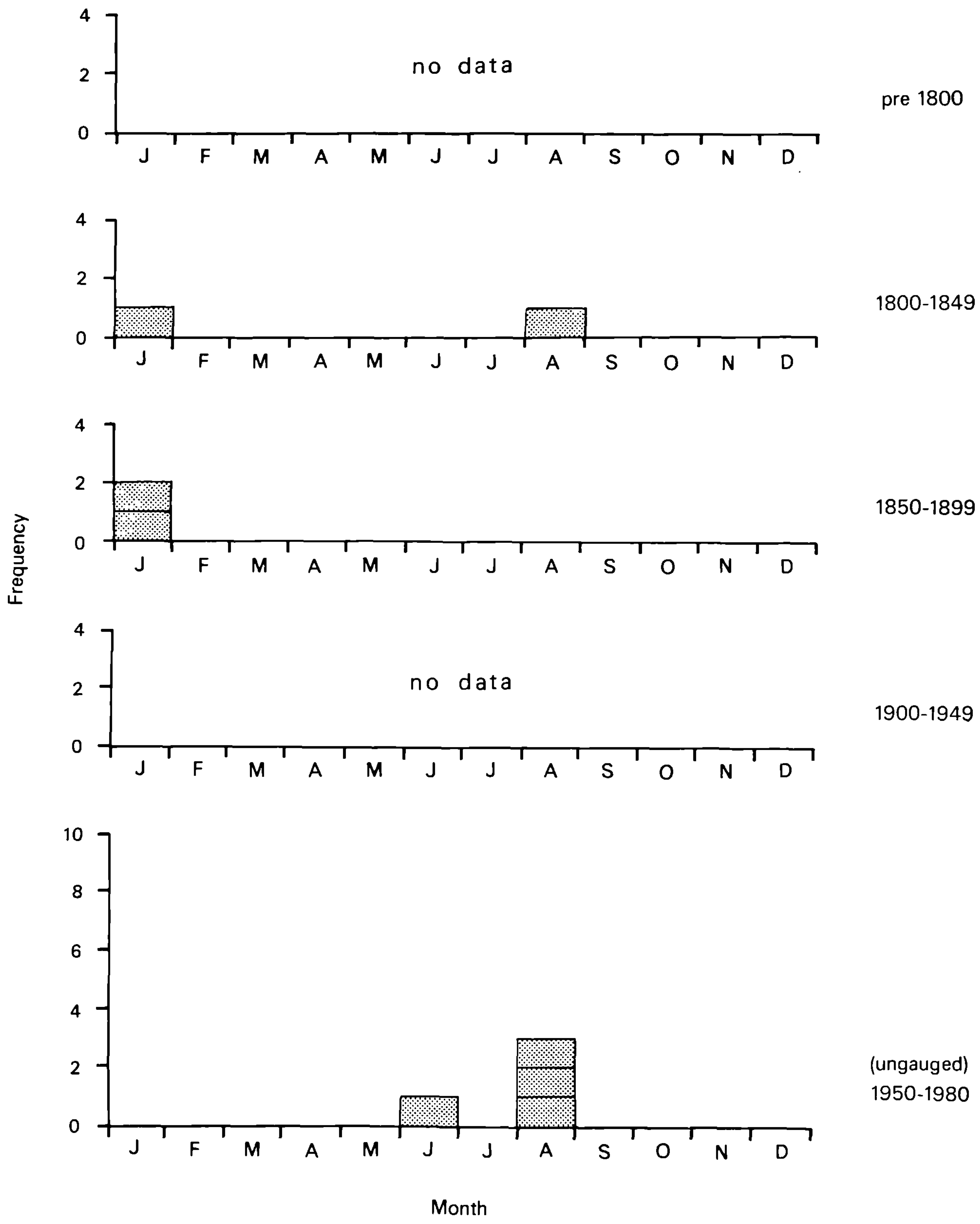


Figure 5.5.2.(vi)

The seasonality of flooding on the River Nethy

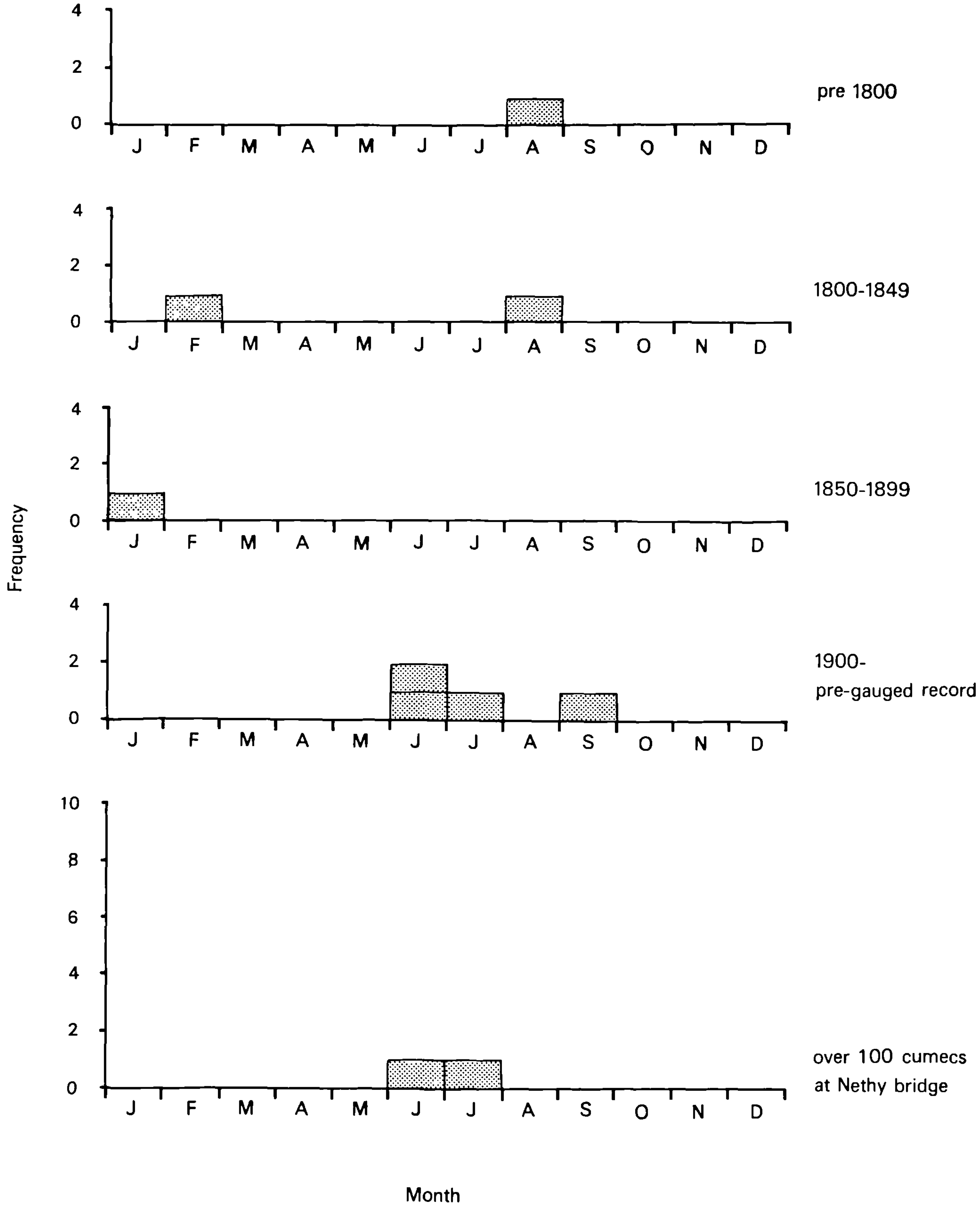


Figure 5.5.2.(vii)

The seasonality of flooding on the River Avon

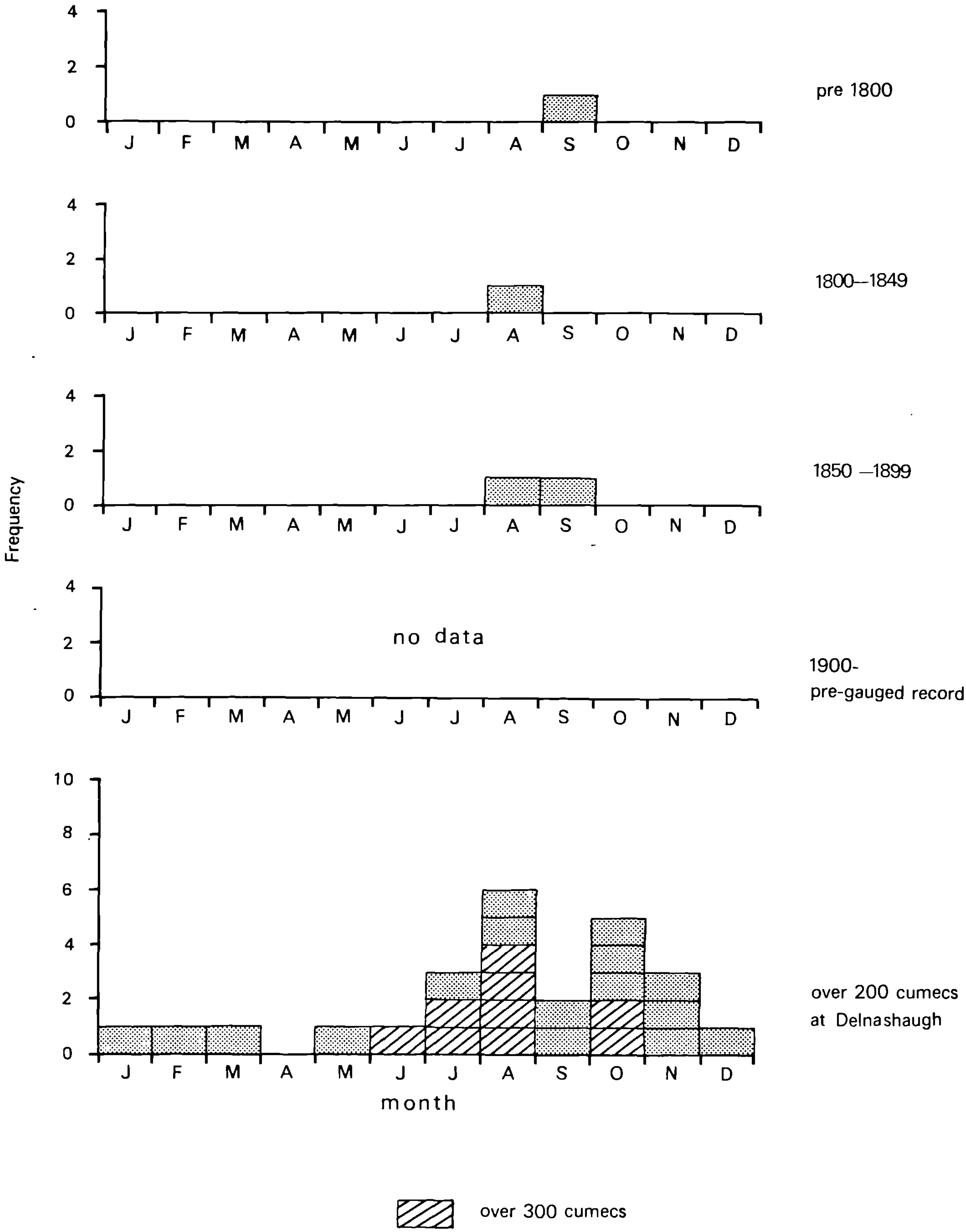


Table 5.5.1.(i)

Hydrometeorological characteristics of historical flood events within upper Speyside

A Regional storm

B Convectional storm

C Intense rainfall

D Thunder and lightening

E Wind

F Wind direction

G Snow melt contribution

H Ice break

I Temperature rise

J Ground frost

K Pre-flood ground saturation

Event	A	B	C	D	E	F	G	H	I	J	K
28/7/1779			*	*							*
4/8/1829	*		*	*	*		(*)				
13/8/1829	*										*
5/8/1846		*	*	*							
24/12/1849	*										
19-21/10/1864	*				*	N					*
30-31/1/1868	*				*	W	*				
30/1/1892	*				*		*			*	
14/8/1894	*										
19/10/1906			*				*				
24-26/9/1915	*				*	NE					
7/7/1923		*	*								
28/1/1927		*	*		*						
9/2/1928							*				
10/4/1934		*					*				
20/10/1935	*										
17/10/1944							*				
17/8/1948	*										
25/9/1950	*										
11/6/1956	*		*			NNE					
19/22/1966	*						*				
6-7/6/1970	*			*	*						
18/8/1970	*					NNE					

hydrometeorological origins, pre-1900 only two storms (30th Jan, 1892 and 31st Jan, 1868) had a substantial snowmelt contribution when rain fell on accumulated snow (Appendix 1.3). In contrast post 1900, although large snowmelt floods have occurred, they did not seem to be associated with heavy rainfall and did not represent the most extreme flood events. The exception was on 18th Dec, 1966 but again rainfall at Ballindalloch was low.

5.5.2 Spey study area: Analysis of discharge records

The available discharge records for the Spey catchment and the study region specifically are tabulated in Table 3.3.5.(i). Identifying the main flood events within the discharge record for the mainstream Spey was achieved by comparing the peak over threshold values for each station (Source: Acreman, Ph.D. in preparation). Unfortunately, as can be seen from Table 3.3.5.(i), the longest record within the Spey study area extends back only to 1951 but Aberlour (with a catchment area of 2640 km^2) has a record from 1938, and this allowed identification of regional flood events over a longer period. Flood marks estimating the maximum stages of historic events were not found in this area, though there may be some at Aberlour.

The most extreme event on the Aberlour record was on 17th Aug, 1970 (1218 m^3 s^{-1}) and this was also registered as a POT on most of the post-1951 records. Specific runoff rates ranged from 0.13 m^3 s^{-1} km^{-2} at Boat of Garten and Kinrara to 0.77 m^3 s^{-1} km^2 at Delnashaugh on the Avon. The different magnitudes of discharge for varying recurrence intervals are shown in Table 5.5.2.(i). Within these records, the approximate RI of the 1970 event at Aberlour was just below 100 years. Upstream however, the RI varied from 2 to 3 years at Kinrara and Boat of Garten, to under 20 years at Delnashaugh. This event was centred mainly on the middle to lower river and demonstrates clearly how discharges from the same event may have markedly different RIs at different locations downstream. From the gauged record, the Q_{max} associated with summer storms decreases as one goes upstream on the Spey (Acreman, pers. comm.).

The most extreme gauged event in terms of the upper Spey was on 18th Dec, 1966 (associated with rain and snowmelt) with specific runoff varying from 0.21 $m^3 s^{-1} km^{-2}$ downstream at Aberlour to 0.64 $m^3 s^{-1} km^{-2}$ at Invertruim. The RI of this event was consistently c. 50 years at the gauges within the study area but at Aberlour and Boat o' Brig, the RI was reduced to only 5-7 years. A third event occurred on 11th Feb, 1962 and although at Ruthven Bridge this had a discharge RI of under 5 years, more generally the value was between 20-30 years. It is notable that post-1900 on all records, there was no flood event that exceeded 50 years RI, in marked contrast to the major floods on the Spey pre-1900.

On the gauged tributaries, the most extreme event on the Avon was on 2nd Oct, 1981 with a discharge of 521.6 $m^3 s^{-1}$ (0.96 $m^3 s^{-1} km^2$) and again this had a estimated RI of less than 50 years (see Appendix 2.2). The second largest was 21 years earlier on 25th Aug, 1960 with a discharge of 435.7 $m^3 s^{-1}$ (0.80 $m^3 s^{-1} km^{-2}$) and with an estimated RI of 20 years. A similar pattern was found on the Tromie (Appendix 2.2), with the maximum recorded discharge at 155.1 $m^3 s^{-1}$ on 6th Sept, 1958

Table 5.5.2.(i)

Speyside: Magnitudes of discharge for different recurrence intervals, fitting the EV1 distribution to the annual maximum series

River	Station	Q(2)	Q(5)	Q(10)	Q(20)	Q(50)	Q(100)	Q(100)*
Spey	Kinrara	125.0	174.4	219.4	273.2	363.3	450.4	349.9
Spey	Ruthven Bri.	098.0	129.0	154.0	180.9	223.0	259.6	267.1
Spey	Boat of Gart.	157.4	215.8	263.6	318.4	403.3	479.4	438.7
Spey	Boat o' Brig	435.1	586.4	695.2	813.4	978.9	1116.0	1162.1
Spey	Invertruim	091.3	133.8	168.9	208.2	269.7	326.0	263.4
Spey	Grantown	229.5	301.9	356.8	414.2	494.0	561.4	618.5
Spey	Aberlour	402.7	567.4	700.0	846.6	1070.8	1272.1	1068.6
Avon	Delnashaugh	208.9	299.4	366.7	438.7	538.5	619.7	576.1
Tromie	Tromie Bri.	063.1	092.5	116.7	144.9	186.4	225.3	182.0

Q(i) discharge with a recurrence interval of i years

All values are in $m^3 s^{-1}$

* calculated using the regional curve ordinate.

(1.19 m^3 s^{-1} km^{-2}), in excess of the 20 year RI. In this 30 year period, there were 5 events with an estimated RI of between 10 and 20 years. How typical or atypical this period was, not only in extremes but also in more moderate flows, must be examined in context of the previously collated flood history information and the following rainfall analysis.

The available discharge records for the Feshiebridge and Nethybridge gauges did not have good enough rating curves to be included in the FSR (NERC, 1975). However, major flood events were extracted from the DAFS microfiche and raw data sheets (Feshie and Nethy, respectively) and their relative ranks are annotated in Figures 5.5.1.(vii) and 5.5.1.(ix). Within both catchments, clearly the major flood events occurred before the gauged record, although the Sept, 1961 event on the Feshie must have a RI of over 30 years and 20th Sept, 1981 flood event on the Nethy was associated with specific runoff rates of 2 m^3 s^{-1}. In terms of the seasonality of flood events, the most major, as within Deeside, were the summer frontal storms which have sufficient rain to bring the ground up to saturation level and more. The most consistent events however were the winter cyclonic events, particularly in January and December.

5.5.3 Spey study area: Analysis of rainfall records

Unfortunately in the Spey study area, there are only 2 long length rainfall records, as seen in Table 5.5.4.(i). There is however a much longer record extending back to 1860 at Gordon Castle (Fochabers) but this is much further down the Spey. It would however be expected that general patterns of annual total rainfall would be similar downstream but not as large. A composite record for Grantown-on-Spey was constructed using annual ratios for shorter lengths of record at different sites. The accumulated totals for 1867-1939 from the Grantown and Gordon Castle records were also correlated and no significant breaks were found on the double mass plot.

To understand firstly the basic trends of the data, the annual totals were plotted against years of record for Grantown, Ballindalloch and Gordon Castle and are shown in Figure 5.5.3.(i) to 5.5.3.(iii). From the Grantown record, which extends back the furthest, the wettest year like the Deeside Braemar record was 1872 with 1082.8 mm. Other notably high annual totals occurred in 1916 (1038.6 mm) and 1927 (1027.4 mm). This pattern was confirmed by the Gordon Castle and Wester Elchies records, with 1109.2 mm and 1049.3 mm respectively in 1872. The wettest years on the Ballindalloch record were 1923 (1075.4 mm), 1927 (1149.1 mm), 1970 (1023.4 mm) and 1980 (1005.3 mm).

When a 5-year running mean was put through the annual total data, there was much oscillation around the long-term average annual total, although the period 1967-1977 stood out as having well below average values. However, if the years 1951-1982 were studied ie. the period of

Table 5.5.4.(i)

Studied long rainfall records within the Spey catchment

(short records used in brackets)

Station		N.G.R.	altitude	Length of record
Ballindalloch		3185 8369	192	1920-1982 (monthly)
				1923-1982 (daily)
Grantown-on-Spey		3028 8274	219	1867-1939 (monthly)
Muckrach Lo.	c.	2989 8251	229	1897-1921 (monthly)
Grantown Stat.	c.	3023 8266	229	1873-1889 (monthly)
Gordon Castle		3350 8595	32	1866-1974 (monthly)
				1891-1974 (daily)
Glenmore Lo.		2986 8095	341	1950-1970 (daily)
Wester Elchies		3255 8432	168	1872-1877 (daily)
[Moniack		3028 8274	213	1950-1957 (daily)]
[Bieldside		3035 8274	229	1944-1949 (daily)]
[Heathfield		3039 8285	229	1967-1980 (daily)]

c. approximate N.G.R. and height

Figure 5.5.3.(i) BALLINDALLOCH: ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN: (1920-1982)

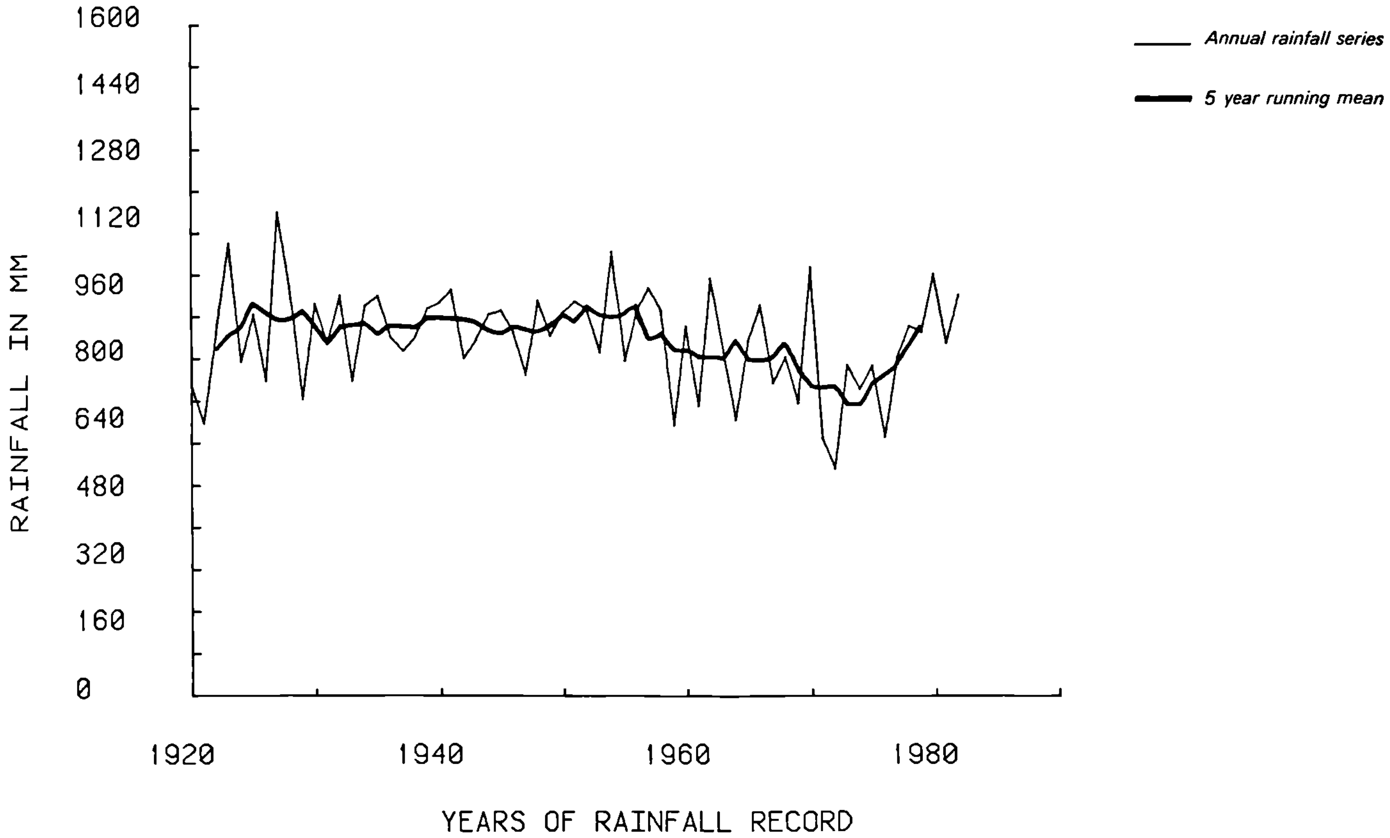


Figure 5.5.3.(ii) GRANTOWN: ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN: (1868-1939)

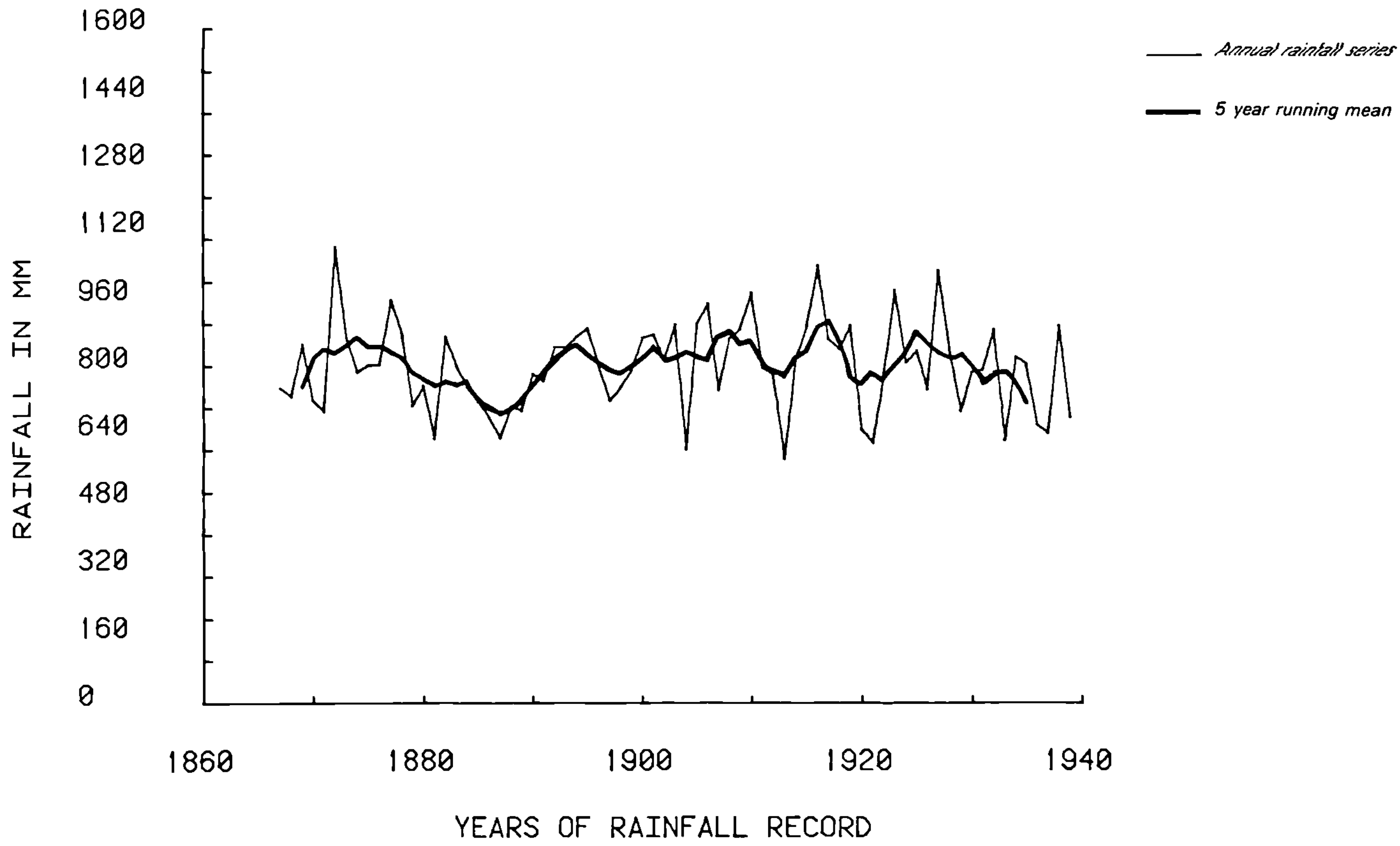
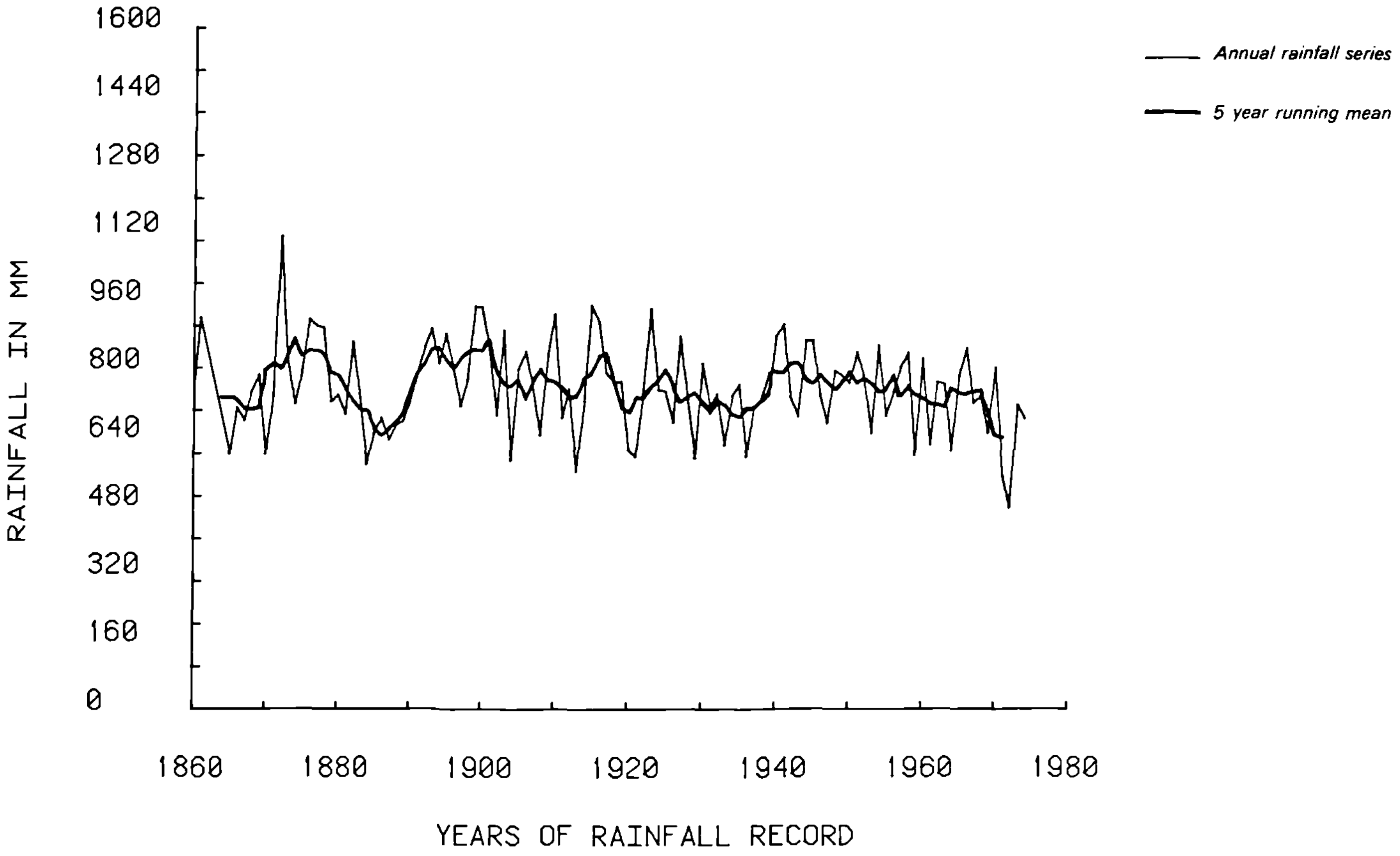


Figure 5.5.3.(iii)

GORDON CASTLE: ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN: (1860-1974)



the gauging record, 13 years were above average and 19 below. When periodicities within the annual total data set were assessed, there was no predominant cycle of "n" years, however possible periodicities should be mentioned. Within the Grantown record, the only positive significant autocorrelation (AC) was at 33 years. Fourier analysis showed no exceptional extreme peaks although there were smaller increases in spectral density at approximately 8, 6 and 4 years. Ballindalloch in contrast, had high positive AC at 4 (0.256), 12 (0.178) and 47 (0.186) year lags.

The Gordon Castle record allowed Box-Jenkins analysis to be carried out on 110 years of annual total data though there were no significant positive AC. The 22-24 year lags did however have a run of positive AC and Fourier analysis showed the highest peak of spectral densities again occurring between 21-23 years with a value of 7.91. Thus, the only possible longer cycle suggested regionally by the three data sets was at 21-24 years, although there were other less pronounced shorter term cyclicities. This included a persistent shorter cycle at around 4 years. This would be consistent with the Deeside results, though at these particular Speyside stations the periodicity was not as significant. Obviously, there may be other cyclicities but those mentioned were the most pronounced.

5.5.4 Spey study area: Analysis of rainfall POT series

In general, the frequencies of POT for different magnitudes (eg. 24 hour rainfall > 25.4 mm) and durations within the available Spey record were much lower than on the Dee, probably as a reflection of the altitudinal difference. At comparatively higher altitudes, rainfall values are probably more similar, as seen by the Glenmore Lodge record. At Ballindalloch, for example, the highest frequency of 24 hour POT was only 5 (in 1927 and 1957). Only 6 years over the period 1923-1982 had a fall in excess of 2 in (50.8 mm), namely 1927, 1950, 1953, 1960, 1972 and 1982 (see Figure 5.5.4.(i)).

As can be seen from the Ballindalloch record, the maximum number of 48 hour POT (> 1.5 in (38.1 mm)) was 4 in 1924, 1954 and 1966. Only 3 POT exceeded 3 in (76.2 mm) in 48 hours and these occurred in 1950, 1956 and 1970; however the 1956 and 1970 falls were associated with major floods. Thus, no year or runs of years were exceptional within the gauged record.

5.5.5 Spey study area: Seasonality of POT

The seasonality of 24 hour and 48 hour POT at Ballindalloch and Grantown is shown in Figures 5.5.5.(i) to (iii). All the large falls in excess of 2 in (50.8 mm) occurred between July and September, while the earlier part of the year, especially March and April, had comparatively few large POT. If SMD was high then much of this rainfall would be used in bringing SMD up to zero and then the remainder would produce flood

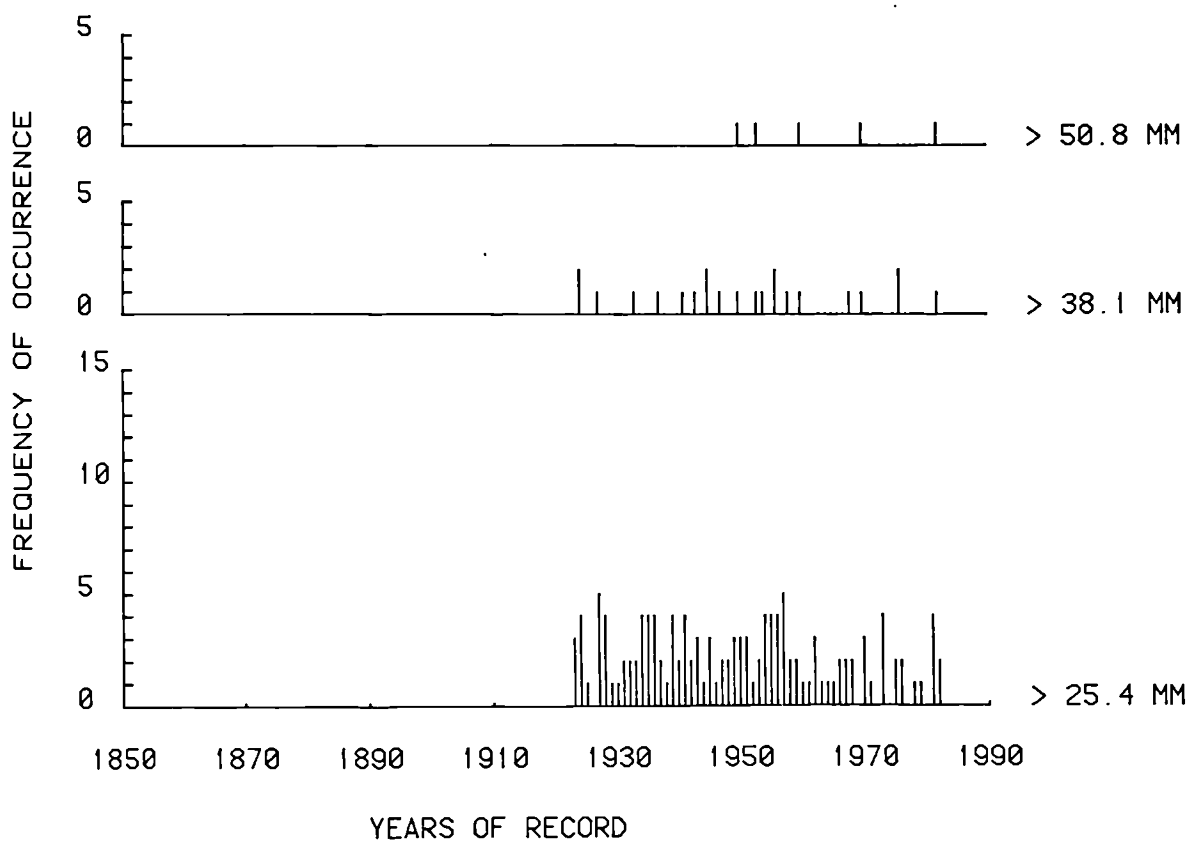


Figure 5.5.4.(i) BALLINDALLOCH: FREQUENCY OF 24 HR POT ABOVE DIFFERENT THRESHOLDS

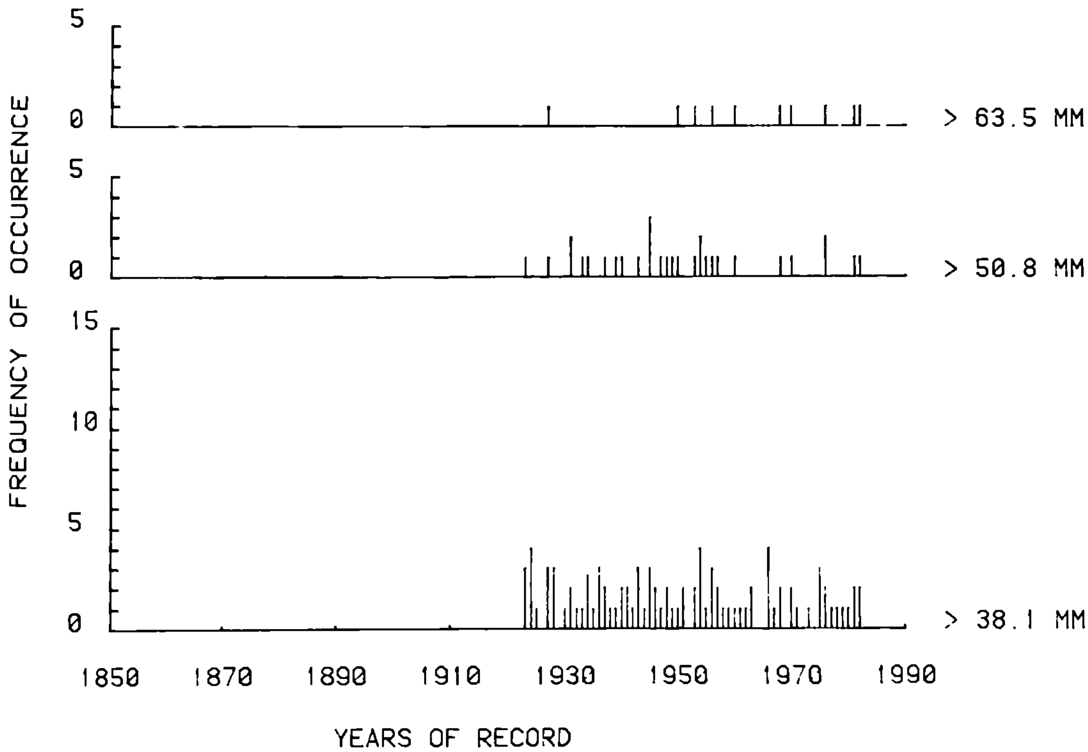


Figure 5.5.4.(ii) BALLINDALLOCH: FREQUENCY OF 48 HR POT ABOVE DIFFERENT THRESHOLDS

Figure 5.5.5.(i) BALLINDALLOCH: SEASONALITY OF 24 HR POT: 1923-1982

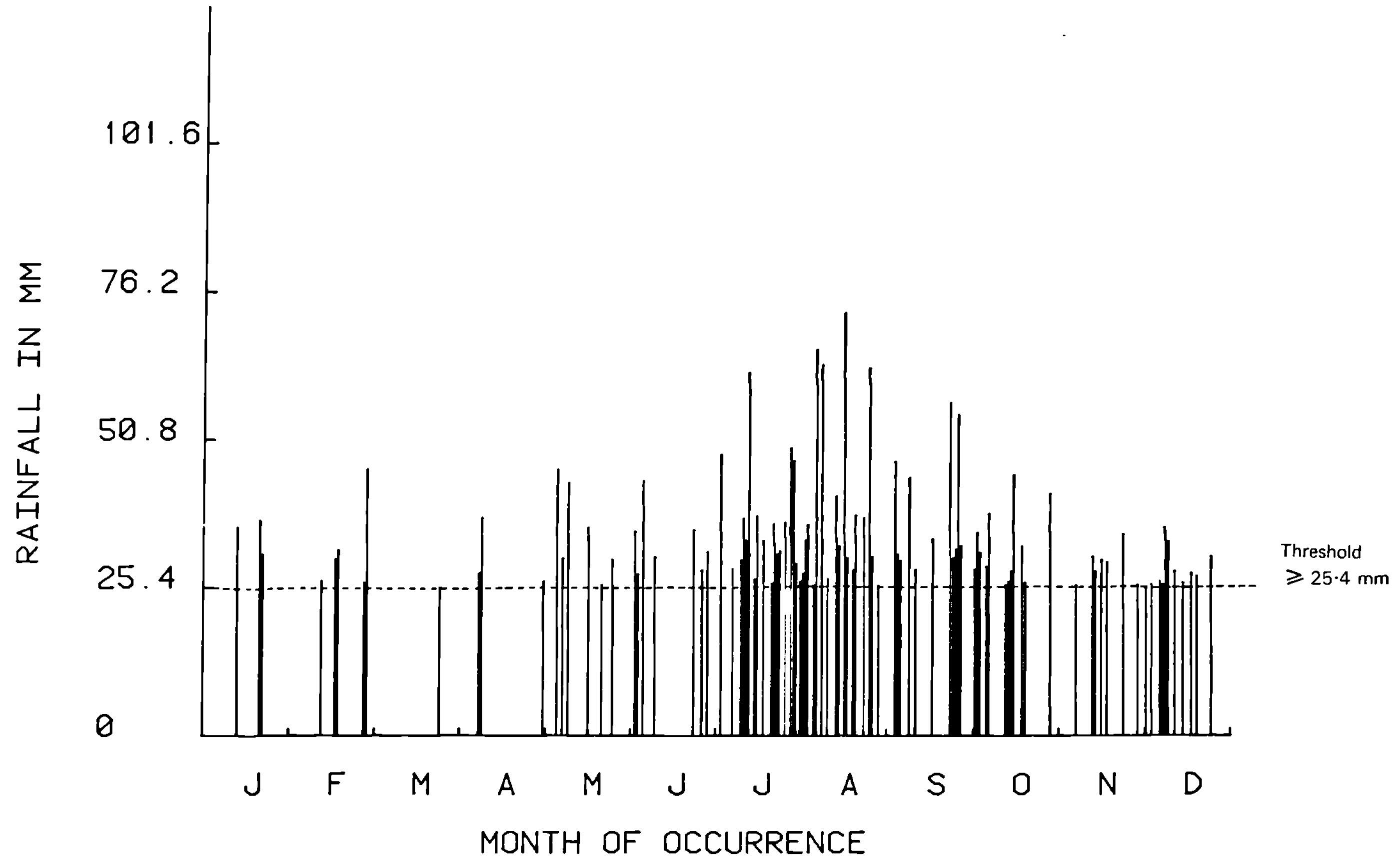
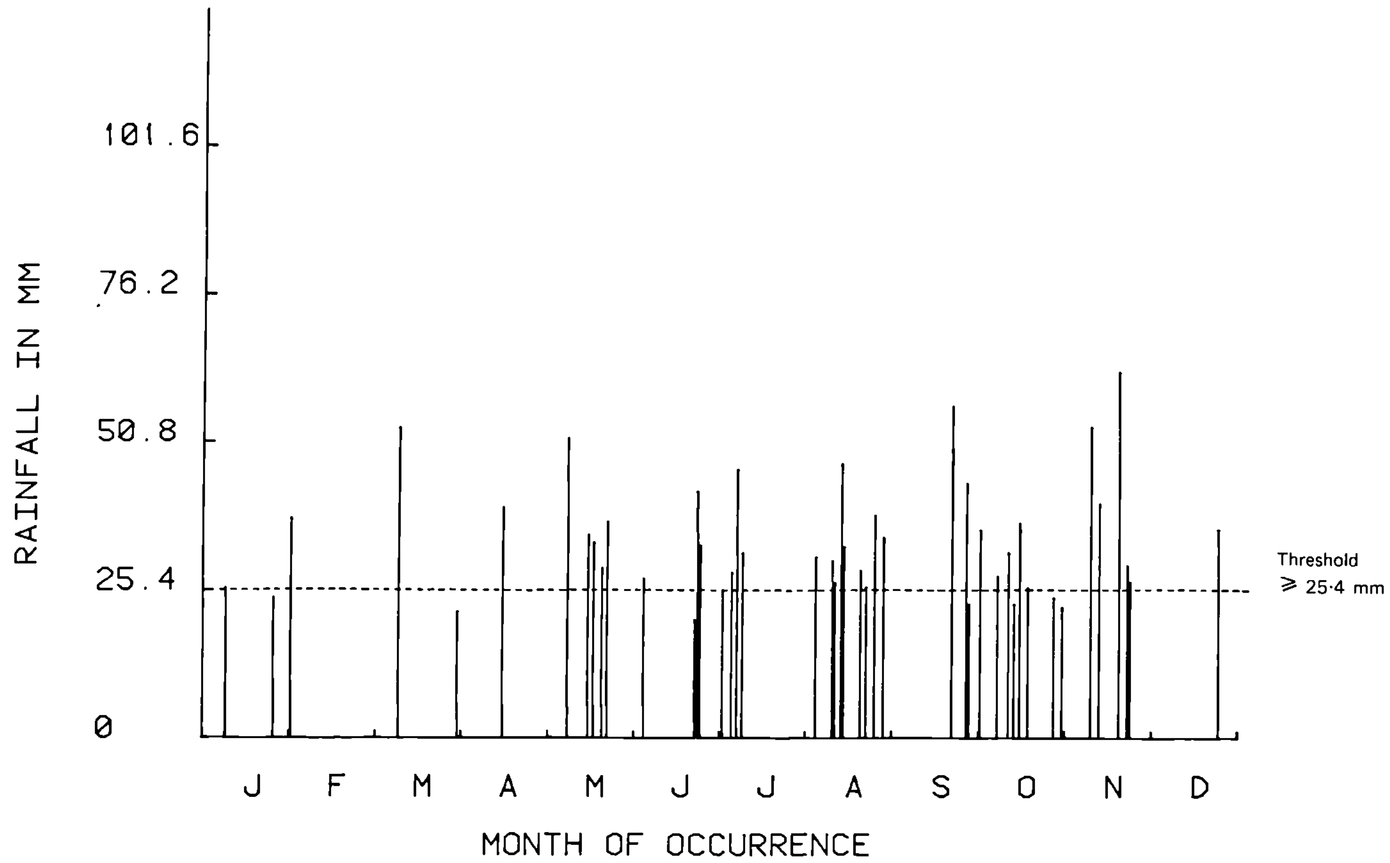


Figure 5.5.5.(ii) GRANTOWN (COMPOSITE): SEASONALITY OF 24 HR POT: 1922-1961



runoff. The same precipitation in November would produce more runoff and a larger flood event. This situation is strikingly at variance with the majority of upland Scotland but corresponds with Bleasdale's maps of percentage of average total rainfall occurring at different times of the year (Figure 5.2.(ii)). Much higher percentages fell on the east coast between May and September than did on the west coast. In terms of 48 hour POT, a similar pattern persisted (Figure 5.5.5.(iii)).

5.5.6 Spey study area: Analysis of annual maximum series

The most extreme 24 hour rainfall event on the Ballindalloch record occurred on 17th Aug, 1970 with 2.85 in (72.4 mm) ie. much less than the FSR (NERC, 1975) maximum probable precipitation of 250-300 mm. This is not to say that larger falls have not occurred within the Spey study area (see Table 5.5.6.(i) for examples). The 24 Sept, 1915 had the highest recorded 24 hour rainfall in British Rainfall or The Meteorological Magazine for the study region, with 3.33 in (84.6 mm) in 24 hours at Aviemore (Rothiemurchus). On the 9th Nov, 1915, 3.80 in (96.5 mm) was recorded at Loch Alvie Manse. It seems likely that the adage that 4 in (101.6 mm) in 24 hours can fall in any area of the British Isles (British Rainfall, 1910) applies as much to the Speyside region, especially when two large falls occurred within one year. The rainfall record at Huntley for the 1829 flood event (3.75 in (95.3 mm) in 24 hours; cited Lauder, 1830) suggested rainfall in excess of that value must have fallen in much of the Spey valley, especially at higher altitudes. This record was however in a non-standard gauge. The recurrence intervals of rainfall magnitude are shown in Table

Figure 5.5.5.(iii)

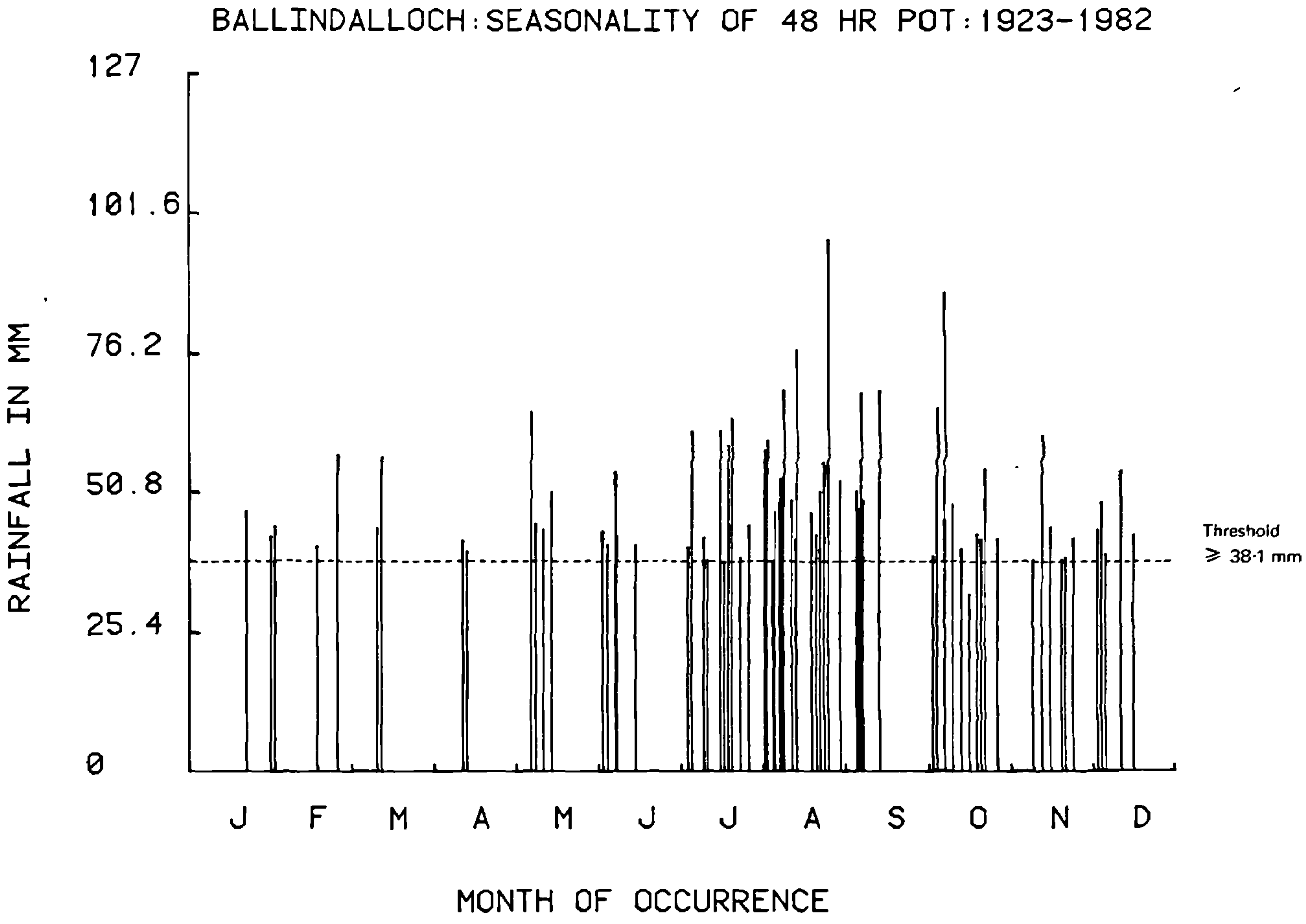


Figure 5.5.6.(i) AVON CATCHMENT AT BALLINDALLOCH 1923-1982
24 hour annual maxima

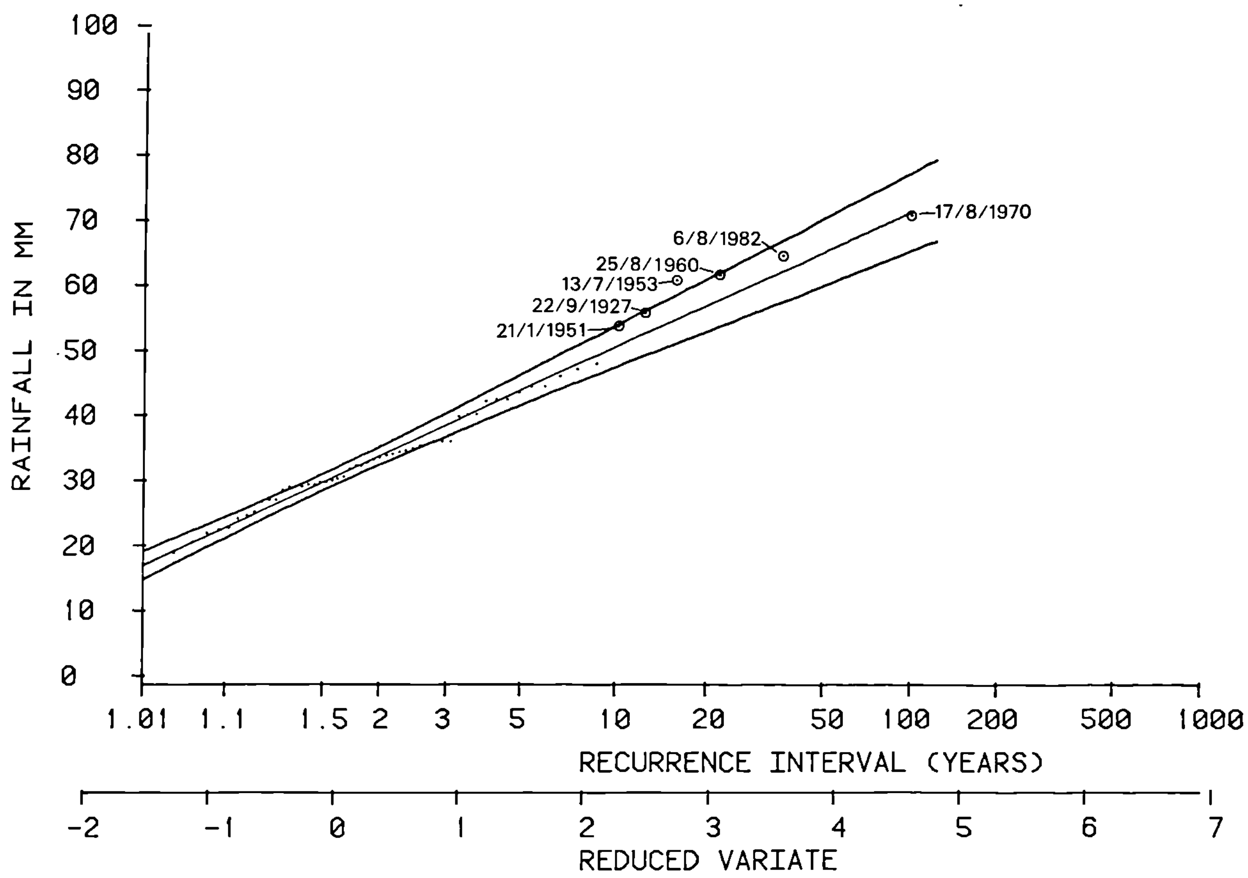


Figure 5.5.6.(ii) SPEY CATCHMENT AT GRANTOWN 1867-1939
24 hour annual maxima

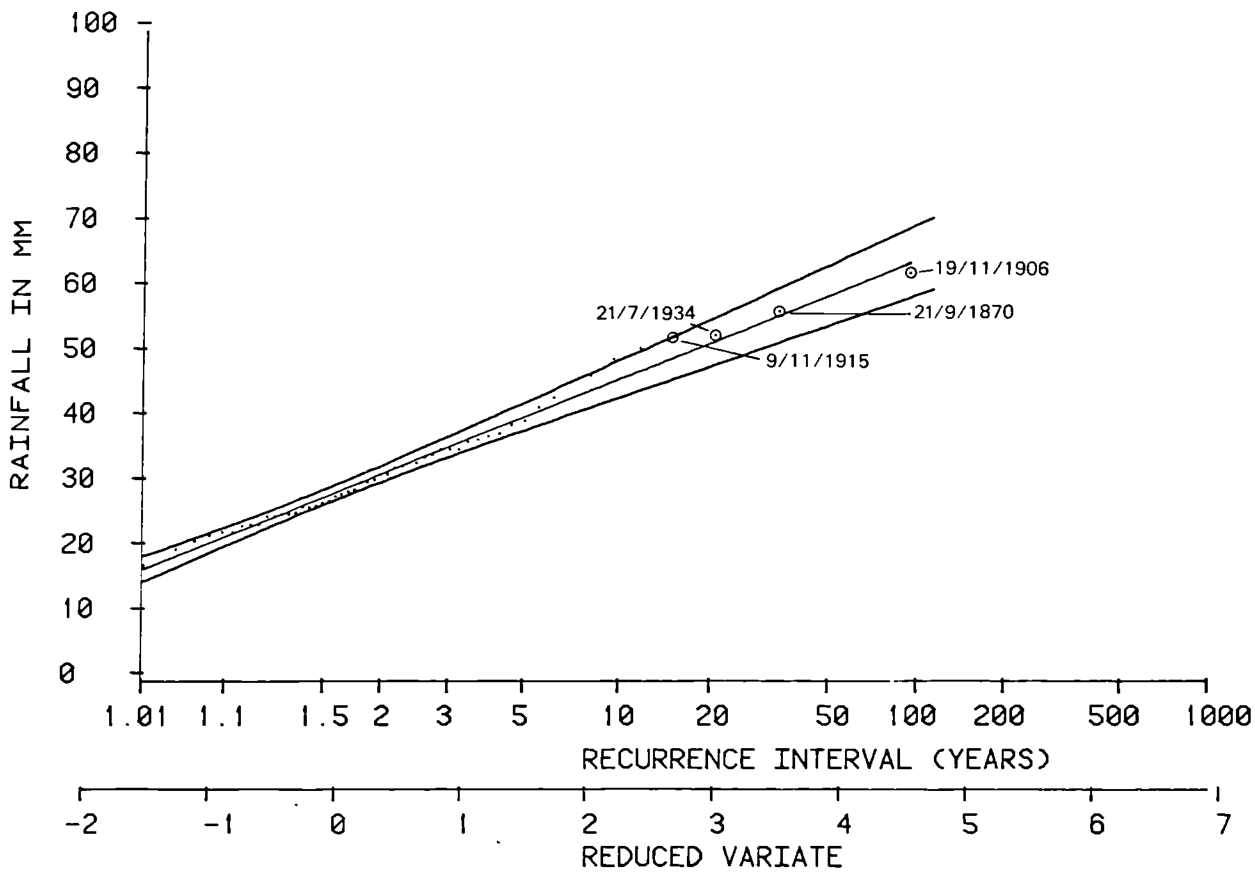


Figure 5.5.6.(iii) SPEY CATCHMENT AT GORDON CASTLE 1891-1974
24 hour annual maxima

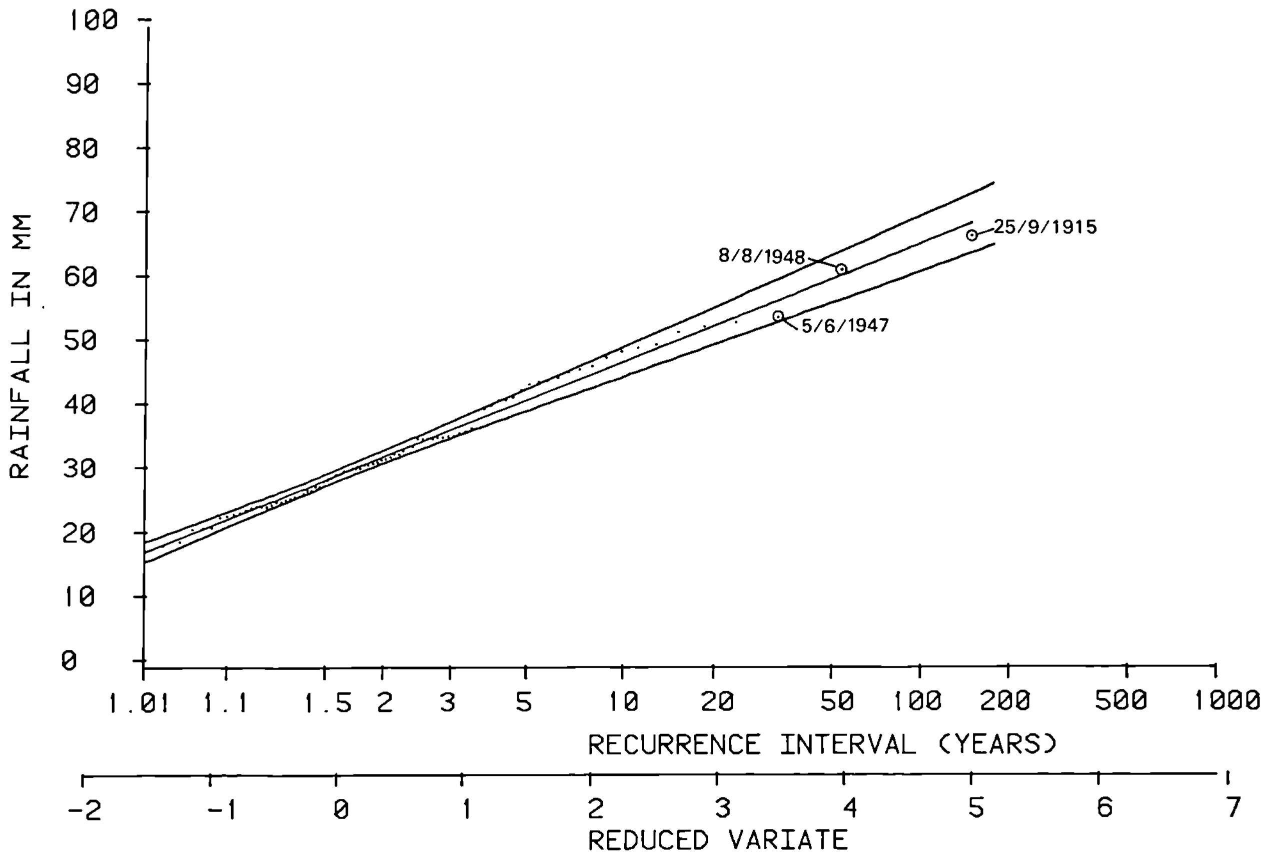


Table 5.5.6.(i)

Examples of large 24 hour rainfalls recorded within

within the Spey study area

Rainfall (mm)	Date	Station
85.1	17/7/1908	Craigellachie
84.6	24/9/1915	Rothiemurchus
77.0	25/8/1960	Glenmore Lodge
75.4	13/7/1953	Glen Livet (Clashnoir)
71.1	28/8/1960	Loch Seilich (Tromie Dam)

Table 5.5.6.(ii)

Magnitudes of rainfall annual maxima (AM) for different recurrence intervals, fitting the EV1 distribution (and GEV) for the Speyside Stations

Station	AM(2.33)	AM(10)	AM(25)	AM(50)	AM(100)
24 hour					
Grantown	33.7	46.8	54.2	59.7	65.2
				(59.3)	(64.4)
Ballindalloch	37.0	51.9	60.3	66.6	72.8
				(67.8)	(75.0)
Gordon Castle	34.6	47.7	55.1	60.6	66.1
				(61.7)	(67.8)
48 hour					
Ballindalloch	50.8	63.8	79.0	86.6	94.1
				(86.0)	(93.0)
72 hour					
Ballindalloch	56.0	77.6	89.9	99.0	108.0
				(92.9)	(97.7)

AM(i) annual maximum rainfall with a recurrence interval of i years

All values in mm

5.5.6.(ii). The rainfall of 17th Aug, 1970 on the Ballindalloch record had an estimated RI of over 100 years. It is interesting to note that none of the major flood events on the Spey were coincident with a high annual maxima on the Grantown record, though this may be due to the situation of the gauge.

When the Ballindalloch record (1923-1982) was subdivided into 22 year sub-sets with a five year overlap, there was again considerable fluctuation in rainfall amount for a set RI although the record was comparatively short. For example, studying rainfall of RI(100), there was a relative trough from 1930-1940 (1931-1952 had a RI(100) equal to 57.6 mm). In contrast between 1950 and 1972, the estimated RI(100) was 82.2 mm. The results are broadly comparable in pattern to the overlapping part of the Braemar record.

The *highest 48 hour rainfall* at *Ballindalloch* was again 16-17th Aug, 1970 with 96.8 mm and over 100 year RI. However, apart from this outlier, unlike Deeside where the ranking of 24 hour and 48 hour events was broadly similar, the Ballindalloch *record showed a shift in the* relative importance of specific events with increased duration. Comparing these rainfall values with the discharge record at Delnashaugh, 24, 48 and 72 hour events could all produce major flood events. The largest gauged flow on 12th Oct, 1981 was in fact associated with a 72 hour event (1st-3rd Oct, 1981), during which the rainfall would have easily exceeded local SMD values. While this rainfall had a low RI for 24 and 48 hour durations, its RI increased dramatically to over 50 years when considered as a 72 hour fall. This again emphasises the necessity to study rainfall recurrence interval

Figure 5.5.6.(v) AVON CATCHMENT AT BALLINDALLOCH 1923-1982
48 hour annual maxima

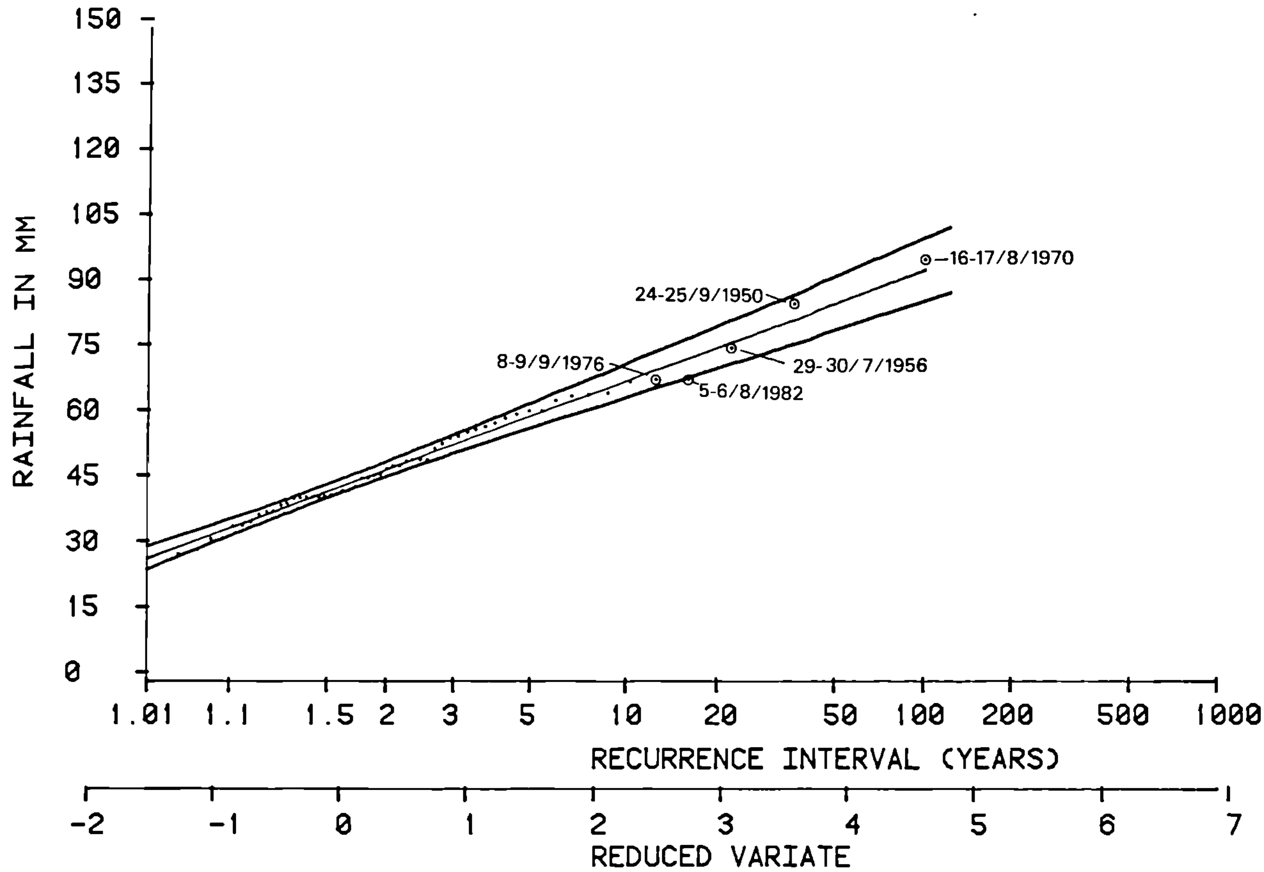
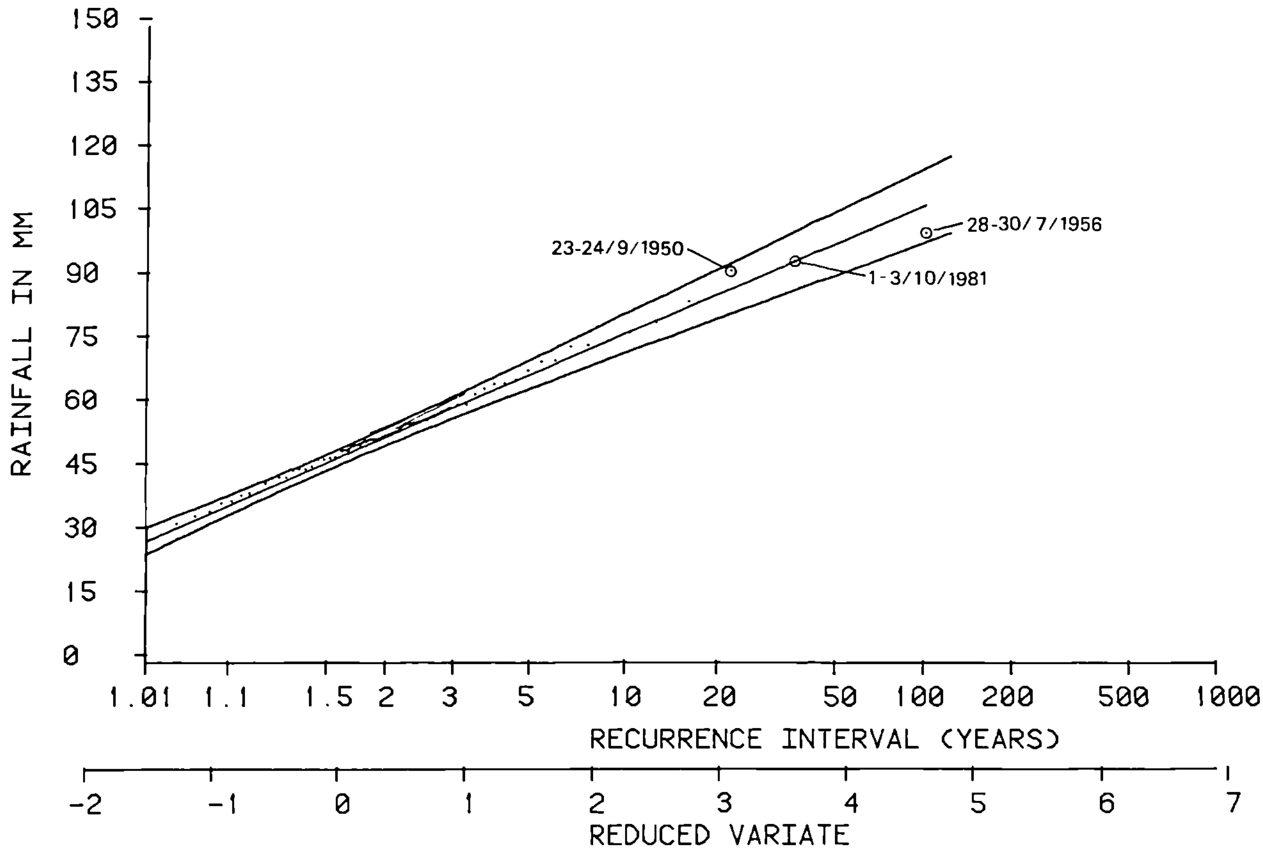


Figure 5.5.6.(vi) AVON CATCHMENT AT BALLINDALLOCH 1920-1982
72 hour annual maxima



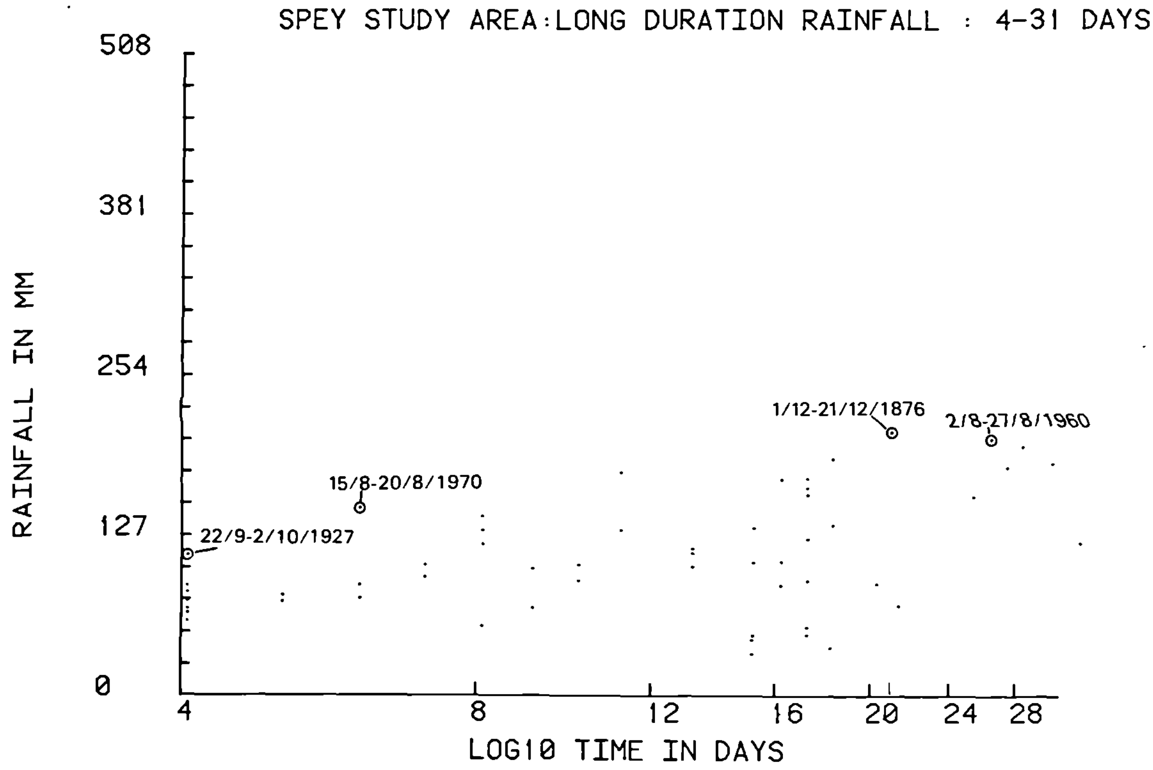
over a range of durations, especially when attempting to identify possible flood events.

5.5.7 Spey study area: Longer and shorter duration events

Available, long duration rainfall data from the study stations and extracted from The Meteorological Magazine were plotted against $\log_{10}$ time (see Figure 5.5.7.(i)). The major longer duration events occurred in 1927, 1956, 1960 and 1970. However, an envelope curve was not fitted to the data set because the limitations, due to the low altitude of gauges, were considered too great to be meaningful. The largest 4 day event was 22-25th Sept, 1927 with 111.5 mm, involving one 24 hour POT in excess of 2 in (50.8 mm) and 3 days later another in excess of 1 in (25.4 mm). This fall was only 92.5 mm lower than the 26 day rainfall of Aug, 1960. As well as these data, there is also documentary historic evidence of the geomorphic role of long duration events. For example, a major flood event reported in the Seafield estate papers took place on the Dalvey Burn on 28th July, 1779 and was associated with a long period of continuous rain (20-29th) (see Appendix 1.3 for report).

Historic short rainfall events were extracted from British Rainfall and plotted duration against $\log_{10}$ time, are shown in Figure 5.3.7.(ii). Probable maximum precipitation was similar to the Dee study area with 250-300 mm for 24 hour rainfall (130-150 mm for 2 hour rainfall), increasing with altitude. The most exceptional within the study area was 3.07 in (78.0 mm) in 65 mins at Loch Morlich within the Druie catchment on 11th June, 1956. This was associated with a major

Figure 5.5.7.(i)


SPEY STUDY AREA:LONG DURATION RAINFALL : 4-31 DAYS
508
381
254
127
0
RAINFALL IN MM
22/9-2/10/1927
15/8-20/8/1970
1/12-21/12/1876
2/8-27/8/1960
4
8
12
16
20
24
28
LOG10 TIME IN DAYS

flood event and channel change on the Allt Mor (McEwen, 1981; Appendix 1.6). A localised fall of 1.60 in (40.6 mm) of 30 mins duration, associated with a convective storm, was also recorded at Tromie Dam on 4th Aug, 1960.

There is a paucity of information about the frequency and geomorphic impact of intense rainfall events in this area perhaps because the catchments are considerably larger than on Deeside. It is difficult to assess whether this is because events are less frequent or because they have not been documented; the latter is much more probable. There are occasional historical and present day accounts of both regional frontal storms with cells of very high intensity rainfall and also much more localised convective storm events.

It is clear that during the 4th Aug, 1829 event, there were cells of very intense rainfall within the upper Avon catchment:

> "We may form some idea of the tremendous nature of the fall of rain, high up among the mountains, from what was observed even at Tomintoul, where the farrier told us that, from two slates having been blown out within two rows of the ridge of his roof, which created a hole of something less than 18 inches square [116 cm^2], 80 scotch pints, or about 40 gallons [182 litres] of rain water entered during the 12 hours between the morning and evening of the 3d of August. (Lauder, 1830 p188)

At the other extreme, British Rainfall (1923) reports that the Rev. R.P. Dansey on visiting the Cairngorms on 8th July, 1923 noticed certain features on "a side stream in Glen Einich."

> "The main stream above the junction was not affected and the bed was old and undisturbed (though it had had the storm), but I found a huge hole some way up a side stream 10 feet [3.1 m] deep and 20 feet [6.1 m] wide, evidently caused by a waterspout, as 50 yards [46 m] above this point, the stream was so insignificant that the heath met over it. Yet it had sent down hundreds of tons of sand, rock etc. into the mainstream half a mile distant. One can hardly speculate as to the mechanism by which such a crater could be dug out." (British Rainfall, 1923)

This was the same day as there was tremendous flooding at Carrbridge on the Dulnain, caused by 1.64 in (41.7 mm) rainfall mostly in half an hour.

Other references confirm the importance of these localised events within the upper catchments:

> "More than one old witness in the year 1770 claimed that the Alltnagillevichael [an upper tributary of the Avon] had formerly run into the Burn of Rorack and thence to the Avon. Janet Farquharson, who had drawn water from the burn for twenty years, "minded to see the burn run into the water of

Avin [Avon]" but since...the year 1714, "by a spait of rain and thunder, it was turned into the Don." An examination of the terrain shows this to have been perfectly feasible and there is an old course with deeply undercut banks to support the contention." (Gaffney, 1960 p9)

More recently, McEwen (1981) reports an intense convectional rainstorm over the Allt Mor catchment on 4th Aug, 1978, when 36 mm rain fell in an hour. This was associated with a high antecedent rainfall, thereby causing a major flash flood within this small catchment.

5.5.8 Spey study area: Storm case studies

The following major storm events were reconstructed for Speyside:

(1) 25-26th Sept, 1915

(2) 28-30th July, 1956

(3) Floods of July/ Aug, 1970

(1) 25-26th Sept, 1915

This fall was cited both within the FSR's (NERC, 1975) list of major historic rainfalls and Glasspoole's (1929-30) study of long duration rainfall events. A rough estimate for the storm's duration was 40 hours, the storm being caused by a shallow depression, which had developed over the south of England and which increased in intensity as it moved in an almost northerly track. On the morning of the 26th, it

Table 5.5.8.(i)

Rainfall recurrence intervals for the 9/1915 flood event *

Station	24 hour (25/9/1915)	48 hour (25-26/9/1915)	72 hour (25-27/9/1915)
Grantown	37.1 mm (4 yr)	-----	-----
Gordon Castle	67.3 mm (130 yr)	-----	-----

Table 5.5.8.(ii)

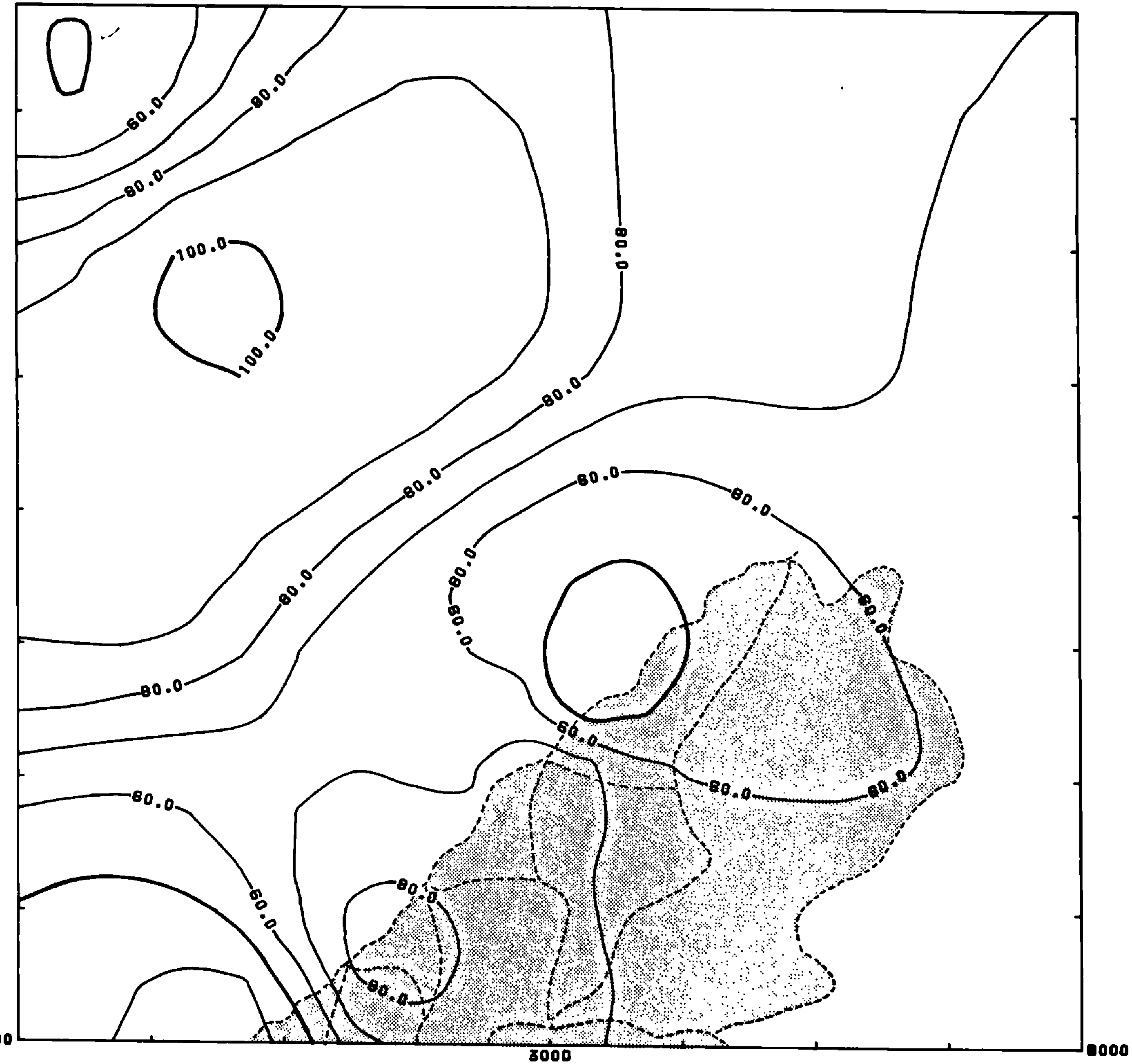
Rainfall recurrence intervals for the 7/1956 flood event *

Station	24 hour (29/7/1956)	48 hour (28-29/7/1956)	72 hour (28-30/7/1956)
Grantown (Moniack)	45.0 mm (-----)	67.3 mm (-----)	85.9 mm (-----)
Ballindalloch	47.2 mm (8 yr)	76.7 mm (23 yr)	101.9 mm (65 yr)
Gordon Castle	40.1 mm (6 yr)	77.2 mm (-----)	102.9 mm (-----)

----- data unavailable

* as an annual maximum

Figure 5.5.8.(i) Rainfall 25-26th September, 1915 Spey study area



RAINFALL 25-26 SEPT. 1915 SPEY STUDY AREA
PLOT NO. 1 DATE 03/09/85 TIME 18.14.50

was centred off the N.E. of Scotland and then moved abruptly eastward. The reconstructed storm profile is shown in Figure 5.5.8.(i). Watt (1917) showed that on the Nairn, Findhorn, Lossie and Spey in Sept, 1915, the high water marks of 1829 were closely approached at various locations but the area of destructive flooding was less extensive. From the profile of Figure 5.5.8.(i), several cells of high rainfall seemed to occur and one was centred within the study area over the Feshie and Druie catchments. Unfortunately, estimates of RI could only be made from the Grantown record, which was outside the zone of most intense rainfall. It was however the highest 24 hour rainfall on the downstream Gordon Castle record (67.3 mm), with a RI of 130 years.

(2) 28-30th July, 1956

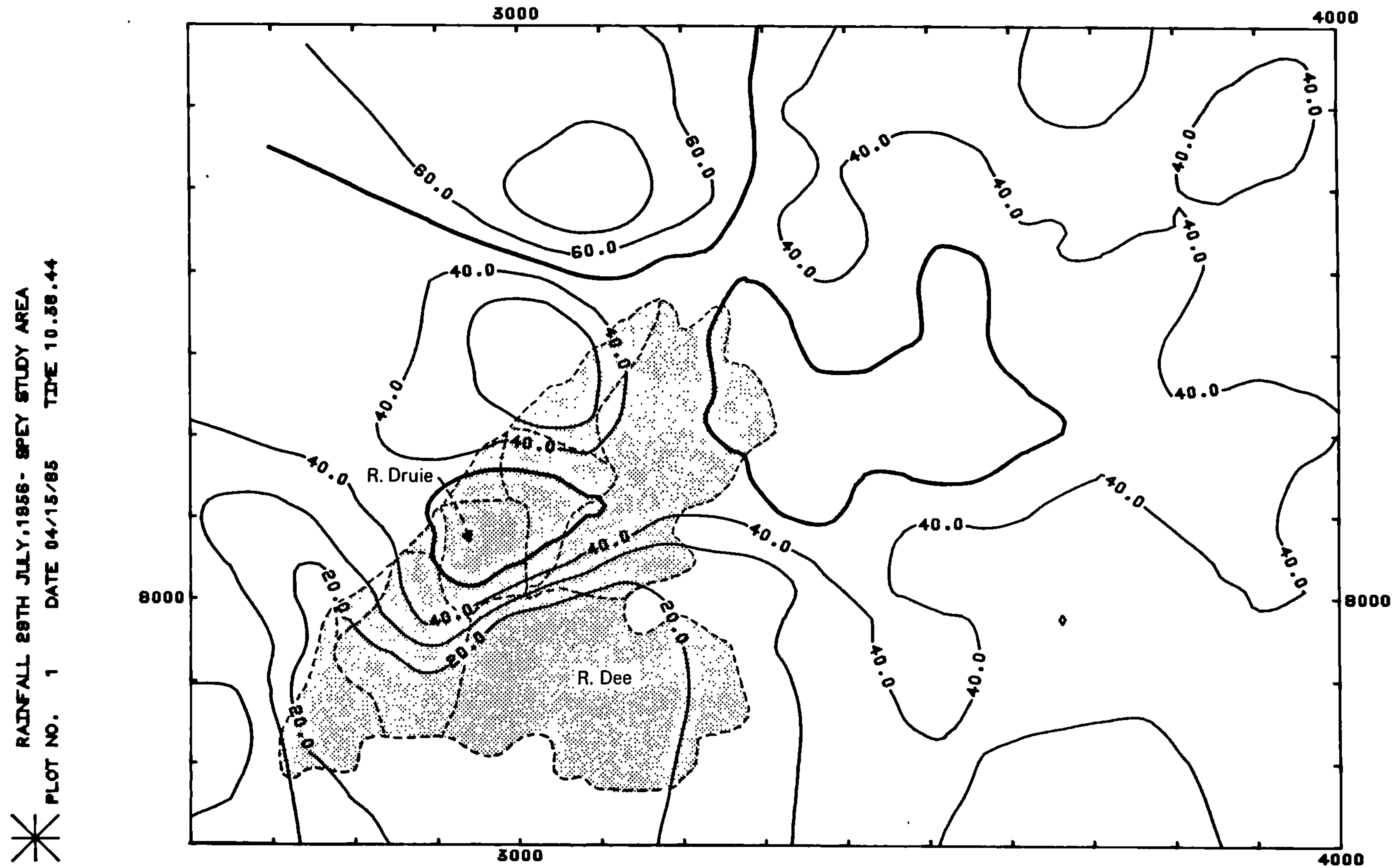
The synoptic situation generating the widespread flooding within the middle Spey catchment was described as follows:

> "The prime cause of the flooding was heavy rain associated with the occluded front of a small depression which moved N.N.E. over Britain between the 28th and 30th July."
> (Green, 1958 p48)

Falls of over 3 in (76.2 mm) occurred over the whole of the Monadhliath Mountains, though the 24 hour rainfall of 47.2 mm at Ballindalloch only had a RI of 8 years. If however the duration was extended to 72 hours, 4.01 in (101.9 mm) fell at Ballindalloch with a RI of 65 years (see Table 5.5.8.(ii)).

Figure 5.5.8.(ii)

Rainfall 29th July, 1956 Spey study area



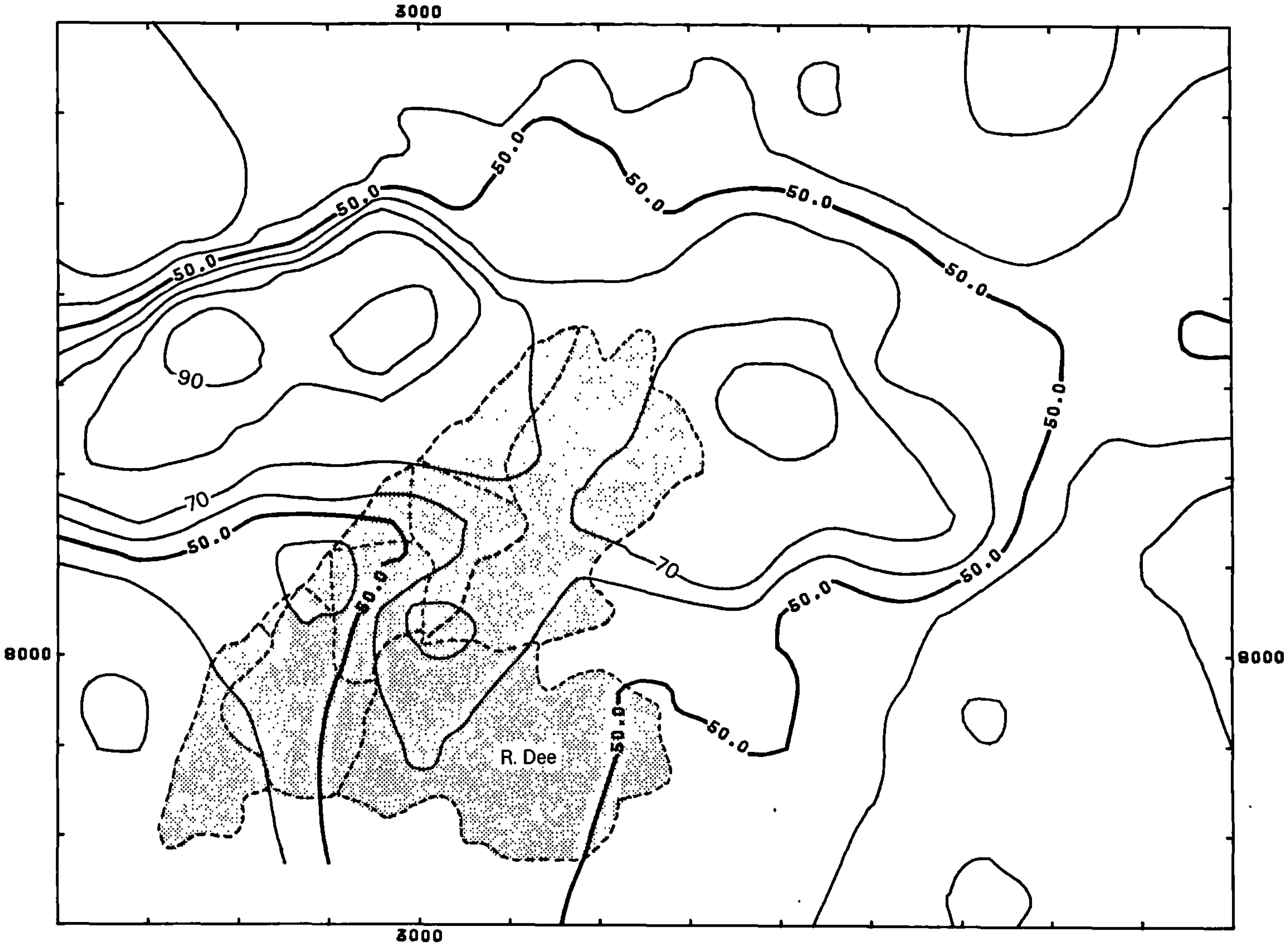
<u>(3) Storms of July/ Aug. 1970</u>

Two large storm events occurred in the summer of 1970, illustrating the importance of antecedent moisture conditions. In terms of synoptic origin, the 6-7th June, 1970 event was caused by intense convectional thunderstorms in a slow moving storm. This was due to the forced ascent of unstable air when a cold pool of upper air passed over the Central Highlands (Meteorological Office Report, 197~~8a~~5). When the 24 hour rainfall distribution was considered, the highest falls in the study region occurred over the upper Druie catchment, with 70.9mm in 24 hours on Cairngorm summit. Other areas affected included the Feshie drainage basin.

Just over a month later on 16-17th August, there was another major rainfall event, associated with a slow-moving frontal storm. In terms of 24 hour rainfall, the study region was outwith the principal cell of high intensity rainfall centred over the Findhorn catchment, but falls in excess of 70 mm were again recorded on Cairngorm summit. Catchments with heaviest falls included the Druie, upper Nethy and the Avon. The 24 hour fall of 72.4 mm at Ballindalloch had a RI of 100 years. The distribution of high 48 hour falls affected a similar area but the relative intensity was higher in the study area with 118.8 mm falling on Cairngorm cf. 120.4 mm falling at Moy in the Findhorn valley (Figure 5.5.9.(iv)). The 48 hour fall of 96.8 mm at Ballindalloch increased the storm's RI to 130 years (see Table 5.5.8.(iii)). Finally, when the storm was considered as a 6 day event (Figure 5.5.9.(v)), its extreme nature can be seen.

Figure 5.5.8.(iii)

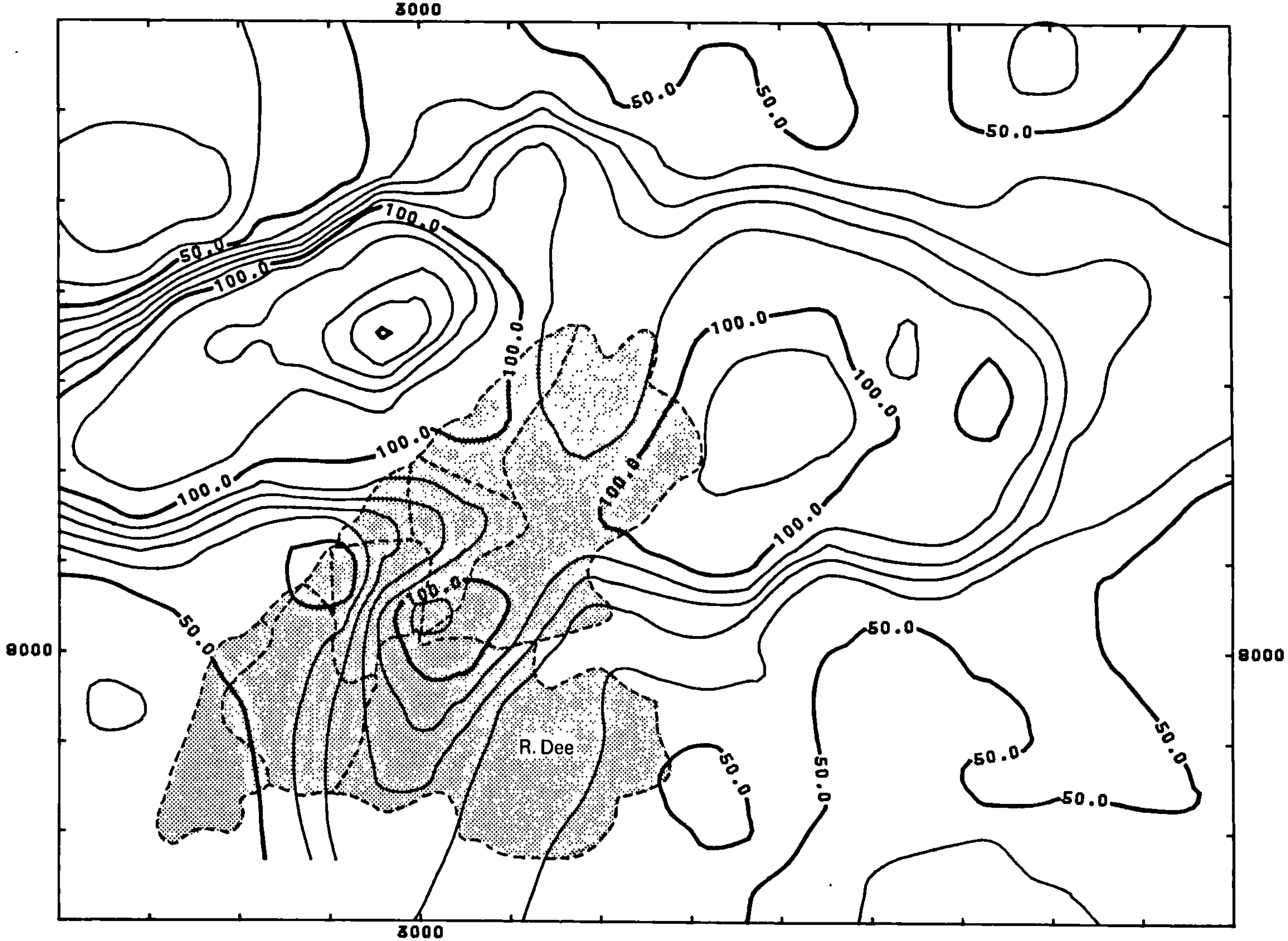
Rainfall 17th August, 1970 Spey study area



RAINFALL 17TH AUGUST, 1970 SPEY STUDY AREA
PLOT NO. 1 DATE 03/10/85 TIME 21.16.28

Figure 5.5.8.(iv)

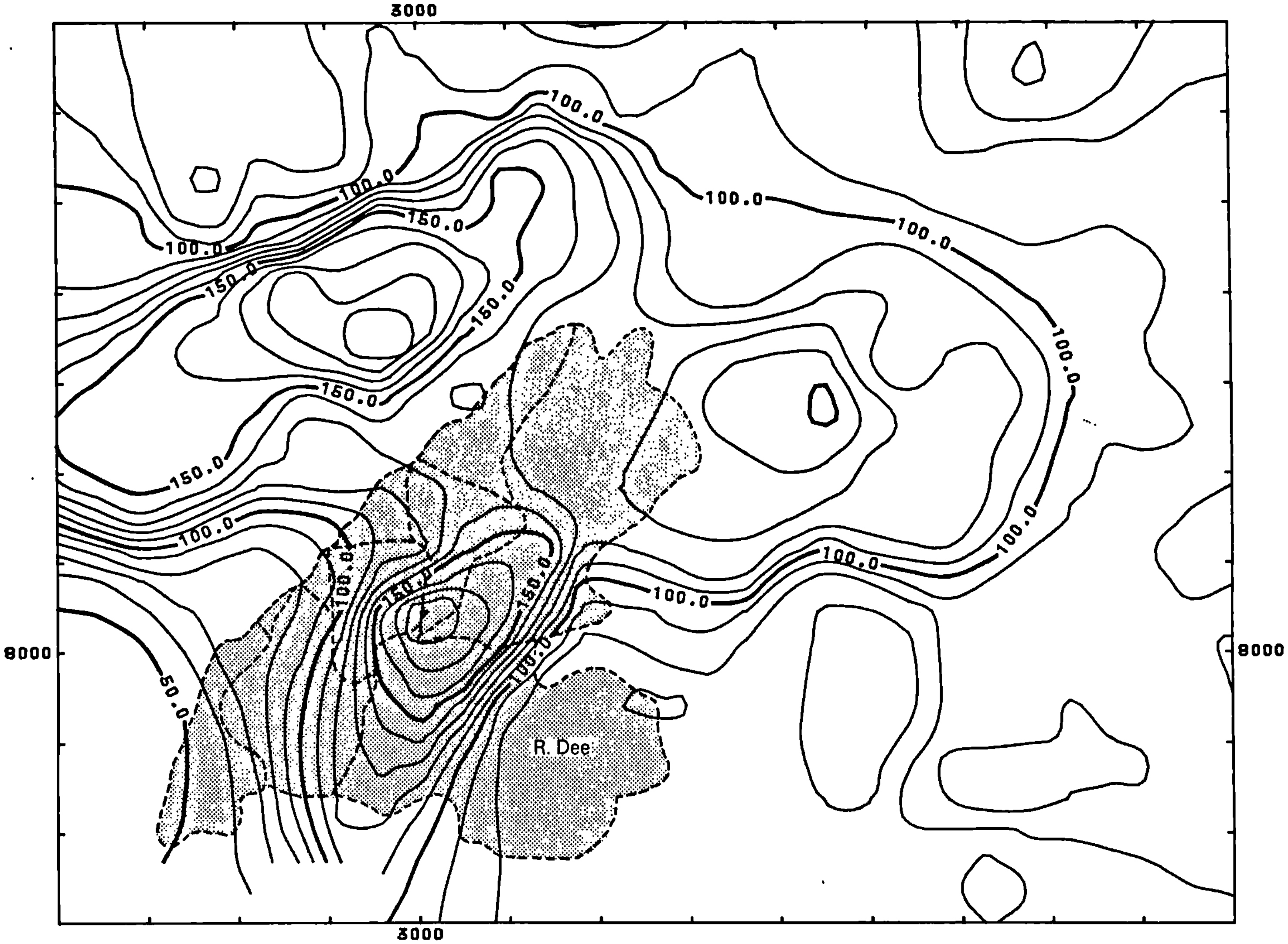
Rainfall 16-17th August, 1970 Spey study area



RAINFALL 16-17TH AUGUST, 1970 SPEY STUDY AREA
PLOT NO. 1 DATE 03/10/85 TIME 21.23.08

Figure 5.5.8.(v)

Rainfall 16-21st August, 1970 Spey study area



RAINFALL 16TH - 21ST AUGUST, 1970 SPEY STUDY AREA
PLOT NO. 1 DATE 03/10/85 TIME 21.06.12

In summarising these storm profiles, it is notable that both 1956 and 1970 had more than one major event of high RI affecting overlapping areas and thus the temporal incidence of such events can not be entirely random. As Green (1971) points out, in both the July, 1956 and Aug, 1970 events, heavy continuous rain was caused by back-bent occluded fronts associated with depressions, moving in a N.N.E. direction. It should be observed however that on both occasions, the most intense rainfall fell to the north within the Findhorn valley, rather than within the study area. The areas especially prone to cells of heavy rain are the upper catchments, due to orographic effects. Discharge records above the confluence of the tributaries thus may give gross under-estimates of specific runoff rates and stream powers within the upper reaches during such events.

5.5.9 Summary

The pre-1900 rainfall/runoff events were exceptional in both magnitude and frequency. Following the catastrophic runoff rates associated with the Aug, 1829 flood, there were 4 subsequent winter cyclonic floods with at least 50 year RI. Although these storms were much more localised in regional extent, at individual sites the resulting floods were comparable in stage with the 1829 event, despite the fact that they occurred on average once every 20 years. Coupled with this, there was also an increase in more moderate flood flows during the 1870s and 1880s.

In comparison with the Dee and Tweed records, the Spey study area had rather a quiescent discharge record within the 20th century, with no floods in excess of a 50 year RI. Although in 1956 and 1970 there were major summer frontal storms, there has been nothing to compare with the pre-1900 flood record on the mainstream Spey. The catchment conditions required to generate the high runoff rates have not been repeated. Within individual catchments however major flood events (RI 20-50 years) have occurred post 1900 but none comparable with the magnitude of the Aug, 1829 event. Localised convective storms only generated high runoff rates when centred over the steep-sloped, upper catchments eg. the Aug, 1978 flood on the Allt Mor.

5.6 Tweed study area: Magnitude and frequency of extreme runoff

The reconstruction of the magnitude and frequency of extreme rainfall and runoff within the Tweed study area will now be presented.

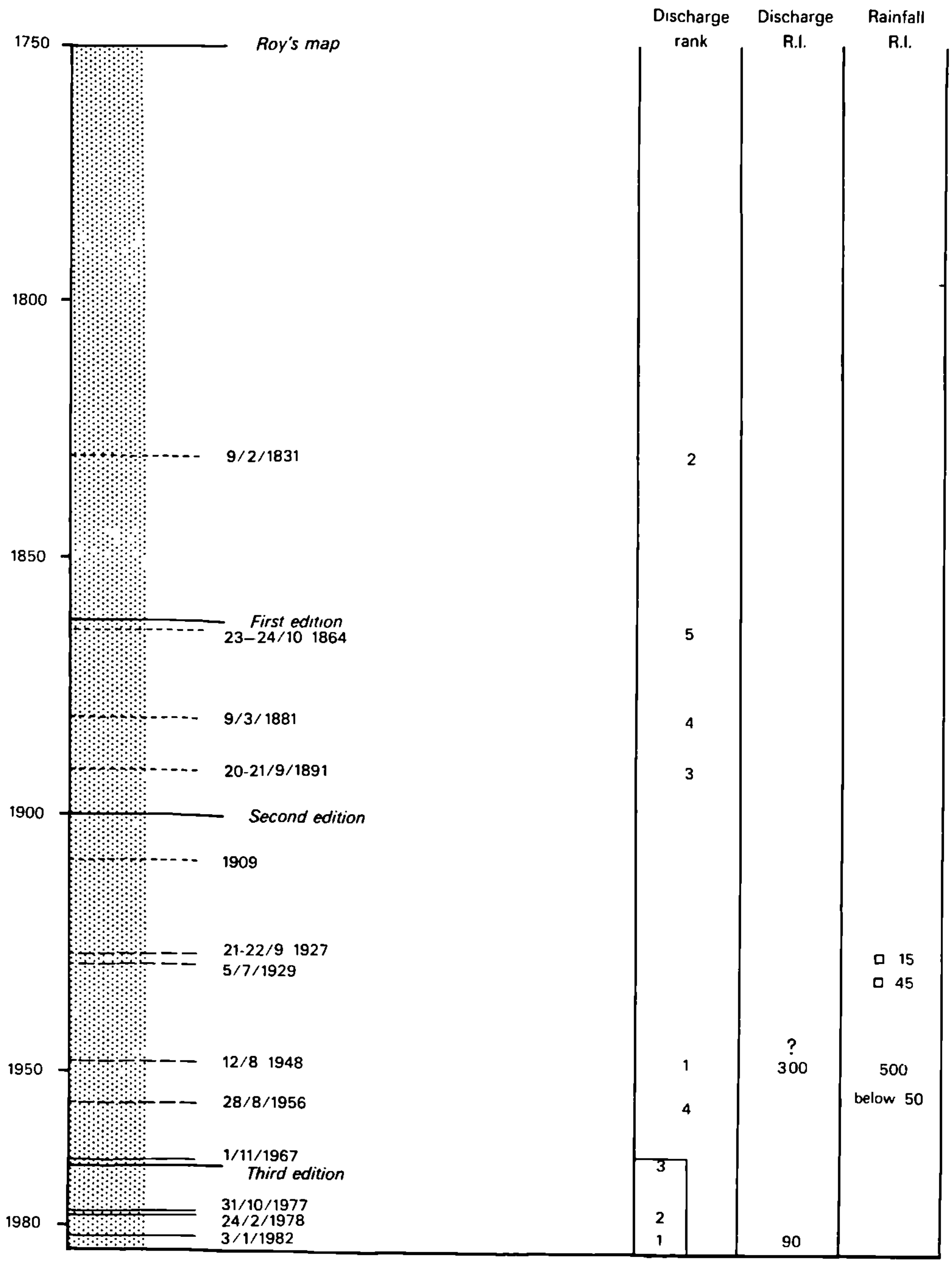
5.6.1 Documented flood history within the Tweed study area

The collated flood histories for the Tweed and the tributary catchments are shown in Appendices 1.7 to 1.12, for both before and after 1800. The largest event pre-1800 appeared to be on 1st Aug, 1294 (subsequent to Brookes and Glasspoole's (1928) rainfall maximum around 1200-1250, related to North sea storms). Two of the more recent extreme flood events have been documented in both the academic and applied literature. The Aug, 1948 flood was studied by Glasspoole and Douglas (1949), Learmonth (1950) and Baxter (1949) while the 1956 flood was documented by Common (1958). When this record was studied in terms of the frequency of historic flooding, the lower Tweed, Whiteadder and Teviot catchments all seemed to have a higher incidence of flood events 1850-1899 (Figures 5.6.1.(i) to 5.6.1.(v)). On the Leader, between 1857 and 1900, flood gates were constructed at Birkwood Haugh and above Earlston, implying that over this period one or more major destructive flood must have occurred.

Attempts to rank the flood events are shown in Figures 5.6.1.(i) to (v). Thus, several events are given relative to their stage at the "Tweedometer" at Kelso (see Figure 5.6.1.(vi)), the "Tweedometer" being a stage board near Kelso Bridge. At this location, the 12-13th Aug,

Figure 5.6.1.(i)

The flood history within the Leader catchment



Discharge gauge: Earlston   Rainfall gauge: Cowdenknowes

Figure 5.6.1.(ii)

The flood history on the Whiteadder

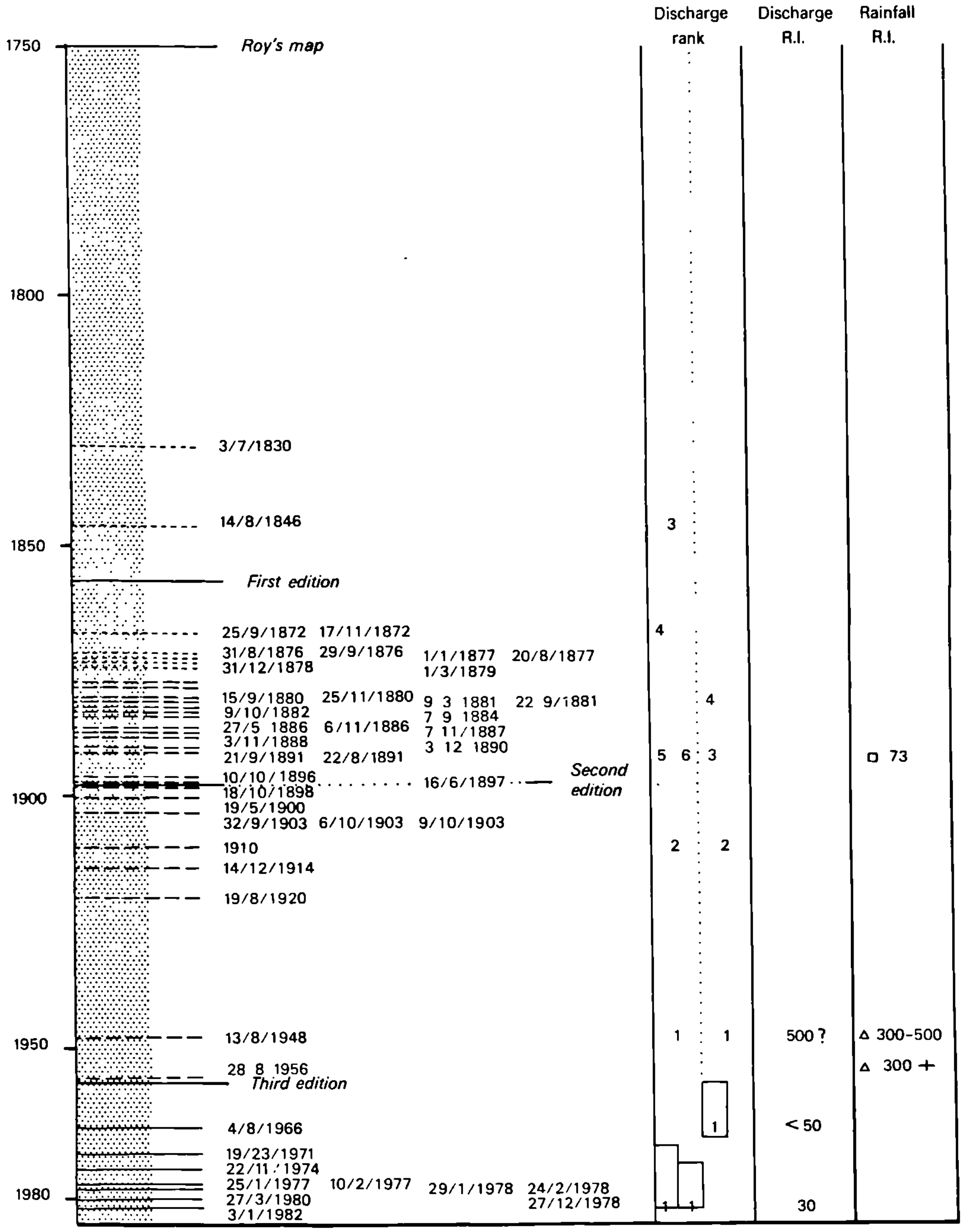
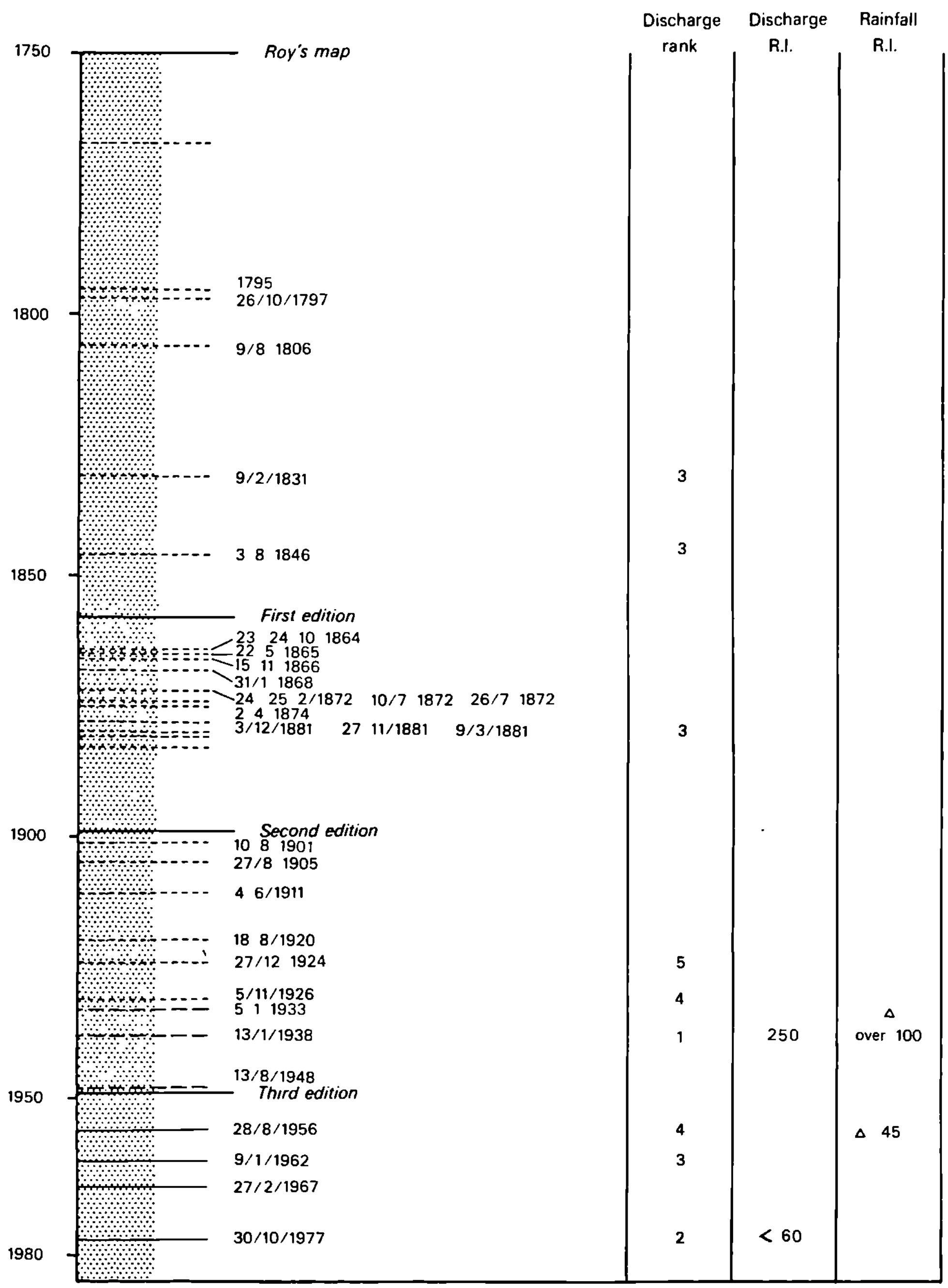


Figure 5.6.1.(iii)

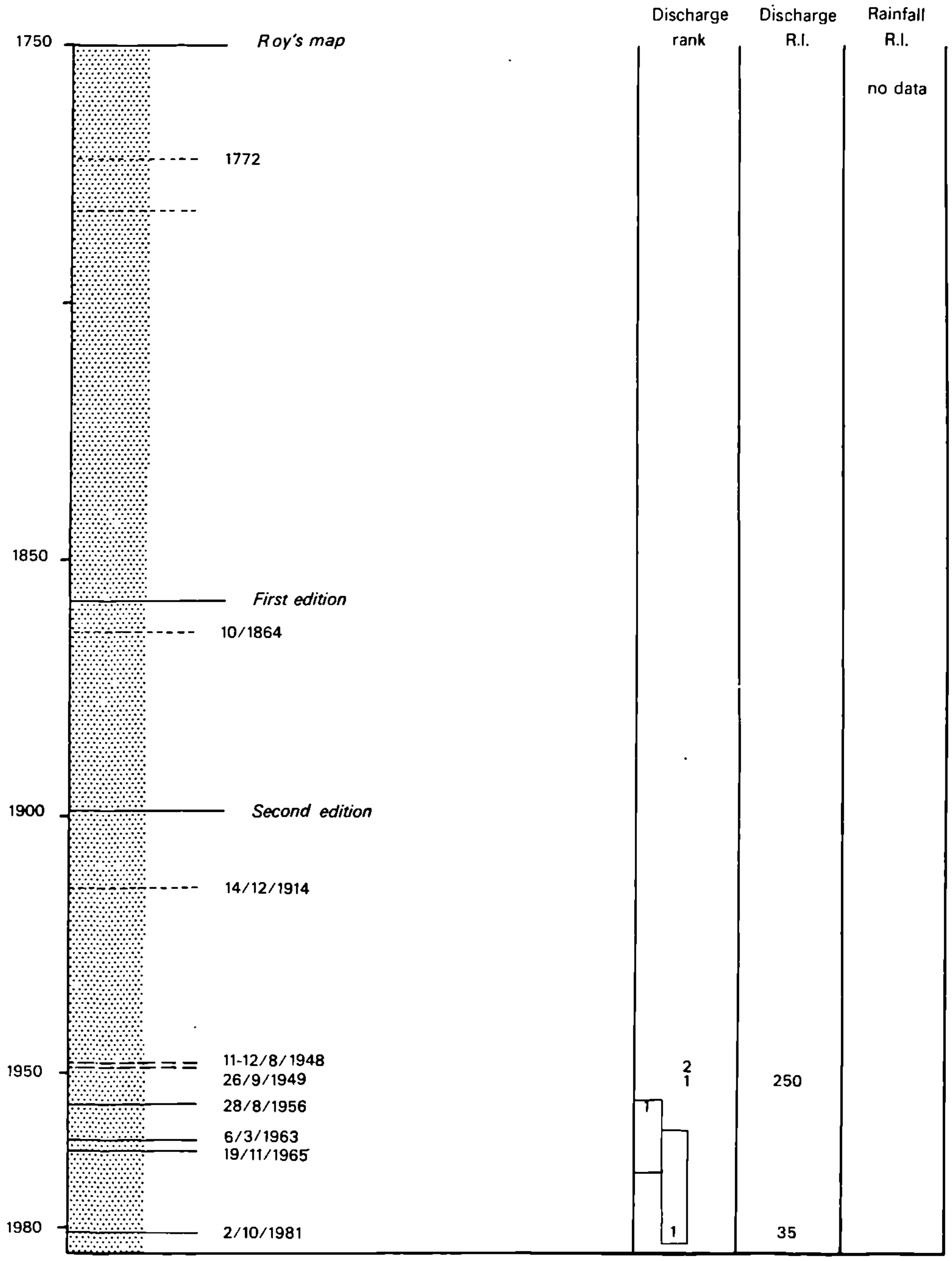
The flood history of the Teviot catchment



Discharge gauge: Hawick *Rainfall gauge: Hawick*

Figure 5.6.1.(iv)

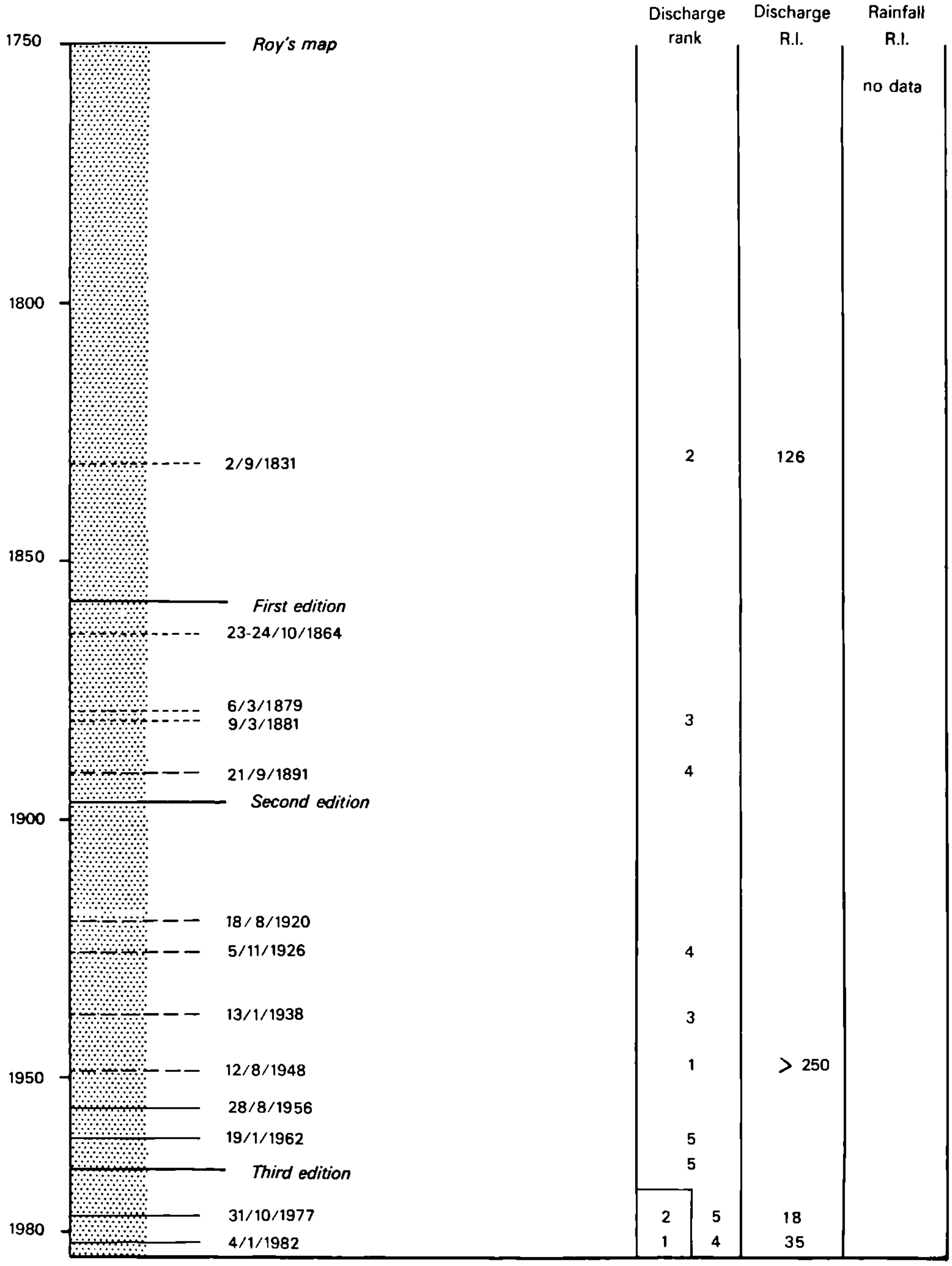
The flood history within the Bowmont Water catchment



Discharge gauge: Kirknewton and Glen

Figure 5.6.1.(v)

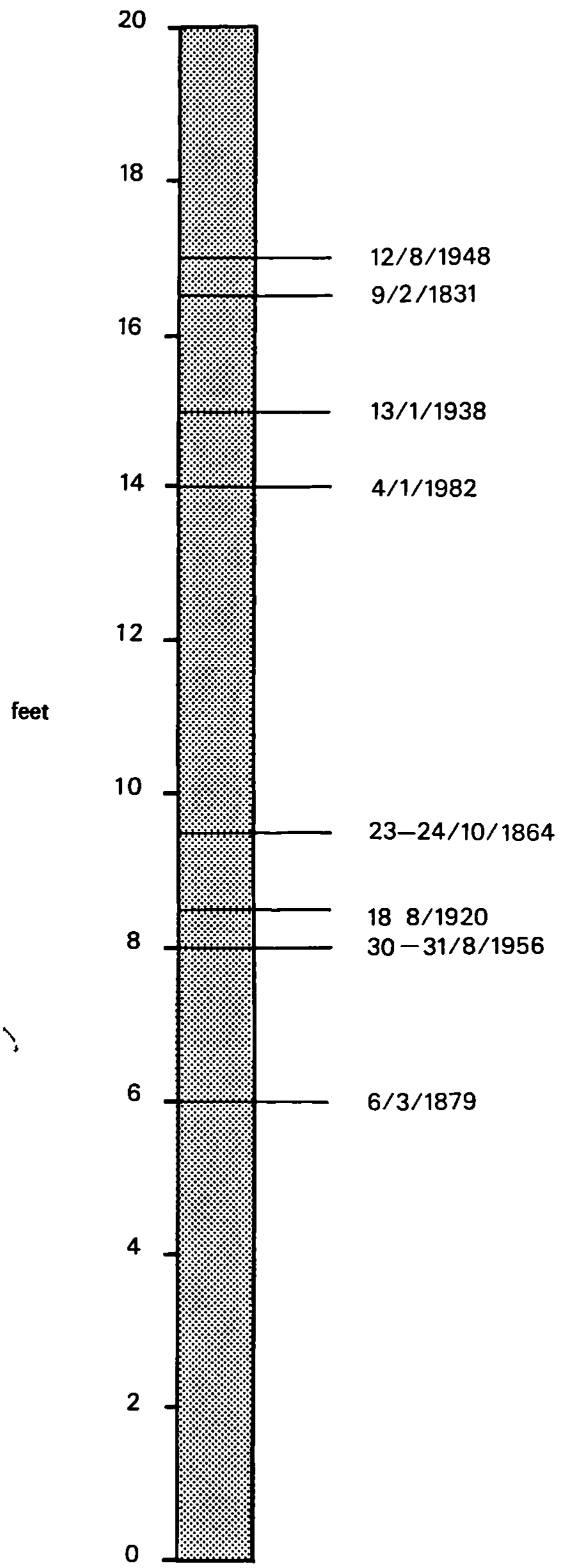
The flood history within the Tweed catchment at Kelso



Discharge gauge: Sprouston

Figure 5.6.1.(vi)

Flood stages at the Tweedometer at Kelso



1948 event was approximately 17 feet (5.2 m) above the normal summer level cf. 9th Feb, 1831 event which was 16.5 feet (5.0 m) above normal summer level. The flood stones marking maximum stage at Floors Castle are however at a similar height (2 pm 9/2/1831 and 13/8/1948). The hydrometeorological origins of these two events were however very different (see Table 5.6.1.(i)), with the 1831 event being a major snowmelt while 1948 was a summer frontal storm. Both were extreme events within the separate population of floods determined by their different genesis. This suggests the recurrence interval of these extreme events is much longer than twice in 150 years.

When the seasonality of historic flooding (Figures 5.6.1.(vii) to 5.6.1.(x)) and available hydrometeorologic information (Tables 5.6.1.(i) to (ii)) were assessed, the following results were found. The Tweed and Teviot especially had several snowmelt or snow storm associated events between 1850 and 1900. For example, it is documented that the channel change on the Teviot by Hassendean was caused by an ice flood in 1795 (Trans. Haw. Arch. Soc., 1865). The Little Ice Age cold years of 1880-1881 led to widespread flooding, with a substantial snowmelt contribution in March and April (see Appendices 1.8 and 1.9). In general, the mainstream Tweed flood frequency was affected by longer duration discharge events while the other study tributaries were more susceptible to more localised storms, although longer duration events were still important.

Figure 5.6.1.(vii)

The seasonality of flooding on the River Leader

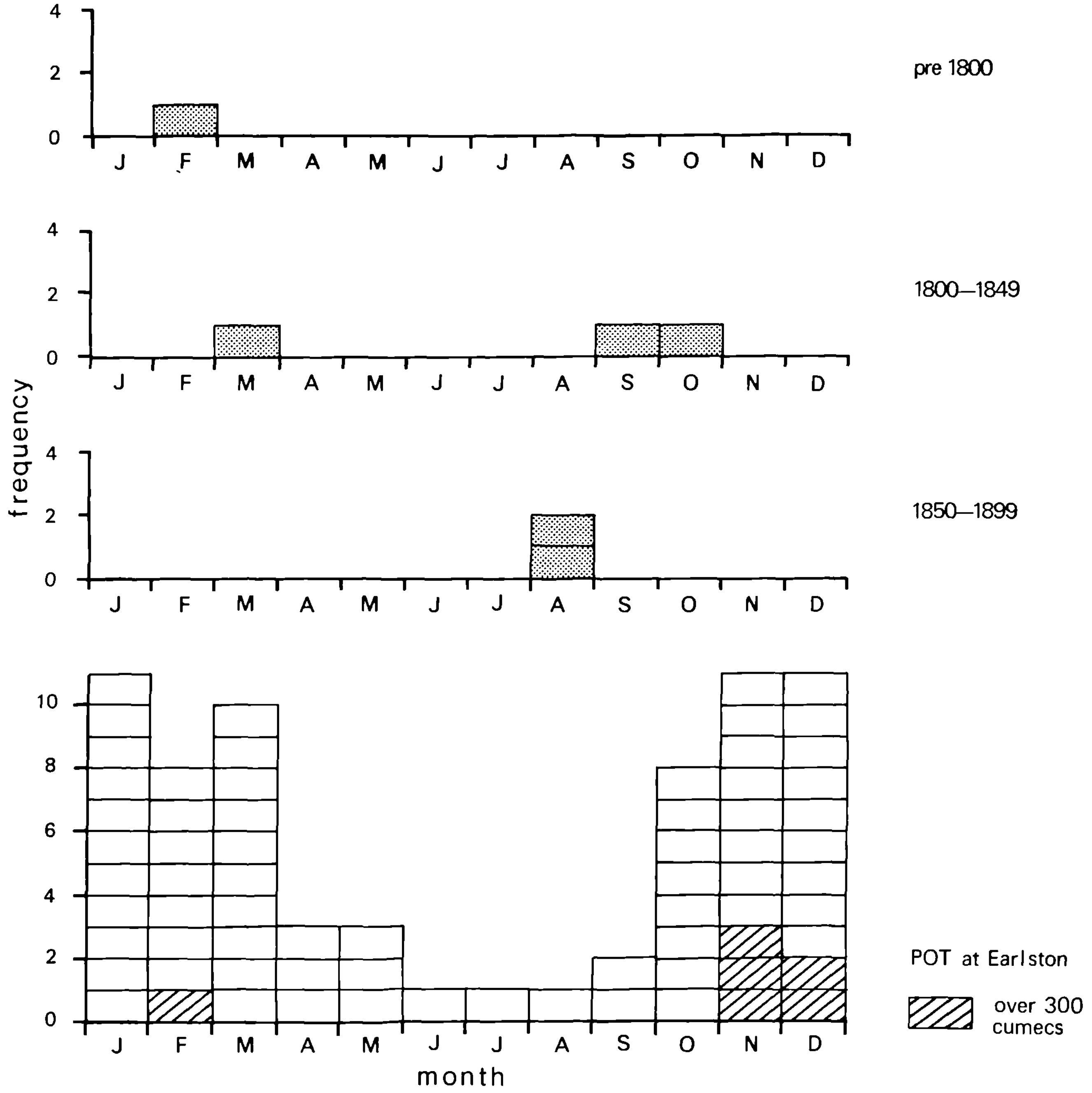


Figure 5.6.1.(viii)

The seasonality of flooding on the River Whiteadder

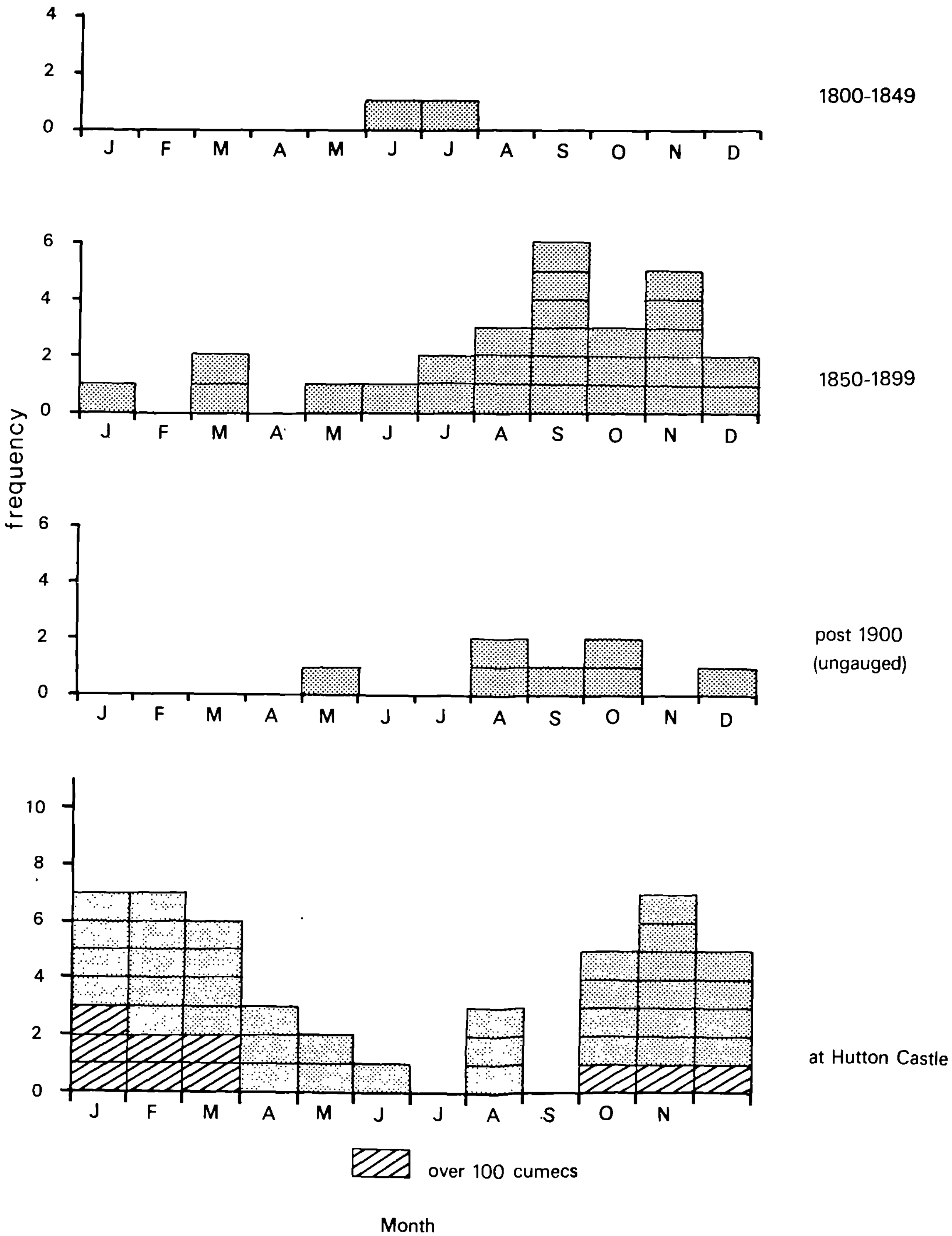


Figure 5.6.1.(ix)

The seasonality of flooding on the River Teviot

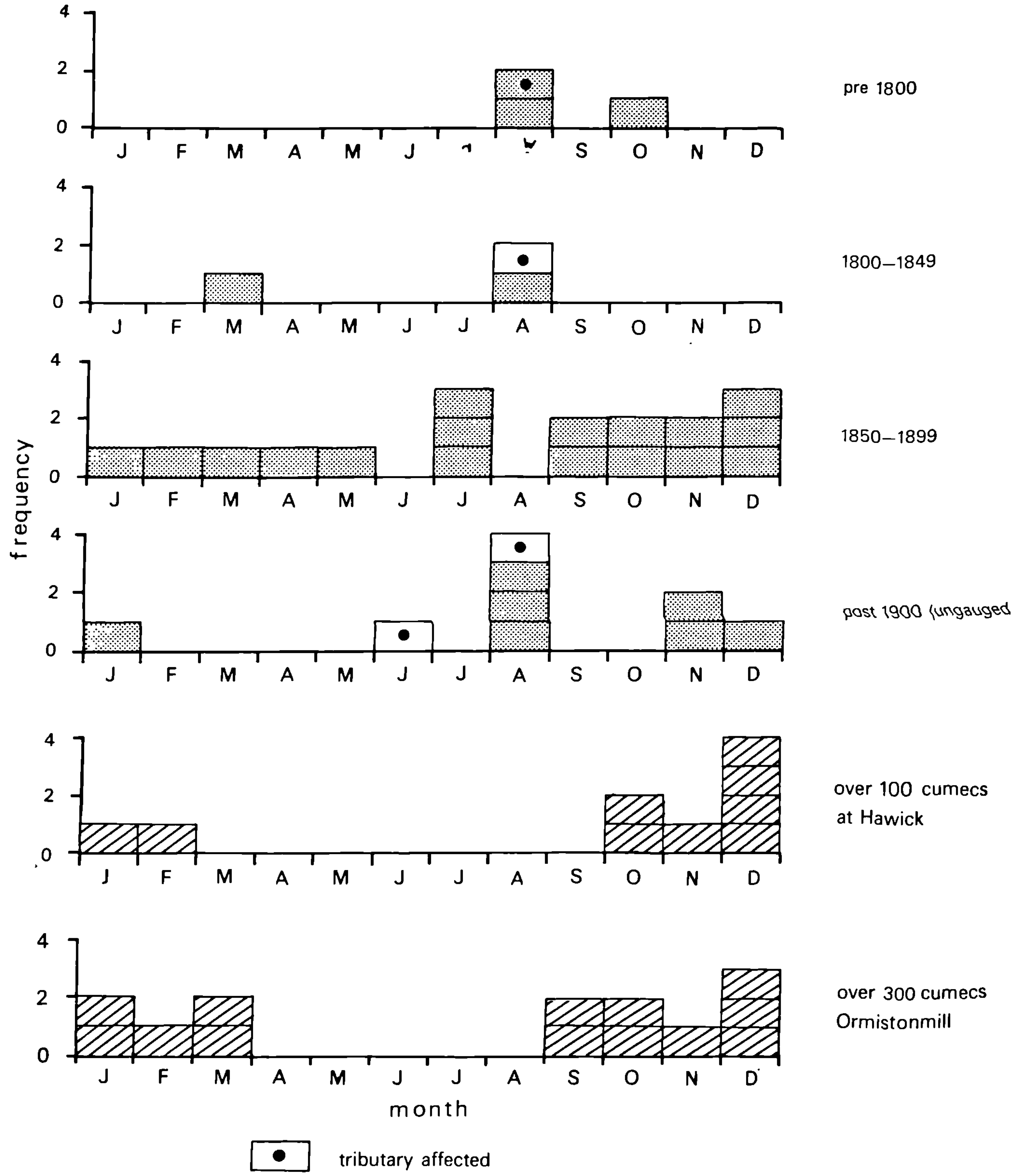


Figure 5.6.1.(x)

The seasonality of flooding on Bowmont Water

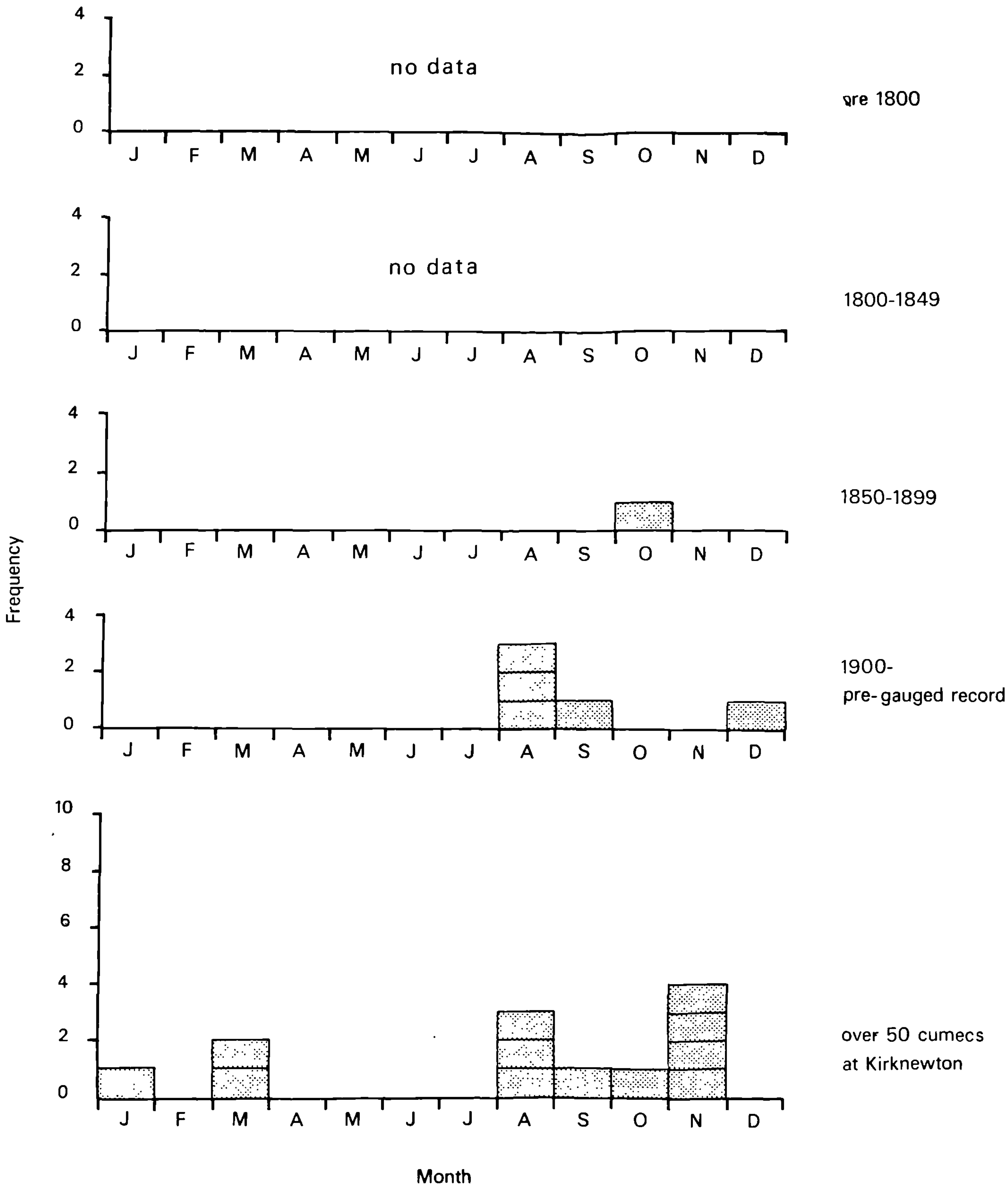


Table 5.6.1.(i)

Hydrometeorological characteristics of historical flood events on the middle mainstream Tweed

A Regional storm

B Convectional storm

C Intense rainfall

D Thunder and lightening

E Wind

F Wind direction

G Snow melt contribution

H Ice break

I Temperature rise

J Ground frost

K Pre-flood ground saturation

Event	A	B	C	D	E	F	G	H	I	J	K
10/08/1829	*				*						
09/02/1831							*				
00/00/1845	*										*
23/10/1864	*		*								*
25/09/1872	*										*
00/04/1874		*									
05/03/1879					*	W	*				
24/11/1880	*						*				
09/03/1881	*						*		*		
20/09/1891	*				*	NNE	*				
14/12/1914	*										
18/08/1920	*		*								
12/08/1948	*		*			NE					*
21/11/1963	*										
04/01/1982							*	*			

Table 5.6.1.(ii)

Hydrometeorological characteristics of historical flood events within the Teviot catchment

A Regional storm

B Convectional storm

C Intense rainfall

D Thunder and lightening

E Wind

F Wind direction

G Snow melt contribution

H Ice break

I Temperature rise

J Ground frost

K Pre-flood ground saturation

Event	A	B	C	D	E	F	G	H	I	J	K
00/08/1294	*										
05/08/1767		*	*								
00/00/1795								*			
09/02/1831							*				
03/08/1846		*	*	*							
23/10/1864	*										
22/05/1865	*		*								
15/11/1866	*										
10/07/1872		*									
26/07/1872		*		*							
02/04/1874	*										
03/09/1875	*										
21/12/1875	*										
10/07/1880		*						*	*		
27/10/1880	*				*						
27/11/1881					*		*				
03/12/1881											
25/09/1883	*				*						*
10/08/1901		*									
27/08/1905		*		*							
13/11/1938	*										
13/08/1948	*										
28/08/1956	*										
09/01/1962	*				*						

5.6.2 Tweed study area: Analysis of discharge records

Within the study area, the mainstream Tweed has 4 continuous discharge gauges with at least one on each of the studied tributaries (Table 5.6.2.(i)), all being maintained by the Tweed River Purification Board (TRPB). At each of these stations, the magnitude of discharges for set RI are shown in Table 5.6.2.(ii).

Although the earliest gauged record on the mainstream Tweed was in 1949 at Peebles, occasional flood reconstructions are found in the literature, which aid in the ranking of flood events. For example, in The Kelso Mail, reporting the flood on the 14th Feb, 1831, it was stated that there:

> "Passed through the arches of Kelso Bridge in a single second, a body of water equal to not less than 100,000 cubic feet or about 630,000 imperial gallons." (Kelso Mail, 1831)

This gave a peak discharge of 2837 m^3 s^{-1}, which although very approximate, provided an order of magnitude estimate. However, there was no indication of how reliable this estimate was. The snowmelt flood of 9th Mar, 1881 on the Tweed is documented by Reid (1882), the flood peak being estimated (post-flood) at Abbotsford Ferry, one mile below where the Tweed is joined by the Ettrick (1465 km^2). At that point, the river had risen 12 feet (3.7 m) above its ordinary level and maximum discharge was not less than 865 m^3 s^{-1} (0.59 m^3 s^{-1} km^{-2}). Unfortunately, there was no comparative discharge record for the Aug,

Table 5.6.2.(i)

Tweed: Magnitude of discharges for different recurrence intervals, fitting the EV1 distribution to the annual maximum series

Catchment	Station	Q(2)	Q(5)	Q(10)	Q(20)	Q(50)	Q(100)	Q(100)*
Whiteadder	Hutton Castle	110.7	166.7	207.7	248.8	307.3	354.5	300.9
Whiteadder	Hungry Snout	18.7	26.3	34.4	45.5	68.3	94.1	15.0
Blackadder	Mouth Bridge	38.3	56.1	71.2	88.1	115.7	140.6	117.0
Leader	Earlston	58.2	77.0	88.5	98.8	110.9	120.0	159.9
Teviot	Hawick	187.5	228.1	255.2	278.4	305.4	326.7	508.4
Teviot	Ormiston Mill	274.2	369.5	438.1	509.6	607.9	691.4	783.7
Till	Etal	86.9	122.3	157.7	204.9	297.2	398.1	231.1
Glen	Kirknewton	48.1	73.3	87.6	98.1	108.6	119.1	-----
Tweed	Peebles	159.9	236.5	322.2	440.4	670.8	925.7	538.3
Tweed	Boleside	388.3	564.7	701.5	847.1	1067.7	1253.0	1074.4
Tweed	Dryburgh	459.5	627.9	755.5	898.5	1118.0	1301.8	1267.7
Tweed	Sprouston	635.1	963.6	1219.1	1496.5	1898.0	2248.4	1837.3
Tweed	Norham	669.6	907.7	1086.2	1294.6	1599.6	1867.4	1963.8

Q(i) discharge with a RI of i years.

All discharges in $m^3 s^{-1}$

(Source:- Acreman, Ph.D. thesis in preparation)

Annual maximum series and EV1 analysis for the Kirknewton Station were calculated by the author.

* calculated using regional ordinates

---- unknown

1948 event on the Tweed; a great loss in terms of knowledge of possible catchment runoff rates with a basin of this size. However, the 7th Jan, 1949 flood at Peebles was calculated to have a discharge of 1079 m^3 s^{-1} (1.56 m^3 s^{-1} km^{-2}) and a RI of over 100 years. These are much higher rates of specific runoff than gained on stations on either the mainstream Spey or Dee, post-1900.

The most extreme event on the Teviot, since discharge records commenced in 1960, was on 31st Oct, 1977 when a discharge of 310.4 m^3 s^{-1} (0.96 m^3 s^{-1} km^{-2}) was recorded at Hawick, with a RI of under 60 years (see Appendix 2.3). The second largest was 279.5 m^3 s^{-1} (0.87 m^3 s^{-1} km^{-2}) with a RI of approximately 20 years. The calculated specific runoff for the estimated 100 year event was 1.01 m^3 s^{-1} km^{-2}. This 1977 flood event was given a ranking of the second largest flood on record by the TRPB, in comparison with the event of 13th Nov, 1938, claimed by newspaper records to be the largest event for over 100 years (Figure 5.6.1.(iii)). Unfortunately, there was no estimated discharge for this event.

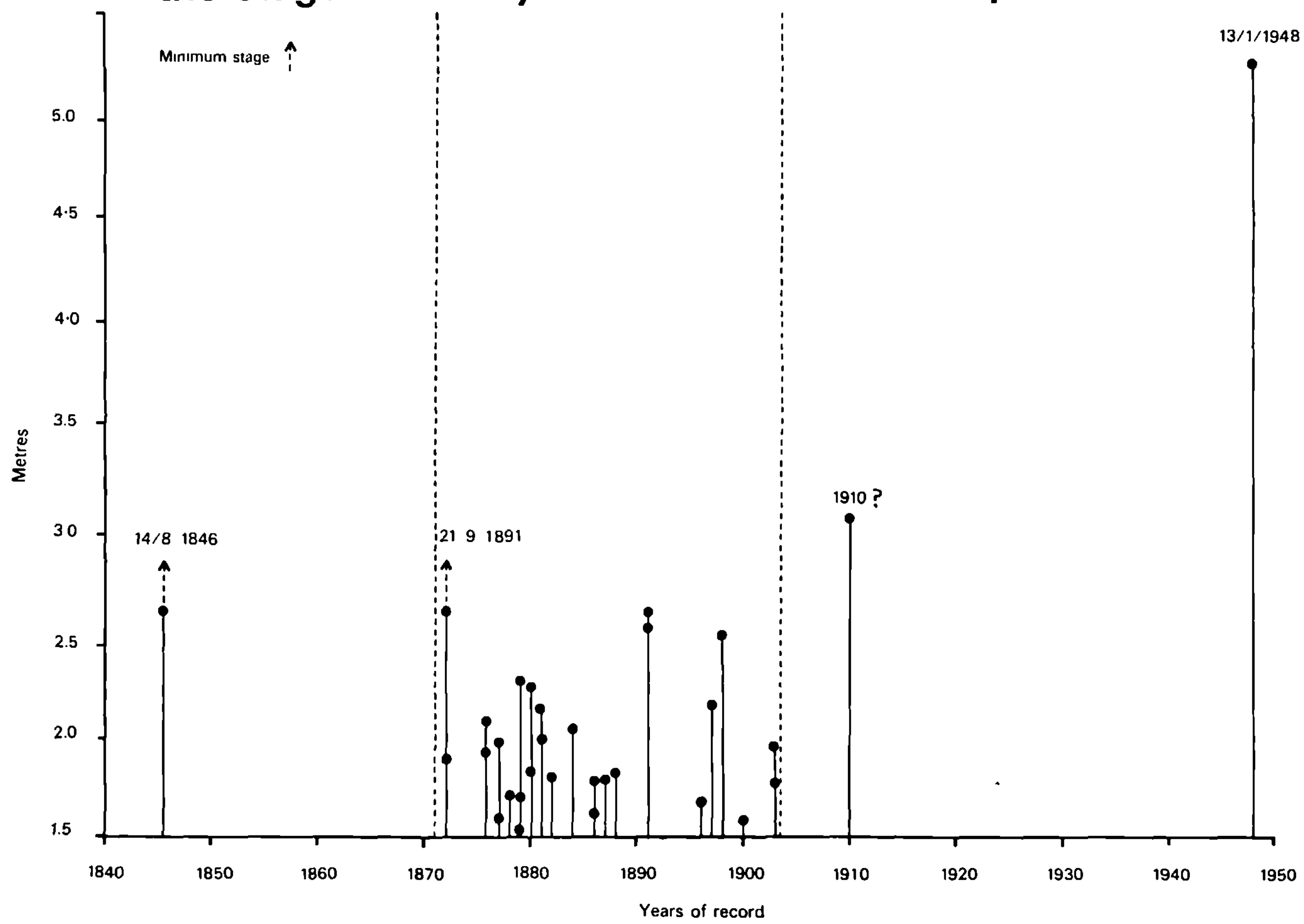
The discharge record on the lower Whiteadder is provided by the Hutton Castle gauge (drainage area 503 km^2). The highest discharge on record was a winter event, occurring on 3rd Jan, 1982 with 265.9 m^3 s^{-1} (ie. 0.53 m^3 s^{-1} km^{-2}) and this represents approximately the 30 year event. Over the period of record 1962-1982, 10 peaks exceeded 100 m^3 s^{-1}, which is less than the mean annual flood.

Upstream on the Whiteadder, there was a gauge maintained from 1958-1968 at Hungry Snout, which has now been drowned by the Whiteadder Reservoir. The largest event on this record was 63.1 $m^3\ s^{-1}$ on 4th Aug, 1966 (1.38 $m^3\ s^{-1}\ km^{-2}$), with a RI of below 50 years. Clearly, much higher runoff rates can be maintained with localised storms in smaller catchments. However, the discharge record on the upper Whiteadder could be ranked in detail beyond this. On the church steps at Abbey St Bathans, on the Whiteadder, the heights are marked of the major floods 1872-1903. These flood marks were found to be degraded but the following list of events and their *stage at this cross-section is found* in The History of the Berwickshire Naturalists Club Vol XXIV (Appendix 1.10). This list showed the height attained above the summer level of the stream (actual height not recorded), at a cross-section which has "a width of 60 feet (18 m), increased to almost 150 feet (46 m) in the highest floods." (Hist. Berw. Nat. Club, XXIV p266). From the same source, it was recorded that on the 25th Sept, 1872 occurred the highest flood on the Whiteadder since 1846. Thus, the stage and ranking are shown in Figure 5.6.2.(i).

To put more recent events in context, Holford (1976) states that in 1910 the maximum flood stage attained was 10 feet (3.05 m) and the 1948 flood reached 17 feet (5.19 m) at the same reach. The source of this information is not known but from the Farmer's Weekly (British Rainfall, 1948), it states that the Whiteadder rose 16 feet (4.88 m) above normal. The 1948 stage was twice the previous maximum level of 8 feet 6.5 in (2.46 m) recorded on 21st Sept, 1891 and the highest stage over the period 1872-1903. The associated rates of runoff in 1948 thus must

Figure 5.6.2.(i)

Flood history on the upper Whiteadder as recorded by the stage at Abbey St Bathans church steps



have been much higher in magnitude than in the other known flood events, despite the high frequency of discharge events in the 1870s and 1880s.

On the Blackadder, the discharge record is short (1973-1982); the largest flood on record as on the Whiteadder was 92.8 m^3 s^{-1} on 3rd Jan, 1982 (0.58 m^3 s^{-1} km^{-2}), approximately a 25 year event. Presumably, a similar flood history to the Whiteadder could be assumed to apply, especially in terms of regional events.

Again, the most extreme event within the gauged Leader record was on 3rd Jan, 1982, when a discharge of 117.6 m^3 s^{-1} was recorded (0.49 m^3 s^{-1} km^{-2}; see Appendix 2.3). The estimated RI of this event was approximately 90 years and only three events have had a discharge exceeding 70 m^3 s^{-1} (< 5 years RI), within the gauged record. In terms of the seasonality and hydrometeorologic conditions, the highest number of flood events (27) occurred in the months Dec-Feb suggesting a predominance of winter cyclonic events. On the other hand, only 3 events were recorded between June and August, in many areas a period distinguished by summer frontal or convectional storms. However, according to the ranking of the TRPB, none of these gauged events rank in the first 5 on the Leader since 1831, and thus the gauged record is rather atypical of possible discharge extremes.

5.6.3 Tweed study area: Analysis of rainfall records

There are numerous long daily rainfall records within the Tweed study area and data collection and analysis had to be limited by the time available for this study (see Table 5.6.3.(i)). Certain periods, especially in the daily record, have unavoidable gaps due to breaks in data collection and shifts in the location of the rainfall gauge. The long-length Hawick record is a composite one, correlated with the closest long-term record at Jedburgh. The correlation between the three segments of record from each station varied as can be seen in Table 5.6.3.(ii). A composite record was produced using annual ratios as described in Section 5.2. A similar process was carried out to correct the annual maxima though it was deemed as accurate as using the whole unaltered record.

In terms of annual totals in the Whiteadder catchment, the wettest year on the Marchmont House record was 1872 with 1402.1 mm (Figure 5.6.3.(i)). Other years with annual totals above a threshold of 1000 mm were 1876, 1877, 1882, 1900, 1906, 1916 and 1928. Unfortunately, the Marchmont House record ceases before 1982, which from the Deeside/ Speyside records ranked second below 1872. The high totals for 1900 and 1916 are confirmed by other records eg. Duns Castle and Manderston House, with 1948 also exceeding 1000 mm annual total (Figure 5.6.3.(ii) and (iii)). The highest annual total in the Leader catchment was again 1916 (1034 mm), the only year above the 1000 mm threshold (see Figure 5.6.3.(iv)). However, Thomson (1902) reports that there were several very wet years in the Leader catchment, over the period 1855-1898 (Figure 5.3.(ii)). Within the Teviot catchment, on the Hawick record,

Table 5.6.3.(i)

Long length rainfall records within the Tweed study area

Catchment	Station	N.G.R.	alt. (m)	Years of record
Whiteadder	Marchmont	3743 6484	152	1/1/1871-31/12/1980
Whiteadder	Manderston	3809 6547	108	1/1/1901-30/11/1975
Blackadder	Duns Castle	3775 6538	137	1/1/1889-30/11/1978
Leader	Cowdenknowes	3575 3372	91	1/1/1898-31/12/1970
Teviot	Wolfelee	**** ****	**	1/1/1921-30/10/1948
Teviot	Westerdykes	**** ****	**	1/1/1947-31/8/1951
Teviot	Hawick S.W.	3512 6156	96	1/1/1956-31/12/1982
Jed	Jedneuk	3656 6234	83	1/8/1919-31/12/1975
Teviot	Silverbut Hall	c. 3503 6157	153	1/1/1866-31/12/1883

**** unknown

c. approximate N.G.R. and height

Table 5.6.3.(ii)

A composite annual total record for Hawick

Station	Years	Years in common	Ratio	Correction
Wolfelee	1866-1948	1921-1948	0.692	0
Westerdykes	1946-1950	1946-1950	0.735	1.062
Hawick S.W.	1956-1982	1956-1975	0.821	1.19

Figure 5.6.3.(i) MARCHMONT HO: ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN: (1868-1978)

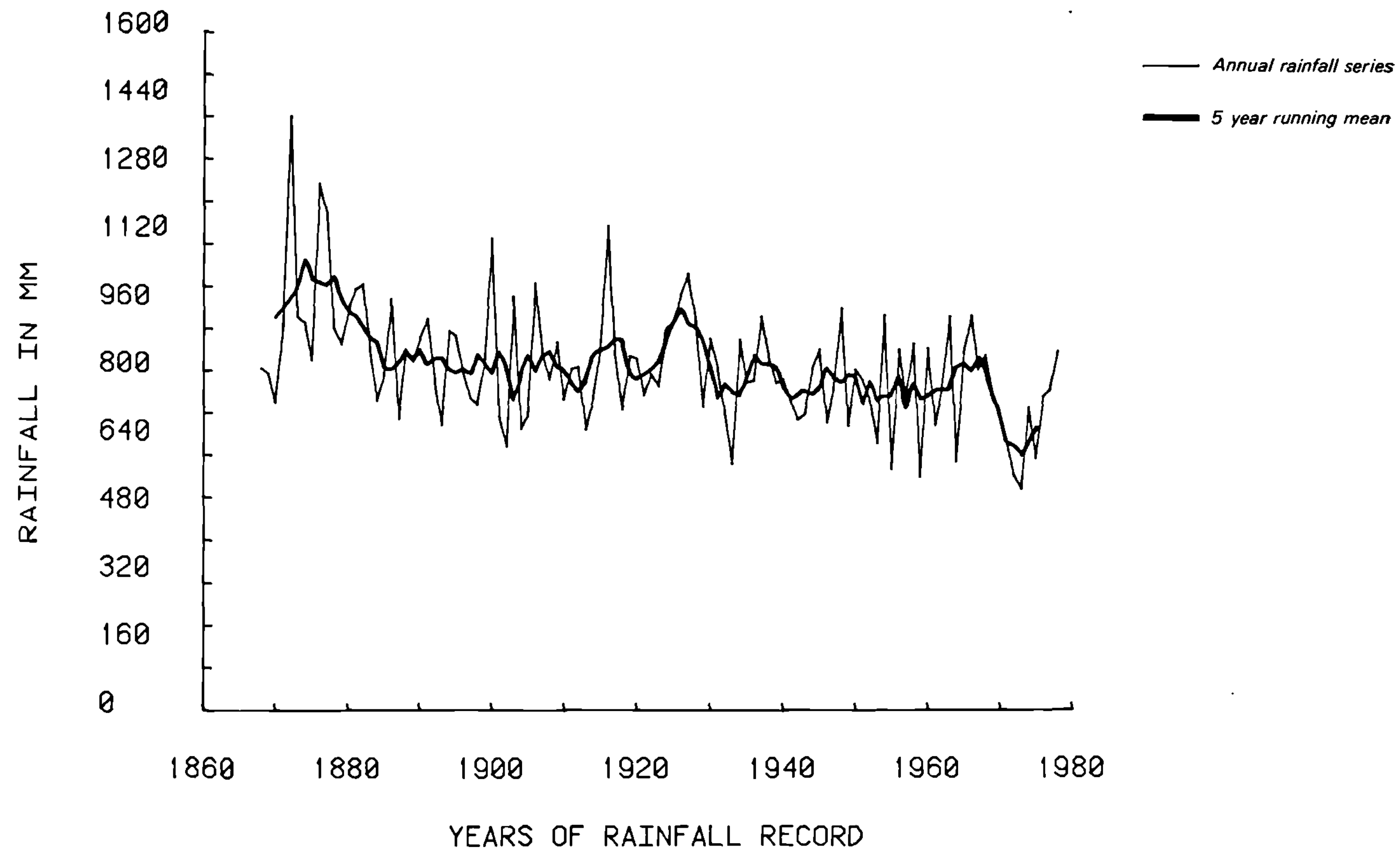


Figure 5.6.3.(ii) MANDERSTON HO: ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN: (1900-1975)

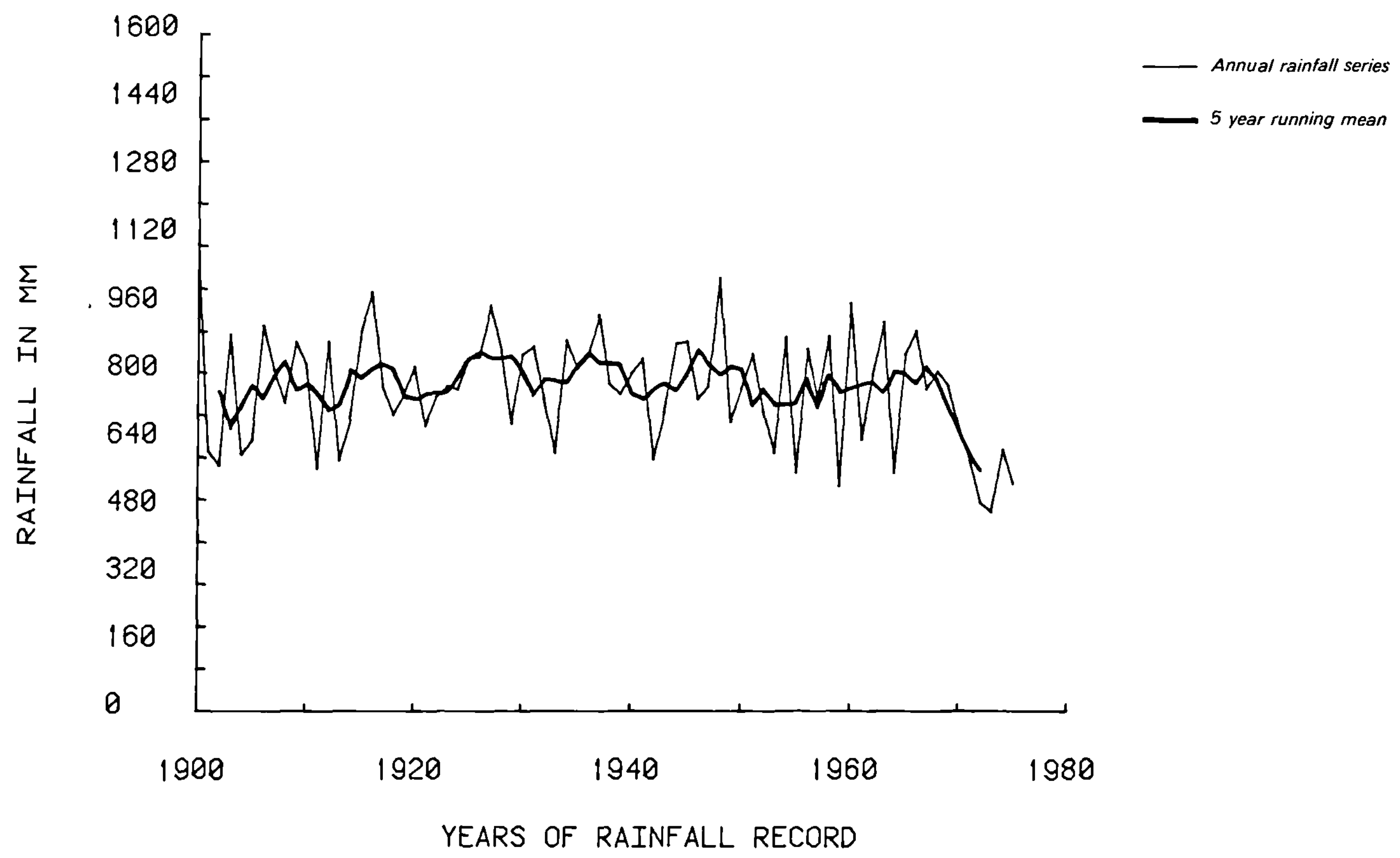


Figure 5.6.3.(iii) COWDENKNOWES:ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN:(1898-1980)

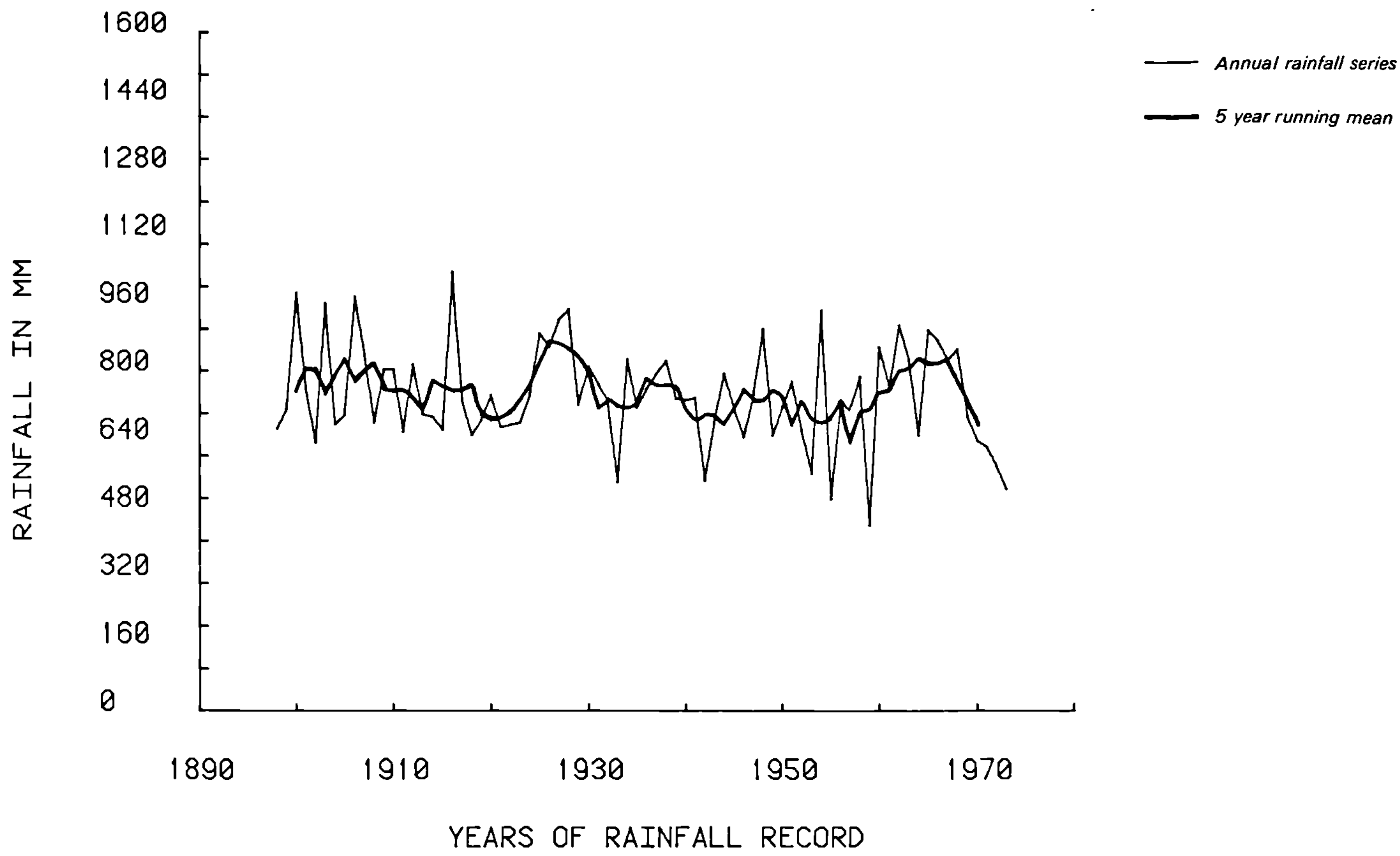
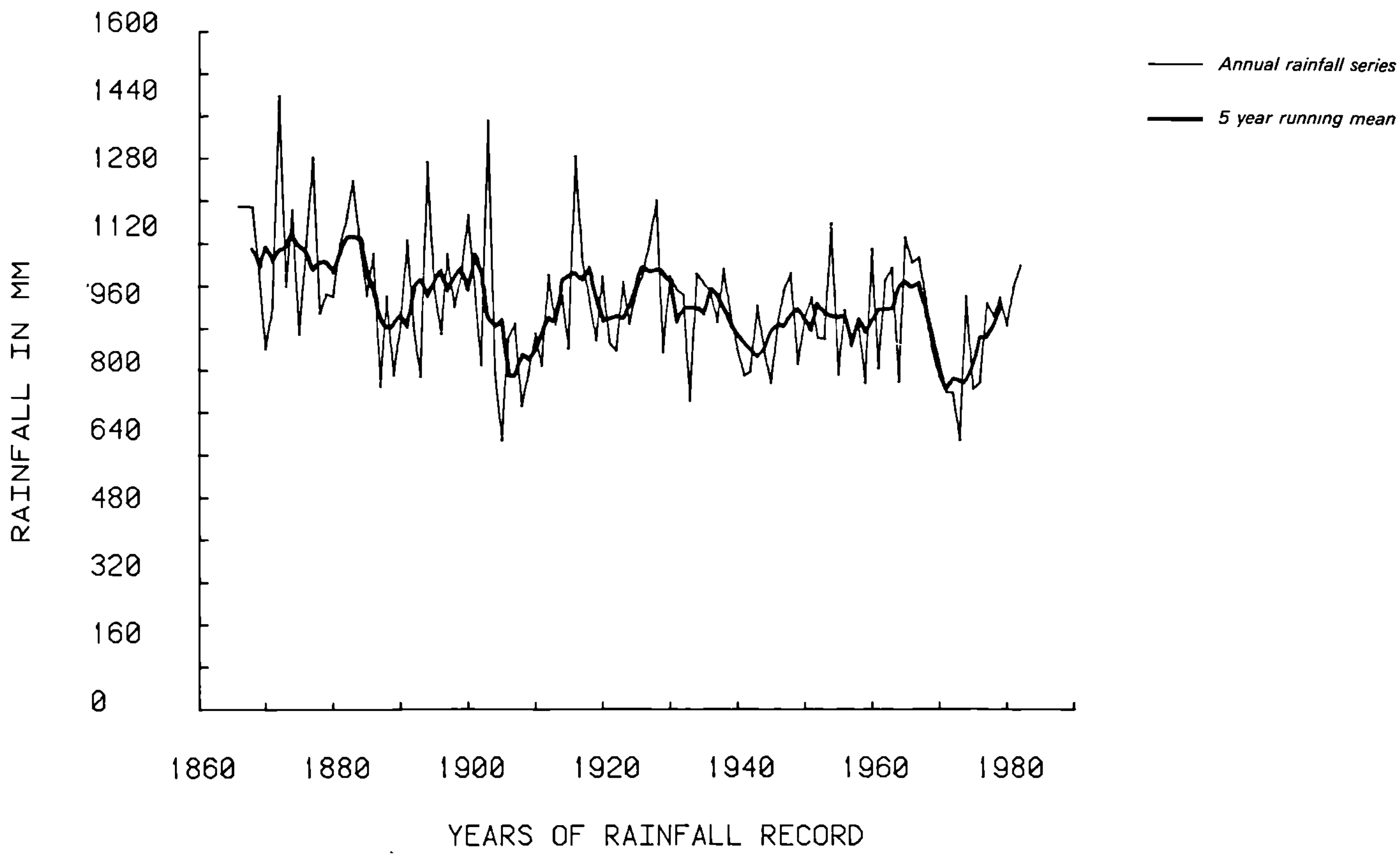


Figure 5.6.3.(iv) HAWICK:ANNUAL RAINFALL TOTALS : 5 YR RUNNING MEAN:(1866-1982)



two years stand out as being exceptionally wet, namely 1872 (1448.8 mm) and 1903 (1390.1 mm). It is notable that most of these years recording high annual totals occurred before 1930. This may have implications for fluctuations in longer term soil moisture deficits.

When a 5-year running mean was put through these data sets, there were obvious temporal fluctuations, which were clearest on the longer annual total records of Hawick and Marchmont House (see Figures 5.6.3.(i) to 5.6.3.(v)). Thus, on both records there was a peak pre-1900 followed by a gradual decline in annual totals, broken by other shorter term fluctuations. According to Brooks and Glasspoole (1928), the rainfall at Marchmont House during the 1870s exceeded the average by 20% and the 1870-1880 peak was the most prominent feature on the rainfall graph. However, when the Kolmogorov-Smirnov and runs tests were carried out on the Marchmont House annual total series, no significant deviation from a normal distribution was found.

The annual total series was then submitted to Box-Jenkins analysis to assess any periodicities in the data. Each stations results will be presented and then any general pattern will be assessed. For the Marchmont House record, 3, 4 and 9 year lags had a significant positive autocorrelation (AC). Although other years had fairly high AC, eg. 16-18 years, they were not significant to 2 S.E. Fourier analysis showed no major periodicities but the highest peak was at 9-11 years. At Duns Castle, only a 3 year lag showed a significant positive AC ie. no longer term trends. In the Leader catchment at Cowdenknowes, there was a significant high positive AC at 38 years (0.226) and a high negative AC at 43 years (0.253). In the Teviot catchment, highly significant positive AC were found at 3, 9 and 31 year lags. This 31 year lag also showed a reasonably high (0.105) positive correlation on the Jedburgh record. There thus does not appear to be a regionally consistent long-term periodicity within the six raingauge records, although shorter term periodicities (eg. 3, 9 and 16 years) are more prominent.

5.6.4 Tweed study area: Analysis of the rainfall POT series

In terms of the temporal frequency of the 24 hour POT series, at Marchmont House the highest number per year was 1876 with 8 peaks over 1 inch (25.4 mm; Figure 5.6.4.(i)). If however the threshold was increased to 1.5 in (38.1 mm), the exceptional nature of this frequency was seen; out of these 8, 6 were greater than or equal to 1.5 in (38.1 mm). The maximum number of peaks over 2 in (50.8 mm) in any year was 2. From Figure 5.6.4.(i), the pre-1890 period certainly gave the impression of having a greater frequency of POT, especially when the threshold was increased to 1.5 in (38.1 mm) and above. Pulses of high frequency peaks were apparent between 1871-1879 (Figure 5.6.4.(iii)), where 35 peaks > 1 in (25.4 mm) occurred, cf. 1930s and 1950s where only 17 occurred. From the shorter Duns Castle record, the highest frequency of 7 POT was found in 1900 with 6 in 1891 and 1909. If the threshold was increased to 1.5 in (38.1 mm), 1891 (a major flood year) stood out with 3 peaks in excess of this value (Figure 5.6.4.(iv)). This frequency of 6 POT in 1909 also occurred on the Manderston record. The maximum there was 7 POT in 1934 but 1927 was notable for 3 peaks over 1.5 in (38.1 mm; see Figure 5.6.4.(v)).

Figure 5.6.4.(i)

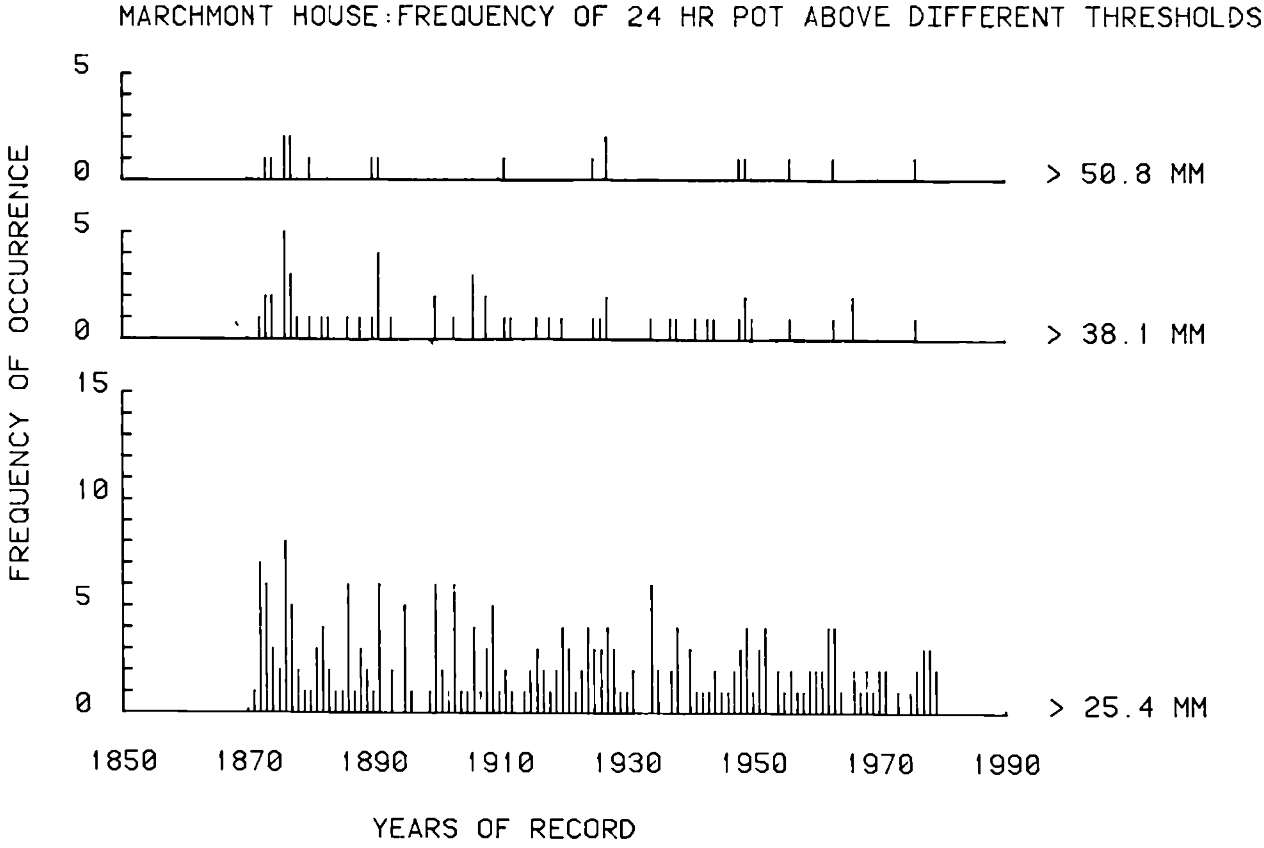


Figure 5.6.4.(iii)

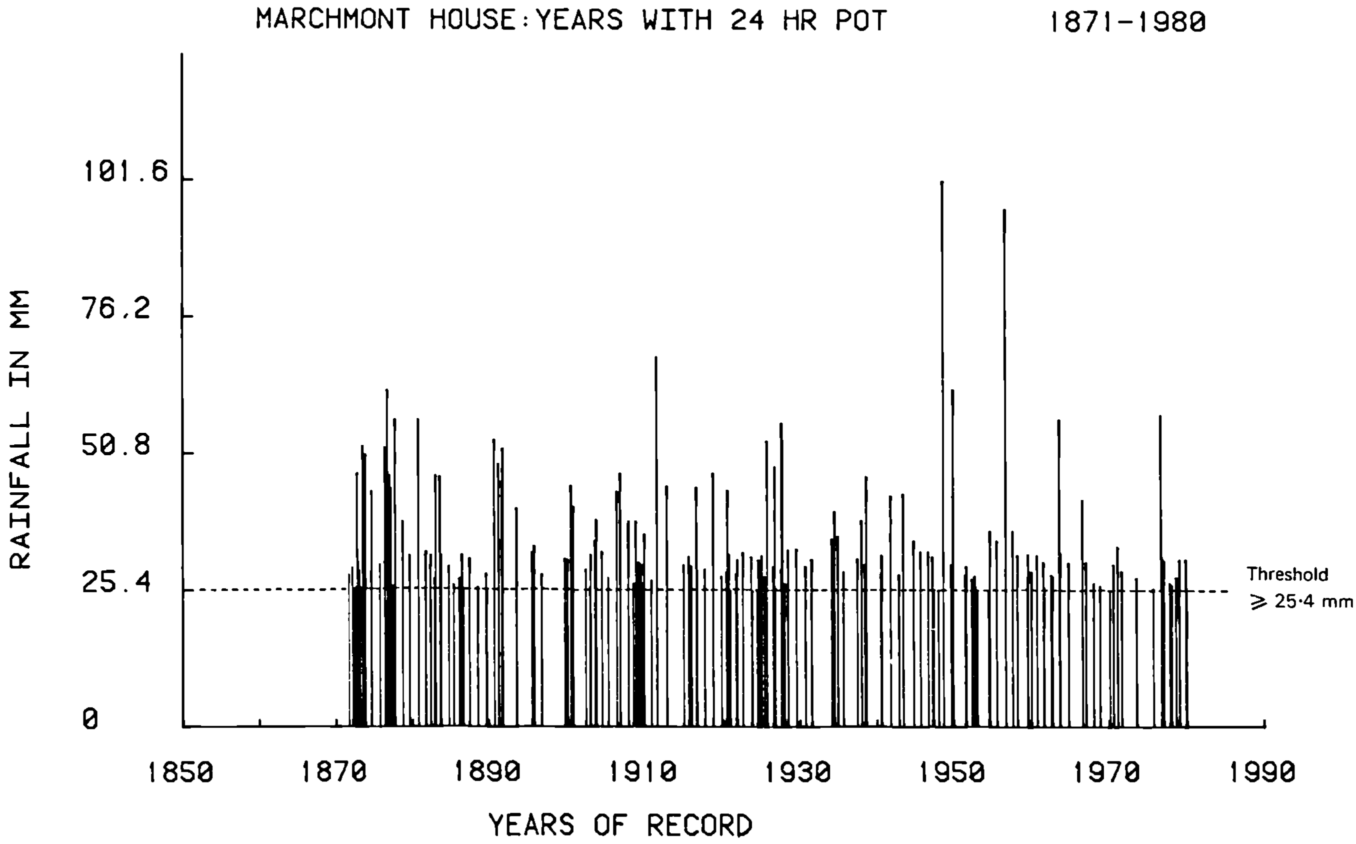


Figure 5.6.4.(ii)

The seasonality of 24 and 48 hour POT (subdivided by decade)

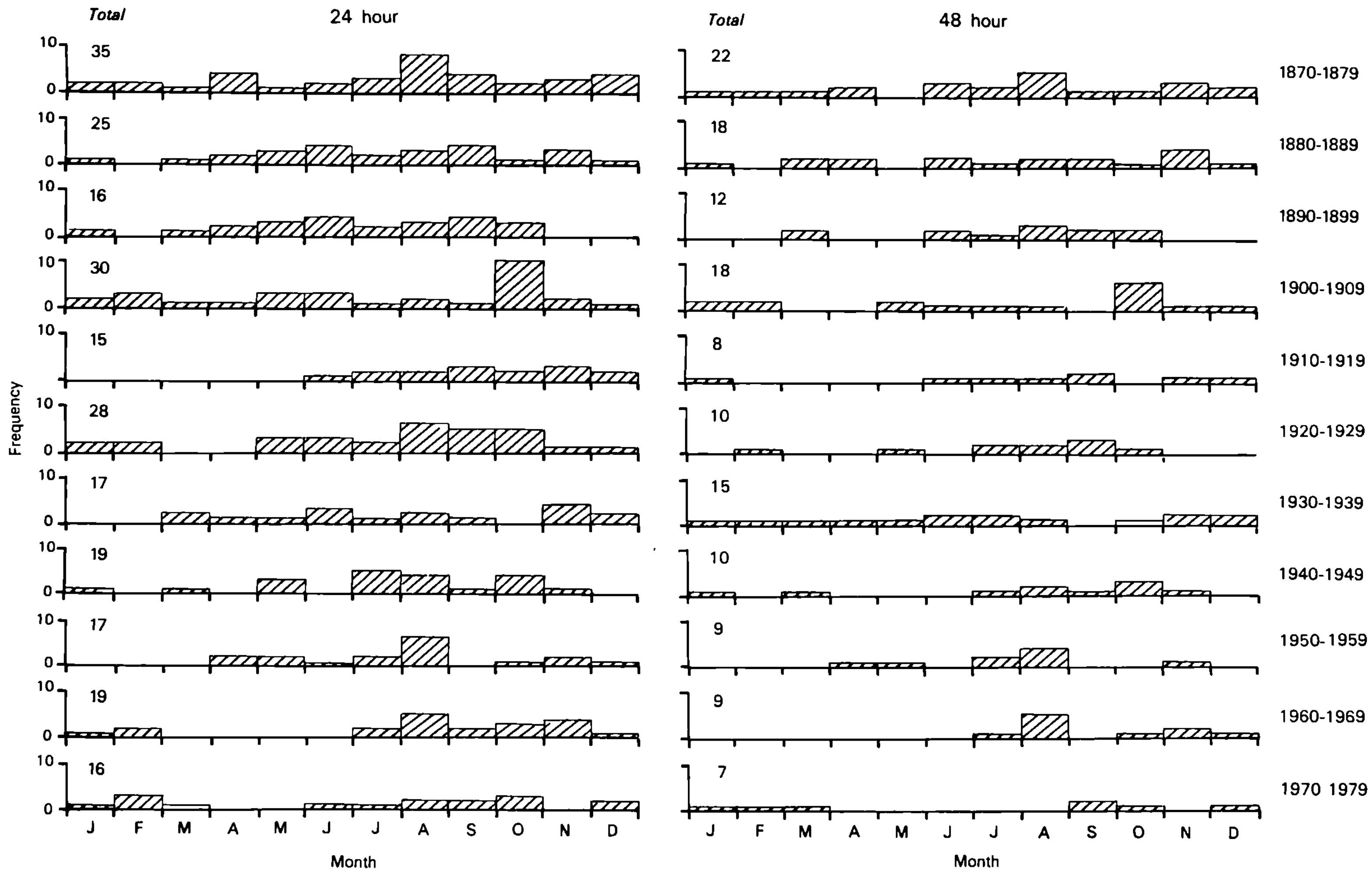


Figure 5.6.4.(iv)

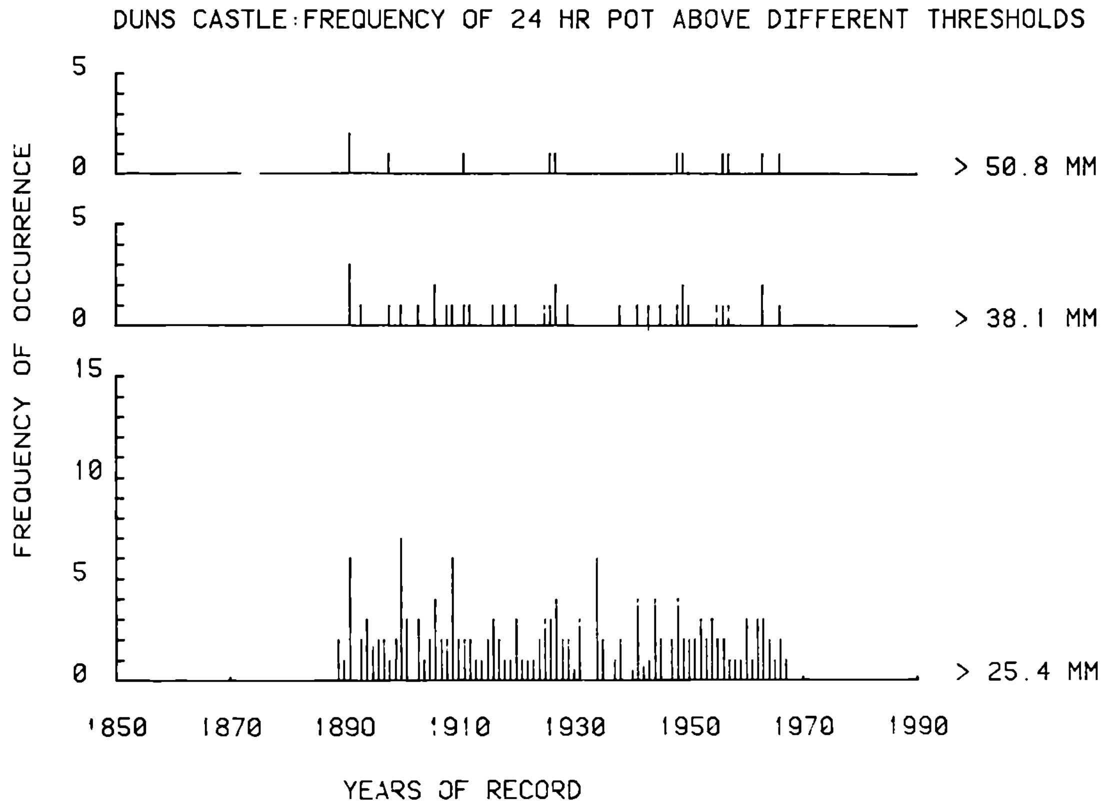
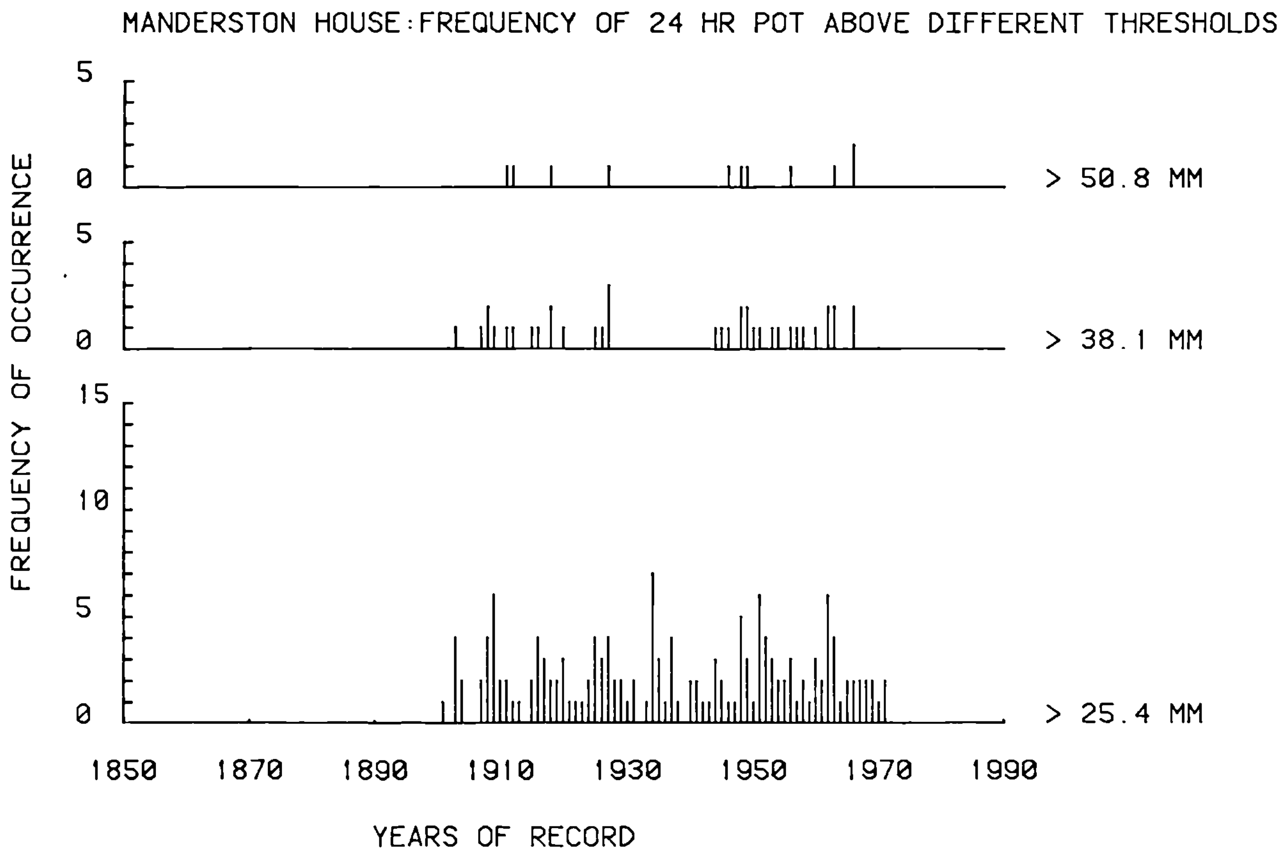


Figure 5.6.4.(v)



At Cowdenknowes, the highest frequency of 24 hour POT occurred on 2 consecutive years (1962-1963) with 5 POT, though no major flood events were recorded. Other notable years included 1927 with 3 peaks above a threshold of 1.5 in (38.1 mm) and the 5 years which had a fall above 2 in (50.8 mm)- 1925, 1929, 1948, 1956 and 1961 (Figure 5.6.4.(vi) and (vii)). From the Hawick record, the highest frequency of 24 hour POT > 1 in (25.4 mm) was 6 in 1962 and 1963. 1947 was exceptional with 4 POT greater than or equal to 1.5 in (38.1 mm; Figure 5.6.4.(viii)). However, when the short Silverbut Hall record (1866-1883) was studied for 1872, only 2 POT occurred. This does not seem to correlate with the reported high incidence of flooding on the Teviot as derived from historical sources. This discrepancy suggests that convectional storms were important but also that soil moisture deficit was low, with increased autumn and winter flooding. On the Jedburgh record, the spatial variation in POT frequency again became apparent; 1962 had 1 POT > 1 in (25.4 mm) and 1947 was without any (cf. Hawick). 1982 was the highest year on record with 4 while 1948 was an unusual year with 3 peaks in excess of 1.5 in (38.1 mm; Figure 5.6.4.(ix)).

When the duration was increased to 48 hours and the threshold increased to 1.5 in (38.1 mm), the following results were found, as seen in Figures 5.6.4.(x) and (xiii). There was considerable spatial variation in occurrence of POT. At Marchmont House, the highest number of 48 hour POT was 6 in 1934, but this year was an exception to the more quiescent post-1900 pattern. Pre-1903, there were several years with high POT frequency. The maximum number of 48 hour POT per annum at Manderston House was 4 in 1958, however other years were more notable in

Figure 5.6.4.(vi)

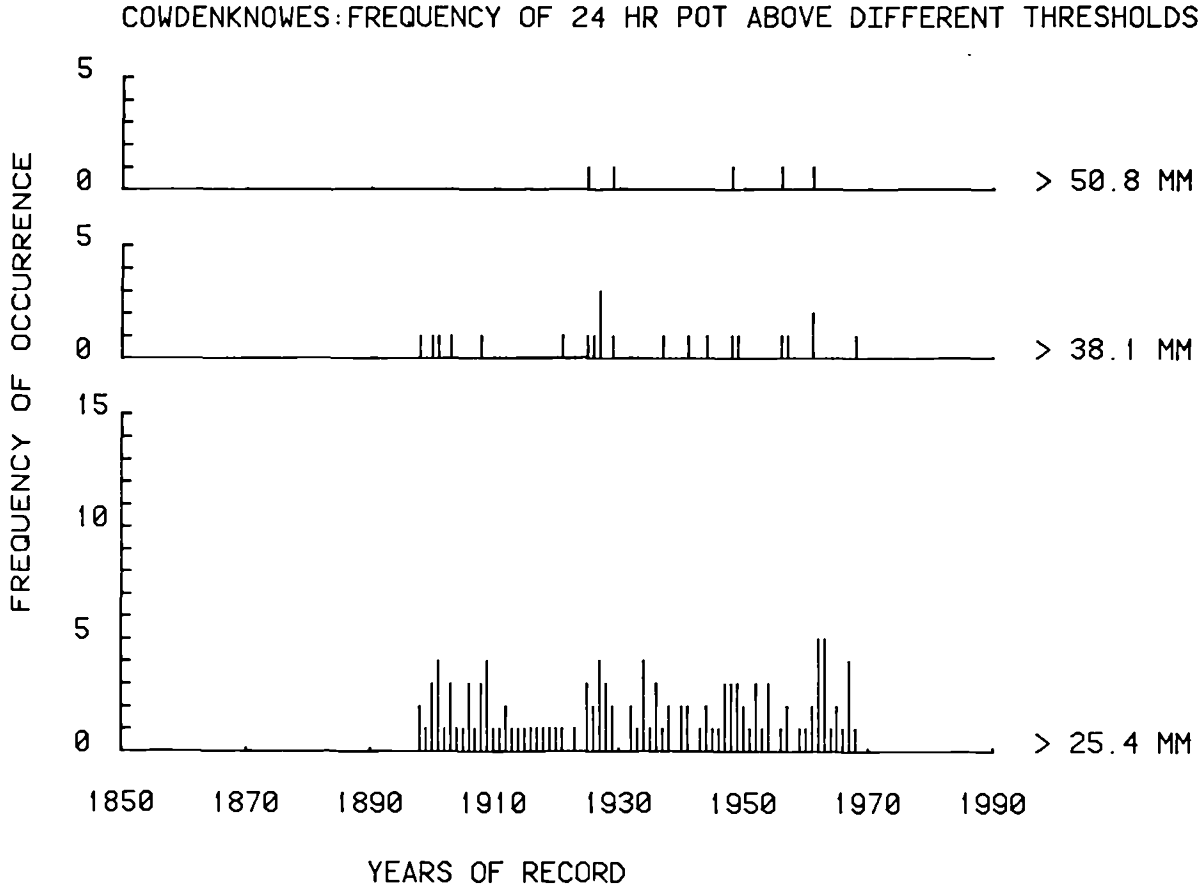


Figure 5.6.4.(vii)

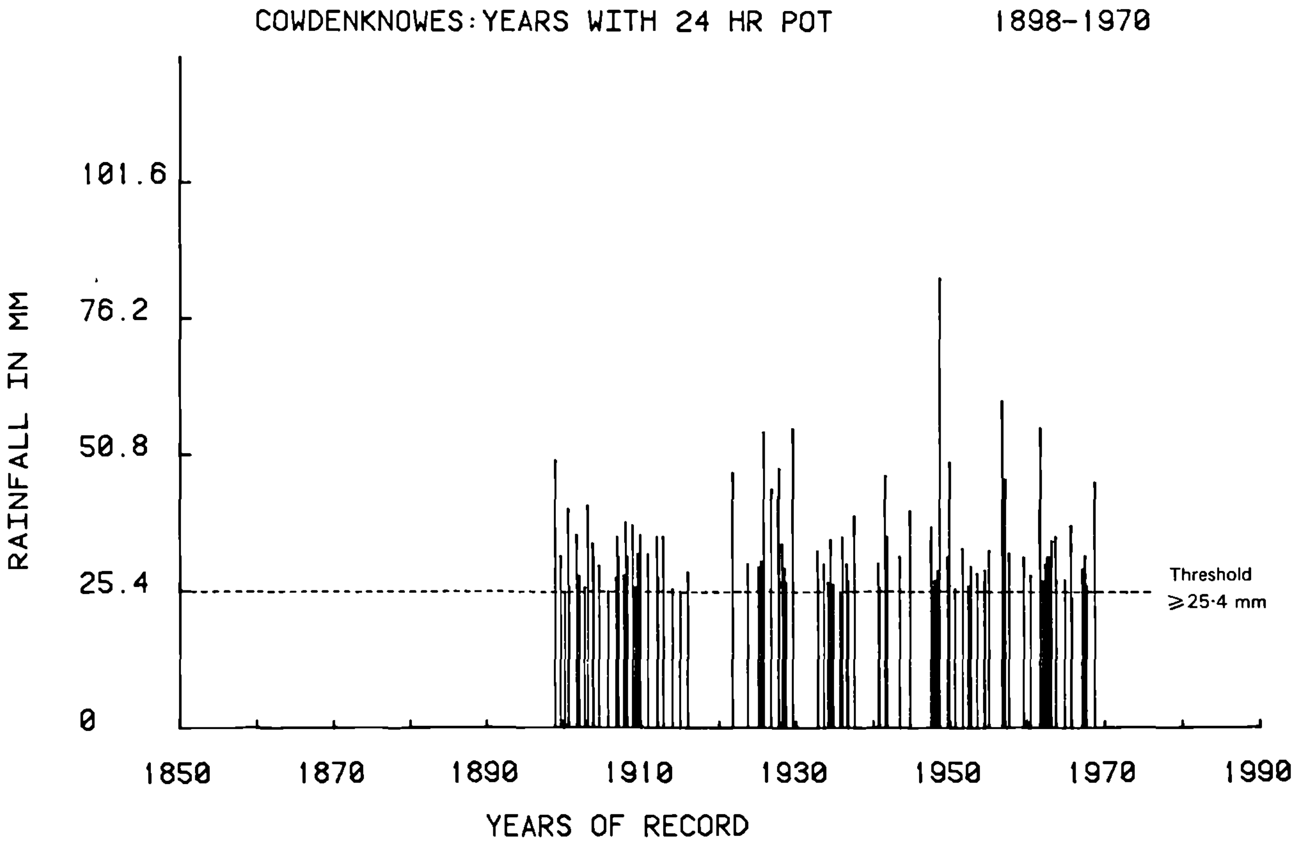


Figure 5.6.4.(viii)

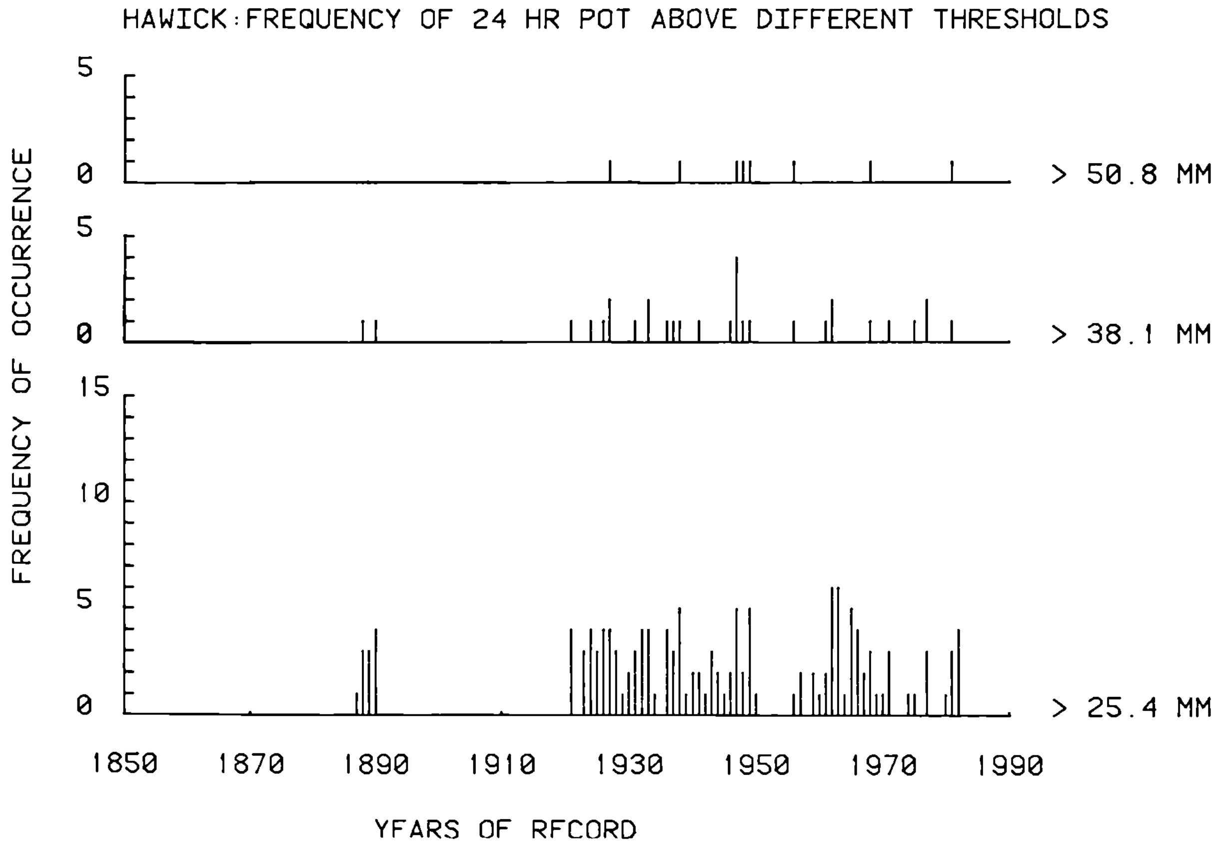


Figure 5.6.4.(ix)

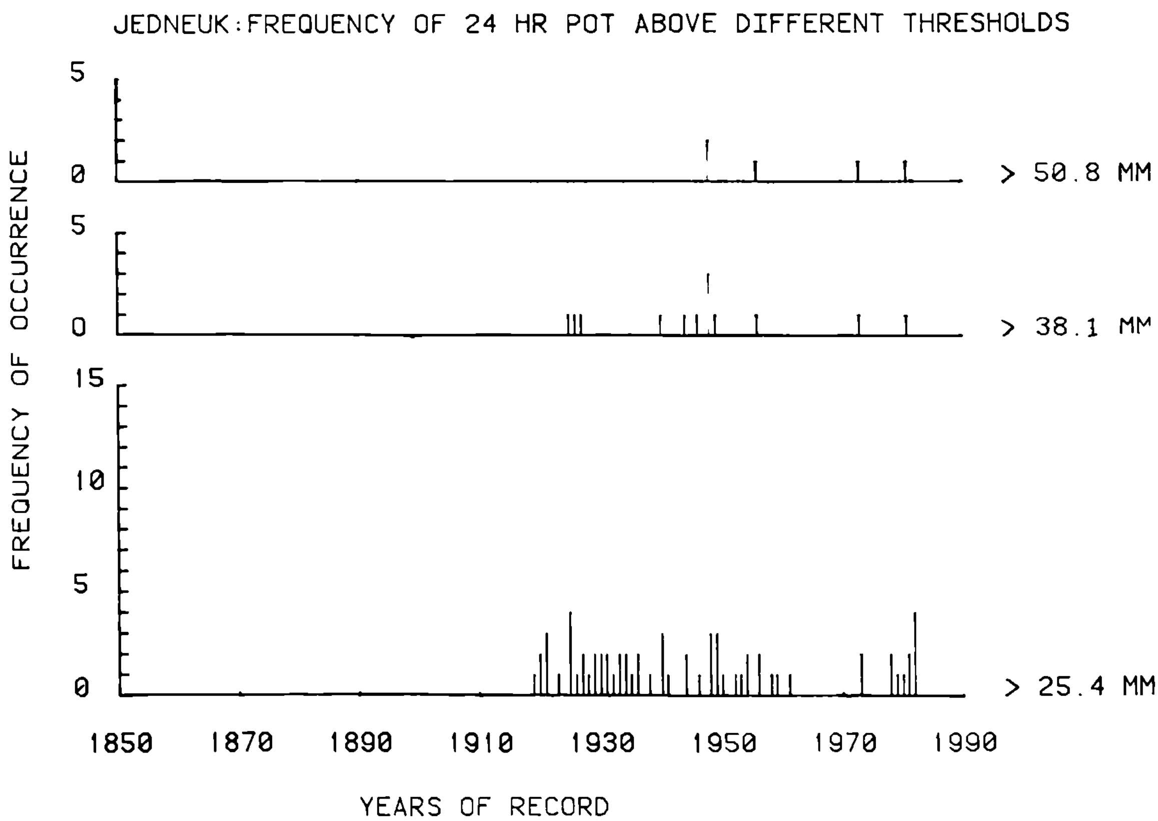


Figure 5.6.4.(x)

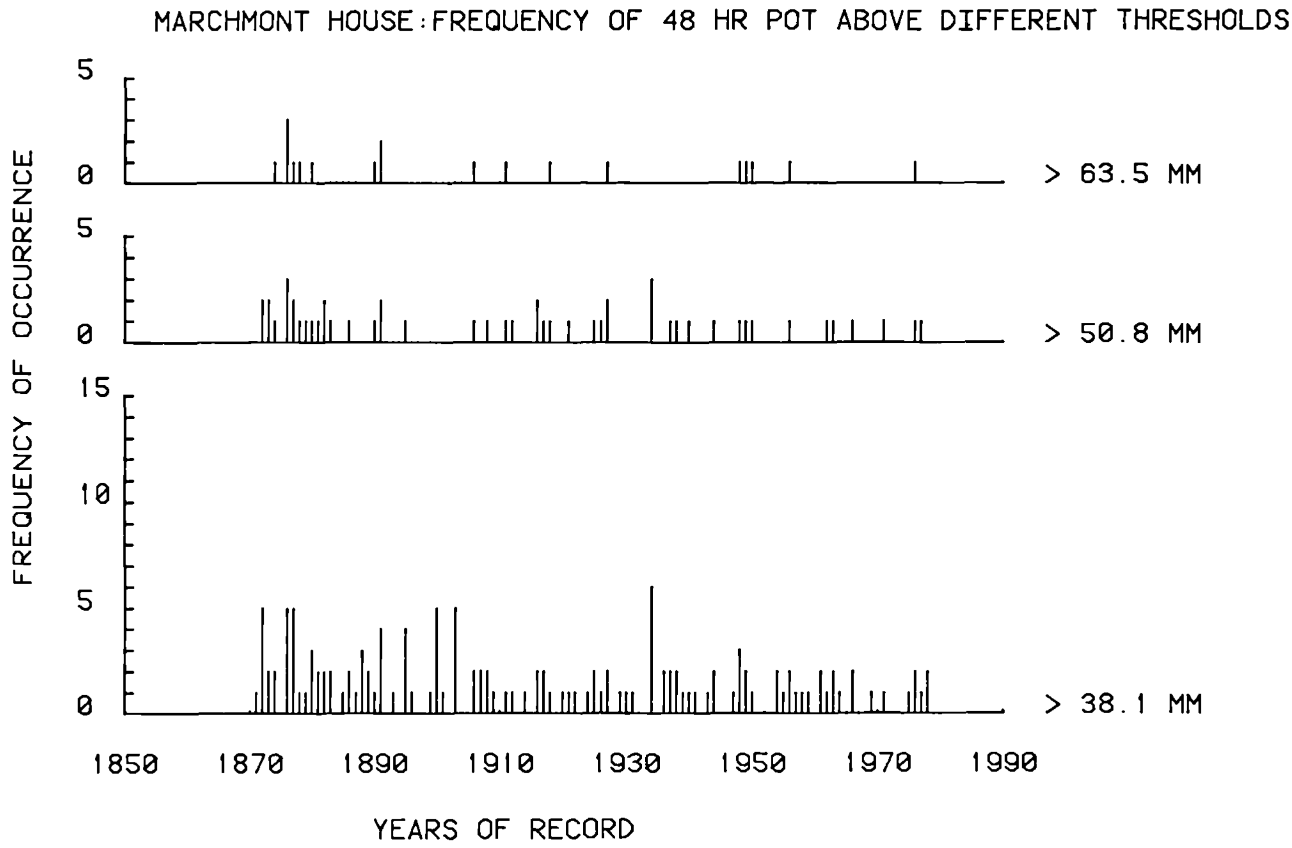
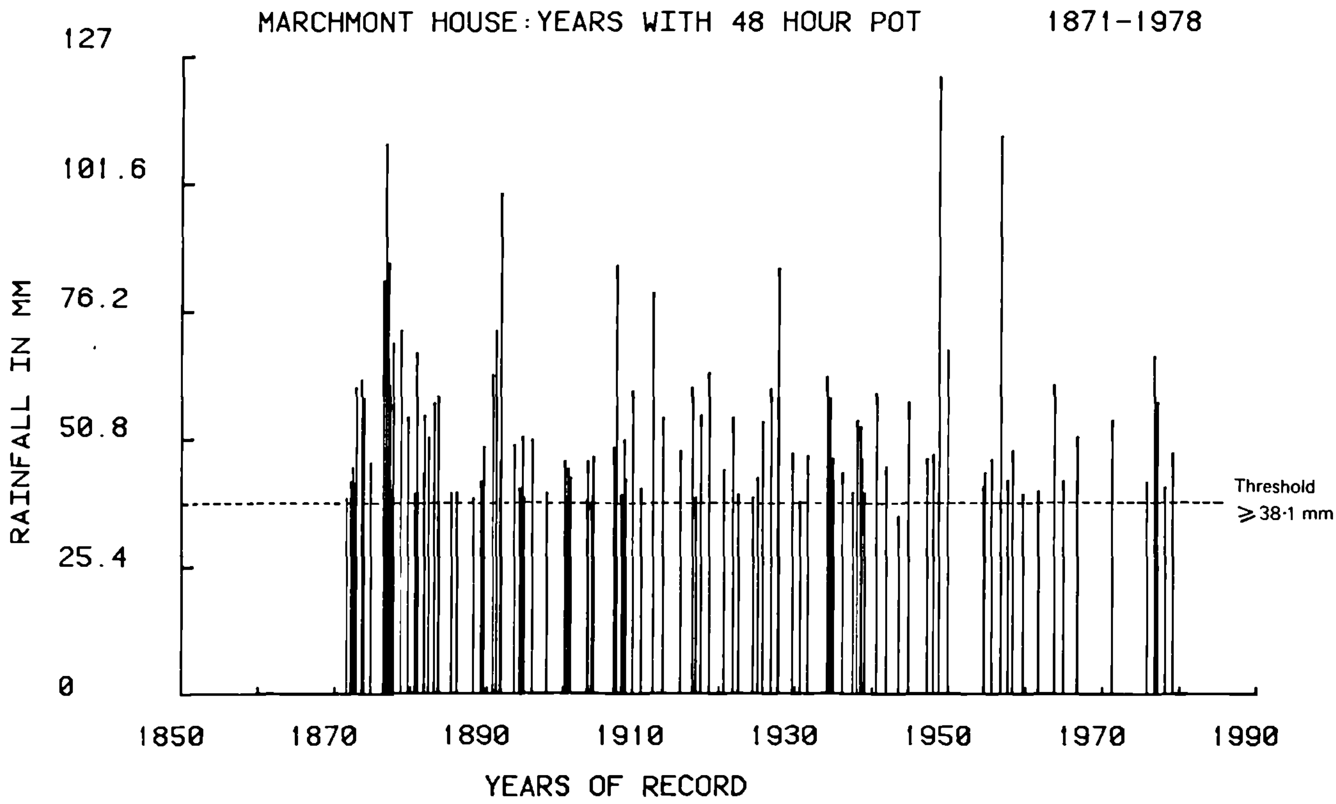


Figure 5.6.4.(xi)



terms of their frequency of POT in excess of 2 in (50.8 mm; see Figure 5.6.4.(xii)). At Cowdenknowes, the highest frequency of 48 hour POT was 3 in 1925 and 1927 (Figure 5.6.4.(xiii)).

5.6.5 Tweed study area: Seasonality of rainfall POT

Seasonality of 24 hour POT > 1 in (25.4 mm), both for the entire data sets and sub-divided by decade for Marchmont House, are shown in Figures 5.6.5.(i) to 5.6.5.(iv). Like within the Deeside study area, the highest density of 24 hour POT occurred from July to October, with much fewer large winter frontal events between November to February. This again corresponds to Bleasdale's (1961-5) diagram (Figure 5.2.(i)). 72% of the POT at Marchmont fell between June and November; 70% at Hawick. Of the 2 most extreme storms on record (1948 and 1956), it is interesting to note that both occurred in August. The exceptional nature of the 1870s was seen at Marchmont House when 23% of the 35 24 hour POT for 1870-1879 occurred in August, which supports the importance of summer frontal storms (Figure 5.6.4.(ii)). Other notable features included the period between 1900-1909, when 33% of all 24 hour POT occurred in October, suggesting this period had an increase in cyclonic events associated with the autumn equinox. Within the Leader catchment at Cowdenknowes, between 1900-1909, 21% of the 24 hour POT fell in either August or October. Again this dominance was frequently associated with summer frontal storms from the east. The frequency of the most extreme of these summer events seems to determine the return period of the most extreme discharges. Seasonality of 48 hour POT showed a similar pattern (see Figure 5.6.5.(v)).

Figure 5.6.5.(i) MARCHMONT HOUSE: SEASONALITY OF 24 HR POT: 1871-1980

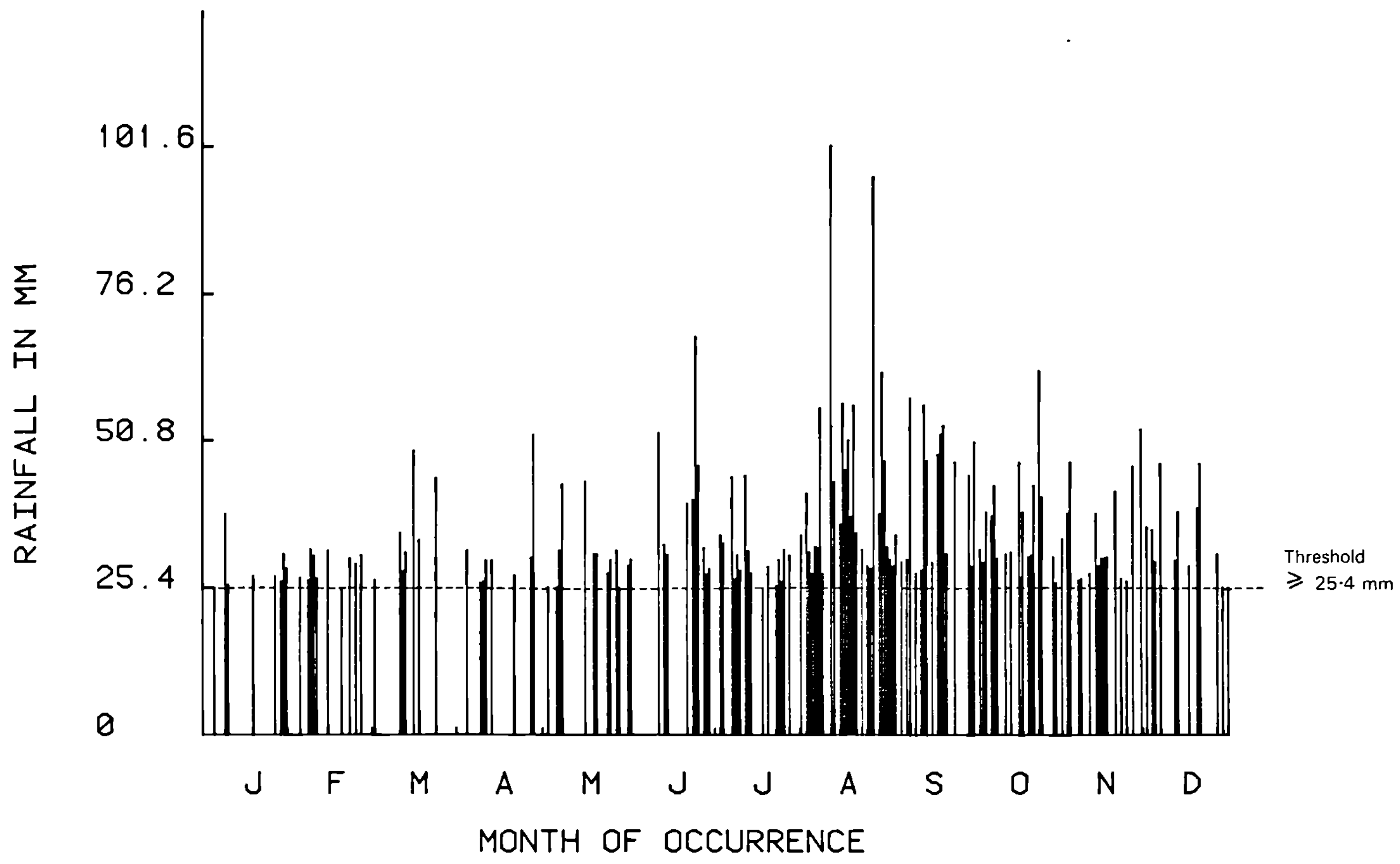


Figure 5.6.5.(ii) MANDERSTON HOUSE: SEASONALITY OF 24 HR POT: 1901-1975

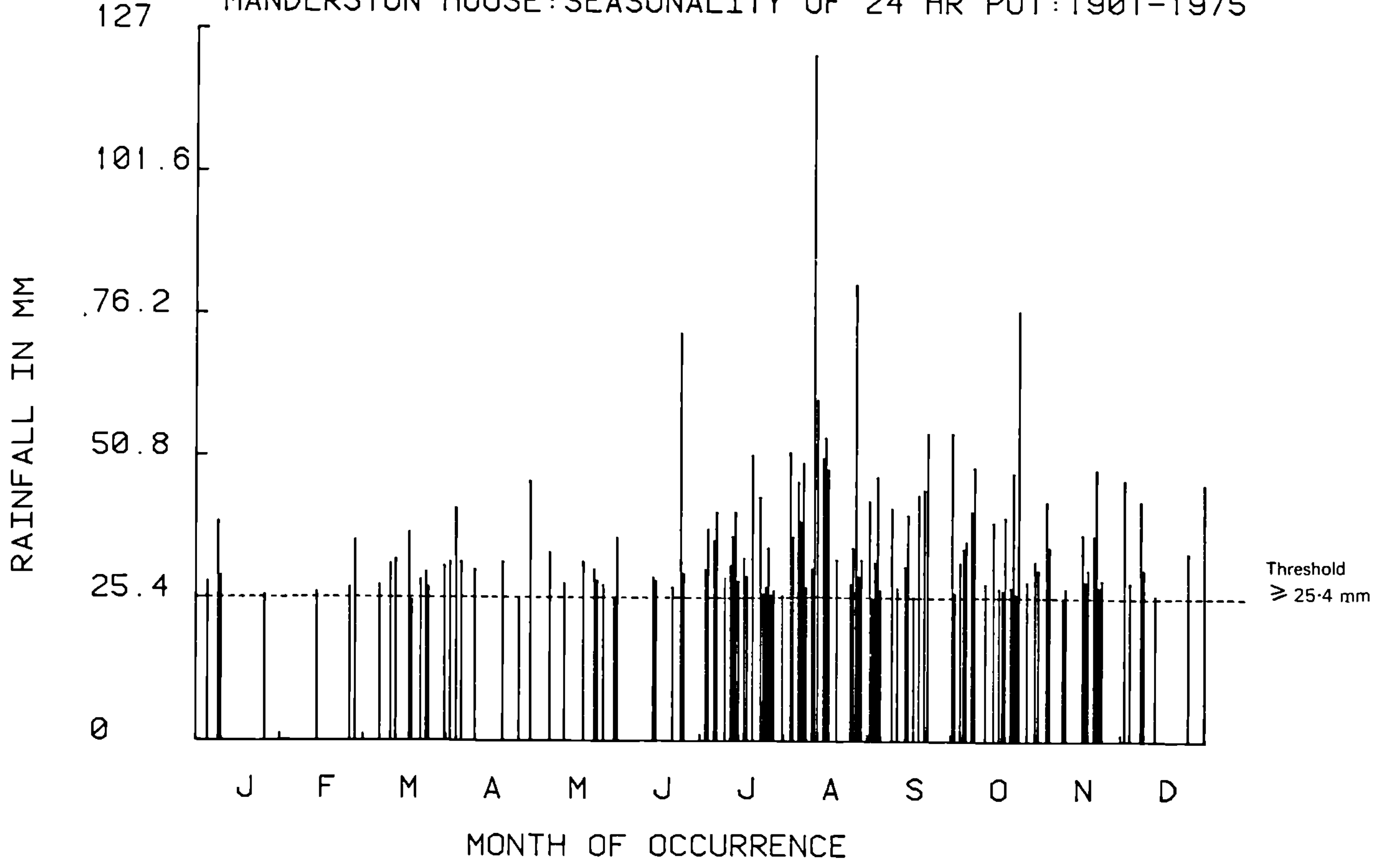


Figure 5.6.5.(iii) COWDENKNOWES: SEASONALITY OF 24 HR POT: 1898-1973

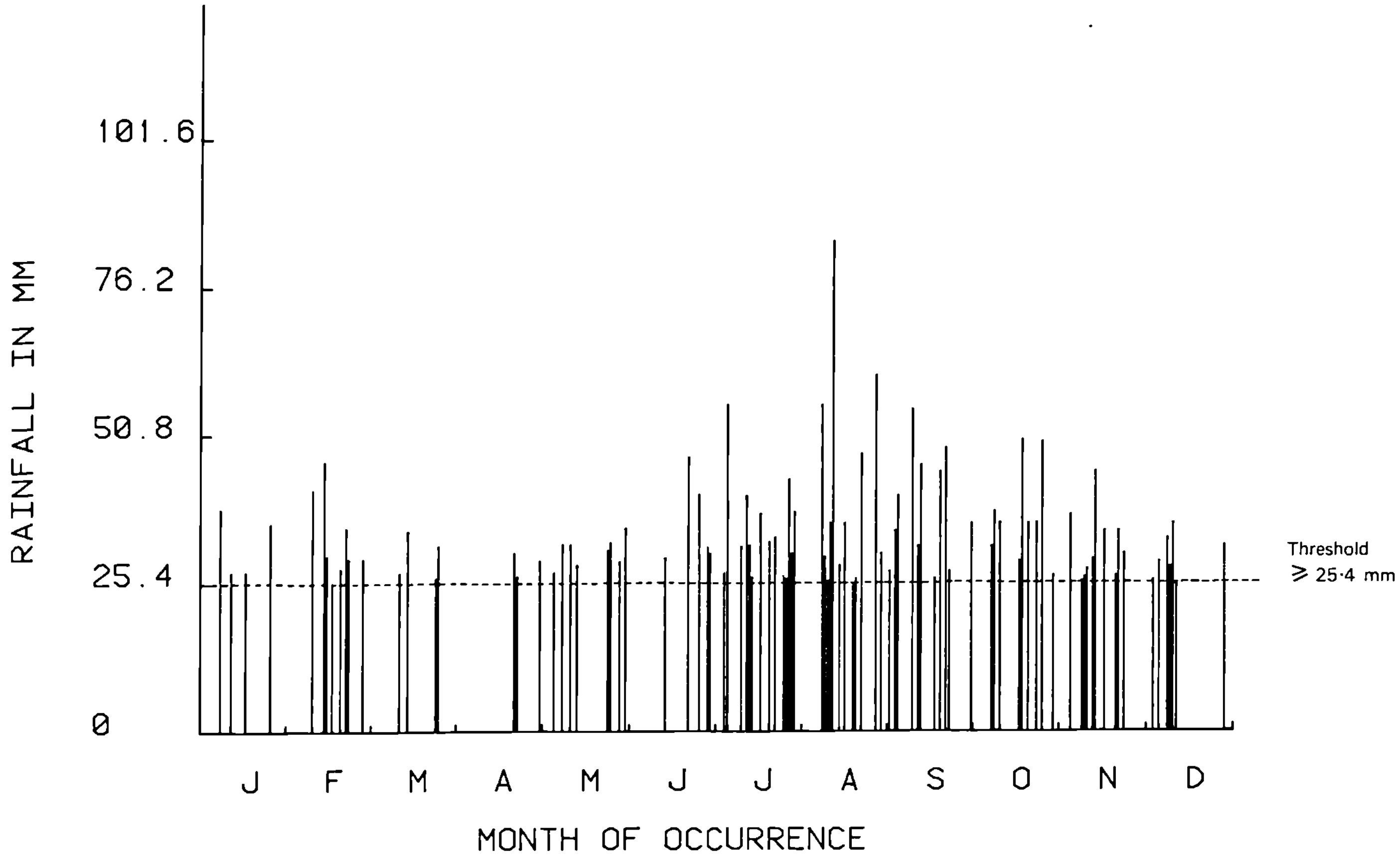


Figure 5.6.5.(iv) HAWICK (COMPOSITE): SEASONALITY OF 24 HR POT: 1921-1982

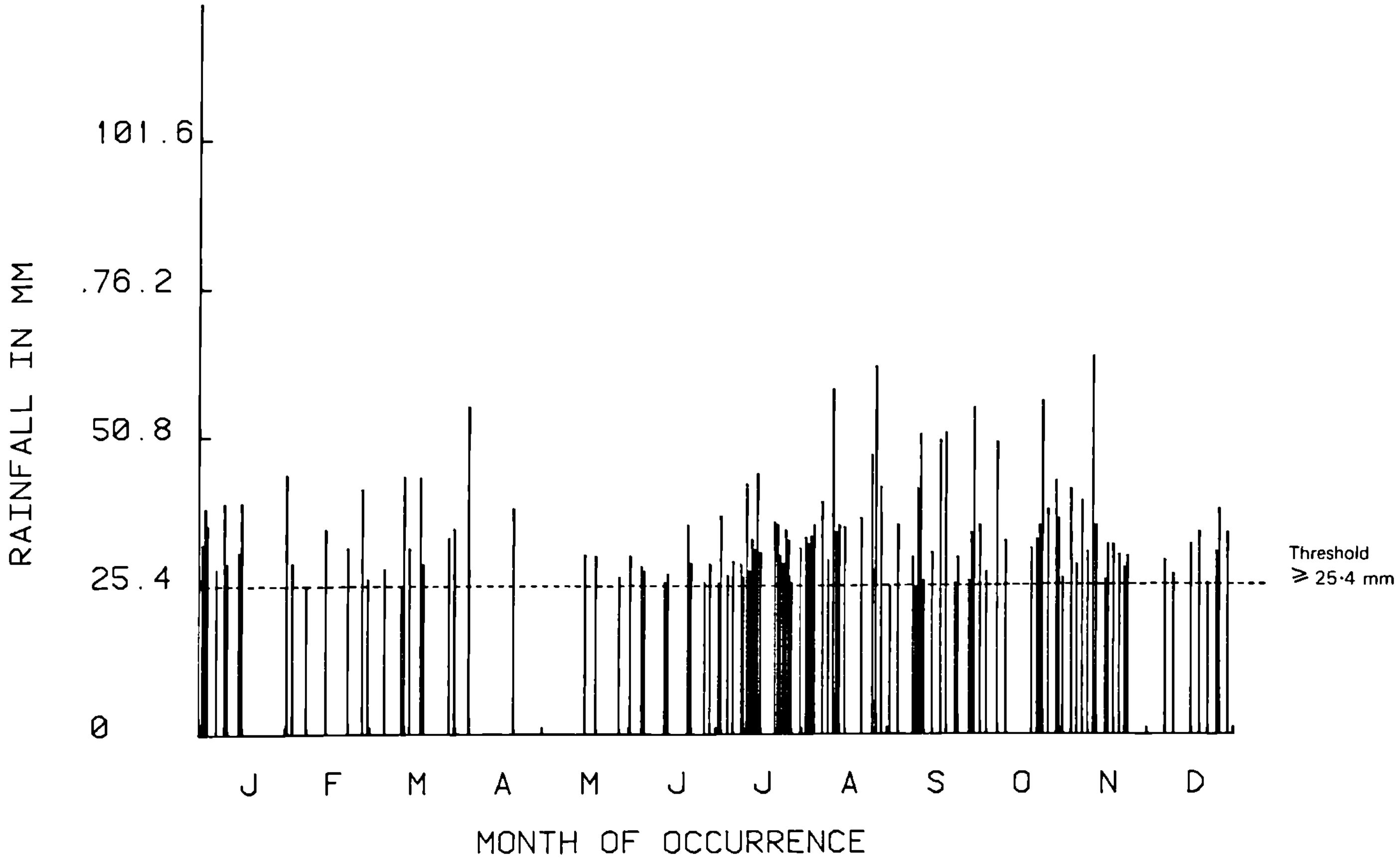
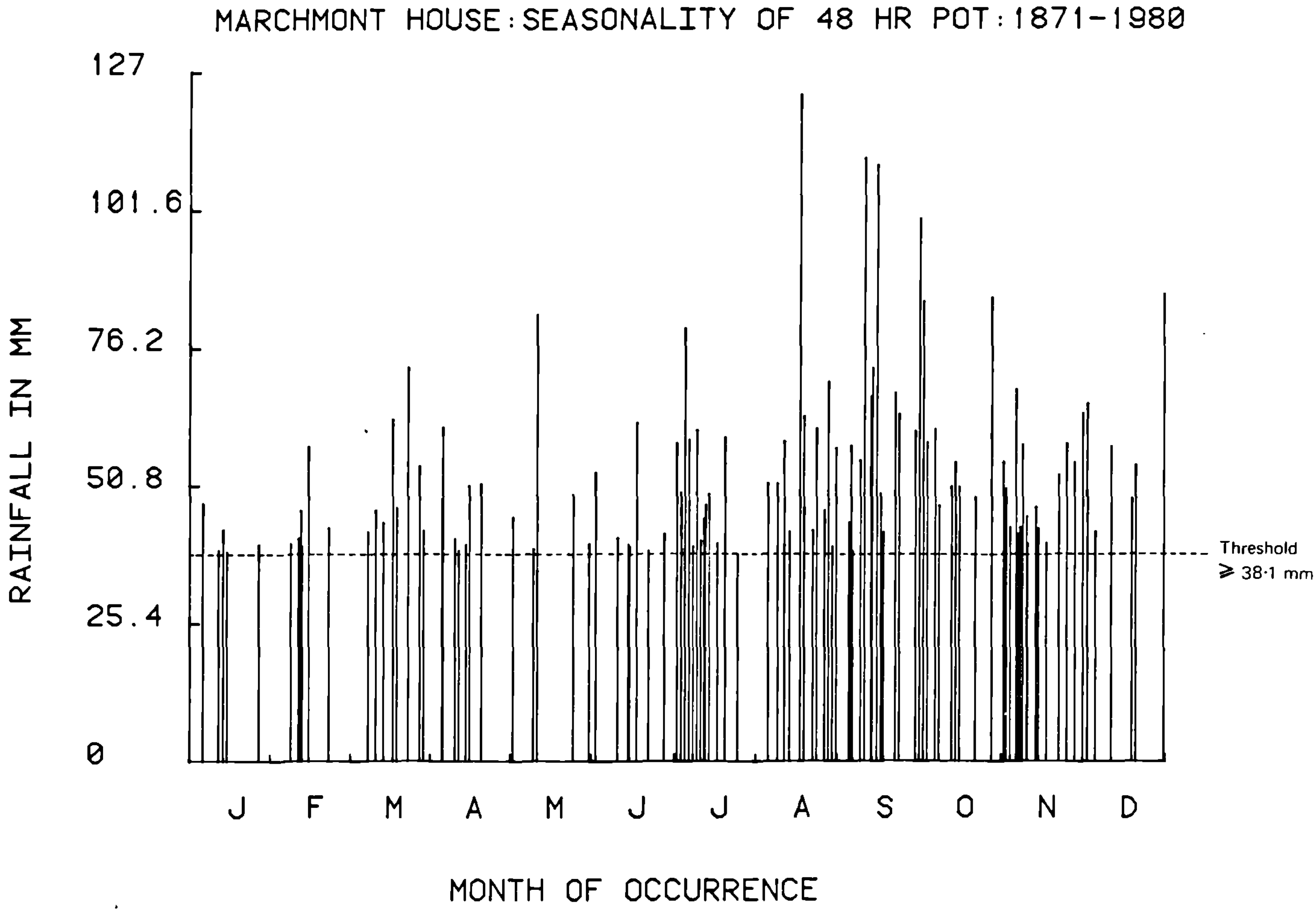


Figure 5.6.5.(v)


MARCHMONT HOUSE: SEASONALITY OF 48 HR POT: 1871-1980
RAINFALL IN MM
127
101.6
76.2
50.8
25.4
0
Threshold
≥ 38·1 mm
J F M A M J J A S O N D
MONTH OF OCCURRENCE

5.6.6 Analysis of rainfall annual maximum series

Plots of the extreme value distribution (EV1) fitted to the 24 hour annual maximum series for each station are shown in Figures 5.6.6.(i) to 5.6.6.(vii), and the rainfall magnitudes for set RI are shown in Table 5.6.6.(i). The highest annual maximum on the Marchmont House record was on 12th Aug, 1948 with 4 in (101.6 mm) and a RI well in excess of 300 years, whichever distribution was fitted (EV1 or GEV) (Figure 5.6.6.(i)). The second ranking annual maximum occurred only 8 years later on 27th Aug, 1956 with 3.79 in (96.3 mm) and again this was in excess of a 300 year RI. Another major event (ranked fourth) occurred on 25th Oct, 1949 with a 24 hour fall of 2.47 in (62.7 mm) and a RI of approximately 23 years. This 9 year period was certainly exceptional in the frequency of high magnitude annual maxima and must be compared with the 1870s-1880s, which was associated with the highest frequency of POT although the magnitude of individual falls was not so extreme.

If we could assume the EV1 distribution accurately predicted the extreme RI then the RI for the 1948 and 1956 rainfall events would be approximately 900 and 500 years respectively. If however the results from the GEV distribution were used, and here the fit only seemed to be better in relation to these two points, the RI would be approximately 300 and 200 years respectively. From the Duns Castle record, the 24 hour RI of the 1948 event as an annual maximum was confirmed with an RI of over 300 years, and the extreme nature of the 1948 event was highlighted on the Manderston record, where a fall of 4.81 in (122.2 mm) fell in one day (Figure 5.6.6.(ii)). Here, we were obviously nearing the centre of a cell of higher rainfall (see Figure 5.6.8.(iv)). The

Figure 5.6.6.(i) WHITEADDER CATCHMENT AT MARCHMONT HOUSE 1871-1980
24 hour annual maxima

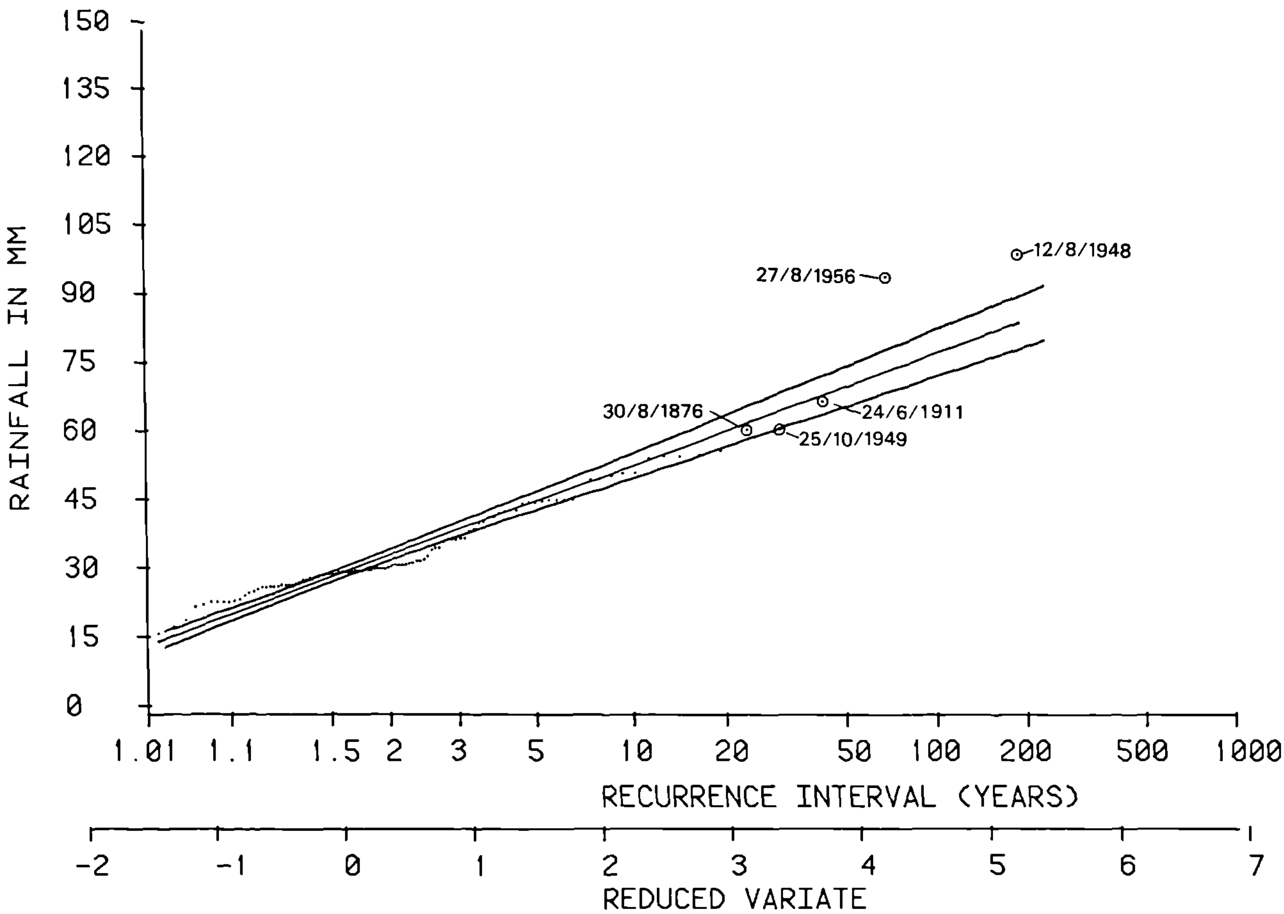


Figure 5.6.6.(ii) WHITEADDER CATCHMENT AT MANDERSTON HOUSE 1901-1975
24 hour annual maxima

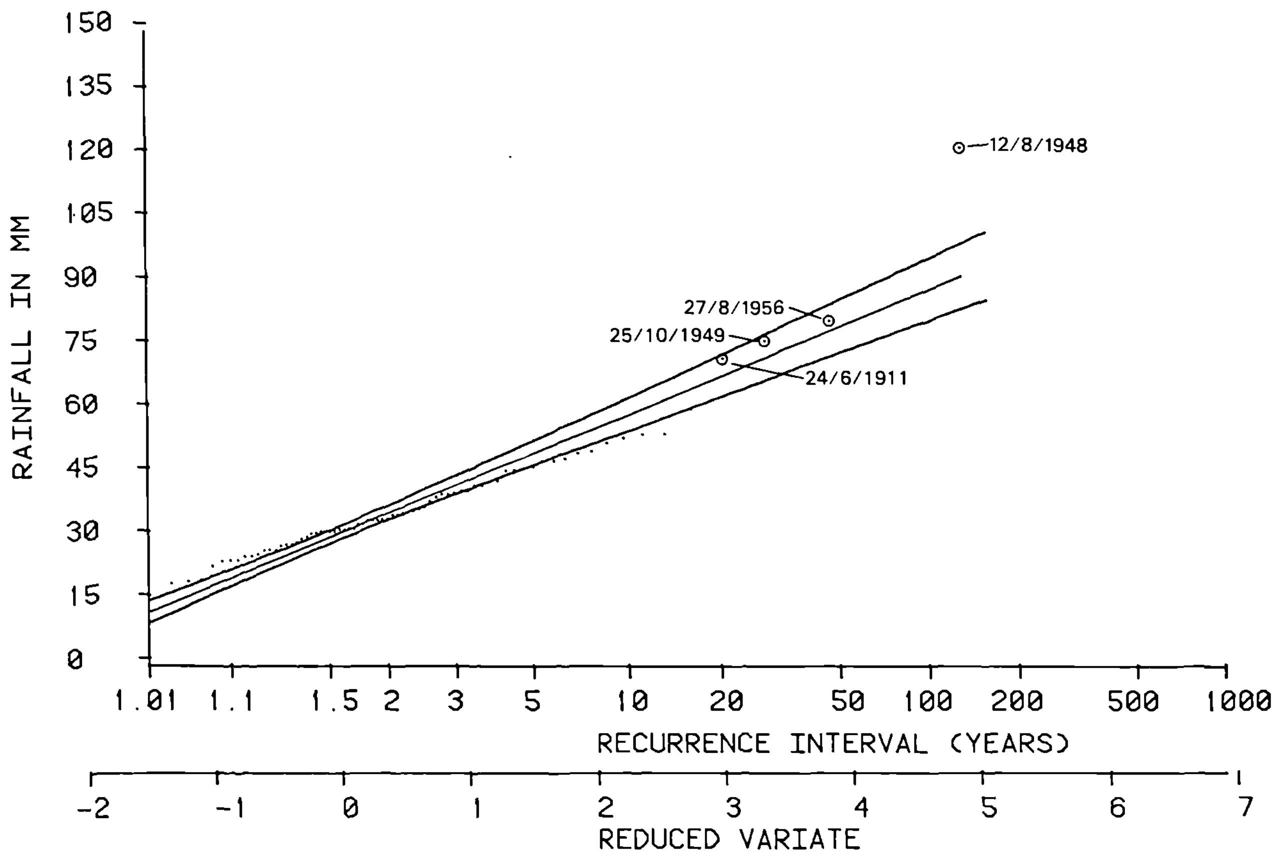


Figure 5.6.6.(iii) WHITEADDER CATCHMENT AT DUNS CASTLE 1889-1977
24 hour annual maxima

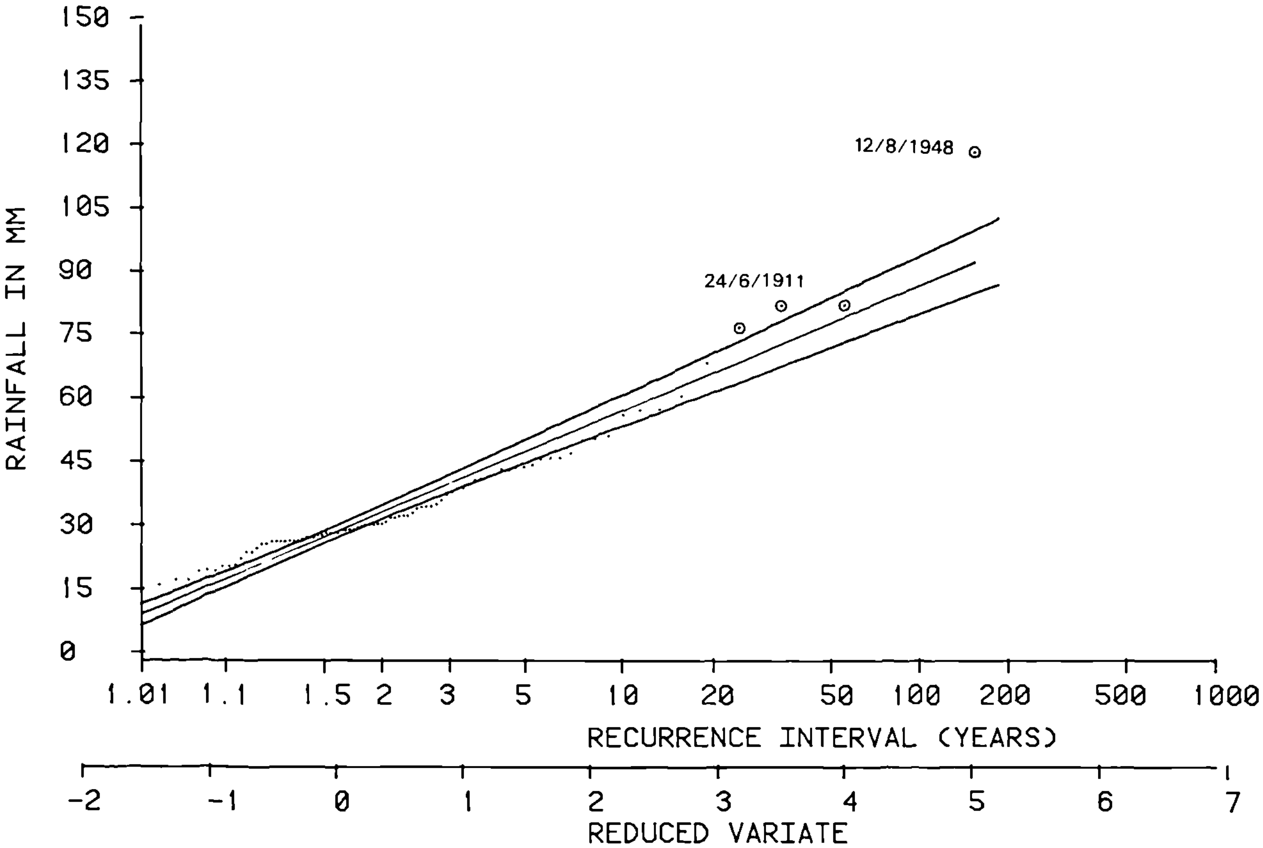


Figure 5.6.6.(iv) LEADER CATCHMENT AT COWDENKNOWES 1898-1970
24 hour annual maxima

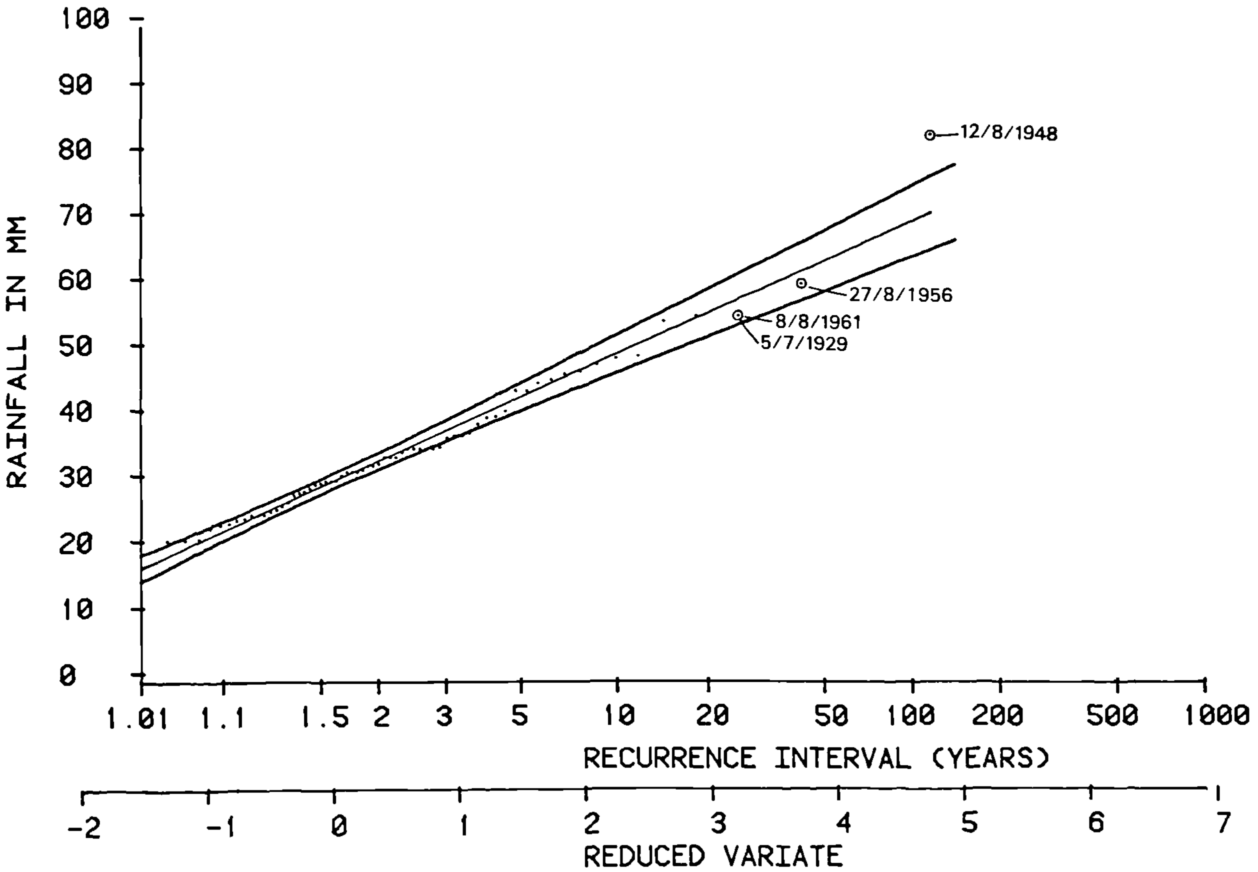


Figure 5.6.6.(v) TEVIOT CATCHMENT AT HAWICK 1921-1982
24 hour annual maxima

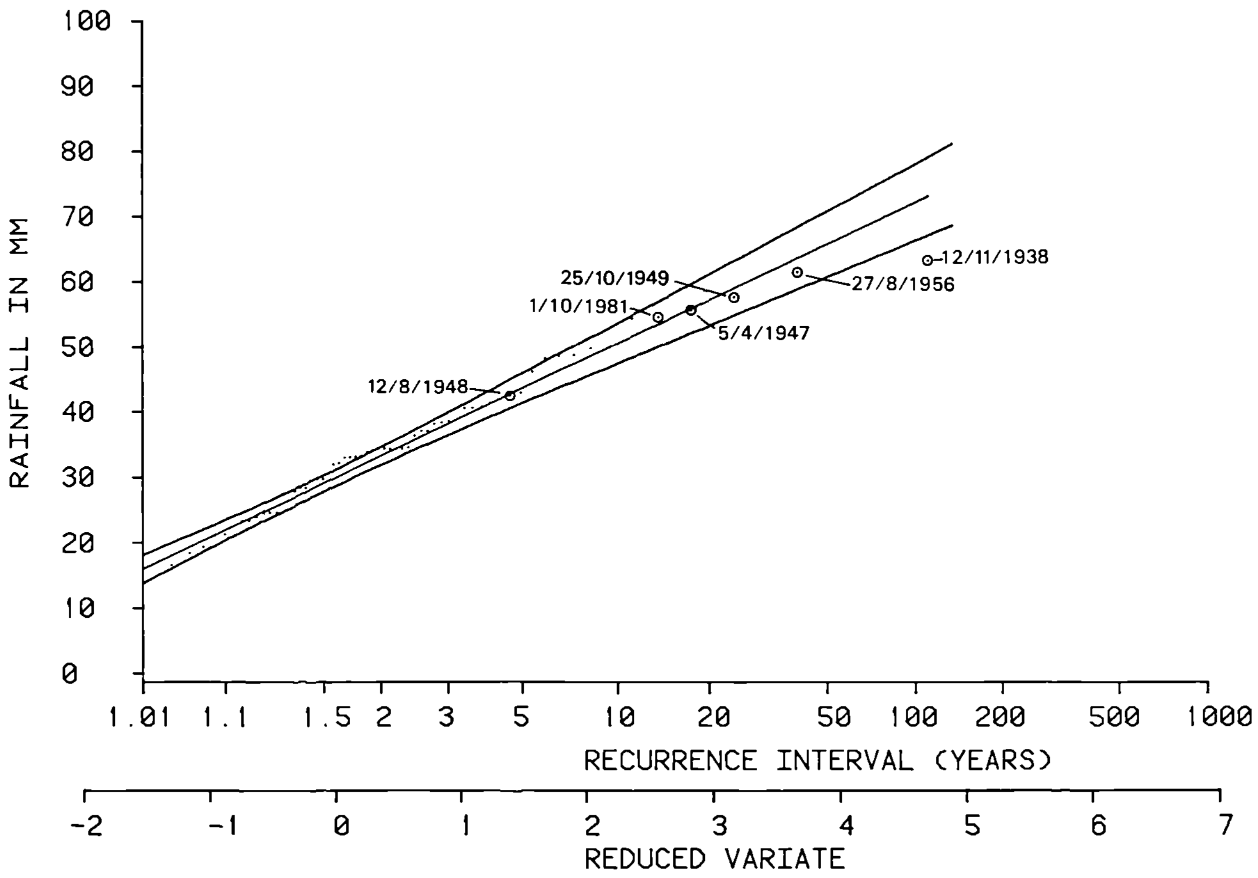


Figure 5.6.6.(vi) JED CATCHMENT AT JEDNEUK 1919-1982
24 hour annual maxima

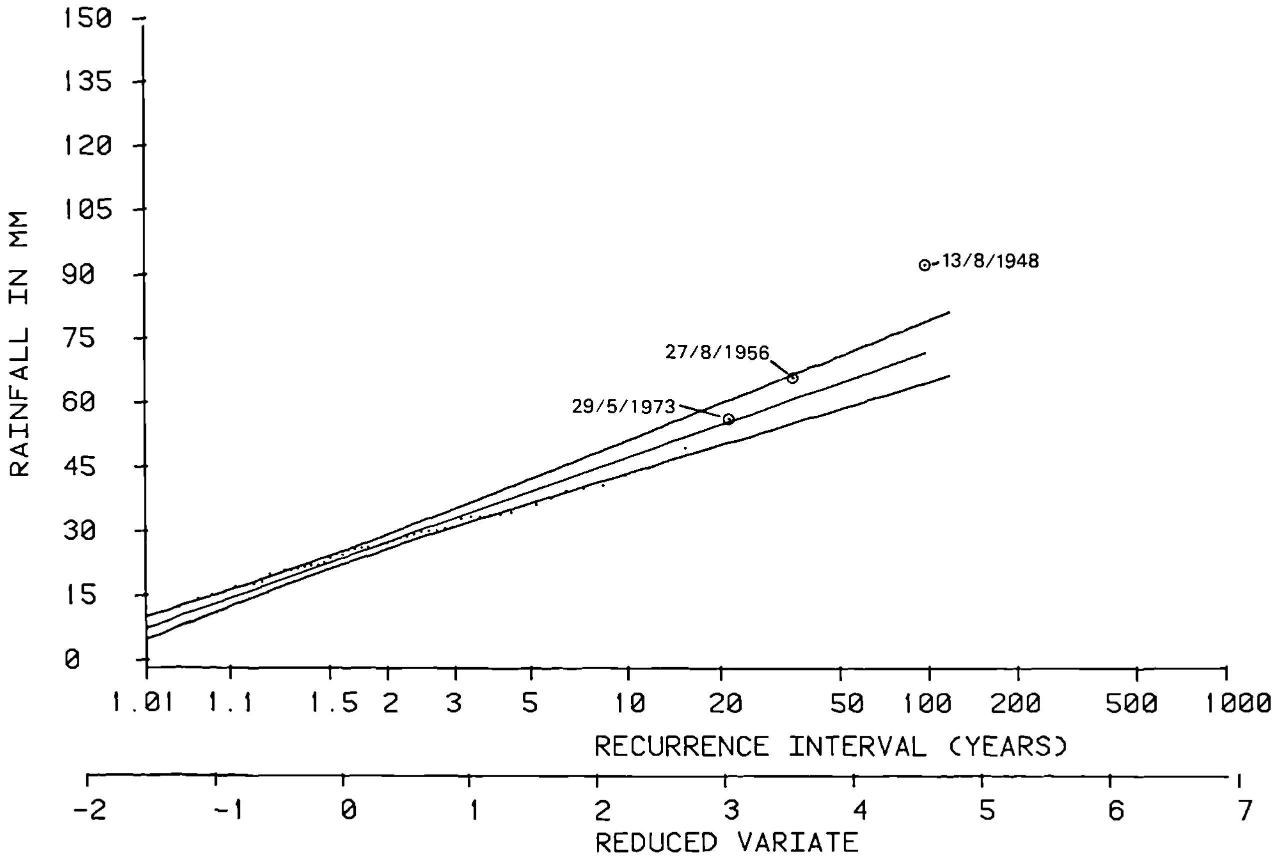


Figure 5.6.6.(vii) TEVIOT BASIN AT SILVERBUT HALL 1866-1883
24 hour annual maxima

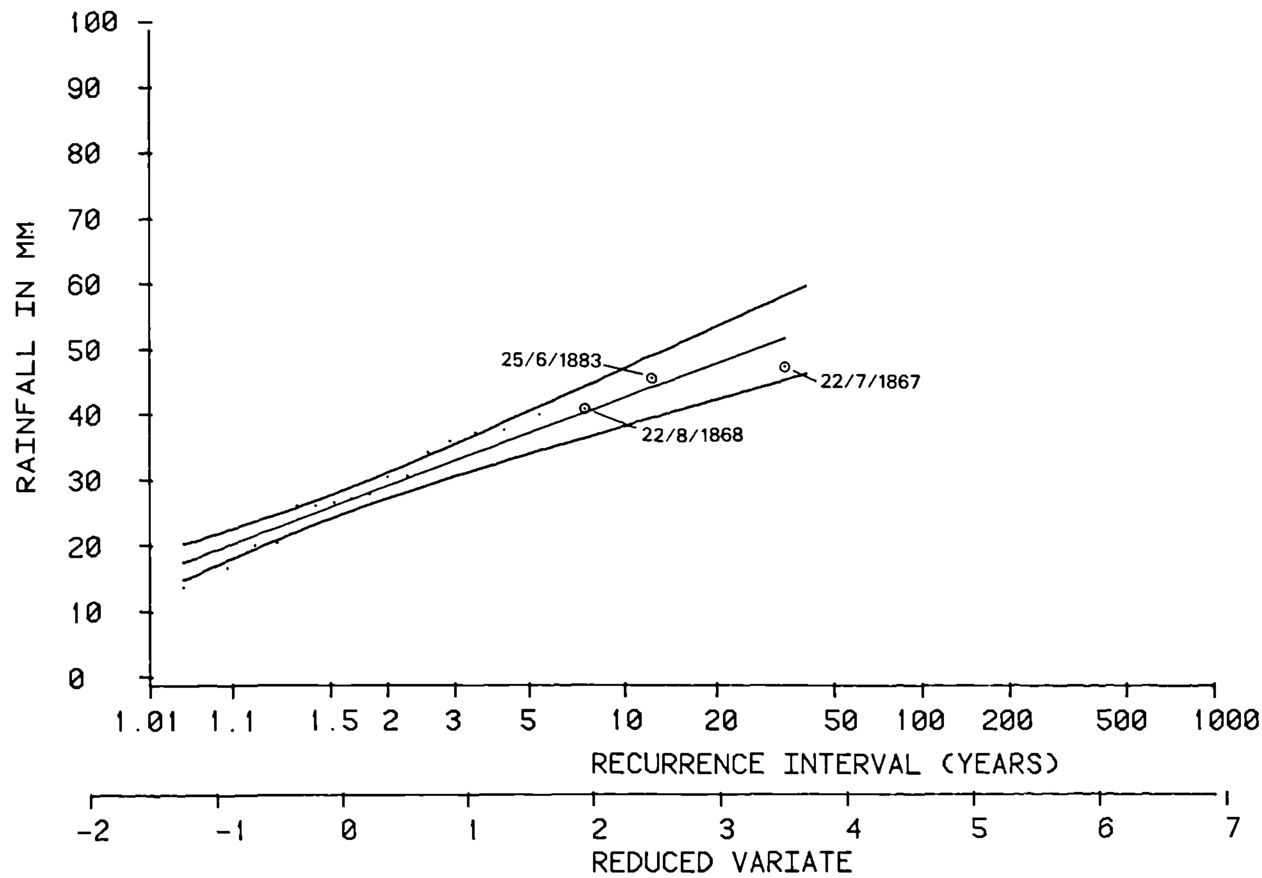


Figure 5.6.6.(viii)

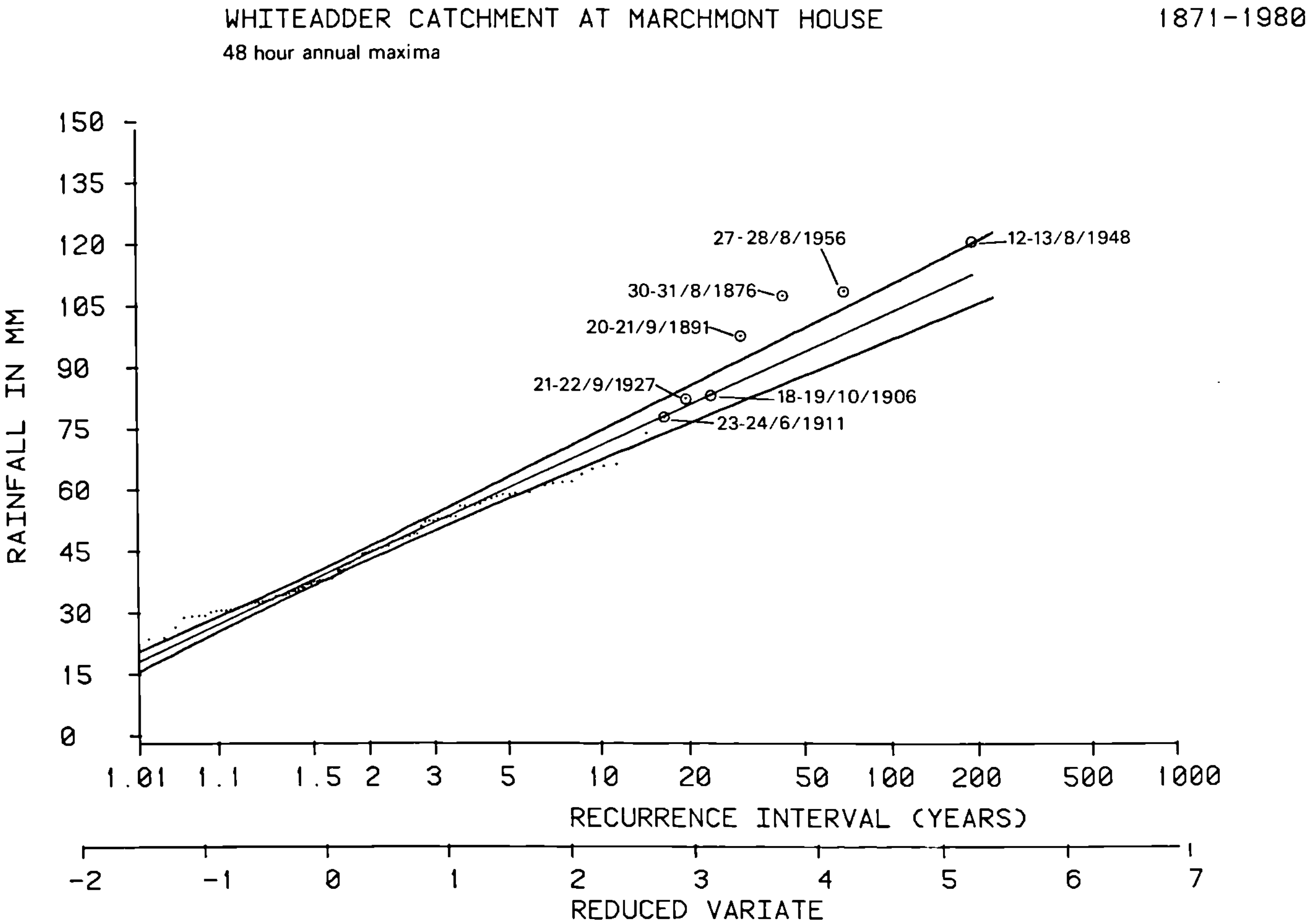


Figure 5.6.6.(x) WHITEADDER CATCHMENT AT MANDERSTON HOUSE 1901-1975
48 hour annual maxima

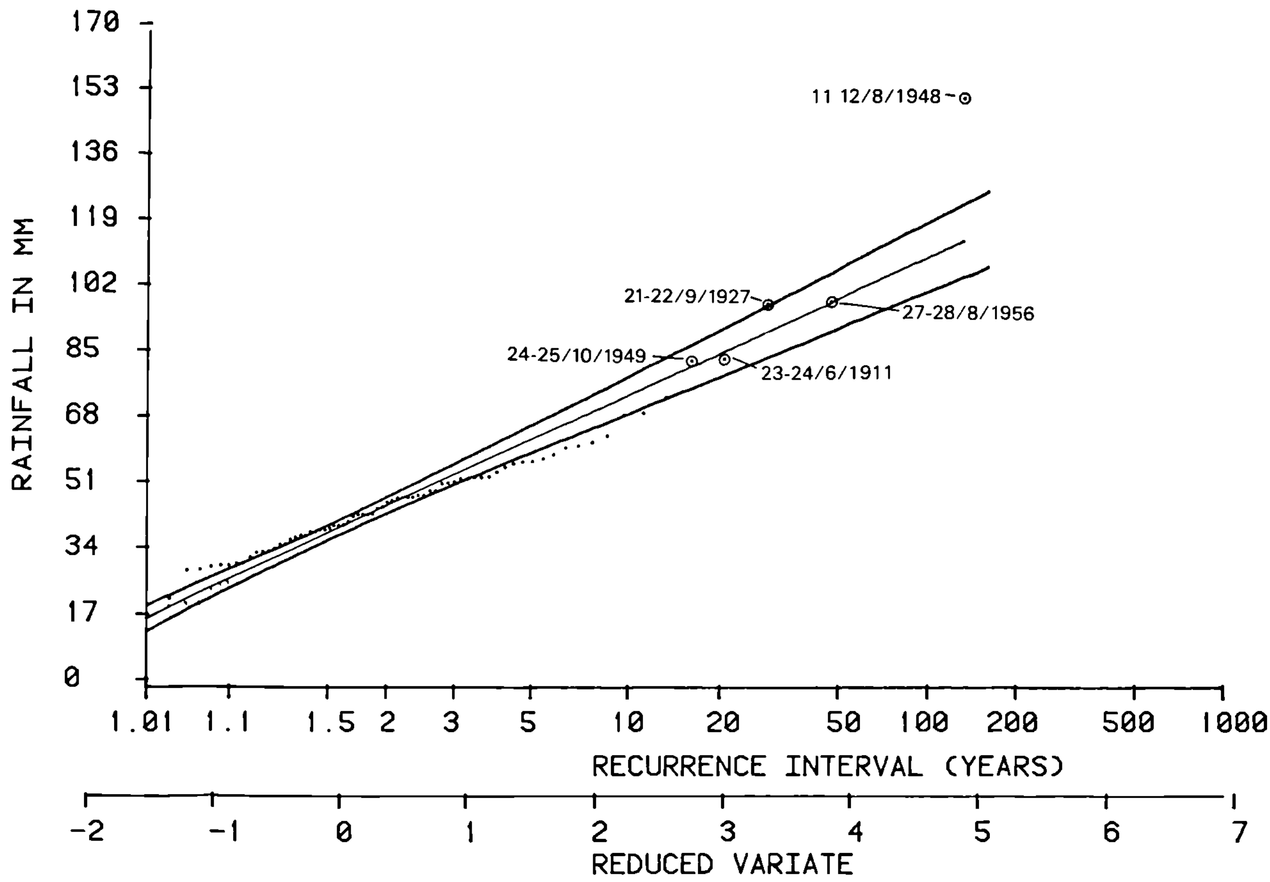


Figure 5.6.6.(ix) LEADER CATCHMENT AT COWDENKNOWES 1898-1970
48 hour annual maxima

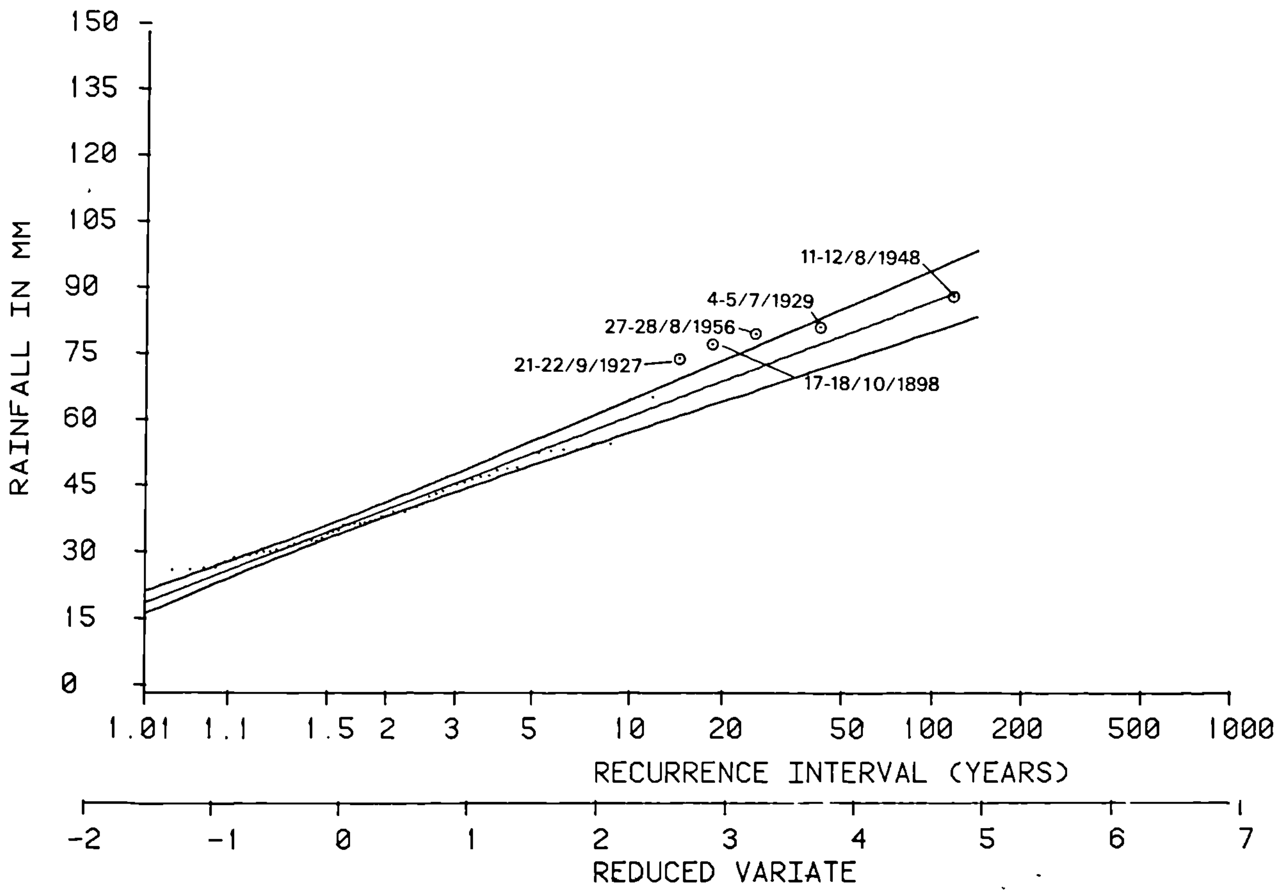


Figure 5.6.6.(xi)

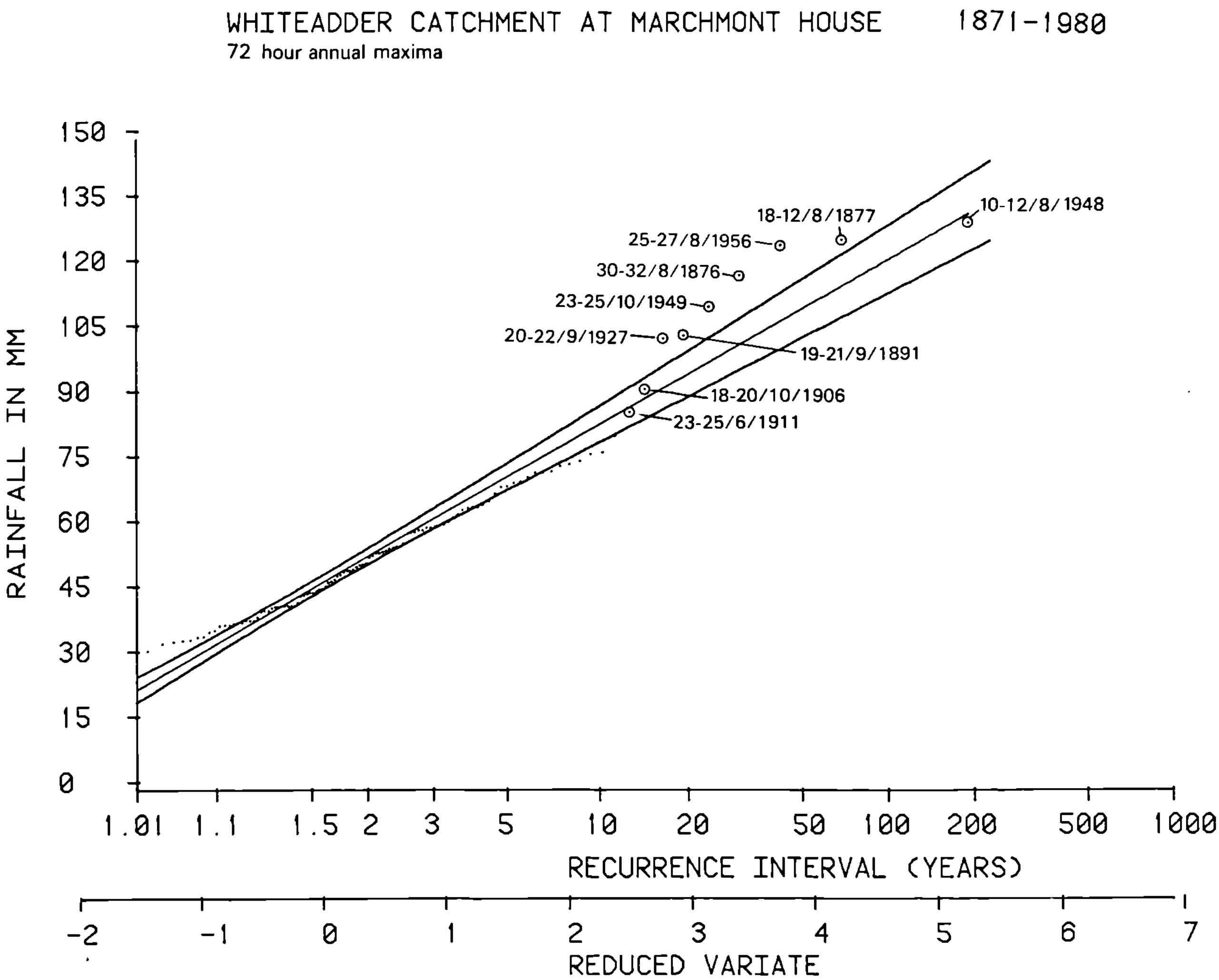


Table 5.6.6.(i)

Magnitudes of rainfall annual maxima (AM) for different recurrence intervals, fitting the EV1 distribution (and GEV) for the Tweed Stations

24 hour	AM(2)	AM(10)	AM(25)	AM(50)	AM(100)
Marchmont	37.1	54.7	64.7	72.1	79.5
				(73.1)	(82.8)
Duns Castle	37.6	58.7	70.7	79.6	88.4
				(83.0)	(97.0)
Manderston	39.0	59.4	71.0	79.6	88.2
				(79.0)	(89.3)
Hawick	40.2	55.9	64.8	71.4	77.9
				(70.9)	(77.1)
Jedneuk	31.6	49.3	59.3	66.7	74.1
				(68.0)	(77.9)
Cowdenknowes	35.6	50.1	58.3	64.4	70.5
				(66.5)	(74.5)
Silverbut Hall	32.1	44.2	51.0	56.1	61.1
				(49.1)	(50.6)

Table 5.6.6.(i) cont.

48 hour	AM(2)	AM(10)	AM(25)	AM(50)	AM(100)
Marchmont Ho.	49.6	72.7	85.8	95.5	105.1
				(102.6)	(118.5)
Manderston Ho.	50.2	75.8	90.3	101.0	111.7
				(99.1)	(110.7)
Duns Castle	49.9	75.1	89.4	100.0	110.6
				(101.1)	(114.7)
Cowdenknowes	44.2	62.9	73.5	81.4	89.2
				(79.4)	(85.1)
Jedneuk	39.2	61.4	74.0	83.4	92.7
				(86.7)	(100.1)

72 hour	AM(2)	AM(10)	AM(25)	AM(50)	AM(100)
Marchmont Ho.	57.2	84.2	99.5	110.9	122.2
				(124.0)	(146.6)
Manderston Ho	57.4	87.3	104.3	116.9	129.4
				(117.6)	(132.5)
Cowdenknowes	47.8	68.2	79.8	88.3	96.9
				(***)	(***)

AM(i) annual maximum rainfall with a recurrence interval of i years.

All values in mm

**** data unavailable

EV1 distribution did not fit this extreme point well on the Manderston record, though again the RI was well in excess of 300 years. In contrast, the 1949 event was over 1.5 in (38.1 mm) in less than 24 hours (see Figure 5.6.6.(ii)).

When the Leader at Cowdenknowes was studied, again the 1948 event was the largest on record (3.30 in; 83.8 mm) with a RI of over 300 years (Figure 5.6.6.(iv)). If the EV1 distribution was extrapolated, the actual RI was nearer 500 years. The 27th Aug, 1956 event was ranked second with an RI of over 30 years. Within the Teviot basin, when the Hawick record was studied, the highest fall was on 12th Nov, 1938 (25.4 in; 64.5mm), and this corresponds with the largest known flood event on record. In contrast to the discharge RI, the rainfall RI was however only 25 years. The SMD thus must have been very low or there must have been much higher rainfall upstream. There was notable antecedent rainfall but this was not exceptionally high (0.85 in; 21.6 mm). Other larger falls corresponded to the known major flood events of 1948, 1949 and 1956. The 1948 storm event seemed to have been less important in the Teviot valley, with 2.32 in (58.9 mm) and a RI of around 15 years. This was confirmed by Figure 5.6.8.(iv) and Glasspoole and Douglas (1949). Thus, the 1948, 1949 and 1956 events were clearly regional events rather than localised ones, though all preferentially affected different catchments. Thus, different discharge recurrence intervals were associated with different tributaries.

There was however considerable variation in the rainfall values that occurred for the same RI when the EV1 was fitted to overlapping 16 year subsets of the total annual series. The length of the subset was related to the dominant periodicity within the annual total record at Marchmont House. On the Marchmont House record (Figure 5.6.6.(xii), in terms of 24 hour annual maxima, the period with the highest peak was 1940-1956 where the RI(100) annual maxima was 4.43 in (112.5 mm). This was in contrast to 1930-1946 when the rainfall for a similar RI was 2.22 in (56.3 mm). This peak was much higher than the increase in annual maxima magnitude with RI in the 1870s-1880s. In the Duns record, there was a similar peak at 1940-1956 where the RI(100) of 4.59 in (116.7 mm) exceeded even that of Marchmont House for a similar period. There was however a secondary peak, highest from 1910-1926, with RI (100) at 3.81 in (96.8 mm). The nadir of the trough at this RI (100) was 1930-1946 with 2.18 in (55.3 mm). There was thus a period of lower annual maxima in the 1920s-1940s before the 1948 event.

The 48 hour annual maxima for set RI for each station are shown in Table 5.6.6.(i). The highest 48 hour annual maxima at the Leader and Whiteadder stations (Figures 5.6.6.(viii) to 5.6.6.(x)) were again on 12-13th Aug, 1948 with 4.83 in (122.7 mm) at Marchmont House and a RI of over 300 years (Figure 5.6.6.(viii)). The fall at Manderston was even higher with 6.02 in (152.9 mm) in 48 hours and an extrapolated RI of over 500 years (Figure 5.6.6.(ix)). The Aug, 1956 event at Marchmont House was also well in excess of a 100 year RI. The most noticeable difference between the 48 hour and 24 hour plots was the occurrence of more high annual maxima from the 1870-1879 period. This could suggest that higher longer duration events rather than high 24 hour events were typical of this time-period. At Cowdenknowes, 4-5th July, 1929 was ranked second with a fall of 3.27 in (83.1 mm) and a RI of over 40 years.

When subsets of the annual maxima data from Marchmont House were compared, the contrast between the 1870s and the 1940-1950s was much less marked when 48 hour annual maxima were studied for different RI (see Figure 5.6.6.(xii) and (xiii)). A RI(100) of 4.85 in (123.2 mm) was found for 1875-1891 while a RI(100) of 5.32 in (135.1 mm) was found for 1940-1956. There was again a considerable trough in 1920-1940. A similar pattern was found at Duns and Manderston for their overlapping time-periods of record.

At Marchmont House, the ranking of 72 hour events was fairly similar to the 48 hour duration with the Aug, 1948 event (5.14 in; 130.6 mm) still exceptional with a RI of around 150 years. Certain events however increased in RI as duration was increased. A good example was the 18-20th Aug, 1877 event, which in terms of 48 hour rainfall (2.99 in; 75.9 mm) had a RI of around 15 years; however when duration was increased to 72 hours with a fall of 4.99 in (126.7 mm), the RI increased to nearly 150 years. When the magnitudes of rainfall for different RI, using overlapping subsets of the annual maximum series, were plotted for the Marchmont House 72 hour annual maximum series, the 1870s and the late 1940-50s were of broadly comparable magnitude. The 1870s peak was much larger in comparison with the 1940-1950s than was the case with the 24 hour peaks (Figure 5.6.6.(xii)). Higher recurrence interval longer duration events were thus more typical of the earlier

Figure 5.6.6.(xii)

Magnitudes of rainfall for different R.I. using overlapping subsets of the annual maximum series Marchmont House 24 hour

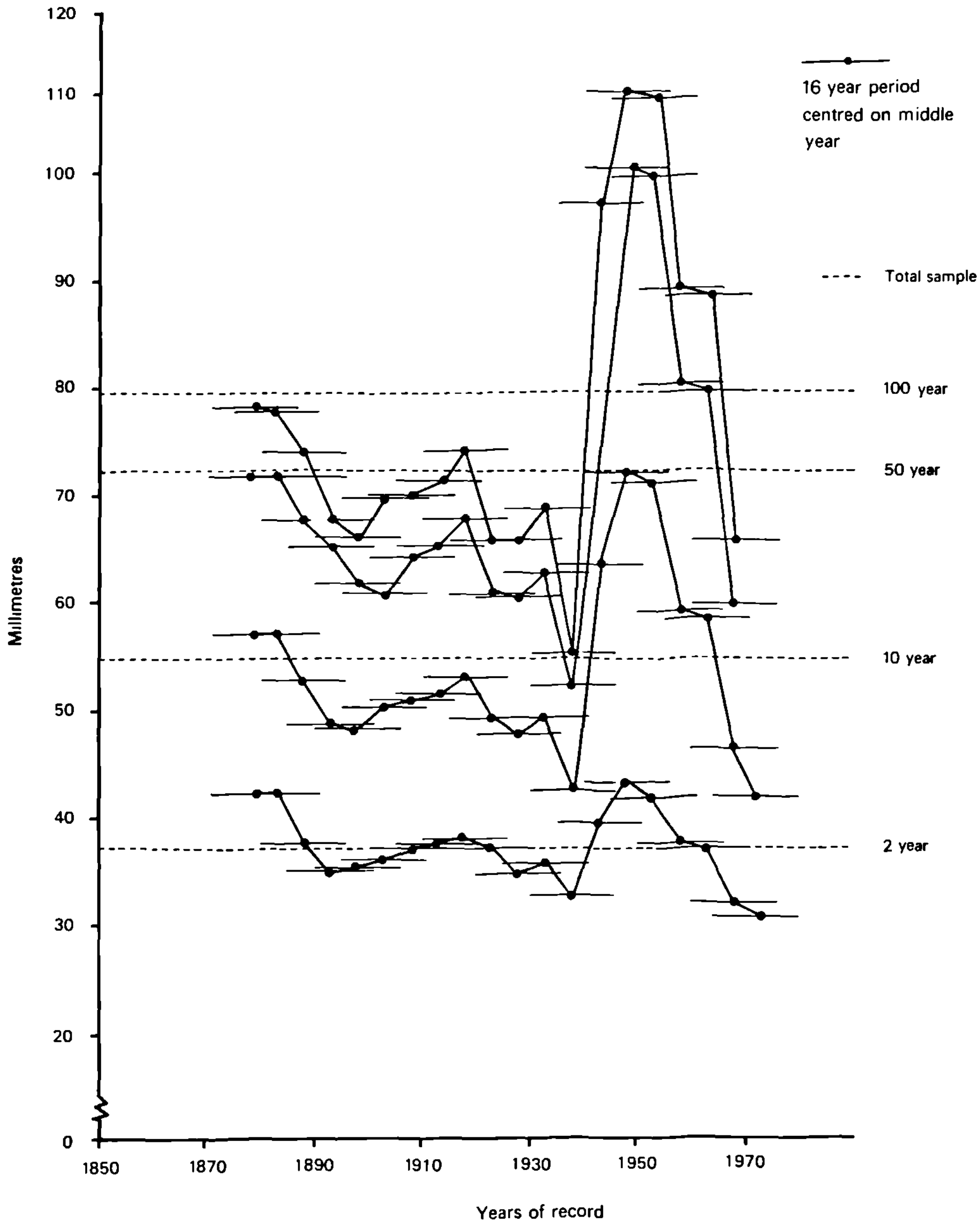


Figure 5.6.6.(xiii)

Magnitudes of rainfall for different R.I. using overlapping subsets of the annual maximum series Marchmont House 48 hour

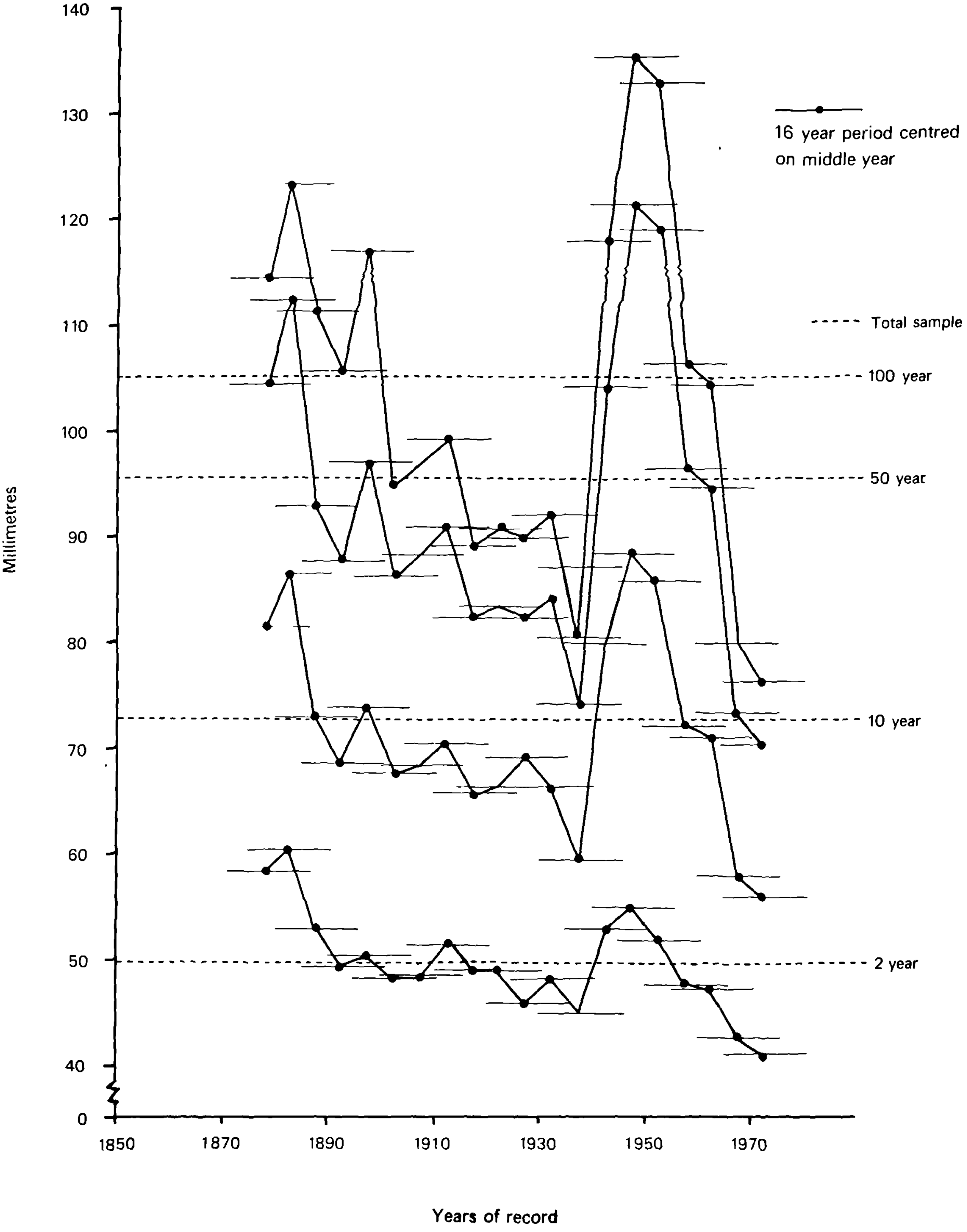
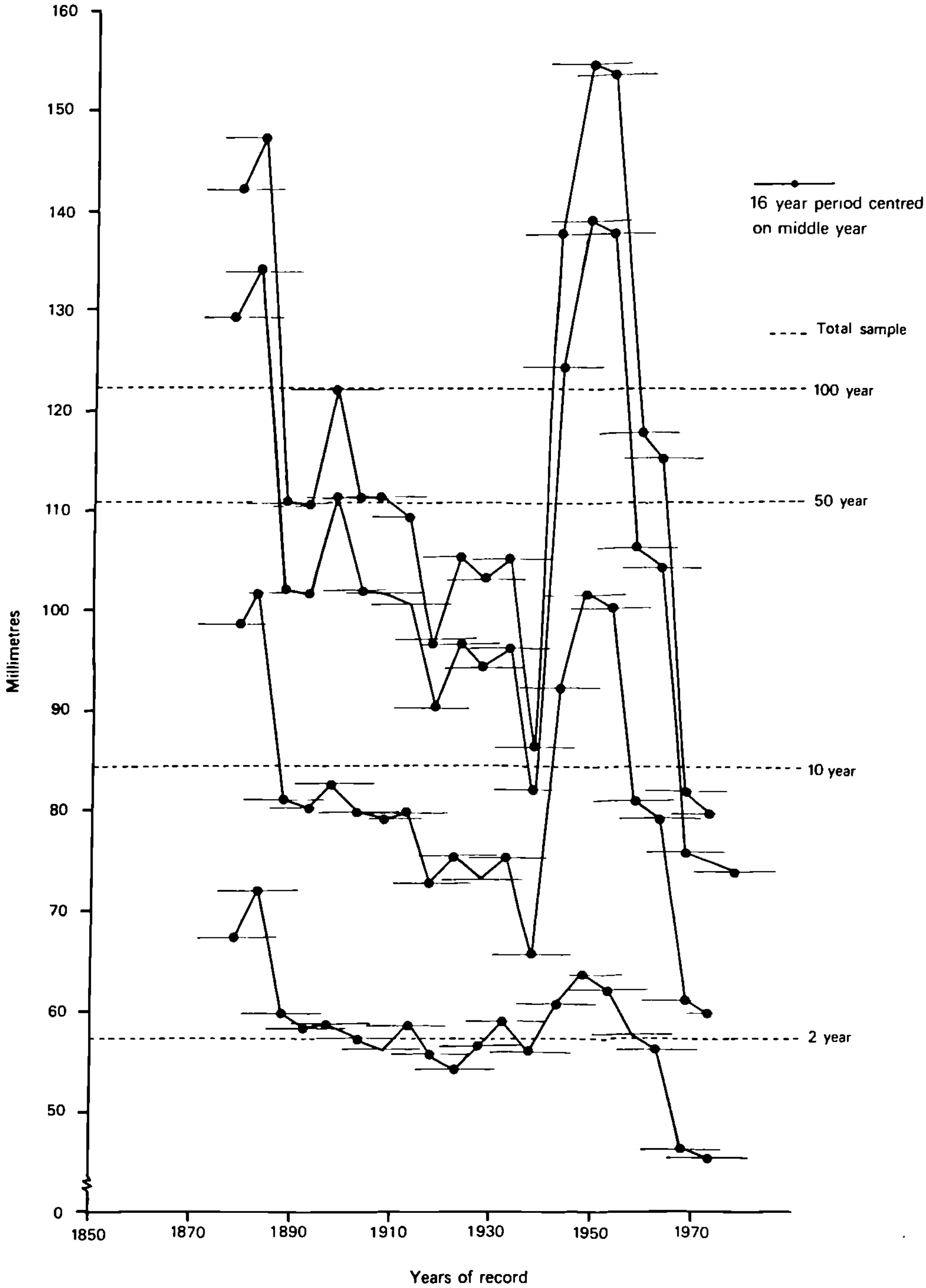


Figure 5.6.6.(xiv)

Magnitudes of rainfall for different R I using overlapping subsets of the annual maximum series Marchmont House 72 hour



period.

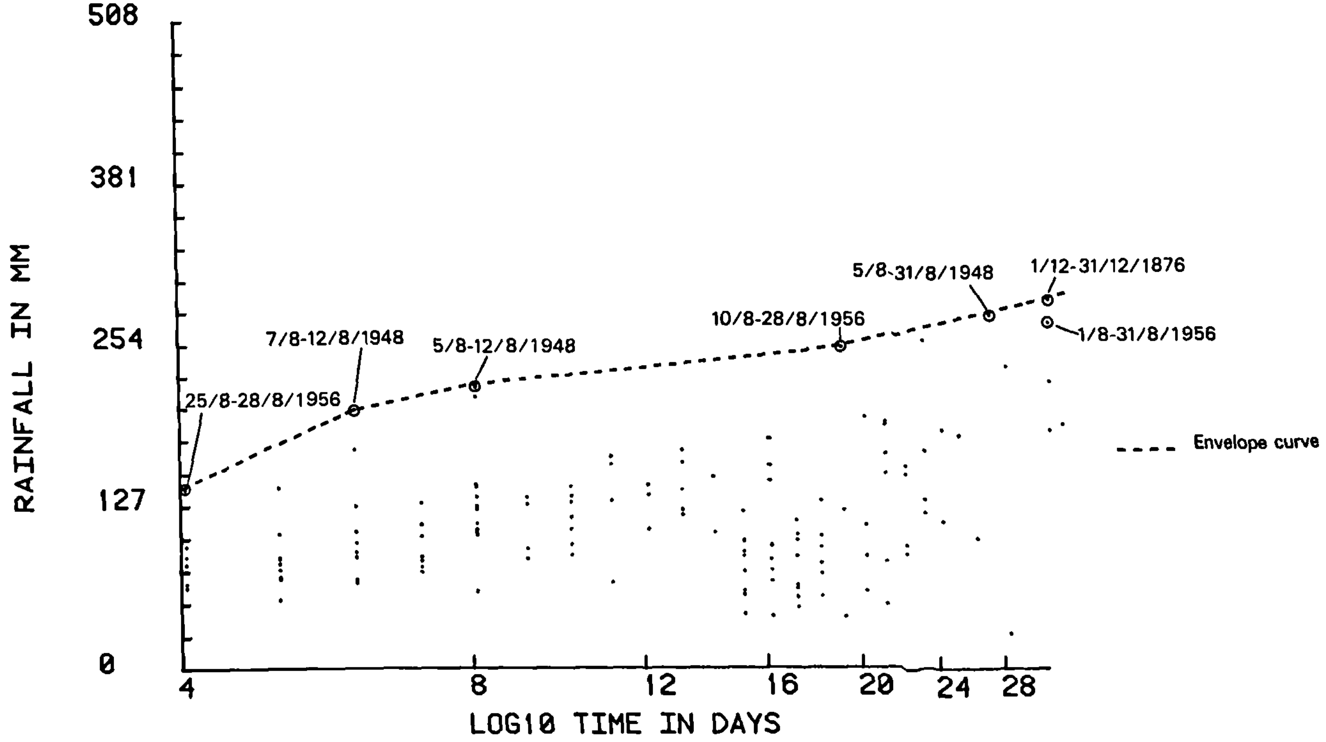
5.6.7 Tweed study area: Longer and shorter duration rainfall

The long duration (4-31 day) rainfall for each of the study stations, plus additional values obtained from British Rainfall, were plotted against $\log_{10}$ duration (Figure 5.6.7.(i)). The envelope curve was really a measure of unusually wet antecedent conditions. It is notable that both the 1948 and 1956 events were exceptional in terms of monthly rainfall as well as 24-72 hour rainfall. In fact, without these two events, the rainfall duration envelope curve would have been considerably lower. In terms of antecedent conditions, the ground clearly in both extreme runoff events must have been highly saturated and this has implications when the RI of the rainfall is compared with that of this discharge. On the Marchmont record, another major continuous rainfall period was December, 1876 when 11.42 in (290 mm) fell in 31 days. Surprisingly, from the known record, this did not seem to be associated with major flooding (as indicated by the Abbey St Bathans record). The importance of major POT within long period rainfall is very important in flood generation.

Available data on short duration rainfall and its documented geomorphic impact were collected and plotted magnitude against $\log_{10}$ time (Figure 5.4.7.(ii)). Again most of these sample sites were at low altitude and at higher altitudes higher intensities may be expected. Certain events were very rare and stood out from the rest of the data set though none approached the maximum probable rainfall (130-150 mm and

Figure 5.6.7.(i)

TWEED STUDY AREA:LONG DURATION RAINFALL : 4-31 DAYS



250-300 mm for 2 and 24 hours respectively), as estimated in the FSR (NERC, 1975). The upper limits according to this source were thus very similar to Deeside or Speyside. The provisional envelope curve was however not as steep as for the other two areas. For example, at Swinton House on 12th Aug, 1948, 5.10 in (129.5 mm) fell in 15 hours. The other major recorded fall within the Tweed basin was in Galashiels on 10th Aug, 1901 with 3.98 in (101.1 mm) in 7 hours (see Figure 5.4.7.(ii)). Frequently, these falls have caused large flood events in large catchments suggesting some of these high falls persisted over an extensive area, the 1948 event being an extreme example.

In addition to such quantitative information, there are also historic accounts of intense rainfall and its geomorphic effectiveness. For example on 5th Aug, 1767, there is record that an exceptional flood took place on the Slitrig, the tributary which joins the Teviot at Hawick:

> "There had been no great fall of rain that day, and the Teviot maintained its ordinary level; but the Slitrig rose with alarming rapidity....The stream began to rise about 4 o'clock and by 6, it had risen 22 feet [6.71 m] above its ordinary level... the inundation was understood to have been caused by the bursting of a waterspout on Windburgh, the hill from which the Slitrig takes its rise....this flood was no solitary experience for the high hills which surround the narrow channel of the Slitrig often send down waters like a dashing torrent. Gawain Douglas [the poet] is supposed to have witnessed a similar flood 350 years before." (Oliver,

1887 p374-6)

On 29th July, 1846, a similar event occurred with the Slitrig rising to between 14-15 ft (4.3-4.6 m) above its usual level (Trans. Haw. Arch. Soc., 1946); the reports stress the intensity of the rainfall with rain in torrents, continuing to fall without intermission. Stream competence was very high, transporting large stones of about one hundredweight in weight. British Rainfall (1905) then reports from Hawick on the 27th June, 1905 that a thunderstorm occurred with heavy rain, causing a flood in the Slitrig valley and sending a huge stream of water through the Whitehope Tunnel on the Waverley route and blocking the line with debris. Thus, four intense localised thunderstorms are known to have affected this catchment in approximately 430 years, with 2 major events in 80 years.

Another incidence of intense localised rainfall within the Teviot catchment occurred from 2 to 9 pm, on 9th Aug, 1806 at Denholm, causing the spate known as the "Denholm flood".

> "The forenoon of the day was usually hot and sultry.....about noon the sky overcast and grew tremendously black...the very clouds seemed to descend in the terrific rain which immediately fell in one whole mass. The Dean Burn was swollen to a great size and rushed down its bed carrying everything along. Trees were carried down on its heaving tide almost like a moving forest..." (Trans. Haw. Arch. Soc. 1863, p12).

Denholm seems susceptible to flood inundation not from the Teviot because the level of the village is well above the bankfull level of the Teviot, but rather because of the small tributary, the Dean Burn. The Scotsman reports in July, 1983 a proposed flood protection scheme for the village as a result of a freak storm which caused serious damage two weeks previously and also the problem of repeated floods in Denholm. This flooding was caused by rainfall of an intensity experienced only twice a century. Again here is an area with a history of at least 2 intense rainstorms in approximately 180 years.

On the Whiteadder, near Abbey St. Bathans, The Scotsman also reports that on 3rd July, 1830 a very localised thunderstorm occurred. It was "the most severe and destructive than any ever known to have happened in that district." Such localised events only maintain high runoff rates when they are centred over smaller catchments and are less important over larger catchments. Their existence is however frequently documented, and the occurrence of similar events over the same catchment appears to occur relatively frequently cf. Deeside and Speyside.

5.6.8 Tweed study area: Storm case studies

The following major storm events were studied in the Tweed study area, both in terms of their peak intensity and also their antecedent conditions.

(1) 20th Sept, 1891

(2) 24th June, 1911

(3) 12th Aug, 1948

(4) 25th Oct, 1949

(5) 27th Aug, 1956

(1) 20th Sept, 1891

The synoptic situation for the 1891 event was a severe gale of wind and rain from the N.N.E., with 3.40 in (86.4 mm) of rain falling in 36 hours (British Rainfall 1891; see Table 5.6.8.(i)). The storm duration was 36-48 hours and the cells of greatest rainfall are shown in Figure 5.6.8.(i). The catchments most affected were the Whiteadder and the Leader with over 4 in (101.6 mm) rainfall, for example 118.9 mm in 48 hours at Duns Castle (175 year RI). The main centre of the storm was however in the Pentland Hills rather than centred over the Tweed catchment. If the centre of high rainfall had been further east, a much larger discharge event would have occurred within the Tweed study area.

(2) 24th June, 1911

The 1911 event was a frontal storm and this time the rainfall event was dominantly 24 hours in duration, with the highest recorded value being 3.30 in (83.8 mm) in 24 hours at Duns castle (RI of 73 years; see Table 5.6.8.(ii)). Falls of 2 in (50.8 mm) in 24 hours were recorded at gauges near the mainstream Tweed eg. Kelso (Springwood) with 2.01 in (51.1 mm). The general pattern of 24 hour rainfall is shown in Figure 5.6.8.(ii), with highest rainfall centred over the Bowmont and

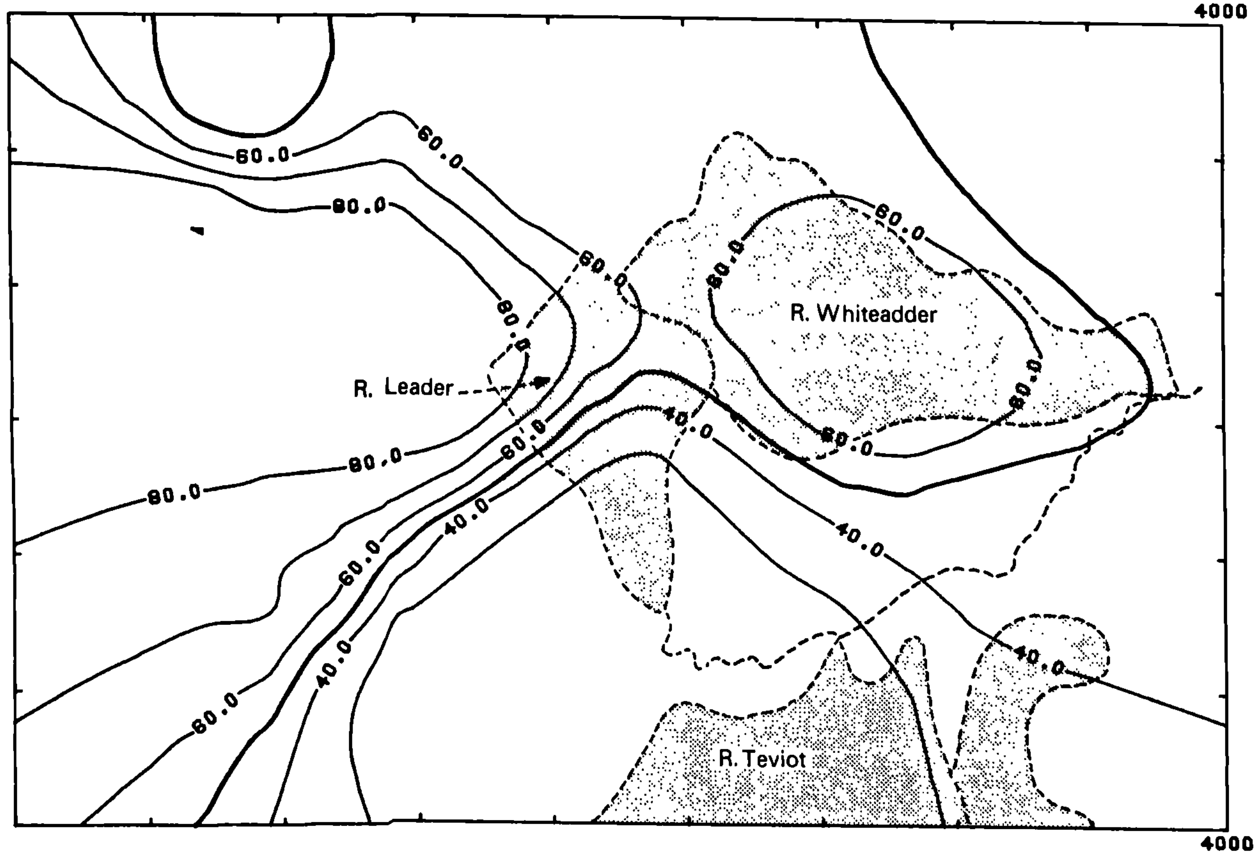
Table 5.6.8.(i)

Rainfall recurrence interval for the 9/1891 flood event *

Station	24 hour (20/9/1891)	48 hour (20-21/9/1891)	72 hour (20-22/9/1891)
Marchmont Ho.	51.8 mm (9 yr)	99.8 mm (73 yr)	104.6 mm (36 yr)
Duns Castle	70.1 mm (25 yr)	118.9 mm (175 yr)	119.1 mm (78 yr)

Figure 5.6.8.(i)

Rainfall 20th September, 1891 Tweed study area



RAINFALL 20TH SEPT. 1891 TWEED STUDY AREA
PLOT NO. 1 DATE 03/11/85 TIME 19.43.40

Table 5.6.8.(ii)

Rainfall recurrence intervals for the 6/1911 flood event *

Station	24 hour (24/6/1911)	48 hour (24-25/6/1911)	72 hour (24-26/6/1911)
Marchmont Ho.	68.8 mm (39 yr)	79.8 mm (18 yr)	86.9 mm (13 yr)
Duns Castle	83.8 mm (73 yr)	95.8 mm (40 yr)	99.1 mm (24 yr)
Manderston	72.6 mm (30 yr)	85.3 mm + (20 yr)	92.5 mm ++ (15 yr)
Cowdenknowes	14.2 mm (< 1 yr)	28.4 mm (< 1 yr)	35.1 mm (< 1 yr)

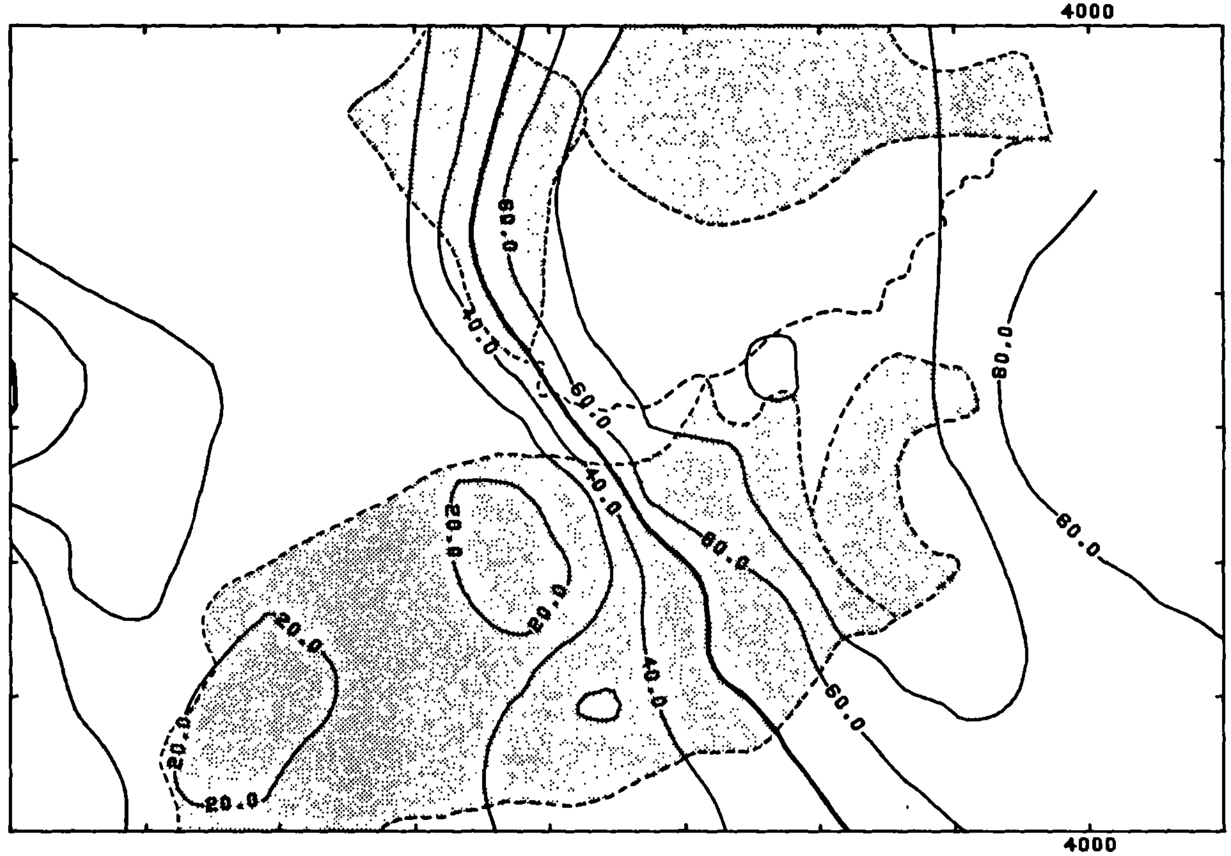
+ occurred 23-24/6/1911

++ occurred 23-25/6/1911

* as an annual maximum

Figure 5.6.8.(ii)

Rainfall 24th June, 1911   Tweed study area



RAINFALL 24TH JUNE, 1911 TWEED STUDY AREA
PLOT NO. 1   DATE 03/11/85   TIME 19.37.59

Whiteadder catchments, thus accounting for the major flooding on the Whiteadder (Figure 5.6.2.(i)). (The 1910 flood reported by Holford (1976) and the TRPB may therefore be an error in date).

(3) 12th Aug. 1948

According to British Rainfall (1948), and the Monthly Weather Report of the Meteorological Office (1948), the synoptic situation was as follows:

> "A shallow depression moved eastwards across the Midlands and the southern north sea on August 11th. Heavy thundery rain fell continuously during that night in the north of England and throughout the 12th in S.E. Scotland." (British Rainfall, 1948).

Total area within the Tweed valley covered by these falls, exceeding specific thresholds in 24 hours, are shown in Table 5.6.8.(iv). This exceptional rainfall was due to the simultaneous occurrence of a combination of favourable circumstances with a very low probability (Meteorol. Office Report, 1978b). Firstly, there was the development of a slow moving trough of low pressure, N.N.W. from the centre of the depression. Secondly, this trough extended to a considerable height (probably around 10370 m) and was accompanied by strong low-level convergence and ascent of moist air, which had become unstable due to the over-running of cold air at higher levels. Thirdly, the resultant instability added further to the rainfall intensity as did the orographic uplift of the strong N.E. air flow by the high ground.

Table 5.6.8.(iii)

Rainfall recurrence intervals for the 8/1948 event *

Station	24 hour (12/8/1948)	48 hour (12-13/8/1948)	72 hour (12-14/8/1948)
Marchmont Ho.	101.6 mm	122.7 mm	130.6 mm
	(300-500 yr)	(300 yr)	(150 yr)
Duns Castle	120.1 mm	145.0 mm	150.4 mm
	(300-500 yr)	(300-500 yr)	(300 yr)
Manderston	122.1 mm	152.9 mm	162.1 mm
	(300-500 yr)	(300-500 yr)	(300-500 yr)
Cowdenknowes	83.8 mm	90.2 mm	90.2 mm
	(300+ yr)	(115 yr)	(61 yr)
Hawick	49.5 mm	(-----)	(-----)
	(7 yr)	(-----)	(-----)
Jedburgh	94.5 mm +	112.3 mm	(-----)
	(300+ yr)	(300+ yr)	(-----)

+ occurred 13/8/1948

* as an annual maximum

------ data unavailable

Table 5.6.8.(iv)

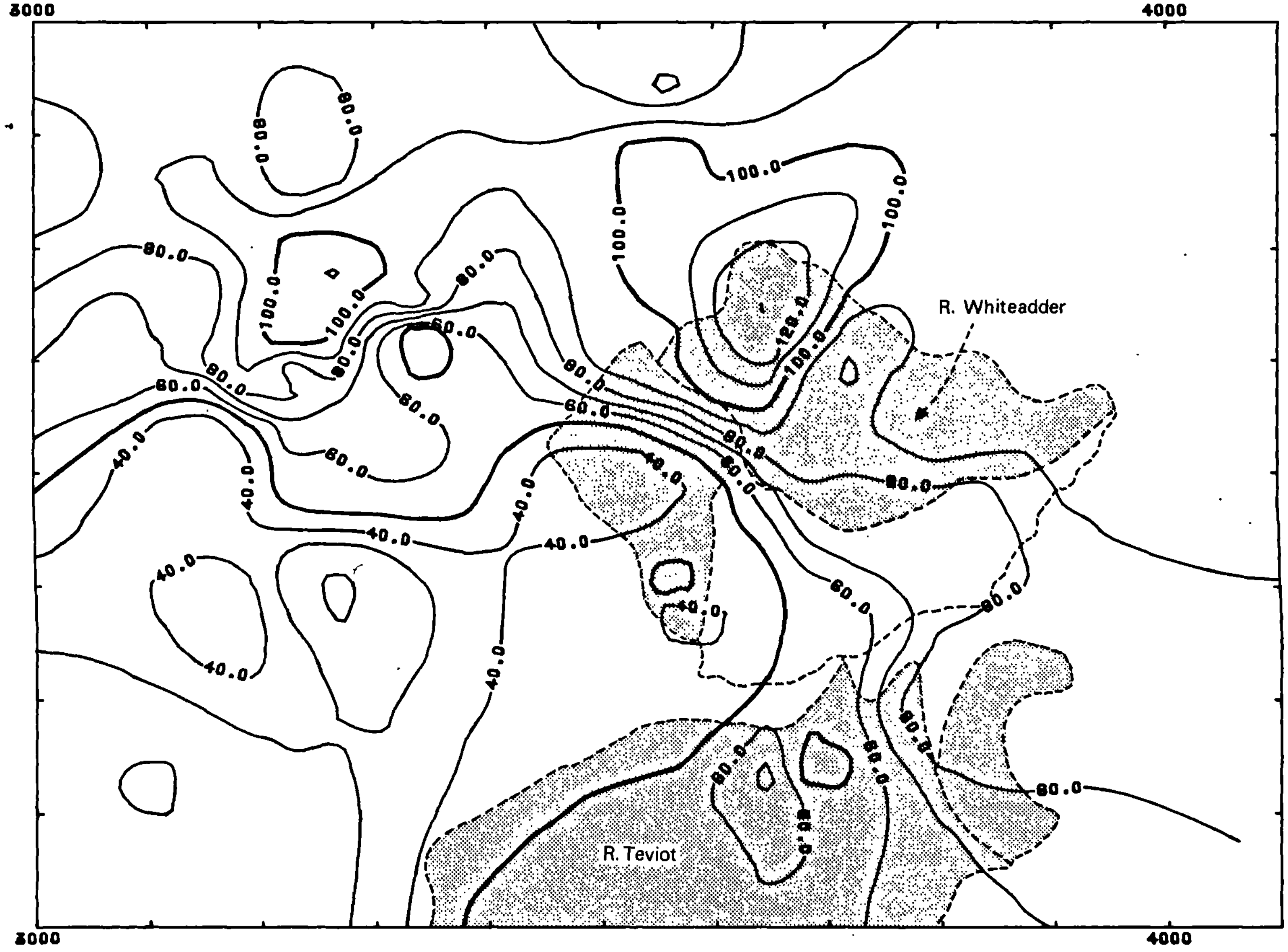
The areal extent of the 12th Aug. 1948 rainfall above given thresholds

Threshold	Area (km^2)
>6 in (152.4 mm)	13
>5 in (127.0 mm)	209
>4 in (101.6 mm)	1287
>3 in (76.2 mm)	2454
>2 in (50.8 mm)	4666

(Adapted from British Rainfall, 1948)

Figure 5.6.8.(iii)

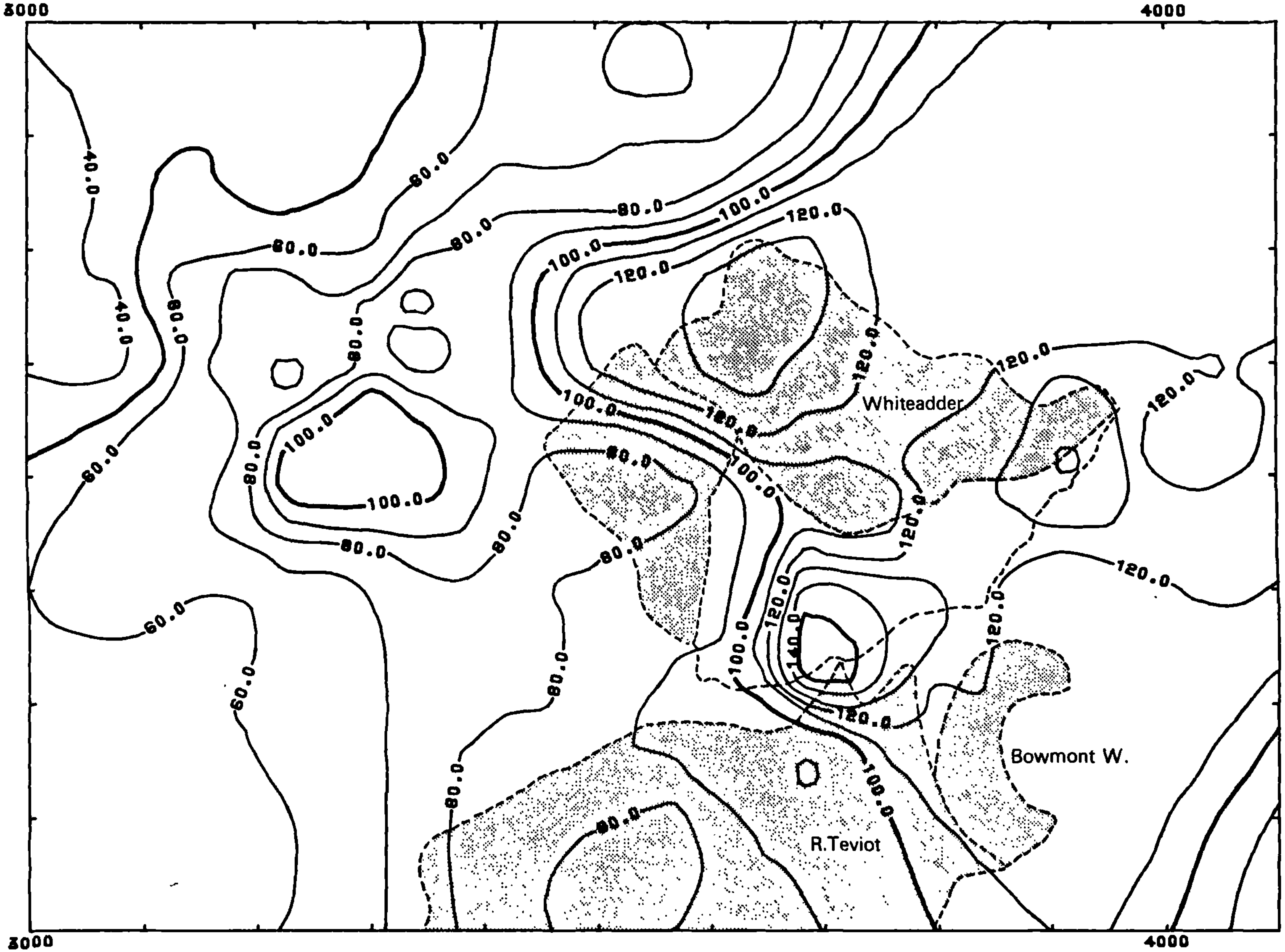
Antecedent rainfall 6-11th August, 1948 Tweed



ANTECEDENT RAINFALL 6-11TH AUGUST, 1948 TWEED STUDY AR
PLOT NO. 1 DATE 03/10/85 TIME 17.25.05

Figure 5.6.8 (iv)

Rainfall 12th August, 1948 Tweed study area



RAINFALL 12TH AUGUST, 1948 TWEED STUDY AREA
PLOT NO. 1 DATE 03/10/85 TIME 17.12.25

As can be clearly seen from Figure 5.6.8.(iv), there were three main cells of intense rainfall. The largest 24 hour fall was the 6.21 in (157.7 mm) at Kelso; however, there were other cells centred around Tweedhill and Kingside, within the Lammermuirs. The Whiteadder catchment fell within the 4-5 in (101.6-127 mm) threshold while the Leader had generally over 3 in (76.2 mm). Another exceptional factor was the high incidence of antecedent rainfall between the 6-11th Aug, 1948 (6 days). Antecedent rainfall amounted to 3-4 in (76.2-101.6 mm) in the Whiteadder catchment and 1-3 in (25.4-76.2 mm) in the Leader and Teviot catchments (Figure 5.6.8.(iii)).

(4) 25th Oct, 1949

The synoptic situation on this occasion was a deep widespread depression, which crossed England to Scandinavia on the 25th, Oct. The actual rainfall duration was really a 72 hour event, for example with an antecedent POT of 1.88 in (47.8 mm) at Manderston House on the 23rd. This was followed by a day of little or no rain. On the 25th, there was another major rainfall event and the distribution of areas with most intense rainfall can be seen in Figure 5.6.8.(v). High rainfall fell on the Cheviot massif with 4.80 in (121.9 mm) in 24 hours at Linhope. By Kirk Yetholm (Bowmont catchment), the rainfall was reduced to 3.00 in (76.2 mm), but the upper catchment must have had falls in excess of 4 in (101.6 mm). Another cell in excess of 3.50 in (88.9 mm) occurred over the Pentlands and upper Lauderdale.

(5) 27th Aug, 1956

Table 5.6.8.(v)

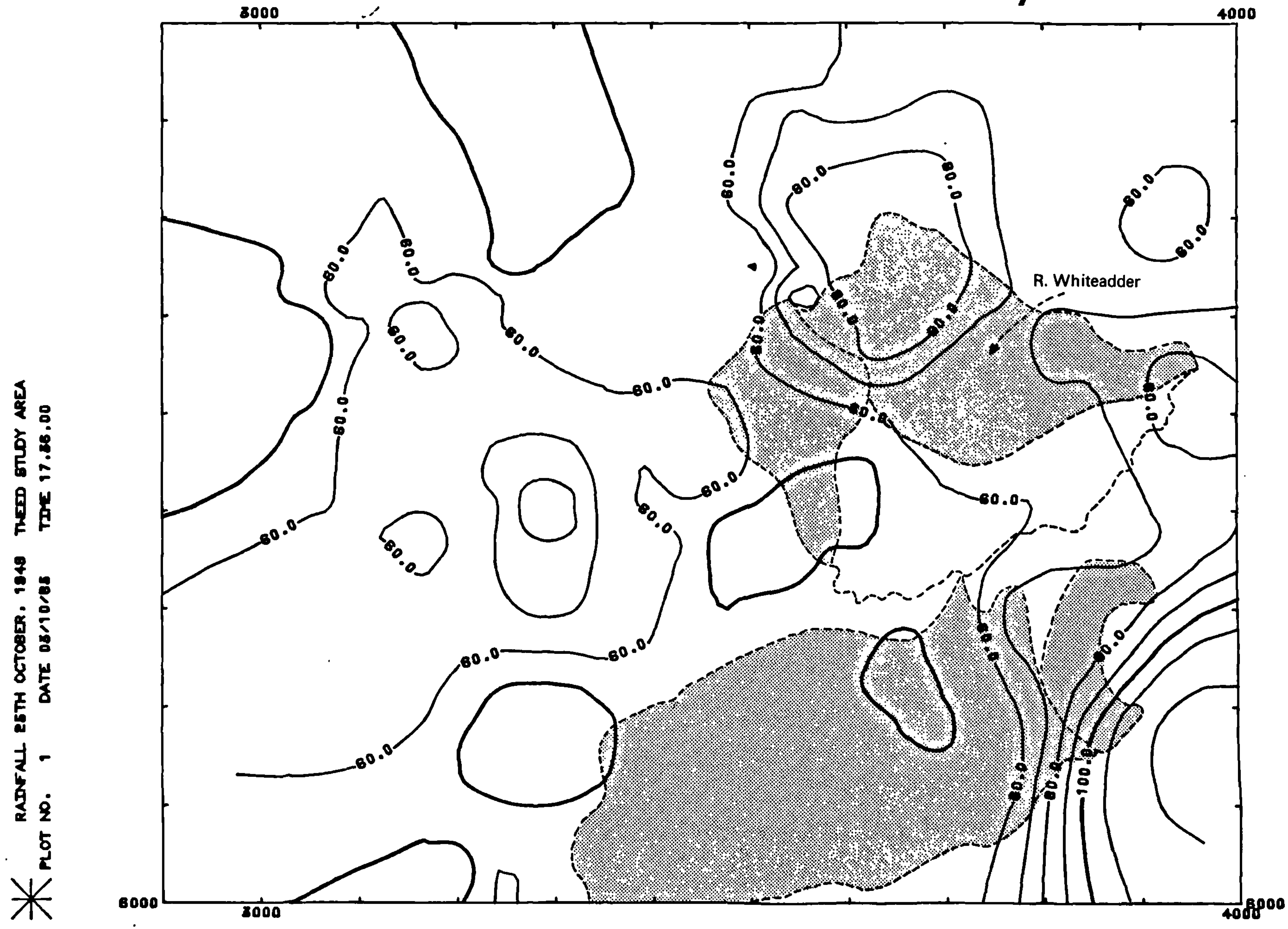
Rainfall recurrence intervals for the 10/1949 flood event *

Station	24 hour (25/10/1949)	48 hour (24-25/10/1949)	72 hour (24-26/10/1949)
Marchmont Ho.	62.7 mm	68.3 mm	111.3 mm
	(23 yr)	(9 yr)	(50 yr)
Duns Castle	62.2 mm	68.3 mm	110.7 mm
	(14 yr)	(8 yr)	(48 yr)
Manderston	76.4 mm	84.8 mm	132.6 mm
	(40 yr)	(19 yr)	(122 yr)
Cowdenknowes	49.5 mm	50.5 mm	77.0 mm
	(10 yr)	(5 yr)	(21 yr)
Hawick	56.9 mm	(-----)	(-----)
	(12 yr)	(-----)	(-----)
Jedburgh	42.7 mm	64.3 mm	(-----)
	(7 yr)	(13 yr)	(-----)

----- data unavailable

* as an annual maximum

Figure 5.6.8.(v) Rainfall 25th October, 1949 Tweed study area



The synoptic situation according to the Meteorological Office was:

"A shallow depression, which developed in a trough of low pressure extending southwards from a low centre over Scandinavia...the cyclonic circulation which extended from the surface to around 30,000 feet (9150 m) produced appreciable convergence over southern Scotland and the *consequent uplift* of moist air resulted in the development of an area of very thick cloud over the region." (Meteorol. Office Report, 1978c)

The system was slow moving with heavy sustained rainfall during the period from late afternoon on 27th to the evening of the 28th. The distribution of 24 hour rainfall (the maximum associated rainfall) for the 27th is shown in Figure 5.6.8.(vi). Again, the main areas affected were the Cheviot massif and the Whiteadder basin.

Table 5.6.8.(vi)

Rainfall recurrence intervals for the 8/1956 flood event *

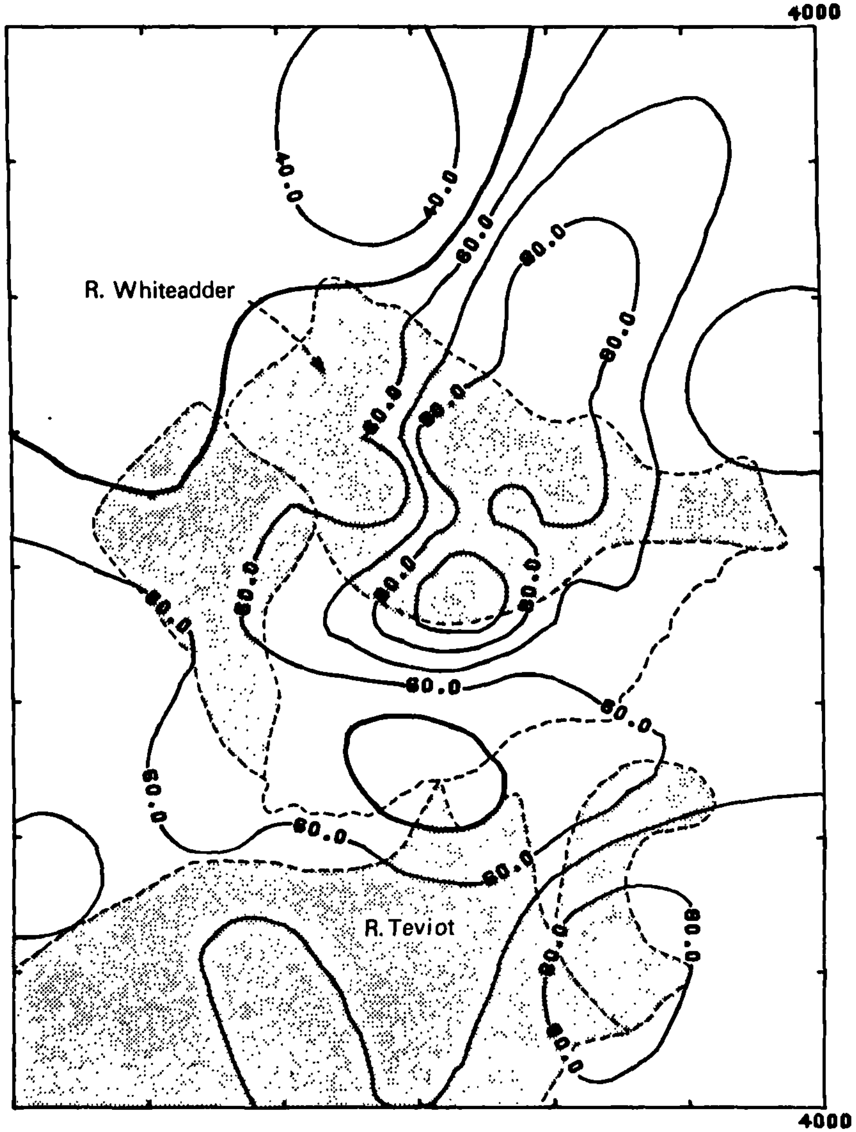
Station	24 hour (27/8/1956)	48 hour (27-28/8/1956)	72 hour (27-29/8/1956)
Marchmont Ho.	96.3 mm	111.0 mm	125.5 mm
	(300+ yr)	(100+ yr)	(125 yr)
Duns Castle	78.7 mm	82.0 mm	110.7 mm
	(47 yr)	(17 yr)	(48 yr)
Manderston	81.3 mm	100.3 mm	111.8 mm
	(60 yr)	(50 yr)	(40 yr)
Cowdenknowes	61.0 mm	81.5 mm	81.5 mm
	(36 yr)	(50 yr)	(30 yr)
Hawick	62.7 mm	(-----)	(-----)
	(22 yr)	(-----)	(-----)
Jedburgh	67.8 mm	86.1 mm	(-----)
	(50 yr)	(65 yr)	(-----)

----- data unavailable

* as an annual maximum

Figure 5.6.8.(vi)

Rainfall 27th August, 1956
Tweed study area



RAINFALL 27TH AUGUST, 1956 TWEED STUDY AREA
PLOT NO. 1 DATE 03/10/85 TIME 17.58.58

5.6.9 Summary

Pre-1900, the most prominent feature was the increase in both rainfall and runoff POT associated with the 1870s and 1880s, within both the tributary catchments and on the mainstream Tweed. These increased rainfall events were frequently associated with summer frontal storms (frequently occurring in August), with rainfall totals capable of exceeding the SMD. SMD would also have been lower than the long term average due to higher annual totals. Pre-1900, there were also extreme floods with a major snowmelt contribution eg. Feb, 1831 and Mar, 1881, and such conditions have not been repeated in the 20th century.

Within the post-1900 record, the most striking feature was the incidence of three, low recurrence interval, summer frontal storms in 1948, 1949 and 1956, affecting similar or at least overlapping areas. The location of the cells of most extreme rainfall however varied. The 1948 storm was by far the most extreme event, both in terms of regional extent and intensity. In intensity, it was closely followed by that of 1956. Apart from these regional storms, individual tributaries had incidences of more localised storms of moderate to extreme RI eg. 1938 storm on the Teviot. However, there were no major climatic fluctuations comparable with pre-1900, except that of the trough in magnitude of annual maxima within the 1960s and 1970s.