MAGNETIC TRACING OF FLUVIAL SEDIMENTS

A study with special emphasis on gravel-bed Rivers

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ABSTRACT

This study examines the application of principles of superparamagnetic enhancement to tracing stream bed-load movement in some gravel-bed reaches of the Rivers Wye and Severn, mid-Wales, and the use of natural mineral magnetic properties to identify dominant contributing sources of suspended material.

The prediction of short term changes in river channel form and process is fraught with problems; their time scales are too short or too long to conform to any established physical principles; and, previous approaches towards describing sediment transport have been too localised whilst processes may be spatially and temporally variable. Sediment tracing may reconcile these problems but few techniques are able to monitor a range of sediment sizes.

This work describes the application of artificial diagenetic heating processes in the enhancement of the natural magnetic properties of stream bed-load to five case studies which typify different scales of the contemporary catchments.

In each case, the main period of tracer movement was associated with one of the longest floods at the onset of winter flows, October 1980. Tracer recovery varied between reaches, from 5% by surface survey to 71% by trapping.

At the smallest scale, tracing in upland drainage ditches, areas contributing significantly increased bed-load yields, exhibited recovery rates of up to 71%. Variations in yield were observed, associated with supply-limitation through sediment storage in shoals even with such small channel systems.

Supply-limitation was equally pronounced within larger upland channels. Within riffle elements, tracer movement occurred at high flow events which were able to overcome the effects of bed armouring and progressive cohesion of the bed-material matrices to release material from the channel bed reservoir. At lower flows, recovered tracer material was coarser in size reflecting supply-restriction by bed material interaction, eg. armouring or induration. However, once incorporated into bedforms, little tracer movement occurred for the range of flow experienced.

Detailed observations of tracer movement within a piedmont reach gravel bar suggested that bed-materials movement occurred as little more than a local flux. Observations confirm the sporadic nature of bar development and their stability which may increase in response to progressive armouring. In the light of the results obtained, it is suggested that more attention should be paid to sedimentological factors when mathematically determining sediment loads.

Likely source areas of suspended sediments were characterised by their natural mineral magnetic properties. Comparing them with the magnetic characteristics of suspended material, a dominant channel-bank source origin was inferred. Source differences associated with water stage were identified but were not consistent probably due to the high frequency of bank-collapse in the region. The dominant source identified is consistent with an immobile channel bed for long periods as noted above. However, in piedmont reaches, such effects may be exacerbated by contemporary flow regulation.

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"Time is like a river made up of events which happen, and its current is strong; no sooner does anything appear than it is swept away, and another comes in its place, and will be swept away too."

Marcus Aurelius Antoninus "Meditations" 121 - 180 A.D.

CHAPTER 1

INTRODUCTION

> "The fact is that, after a century of relatively intense effort to understand the ways of rivers, we remain unable to predict an accurate channel response to a simple change in flow regime at any time scale."
>
> E.J. Hickin, 1983

The ultimate goal in the study of stream channel dynamics and morphology must be to be in a position to understand and, hence, predict the nature of channel changes either as a result of shifting environmental conditions or, as is more recently the case, as a result of human intervention; to be able to predict the consequences of man's use or misuse of drainage basins.

The analysis of channel change has been approached within three broad time scales: Geologic ($>10^6$ years), Geomorphic (10^2-10^5 years) and Engineering ($<10^1$ years). Over geologic time scales it is possible to assume that those processes governing change for example, major tectonic events, are doing so at a constant rate and that therefore the channel is in a form of steady state equilibrium. To some extent, the same may be assumed at Geomorphic time scales where changes over Geologic time are constant and therefore largely irrelevant; changes have occurred in response to profound environmental fluctuations, typically associated with the Pleistocene glaciation. The channel over geomorphic time exhibits periods of episodic change and is in a state of metastable equilibrium. Within engineering time scales, of the order of seconds, hours or days, both Geologic and Geomorphic change are constant and provide a regime state of river behaviour. Short term changes of river behaviour may be modelled with Newtonian physical principles. As such, the river modeller is in such a position to predict certain aspects of channel change over the very short or very long time spans. However, for time spans such as years, or decades to centuries the assumptions made above cannot be readily applied; the time periods are too long to conform to basic physical principles, and too short to achieve steady state equilibrium (Meade, 1982). It is over these time scales that we need to perceive and understand the nature of fluvial problems to be able to predict channel response. River engineers are expected to be able to design structures to last over such periods (Hickin, 1983), and it is over such periods that the effects of contemporary drainage basin use will begin to be realised. The major problems at this scale are that we do not have sufficient knowledge to be able to predict the effects of transient runoff behaviour and the workings of geomorphic thresholds (Pickup, 1981). As a result, we have to rely on empirical studies (see for example, Gregory's (1977) Table 1.2) of the effects land-use changes, river regulation etc.. As regards sediment movement over these shorter time scales, whilst individual eroding sites may be easily identified, the contemporary channel sources are increasingly being recognised as the storage areas between the uplands and estuaries. The nature of the processes by which sediment moves into and out of storage has as yet received very little attention.

At any one point in time, river channel morphology and the activity of channel forming processes may be viewed as the work of the river channel to achieve stability, a dynamic equilibrium of morphology, discharge regime and sediment transport processes. The river channel possesses atleast eight degrees of freedom for change (Hickin, opp. cit.) : width, depth, velocity, shape, boundary roughness, planform, sediment discharge and sediment character. In addition we may consider vegetation, contemporary climate and palaeo-climate and-hydrology (Schumm and Lichty, 1963). Maddock (1976) suggests that some of these parameters tend to resist change or are held constant (minimised variance) for the time periods with which we are concerned. However, the identification of independent and dependent variables is difficult for natural river channels; the governing equations may be easily identifiable under flume experimental conditions, but for the natural condition are largely indeterminate (Hey, 1978). Three particular processes have been the subject of much attention : flow continuity, flow resistance and sediment transport.

Channel morphology and sediment transport are related fundamentally in terms of bed elevation; the determination of rates of aggradation and degradation are central to any determination of change, and, as a result, the calculation of a sediment transport (delivery) rate is vital. Most formulae express sediment transport as a function of the bed shear stress in excess of the critical condition for the initiation of grain motion. This concept was originally put forward by Du Boys and has been consistently used since in formulae such as the Meyer-Peter and Muller, Shields, Schoklitsch and the more recently approved Bagnold stream power approach. The only real departure from this concept is in the Einstein and modified Einstein formulae which relate sediment transport rate to flow turbulence and the probability of a particle being shifted by statistically determined Lift forces.

A major problem with the application of these formulae is their dependance upon empirically determined coefficients based upon data derived from limited experimentation. The latter has all too frequently been based upon flume experiments (to provide some external control over other factors) which operate with a limited range of sediment sizes. Field verification of such formulae is fraught with problems, primarily due to the inadequacy of the available technology (Hubbell, 1964) but also due to the spatial and temporal inconsistencies in transport rate (Leopold and Emmett, 1976). Significant technological advances have been made over recent years, as shown in work by Helley and Smith (1971) and Klingeman and Emmett (1982).

A major weakness of many formulae lies in their assumption of unlimited supply of sediment and in the simplification of the processes of bed-material entrainment which are a direct result of flume experimentation using sediments with narrow size grades and small shape variations. In natural channels, observations of bed-load transport rates have illustrated both temporal and spatial variability. Observed and predicted bed-load transport rates frequently vary as a result of sedimentological and morphological controls on sediment availability for transport operating within the system. Sedimentological controls are primarily determined by the character of the bed material and its response to flow within a given regime. Such influences include the effects of segregated surfaces (armouring, paving etc; see Gomez, 1983; Carling and Reader, 1982), the nature and rate of sediment supply (Nanson, 1974; Newson 1980a and b) and the character of the bed material and bed-forms (Lane and Carlson, 1954, Laronne and Carson, 1976). These factors exert a primary control on the temporal variation of bedload transport rates, typically seasonal and storm hysteresis of bed-load transport. Spatial variations in bed-load transport rate may be related to similar factors such as in the case of individual erosional or depositional sites, although more commonly associated with morphological variations in the channel itself eg. channel shape (Andrews, 1979a; Bathurst, 1979) and channel roughness (Simons and Senturk, 1977). These factors are discussed in greater detail below.

It is apparent from the foregoing discussion that, despite improvements in techniques for measuring bed-load transport rates, the spatial and temporal inconsistencies make estimations of 'at a section' bed-load transport rates highly dubious as representative of the channel reach as a whole. Indeed, this has been a problem with many studies carried out to date. The approach towards quantifying rates of channel change has been largely one of 'trouble-shooting', identifying individual erosional and depositional sites (which may be acting at a much greater rate than many other areas of the system) and using these to characterise major channel reaches of a system. Furthermore, such observations are normally carried out over a limited time period and whilst spatial and temporal inconsistencies are accepted, these studies highlight short-term measures of highly active channel segments. Infact, rates of erosion measured at such sites and rates of delivery at downstream depositional areas such as lake basins or resevoirs commonly produce disparate results. There is, currently, a growing awareness that sediment may be stored within channels for varying lengths of time (Harvey, 1977; Meade et. al., 1981; Meade, 1982). This is particularly so with coarse sediment, which may only be moved infrequently by significantly high flows. In such cases, sediment may move slowly through transporting reaches to be incorporated into medium-large scale bedforms in depositional reaches (Church and Jones, 1982) which may themselves only be subject to episodic disturbance (Ferguson and Werrity, 1983). Whilst sedimentologists have for some time been able to qualitatively describe the process and formation of such bedforms (see, for example: Bluck, 1971), little quantitative effort has been applied, until recently, to relate these to hydrologic-hydraulic factors and to short-term sediment transport processes.

Sediment transport formulae are based upon the mathematical description of loose, freely available sediment entrainment by flow (although some corrections may be inserted such as Einstein's hiding function). However, it is apparent from recent work that sediment storage for significant periods of time may introduce considerable errors in such calculations. In addition, we have yet to realise the full consequence of flow divergence caused by semi-permanent bedforms, particularly, increased suspended sediment loads as a result of increased bank erosion.

In rivers comprising a heterogeneous bed-material matrix ie. gravel-bed rivers, the consequence of fluvial adjustment in addition to the development of quasi-stable bedforms, may include the formation of a segregated surface eg. armoured, paved or censored (Carling and Reader, 1982). Segregated surfaces act in such a way to regulate bed-material availability for transport; general movement of sediment may only begin once the flow is sufficient to cause the segregated surface to break-up (Klingeman and Emmett, 1982). The critical flow at which this occurs will normally be equal to or greater than the flow at which the surface formed, although this depends to some extent upon the time period of adjustment between events and the effects of progressive armouring processes (Day, 1981).

Bed-material segregation is a natural mechanism of adjustment in heterogeneous bedded rivers. However, the effects of segregation may be enhanced by river engineering activity, typically by regulation schemes (Livesey, 1965; Petts, 1977) which reduce the extremes of high magnitude flows, but increase the frequency of lower-mid-range flows for supply purposes. In such instances, the frequency with which finer bed-material may be winnowed from the bed surface, without the disturbance of coarser grains, increases to such a point where the coarse segregated surface limits any further supply and the river capacity is released in the process of bank erosion. The increasing use of piedmont gravel-bed channels for the conveyance of water to major industrialised settlements downstream together with increasing upland land-use activities and the use of the piedmont valley floors for rail/road access and industry makes it vital that natural and enhanced artificial channel planform changes can be predicted. In addition to these physical aspects, there now exists a considerable body of evidence to suggest that man's use of gravel-bed channels may have considerable ecological ramifications (see, for example, Edwards and Crisp, 1982; Milhous, 1982).

The measurement of sediment loads in gravel-bed rivers is extremely problematic (Newson, 1981) and in many cases researchers have resorted to predictive formulae. Operating bed-load traps or sampling devices over a wide catchment area would be prohibitively expensive both in engineering application and labour. Elaborate fixed-point sampling devices such as those described by Klingeman and Emmett (opp. cit) may provide considerable data but lack the necessary spatial recognition of bed-load transport variability and the use of smaller, portable devices which may also be innefficient in sampling gravel material pose practical constraints as to when and where they should be used over the catchment area or at a reach given the spatial variations which may arise. An alternative approach may be to use sediment tracers in association with repeated topographic surveys. However, one immediate problem with this approach lies in the fact that of those techniques available, few may be used to monitor effectively the transport of a range of sediment sizes. Rather, like sediment transport formulae, tracers have been developed to solve particular cases; gravel-bed rivers have only recently received such attention.

This research is based upon the application of enhanced, natural magnetic characteristics derived from artificial diagenetic heating processes. The processes of magnetic enhancement of natural sediment mineralogy has been observed within soils and may be particularly attributed to the mechanisms of burning (Mullins, 1977; Rummery et.al., 1979a). Oldfield et. al. (1981) demonstrated that similar effects could be produced in laboratory furnaces with stream bed-material samples. This work looks at the practical application of such principles to monitored gravel-bed reaches of the Rivers Severn and Wye in mid-Wales, Great Britain, a region of contemporary increase in sediment yields and concern with regards the effects of flow regulation and bank erosion.

CHAPTER 2

APPROACHES TO BED SEDIMENT MOVEMENT

2.1. INTRODUCTION

Sediment transported as bedload comprises that fraction of solid material (usually the coarser grades of the bed material) which is transported, for much of the time, in contact with the stream bed moving in a series of parabolic paths at a velocity generally less than that of the fluid causing its motion.

A complete understanding of the mechanics of bedload transport provides a theoretical basis from which deterministic and probablistic models may be constructed for use in an engineering context to predict channel changes as a result of specific structures or actions (water abstraction, gravel extraction, etc.,), or in a geologic context, to provide an insight into sedimentary morphology and process.

Whilst the description of bedload transport processes is relatively uncomplicated, the quantification of bedload transport rates is extremely problematic. The volume of bedload in transport is highly variable both in a spatial and temporal context as a result of hydraulic, sedimentological and morphological factors. Despite considerable research there exists no single apparatus or procedure which may be considered acceptable (Hubbell, 1964). The interest in sand grades by most American workers and the constraints on size ranges which can be investigated in flumes has meant that many empirical equations do not extend to coarser bed material and extrapolation through grade of several orders of magnitude larger are tenuous.

This chapter considers the mechanisms by which bedload is set in motion and how previous research has used such principles in predicting and assessing sediment transport. It is not within the scope of this analysis to describe each approach in detail but rather to present a discussion which places the experiments described herein in the context of current research. An in depth study of sediment transport technology is provided by workers including Yalin (1977), Bogardi (1974), and Simons and Senturk (1977).

2.2 FLUID MECHANICS AND SEDIMENT TRANSPORT

The erosional and depositional processes occurring in river channels primarily depend upon the interaction between turbulent flow, sediment transport and the nature and occurrance of bedforms (Leeder, 1983). Under turbulent flow (see below) the practical determination of stream velocity under given conditions and its effect within the channel is extremely difficult to achieve (Bray, 1982); problems arise in the evaluation of the resistance to flow imparted by various components of the channel e.g. grain resistance and turbulent dampening, flow structure and resistance related to channel form etc.. Much of our current understanding of the theory of hydraulics is based upon largely theoretical considerations, or upon observations restricted to flow processes through pipes or flumes. Not surprisingly, these restrictions limit the application of many models to a very narrow range of natural channel conditions (Richards, 1982). Furthermore, the lack of field data and instrumentation (Bhowmik, 1982) has restricted the calibration and criticism of these approaches.

Water is a typical Newtonian viscous fluid; it will move in response to an applied stress (here, gravity) and, in so doing, a frictional resistance will be imparted at the fluid-bed boundary resulting (in the simplest case) in laminar flow ie. moving as a series of fluid layers, each successive layer towards the surface flowing progressively faster than the one below (q.v.Richards,1982). The frictional resistance imposed by the channel is balanced by an equal, but opposite, force of fluid drag : the bed shear stress. Flow in natural channels is invariably turbulent; elements of channel roughness and form induce eddying within the fluid, mixing bothslow and fast moving bodies of water.Thus, whilst the overall pattern of shear and velocity remains similar to that of laminar flow, the shear stress at any point may fluctuate over time. The transfer of pockets of high momentum water from the upper flow to the to the fluid-bed boundary results in a steep vertical gradient near to the bed and provides, in addition to fluid drag, hydrodynamic lift (see Below) to incorporate bed material into the flow for sediment transport to occur.

The relative uniformity of the shear stress distribution under turbulent flow is a function of the Reynolds Number (Re* : a function of fluid velocity and viscosity related to the hydraulic characteristics of the channel boundary); channel cross-sectional shape and planform (Bathurst, 1982b). In relation to this, Bathurst identifies several components of resistance. The effect of boundary resistance is related to bed material calibre and its influence on near wall flow providing microtrubulence(Leeder, Opp. cit.). Macroturbulence is related to channel cross-sectional and planform changes providing internal distortion resistance. The chief effect is on the outer layer of the velocity profile and on large scale vortices within the flow. Such effects may be induced by longitud--inal non-uniform flow varaition such as that associated with acceleration/ decelleration related to supercritical flow as predicted by the Froude Number; to variations in flow and flow depth produced by longitudinal non-uniform bed-profile (e.g. riffle and pools); and, to the effects of secondary circulation either in channel bends (Bathurst et. al.,1979) or straight reaches (Leopold,1982) In addition to these, Bray (1982) identifies three further components :sediment transport which, due to the interrelationships involved, provide a dynamic boundary roughness over time; time, related to the long-term in such instances as the provision of sediment to the channel e.g. by debris flows, or to the stabilisung effects of vegetational growth within the channel or on channel bars at low flows; and, stage, which may also provide a dynamic boundary roughness but related to flows in excess of bankfull.

Many of the empirical approaches relate to fairly rigid channel conditions, particularly to a static bad surface with predominantly fine sediment in transport (Richards, opp. cit.). However, it is at high flows that channel change and significant sediment transport occurs and whilst those relations go some way toward explaining flow - sediment transport interrelationships, it is apparent from the above that shear stress is both spatially and temporally variable and for that reason often plot well below that empirically derived.

For a particle resting at the fluid bed boundary, the shear stress applied at the bed provides hydrodynamic forces in two forms. Firstly, as a shear force in the direction parallel to flow, and secondly, as a lift force, caused through the pressure difference created in the vertical profile and by turbulence.

As the intensity of flow and, consequently, the intensity of hydrodynamic force, increases there will be a point at which a particle on the bed will be unable to resist motion; it is dislodged and begins to move. The particle has a resisting force which if less than the resultant hydrodynamic forces will result in particle transport, and if more a particle in transport will become stable on the bed again.

Quantifying the critical conditions for sediment motion (incipient) has been generally approached by analysing time averaged properties of the whole flow, such as mean bed shear stress or velocity. Of the two, velocity is the most readily quantifiable; normally measured at a fixed distance above the bed so as to represent the mean value for the profile. Critical erosion velocities for specific grain sizes are described in detail by Hjulström (1935). This work demonstrated that the two extremes of grain size provide the most resistence to transport; finer material (silt - clay) by cohesion, and coarser material (gravel-boulder) by adhesion. Fine sand, because of its non-cohesive nature was shown to be the easiest to erode. Perhaps the most significant feature of Hjulström's work was in showing that a much higher velocity was required to initiate sediment motion than to sustain it.

However, Hjulström's data were derived from observations of particle transport in the sand to fine gravel range; the two extremes of data were extrapolated from this, and the data itself were collected from flume observations rather than natural stream channels. Both Helley (1969) and Fahnestock (1963) showed that whilst the curves were broadly similar for naturalchannel situations and flumes, the data often plotted well below those empirically derived by Hjulström. Further to this, in summarising those data and others, Novak (1973) showed that the transport of sand-silt size ranges could provide an increase in fluid density which may lower the velocity at which coarser grains are entrained. Novak also suggested that the element of bed roughness also plays an important role in determining motion for different size grains.

The critical mean velocity for particle detatchment from the stream bed varies with shear stress and local sediment characteristics (sorting, consolidation and exposure above the bed surface). Relationships for the initiation and cessation of bed load transport may be more readily made by analysing the variations in shear stress which would include consideration of particle size and, therefore, the condition of bed roughness.

Calculations derived from Shields (1936) and Schocklitsch (1914) incorporate both a shear velocity calculation, representing the intensity of fluctuation in shear stress derived from turbulent flow, and a calculation of the Reynolds Number describing local bed sediment conditions. The immediate availability of transportable material then, depends upon its exposure to the laminar sublayer of the fluid flow. Fenton and Abbott (1977) suggest that this relies on there being two different populations within the sediment matrix: i) coarse grains (large Re*) with small relative profusions into the laminar sublayer; and, ii) small grains (low Re*) where the protrusions are almost equal to the complete grain size. That is, for a smooth bed, grains are exposed well within the laminar sublayer and require a larger shear stress (for a given grain size) to initiate movement on a progressively smoother bed. By contrast, a rough bed means that grains are generally perched well into the flow, the size of the laminar sublayer at high flows is reduced by turbulence and the grains are more readily mobile at much lower values of shear stress. This may be further related to the structural arrangement of the stream bed, which is also crucial for the maintenance of stream competence (see section 2.4.). Large scale roughness elements, as compared to most of the bed, may provide 'safe shadows' where reduced shear stress prevents dislodgement or encourages deposition. This feature is dealt with further in section 2.4.

The critical conditions for sediment motion may also be defined in terms of the Lift Force, that vertical exponent of pressure change created as a result of eddying in turbulent flow and by the vertical pressure gradient induced by the velocity profile of the flow. Experiments using simple plastic spheres (Einstein and El Samni , 1949; Coleman, 1967 and , Watters and Rao, 1971) have demonstrated the significance as well as the variability of efficacy of the lift force. The dependancy upon turbulence means that the lift maybe positive or negative thus providing components of force for saltating grains. Negative lift may also be effective on a smooth bed within the laminar sublayer thus resisting erosion.

Downstream transport of material is achieved by the balance of Lift with fluid drag forces (shear stress per unit grain area). For an exposed grain on the streambed, vertical motion is initiated when the lift force equals or exceeds the grain weight. Clearly, as the component of lift increases, the amount of drag necessary to initiate movement will be minimal (Benedict & Christensen, 1972). The contribution of drag or lift to the initiation of particle motion will depend upon grain size and bed conditions at the time of transport (armoured or paved bed etc.).

Incorporating the effects of turbulent flow in this fashion may provide a more accurate assessment of transport conditions and loads. As suggested above, velocity and shear stress calculations are based upon time averaged properties of the whole flow. When turbulent flow is effective on the stream bed, instantaneous velocities may be two or three times the mean bed velocity (Kalinske, 1947) and the shear stress may be well below the actual value when incorporating the effect of drag and lift (Cheetham, 1979). As yet, however, lift force is a relatively unknown quantity and requires a much more detailed approach in field investigation.

2.3. BED MATERIAL IN MOTION:

A particle forming part of the surface layer of a river bed may begin to move with the mean direction of flow once the critical conditions for transport are surpassed. The mode of transport is related to particle and fluid attributes; hydrodynamic forces acting on the bed material will vary as a result of particle size or due to the variations in lift force or fluctuations in turbulence.

Three modes of transport are commonly described (see Abbott and Francis, 1977):

i) Contact : transport by rolling or sliding in which the particle is always in contact with the bed;

ii) Saltation: movement as a series of low, ballistic jumps in which the particle periodically contacts the bed at the beginning and end of each trajectory;

iii) Suspended : particles are transported high in the fluid in long, wavy trajectories. Additional upward lift impulse may be provided by pockets of turbulent water. Particle contact is only occasional.

At any one point in time, these three modes form a coninuum However, each may be isolated according to its importance in terms of transport stage (the ratio of shear velocity to critical shear velocity; Abbot & Francis (opp cit)). At low transport stages, rolling in the contact mode predominates; velocities are high enough to cause disturbance of the bed fabric thereby initiating motion, but so low as to preclude the vertical movement necessary for saltation. (Bagnold, 1973). The rolling mode also plays a significant role in reducing motion as transport stages decrease towards zero.

Saltation occurs when the transport stage has increased to such a point when the vertical force influences of the fluid become strong enough to pull particles into trajectory paths. As stream velocity increases, so the time spent in a saltation trajectory increases. However, there will always be some particle contact with the river bed.

The suspended transport mode occurs at even higher transport stages when particles are subject to such increased turbulence that the upward trajectory of saltation is extended and the period spent within a single trajectory is increased. By contrast, particles in saltation achieve their greatest acceleration in the downward return trajectory to the bed. The suspended mode differs from the normally accepted definition of suspended sediment load in that there is still some contact with the bed.

In a channel comprising a mix of sediment grain sizes on its bed, the prevailing conditions will determine which particle size range(s) will be transported. For a given transport stage, however, there is likely to be a continual interchange between successive modes as a result of the temporal and spatial variations of hydraulic and sedimentological conditions discussed earlier. The maximum particle size transported by a given flow, prescribes the COMPETENCE of that flow.

2.4. THE QUANTIFICATION OF BEDLOAD TRANSPORT RATES: PREDICTION + MEASUREMENT

2.4.1 PREDICTION

Most predictive equations rely upon similar factors to those described by Du Boys in 1879, equating shear stress at the bed to sediment transport rates at a section. In this way, sediment transport rates will be proportional to the excess bed stress above the critical value (see section 2.2), which will be dependent upon the local sediment size ranges available for transport.

Some thirty different formulae are available for use, and are described in varying detail and application in the following: ASCE (1971), Bogardi (1974), Graf (1971) and Simons & Senturk (1977). The equations may be grouped into three types based on their relationship either to discharge, shear stress or to the statistical distribution of the lift forces available (Graf, opp. cit).

Despite the large number of formulae available, a consistent transport rate is rarely produced (Emmett and Leopold, 1977); individual formulae more often provide an overestimated sediment transport rate (Johnson and Smith, 1977). One reason for this lies in the application of these equations for conditions other than those for which they were developed. As Maddock (1976) suggested: "Other relations have been developed from observations of recirculating flumes. These are very good for describing other recirculating flumes". This statement is likely to hold true for a variety of experimental situations. One immediate problem with those equations derived from laboratory flume experiments is that few have been tested with field data. This is largely due to its paucity, but also due to the associated costs and measuring difficulties (Simons and Senturk, opp.cit.). Moreover, many flume derived formulae rely on data describing a limited range of sediment sizes in transport, normally within the range 0.3 to 7.0mm., and frequently as a uniform size range rather than a heterogeneous composition. Even within such narrow size ranges a broad span of calculated sediment transport rates may occur. For example, Crickmore (1967) examined the transport of radio-labelled tracer sediment (using Au^{198}) within the size ranges 0.1 to 3.0mm and showed that, of five bed load formulae none were comparable excepting in conditions of high, steadv transport rates.

An assumption implicit in many formulae is that of a 'steady state' relationship between discharge and sediment transport rate. However, if this was the case, then observed spatial and temporal variations in bedload transport rates (see later, sections 2.5 and 2.6) and channel morphology would be difficult to resolve. Moreover, the highly unsteady nature of flow-transport relationships may result in actual bedload transport rates being lower than that theoretically estimated. The adoption of any one formula is likely to be a 'hit and miss' affair. Indeed Simons and Senturk (opp. cit.) suggest that in many cases, bedload transport formulae have been selected based upon the particular investigators experience, and usually without having any actual bedload measurements to justify the choice. Nevertheless, there are a small number which repeatedly appear in the literature.

Meyer-Peter and Muller (1948) derived an empirical equation relating bedload transport to the critical tractive force. This work showed that in the case of undulating beds, part of the total available shear stress is used up in overcoming form resistance, the remaining quantity in overcoming the resistance of the grains for bedload transport. The latter is calculated as a mean bed shear stress multipled by a factor $k^{3/2}$, derived from the Manning-Strickler equations. The amount of bedload transport per unit time is calculated from the result of this function in excess of the threshold shear stress as defined for hydrodynamically rough beds, such as by the shields entrainment function (c.f. section 2.2.).

The Meyer-Peter-Muller relation has been used extensively for upland, European streams where the nature of the bed material load has been similar to that used in the original experiments (within the size range 0.4-30.0mm ie. medium sand to gravel). The equation does not perform at all well for material beyond these size ranges (Yalin, 1977). One reason for this is that the Shield's type entrainment function is unreliable, varying with grain exposure or burial. Despite its considerable application in Swiss mountain streams, the absence of a depth function means the equation is not effective when bedload is transported in saltation or suspended modes, both of which are spatially and temporally variable for a given flow volocity.

The first semi-theoretical approach to the quantification of bedload transport was made by Einstein (1942, 1950). Rather than being based upon the concept of a critical stress or tractive force to initiate sediment motion, which is virtually unquatifiable (Sec. 2.2) Einstein assumed that individual particles will move on such occasions when the instantaneous hydrodynamic lift forces (2.2.) exceed the submerged weight of the particle. Once in motion, the probability of the particle being redeposited is assumed equal at all points where the local flow would not cause immediate dislodgement again.

In order to eliminate the natural irregularities of bed material motion due to variations in grain geometry and weight, Einstein assumed a bed of identical grains (similar in geometry and shape) such that their response at rest or in motion would be consistent. In addition, this assumes that the rate, type and distance of movement will be consistent for all particles. The rate of sediment transport is obtained by equating the probabilities of the lift forces exerted by the flow and the particle resistance to movement. The former is a stationary random function. Therefore, since the equations assume a constant particle geometry, the lift forces exerted by the flow possess the same probability for any particle on the bed. The effect of sheltering of smaller particles by larger ones is incorporated in the equations by introducing a correction factor, which is a function of particle size to a critical particle size.

The Einstein relationship was the first theoretical approach to emphasise that the rate of sediment transport is a function of the intensity of flow. However, as a result of the assumptions made about particle geometry, if the bedload is composed of different particle sizes then the rate of transport for each size range has to be computed separately - clearly a cumbersome task. A generalised sediment transport rate may be calculated by using the D_{35} (percent finer) particle diameter of the bedload size distribution. However, it is suggested (Bogardi, 1974) that because the Einstein relation is based upon so many constants, its application is limited to "plain-land" streams (this has been interpreted to mean fine gravel-sand bed streams - Author). Subsequent modifications to the Einstein procedure (Colby and Hembree, 1955) have included measurements of velocity and depth of flow in the calculations, and would appear to provide more consistent results for coarser, gravel-bed rivers (Hollingshead, 1971). In both the modified and original forms of the Einstein equation, however, a practical limitation must be recognised in their assumptions of consistent particle behaviour. In particular, the Einstein approach assumes a constant saltation step length which as seen in section 2.3. may vary with stream power or transport stage.

The most significant advances in the theoretical technology have been made by Bagnold (1960, 1973, 1977) and the similar equation by Engelund and Fredsoe (1976) in relating the physical quantity of work done to the available power and its efficiency. The bedload transport rate and available power are considered by Bagnold to be essentially different states of the same physical quantity and may be defined in the same terms. The unit bedload transport rate (i_b) as determined by Bagnold may be expressed thus:-

Equation 2.4.1

$$i_b=(u_*-u_{*c})\Big/u_*.\frac{W}{\tan\alpha}\left|1-\frac{5.75u_*\ \log\ (0.37\ Y/nD)+Vg}{u}\right|$$

where,

u_* : shear velocity
u_{*c} : critical shear velocity
W : Unit stram power ($W=pgYSu$)
p : density of water
g : acceleration due to gravity
Y : flow depth
S : energy slope
u : flow velocity
$\tan\alpha$: solid.solid friction coefficient
n : nominal distance, in grain diameter, from boundary to centres of fluid thrust
D : Representative grain size
Vg : terminal fall velocity

From previous discussions, it can be seen that the Bagnold equation takes into consideration possible permutations in transport modes by incorporating terms for flow depth and the relation of shear to critical shear stress transferred to saltating grains (see Bagnold, 1973). Bagnold (1966) theoretically derived an inverse relationship for transport rate to flow depth at a given stream power, showing that at large depths the height of saltation for a grain in transport was insignificant as compared with flow depth. Evidence to support this has been published by Williams 1970) showing that, in simple terms, the effect of flow depth on sediment transport rate may be expressed in terms of a depth to grain size ratio Y/D in equation 2.4.1.).

The acceptance of the Bagnold equation as the main advance in sediment transport technology is largely due to its success in relating to field data provided by Leopold & Emmett (1976 and flume data by Williams (opp.cit(see Bagnold (1977))). However, a review by Leeder

(1983) suggests that both the Bagnold and Engelund & Fradsoe approaches may still be overestimating transport rates. Further to this, Emmett (1976) had already shown that the sediment transport rate may not always be in equilibrium with the flow rate as a result of supply mechanisms operating within a reach. Indeed, Leopold and Emmett (1976, 1977) suggest that temporal instability is inherent in bedload transport processes as a result of the variation in hydrodynamic forces and due to the unsteady nature of sediment supply. Changes in the rate of sediment supply or availability for transport have been shown to produce hysteresis responses as discussed by Newson (1980a) Andrews (1979a). Temporaral variations may also be related to bed topography (Andrews,1979b), to the passage of bedforms through a reach (see Emmett, 1980 ; Meade, Emmett and Myrick, 1981), or to the presence of an armoured bed surface (Emmett, 1976). These influences are discussed in greater detail in section 2.5.

It is apparent from the foregoing discussion that to provide a reliable measure of bedload transport in a river, not only must the most appropriate equation be chosen to comply with the on-site conditions, but the measurements used therein should be gathered from more than one reach (or one cross-section in many cases) to minimise the errors introduced by the natural spatial and temporal variations involved.

2.4.2 MEASUREMENT

From the previous section, it is apparent that even with such a profusion of predictive equations field data is required to test and improve them. Comprehensive reviews of the methods available for measuring bedload in transport have been presented by Hubbell

(1964) and by Gomez (1979). However, again as Hubbell states: "no single apparatus has been widely adopted". Conventional devices, such as basket or tray samplers, permanent collectors or slot traps, fail in that many obstruct the flow pattern and thus the transport rate at the bed limiting any operating efficiency. Similarly, the operation of these techniques is such that the sampling pattern is infrequent and may only be related to a short storm period rather than the full range of discharge variations at a point within a given cross-section. Recent developments have gone some way to solve this problem. Working with relatively narrow, shallow reaches (less than 2 metres), Reid, Layman & Frostick (1980) developed the Birkbeck bedload sampler to provide continuous measurement over a full range of flood flows. The sampler consists of a traditional pit slot collector which rides on a rubber pressure pillow; synchronous recording is made of both stage and pillow pressure which can be converted to discharge and sediment load per unit time respectively. Its obvious advantage is in continuity, though the fact that it can be run unmanned/remote is important. This fact also applies to a new generation of 'acoustic' bedload recorders. Again, a continously operating instrument which responds to the acoustic energy generated by particle impact on a metal plate, or on the bed itself. The instrument is potentially quite simple and cheap (Richards and Milne, 1979; Anderson, 1976) with immediate applications to the study of the initiation of motion aswell as temporal variations in intensity. However, the instrument has to be calibrated against a standard measuring technique to relate power output from a hydrophone to rate of sediment transport. The relationship is far from simple varying with particle size but also streamwater turbulence.

The development of the Helley-Smith pressure difference sampler (Helley & Smith, 1971) has provided one of the most efficient techniques applicable to a variety of reach sizes. It is a relatively cheap instrument to build but requires manual operation from a cableway or rod. Provided the manual input, continuous sampling within a reach may be undertaken quite easily; the sampling duration is about 30 seconds allowing successive sampling traverses across a reach to compensate for temporal variations in transport rate. It has been extensively used in the USA and has been compared with more elaborate methods by Klingeman and Emmett (1982). The main drawback of the pressure difference sampler is that it is operable only for sediment in the range coarse sand to fine gravel.For example, Ergenzinger and Conrady (1983) used both this technique and a Muelhofer Basket device for trapping tracer sediment. The latter failed as a result of immense pressure created at high discharge within the instrument. The Helley-Smith device proved difficult to operate under flood conditions and frequently suffered excessive damage to the collecting sack as a result of the discharge energy and sediment size in transport. Extending the design of the basic Helley-Smith type device, to sample material in the gravel and coarser grades has been attempted; the Institute of Hydrology, U.K operate a 15cm version (Newson, pers. comm.). However, it is now thought that the original expansion factor is too great with the result that the sampler actually "sucks in" sediment.

Klingeman and Emmett (opp. cit),and Leopold and Emmett (1976) describe much more extensive and expensive systems for accurate and continuous use on regularly monitored reaches. These are: the conveyor belt sampler operated on the East Fork river; and, the vortex sampler operated on Oak Creek in Wyoming and Oregon respectively. Both require extensive modification of a reach for installation though, once operative, may provide a wealth of data for comparison and calibration with existing technologies. A similar device based on the Oak Creek Vortex samples has been operated in the Torlesse Catchment, New Zealand (Hayward, 1983; Jaeggi and Smart, 1982). However, data collected over a five year period suggest that the sampler works well except at extreme flows (in excess of 0.3m depth) when the bed becomes unstable and results in 'waves' of gravel passing through the sampler flume. This effect, in combination with channel armouring, makes verification of sediment transport formulae extremely complex.

2.4.3 TECHNIQUES FOR TRACING SEDIMENT MOVEMENT

There is a wealth of practical and theoretical techniques available to quantify sediment transport in natural channels. However, their application is limited to specific reaches which satisfy the hydraulic or geomorphological criteria for which each individual technique has been developed. None of the techniques may be applied to a range of situations; installations in large river sections may be costly either in terms of manpower, or in construction and those techniques developed for narrower reaches do not lend themselves to efficient operation in larger channels. The spatial relationships between bed material form and process may not be easily described by such techniques. As such, there still exists a paucity of accurate field data with which to either test and calibrate predictive models, or to monitor for natural/artificial changes.

In an attempt to overcome these problems, research has pursued various avenues to develop techniques to trace sediment and identify transport processes.

Kidson and Carr (1962) and Hubbell (1964) present detailed reviews of tracer technologies avaialable to the early 1960's. A selection

AUTHORS :				
RADIOACTIVITY				
Kidson & Carr	1962	Radioactive plugs of La^{140}, Ba^{140} & Ta^{140}.	Coastal Sands	South coast of Britain & Normandy, France.
Cummins & Ingram	1965	Gold^{198}	Sand	Cape Fear River, USA.
McDowell	1965	Scandium^{46} chloride	0.25 - 0.35 mm	Laboratory flume.
Hubbel & Sayre	1963	Iridium^{192}	Sand	North Loup River, USA.
Thomsen	1980	Chromium^{51}	500 - 595 µm	River Ansager, Denmark.
Crickmore	1967	Gold^{198}	0.1 - 0.3 mm	River Idle, Great Britain.
FLOURESCENCE				
Helley	1969	Red dye	up to 60 cm	Blue Creek, Calif., USA.
Kennedy & Kouba	1970	Red, yellow, green & blue dyes	0.15 - 0.86 mm	Clear Creek, Colorado, USA.
Rathbun & Nordin	1971	"	0.125 - 1.0 mm	Rio Grande, New Mexico, USA.
Rathbun et. al.	1971	"	0.125 - 1.0 mm	"
Rathbun & Kennedy	1978	"	0.125 - 1.0 mm	"
PAINT				
Ritter	1967	Various marine paints	Sand - cobble	Middle Fork Eel River, USA.
Leopold et. al.	1966	"	"	New Mexico, USA.
Laronne & Carson	1976	"	4.0 - 256.0 mm	Seale Brook, Quebec, Canada.
Thorne & Lewin	1979	"	Wolman 100 representative of bed material.	River Severn, Great Britain.
EXOTICS				
Melland & Norman	1969	Ground glass	0.85 - 7.0 mm	Laboratory flume.
Moseley	1978	Limestone aggregate	8.0 - 256.0 mm	Tamaki River, New Zealand.
MAGNETISM				
Butler	1977	Aluminium strip tags	32.0 - 127.0 mm	Horse Creek, Wyoming, USA.
Arkell et. al.	1981	Enhanced natural magnetism	1.4 - 90.0 mm	Rivers Severn and Wye, Great Britain.
Ergenzinger & Conrady	1982	Permanent magnetic plugs	50.0 mm	Buonamico Basin, Italy.
Ergenzinger & Custer	1983	Exotic Magnetic pebbles	16.0 - 64.0 mm	Squaw Creek, Montana, USA.
Reid et. al.	1984	Ferrite magnetic plugs	29.0 mm	Turkey Brook, Great Britain.

TABLE 2.1. APPROACHES TO TRACING BED-MATERIAL TRANSPORT.

of various approaches used since then are presented in Table 2.1 which summarises the approach used to monitor transport of injected or labelled sediment, study site and size range of sediment used in each case.

Ideally, a tracer should be able to satisfy the criteria set out by Nelson and Coakley (1972); the tracer should be:-

a) Stable against premature loss of tag;

b) detectable at low concentrations, thereby providing minimal input for greatest sensitivity;

c) reasonably inexpensive;

d) non-toxic to aquatic and human life;

and,

e) should be tagged in a way that allows repeated tracing within a study area, either by a decay of the tag or by a characteristic signature eg. colour, response to detection instrumentation.

In addition to these criteria the development of any new technology should provide a tracer with comparable density, shape and replication of size range of the host sediment. Many of these characteristics are sadly lacking in those technologies reported to date, and all too frequently, tracing has been limited to specific size ranges. This may well be due to the fact that many tracers have been used or developed with specific problems in mind, rather than having been designed for use in a variety of situations.

Perhaps the simplest and most economic approach to tracing has been in the use of paint or resins to coat the exterior of sediment particles. However, with its cheapness comes a number of problems. The technique is really only viable for larger grain sizes (c.f. Table 2.1); at smaller grain size ranges, the paint or resin can cause coagulation of the particles during the application stage.

Also, as the sediment size range used for a trace decreases, the application may alter its specific gravity and thus any response in transport process. The most important consideration in using paints or resins is their durability and chemical stability. The most frequently cited finishes are with marine paints or epoxy resins. The immediate advantage of this approach is that with the vast range of colours now developed, it is possible to trace from a number of locations within a reach using similar sediment size ranges. Despite this, the major drawback of this approach is in the recovery of the labelled sediment following movement. For example, Laronne and Carson (1976) reported as low as 5 percent recovery of injected sediment as a result of selective burial processes to specific size ranges. As the technique relies purely on visual recovery, any large scale movement or local sediment mixing in this way will make surveying a difficult and laborious task. In addition to this, the persistence of the paint or resin coat will be dependent upon the relative activity of the study area and may well be removed by high abrasion rates. A way round this problem was illustrated by Moseley (1978) by introducing limestone aggregate as an exotic tracer material in the Tamaki River, New Zealand. This overcame the worries of any loss of tag and also provided a tracercovering a wide range of sediment sizes. However, again recovery was limited to as low as 5 percent by volume. This approach also suffers in that the exotic material is unlikely to reproduce the actual characteristics of the local bed material, either in the terms of specific gravity or shape indices. Similar approaches have been adopted using angular fire brick, coal dust and coloured glass though all would be liable to those problems discussed above.

More successful tracing procedures rely upon the tracer sediment retaining a tag 'signal' so that it may be detected insitu or from collected samples. The simplest approach, in this manner, is by using flourescent dye coats applied to sediment with a vinyl binder in much the same way as paint or resin. Indeed, both approaches may be used in conjunction. Fluorescent dyes have been used to coat a wide range of sediment sizes, though more particularly in the sand and finer grades. The main drawback occurs with coarser sediment, typically from a much higher energy system, which cannot retain the dye as a result of abrasion processes. However, in finer sediment matrices this technique may be used to collect depth-integrated samples to give a realistic insight into sediment mixing in transport. Insitu surveys have been carried out using ultraviolet detection equipment at night. The main difficulties here occur with the volume of sediment which needs to be collected for accurate analysis; underwater sampling at depth may be difficult and once the samples are collected, analysis under a flourometer can be very laborious. Typical dyes used for these experiments include Rhodamine B, Primuline, Quartzose, Monzanite, Garnet and Lead.

Coating or treating sediment with a low level radioactive element suffers similar drawbacks to those discussed above. However, Kidson and Carr (1962) showed that the use of radio-active plugs sealed in a drill hole with resin provides a suitable tracer for coarser bed-material. Many of the elements used for such studies have a very short decay time (half life of approximately 40 hours) such that experiments can be carried out safely over a tidal cycle or repeatedly within an area. Such elements include: Tantalum^{51}, Sodium^{24}, Chromium ^{51}, Silver ^{110}, Gold ^{198} and Silicon^{31}. The most immediate advantage of this over other techniques is that the material can be detected insitu on the bed surface or at depth using a rolling sledge carrying a scintillation counter. The disadvantages occur in practise; it is not a very simple technique to handle and all personnel have to undergo repeated health checks. There is also quite considerable cost in using some of the recommended elements added to which, there may also be some sedimentological change in the tracer probably by changing its density or specific gravity. No rates of recovery of tracer material are published for these methods.

The use of magnetic characteristics as a tracing technique has, until now, been restricted to very coarse grain sizes. This is a result of the approach adopted. Butler (1977) monitored transport of bed material in the range 32-127mm using a wrap of aluminium around each clast. The tagged sediment was relocated following each transport event by using a commercially available metal detector. Ergenzinger and Conrady (1982) used a similar approach to this but using permanent magnets embedded in a resin plug within the coarse sediment. Detection devices of a similar design were buried in the channel and operated continuously. In a subsequent study, Ergenzinger and Custer (1983) used naturally magnetic bed material (magnetite and Pyrrhotite) extracted from the Boulder River, Colorado, USA. This was introduced as a tracer to the Squaw Creek catchment near Boulder River and used to monitor sediment transport. Unlike its forerunners, this approach was able to assess the technique for all size ranges of material, though in practise particles less than 32mm could not produce a signal on the detection equipment. The detection equipment also failed in that it only responded by recording a pulse on each occasion that a magnetic tag passed over. The strength of signal was found to be limited by particle velocity, the saltation height from the detector and also the particle orientation to the detection equipment (see chapter 3). Similar problems were identified by Reid et al. (1984), who eventually abandoned any hope of using a range of tracer particle sizes in favour of one standard size at D_{90} (equivalent to 29mm mean clast diameter). Further to this, the Reid et al. tracer was made up of a crushed Barytes and resin mix plugged with a ferrite rod to induce a magnetic signal. The cost of manufacturing tracers in this way proved so high that only 100 particles were used in field experiments.

2.5. VARIATIONS IN BEDLOAD TRANSPORT RATES

From the preceeding discussion it is clear that spatial and temporal variations in bedload transport rates are inherent to the processes operating and may result in considerable scatter when using approaches such as suggested by Bagnold, Shields. Whilst the implications of fluid influences are accepted, such scatter suggests that other factors should be taken into consideration.

In addition to the effects of fluid flow on transport rates, variations in the density and viscosity of the flow may also be important (Colby & Scott, 1965). Variations in viscosity may arise as a result of temperature changes thus affecting sediment fall velocities as described by Stokes Law. The direct effect of such changes is to alter the concentration of the suspended load carried, though this indirectly affects the fluid density resulting in increased friction at the bed and increased bedload transport rates (Beverage & Cuthbertson, 1964). However, to create such considerable change in the fluid properties, relatively large variations in temperature (of the order of 20°C), or suspended sediment concentration. (40-60% by weight) would be required before any measurable differences could be made. More direct and considerable influences may be exerted by sedimentological and morphological characteristics.

Bedload transport rates are a balance between hydraulic conditions and available sediment size ranges. Many equations describing this relationship are based on homogeneous sediment matrices. Variations in transport rates may also be related to particle shape. Lane and Carlson (1954) have described the effects of grain shape on the incipient motion characteristics; rounded grains are more susceptible to movement and, in motion, may be propelled much faster and for greater distances than more angular grains. The effect of particle shape is less significant at high flows (Meland & Normann, 1969) but becomes more important at reducing flow stages when particle settling velocity can be shown to be a function of geometric shape. In channels comprising a range of sediment sizes and shapes, this effect may be compounded by changes in the fabric of the bed material (Laronne & Carson, 1976). This effect is most evident in beds comprising heterogeneous sediment matrices or those characterised by imbricate structure which may directly limit the supply of material for which a streamflow is competent to transport, or provide obstructions to the moving load. Consequently, when field data is applied to the established equations they may plot well below predicted values for the same flows. Calculated bedload transport rates are capacity rates; the calculations assume an unrestricted supply of material. However, as intimated above, variations in sediment availability means that most natural flows transport much less material than theoretically capable. As a result, seasonal and storm variability of bedload transport rates follow similar hysteretic patterns as observed for suspended load (see for example: Walling (1974); Hem (1970); Heidel (1970)) but it is only recently that these effects have been documented. Field experimental data for the East Fork River, Wyoming, USA (Leopold and Emmett, 1976) illustrated that bedload transport rates may be greater, for a given stream power, during the initial rise of the hydrograph at the beginning of a flood season than during any subsequent event. Further studies (Klingeman and Emmett, 1981) related these changes to progressive armouring of the channel bed as a result of selective removal and deposition during subsequent events from an originally loose bed.

Progressive declines in seasonal transport rates may also be related to the geomorphic processes which supply sediment (Nanson, 1974). Nanson and McGreal and Gardiner (1977) illustrated how seasonally active subaerial erosion was crucial in determining the amount of sediment available for transport before each storm event. Thus, transport rates may be dependent upon two processes: the period of preparation of the supply, and, the magnitude and frequency of storm events causing subsequent exhaustion. Further to this, it is possible to identify storm sequences which are supplying sediment to the system becoming limited only by the transporting capacity of the flows, and storms which are supply limited following exhaustion of available channel and slope material and restricted supply of transportable material by armouring processes (Newson, 1980a).

Transport rates may also vary with the composition of the bed material. An assumption implicit in much of the work previously cited is that material on the bed is the same as that being transported (Pickup, 1982). For considerable periods, bedload may be characterised by much finer grain size distributions than those of the bed material (Carling and Reader, 1982) and may only approach that of the latter at relatively high stages. In the case of heterogeneous bed sediment, the bed acts as a coarser fraction of the load through which the river adjusts by the entrainment and deposition of coarse material. When the transporting capacity of a flow is less than that capable of transporting the complete range of available bed-material, differential (or preferential) erosion or deposition may occur to produce a lag surface layer of the coarser size fractions. Whilst distinct form-process interrelationships may be identified in the productions of these "segregated" surfaces, all too often the terms describing them (pavement, armour) have been used synonymously.

In predominantly sand-bed channels, gravel segregation may occur as a result of prolonged channel degradation. The coarser, gravel fractions are worked down to the base of the active bed layer whilst finer sand ranges are preferentially removed. This occurs until such a point that the coarse sediment becomes concentrated into a quasi-continuous layer which is able to resist further degradation. This process is normally associated with a change in hydrologic regime such as reservoir regulation (Livesey, 1965) which results in the reduction of the frequency and magnitude of high flows. As a result, a surface once formed in this manner is rarely disturbed. Segregated surfaces of this type are termed 'pavements';they are characterised by two distinct grain size populations of coarse over predominantly fine fractions.

Armoured, segregated surfaces are more commonly associated with coarser, gravel-bed reaches. Armour surfaces differ fundamentally from 'pavement' or 'paved' surfaces in that they may be periodically disrupted by high flows and reformed by segregation at low to intermediate flows either by winnowing of fines or deposition of the coarser fractions. Transport rates are related to the nature of the heterogeneous bed-material composition and the stage at which the armour layer exerts most control over the available bed-sediment supply, (Klingeman and Emmett, 1982). In streams with a monomodal bed-material composition armouring occurs as a result of segregation by winnowing. Downstream winnowing may occur at low to intermediate flows, removing finer particles from the bed surface and resulting in an accumulation of a coarse gravel layer which is immobile at all but extreme flows. Vertical winnowing within the bed-material matrix may occur at much higher flows when the full range of available bed-material is mobile. As coarser particles are moved, smaller grains fall into the vacant gaps and are effectively restricted from further movement by the slower moving, coarse bed surface. Monomodal beds are characterised by a gradual change in bed-material availability and transport rate for given changes in discharge because there are no difficiencies within any one size range of the bed material.

In streams with bimodal bed-material, segregation by winnowing may still occur. However, transport rates are most clearly affected on the recession of flows where because of the large proportion of coarse gravel in the bed-material load, as the stream loses competence and the coarse material load is deposited, the proportion of the armoured bed increases rapidly for small changes in flow.

The effect of surface segregation, as described above, is that any flow subsequent to its formation must be great enough to disturb its fabric before any sediment motion occurs. Emmett (1976) distinguishes two flow regimes for an armoured bed; one at low flows where coarse particles are not moved and the availability of finer material is limited, and another, at substantially increased flows where the increased capacity can move all available bed-material sizes. For the latter situation, the Bagnold stream power-bedload transport relationship holds true. However, the efficiency of the equation decreases when, at reduced flows, the armour surface restricts the supply of finer material and consequently requires much higher velocities to initiate motion. The effects of heterogeneous bed material are discussed further in section 2.6.

Variations in bedload transport rates may also be caused by morphological changes within a channel reach. Leopold and Wolman (1957) discussed the apparent variations in bed load transport in relation to channel shape. It is suggested that for the same stream power, wide, shallow reaches are capable of transporting more material than their narrow, deep counterparts (Mackin, 1948). It was shown earlier that this relationship may be mathematically derived also, (Bagnold, 1973). However, this contrasts to field data collected by Andrews (1979) indicating that increased scouring of bed material occurred in narrower, deeper sections as compared to the main reach, whilst the transported sediment was deposited within the wider, shallower reaches.

Bathurst (1979) and Hooke (1975) indicate that the distribution of shear stresses withina meander bend may also influence sediment transport rates across a section, and Bathurst, Thorne & Hey (1982) suggest that longstream increases in shear stress through meander bends may be associated with increased sediment output over input, thus scouring the bed. Leopold (1982) associates down and across stream variations in sediment transport rate and associated alternate bar formation with the existence of secondary circulation cells and the local water surface topography. Observations suggest that secondary circulation cells may exist in straight reaches aswell as meander bends; within the former, two cells of unequal and alternate strength combine to alter a topographic ridge within the channel and associated bedload transport.

Channel morphology and transport rates may provide further temporal and spatial inconsistencies when sediment is derived mainly from channel sources. In such cases, sediment is derived from storage features such as channel bars which provide key controls on the release and distance of transport (Meade, Emmett and Myrick, 1981 . Further to this, the presence of large scale storage features in a channel reach may provide effective flow resistence elements (Bathurst, 1982b aswell as potential sources of supply at much higher flow stages. Extensive bar assemblages of this type are typical of rivers carrying considerable coarse sediment load. Their presence serves as an energy dissipator to provide distinct 'sedimentation zones' connected by reaches that are more or less 'transport reaches' (Church & Jones, 1982).

2.6. BEDLOAD TRANSPORT IN GRAVEL BED RIVERS

It is clear, from much of the foregoing discussion, that a great deal of research has been carried out within the confines of sand-bed or alluvial channels, and not surprisingly so; rivers flowing in fine alluvial material exhibit rapid responses to changes in the fluvial system whether by land-use, flooding or management schemes. Increasingly, today, gravel-bed channels are being used for water resource schemes, typically for water regulation or hydro-electric schemes. In British rivers, this means that there is an increasing tendancy to use piedmont river channels (Newson,1981) to convey water from upland regulating reservoirs to population centres in the lowlands. Thus, an understanding of any changes in the flow regime, and consequent effects on morphology and process, is vital for river management. The effects of regulating releases, apart from adjusting the hydraulic characteristics of the river, are primarily in restricting upstream sediment supply and consequently, increase erosion in the area immediately downstream of a dam and channel armouring for considerable distances downstream (Petts, 1977). As a result, during subsequent releases, it is suggested that a stable paved channel is able to resist flow stresses and the less stable river banks may be undercut by erosion. Bank erosion may also be increased where, as a result of changes in upland management practices, there is an increased supply of coarse sediment to the piedmont channel (Newson, 1980b). In such instances, the increased gravel supply is deposited in large scale semi-permanent bed features which divert flow against more easily erodible bank areas. The loss of valuable agricultural land in this fashion is just one problem. Gravel bed rivers are also important for fisheries; it is essential that where regulated, releases are not so low as expose the gravel shoals or so high as to cause their erosion (Carling, 1979). Water quality changes may also occur through regulating releases (Neill and Hey, 1982).

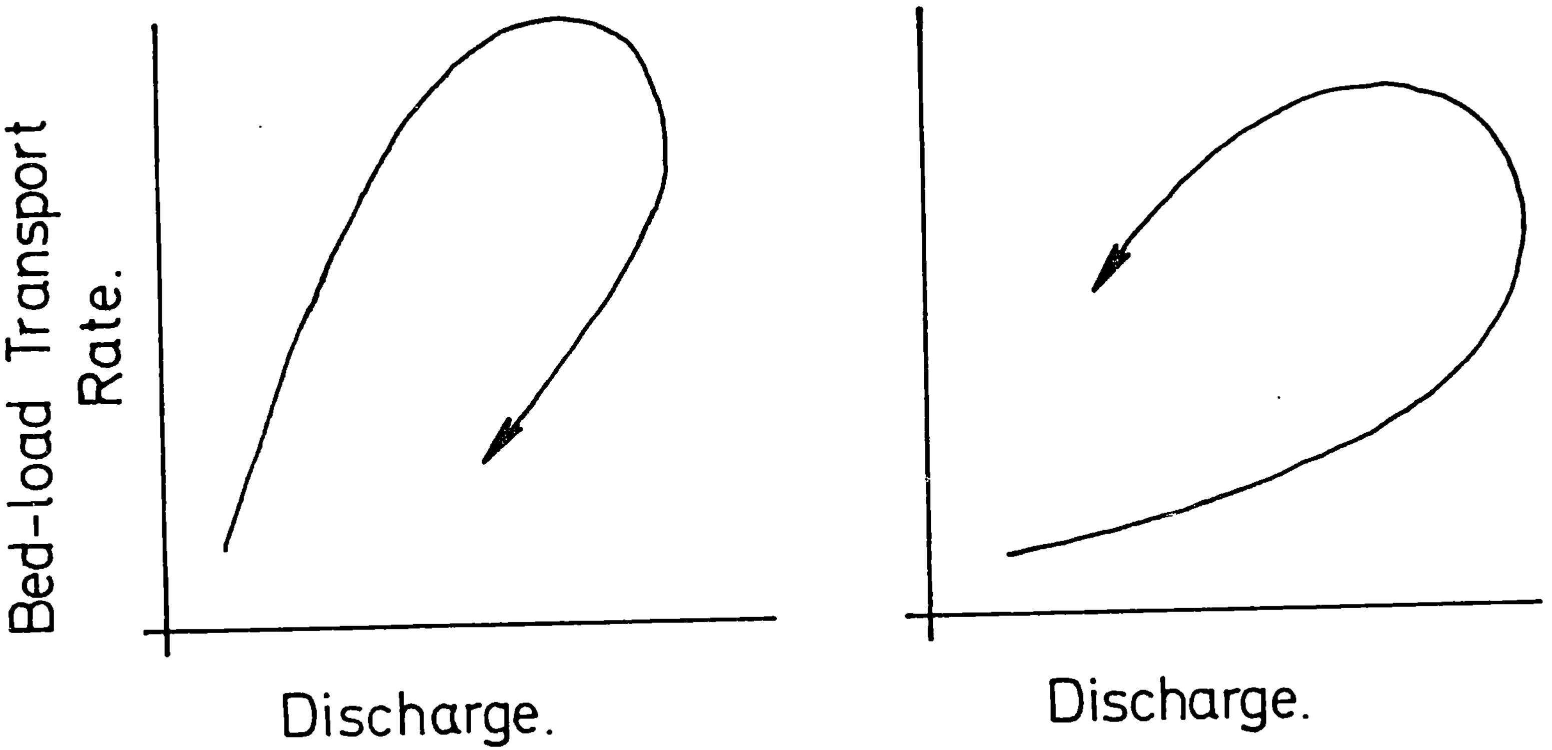
Gravel-bed rivers typically have a wide range of bed material grain sizes, frequently in excess of four orders of magnitude (Klingeman and Emmett, 1982). Bed material heterogeneity of this nature is frequently expressed in two characteristics of the gravel bed: vertical grain size differentiation (Armouring), and bimodal grain size variations. The hydrology of these reaches is typified by transient extreme streamflow events.

The effects of bed material heterogeneity on transport rates are discussed in detail by Klingeman and Emmet (opp.cit), Milhous and Klingeman (1973) and Emmett (1976). The main feature which emerges from this work is the role of the armour layer in inhibiting the incipient motion of the bed sediment, in that, the armour layer effectively controls the availability of bed material for transport. At high flows, the armour layer may break up to provide the whole range of sediment sizes for transport. However, at the lower flow stages of the rising and recession limbs of the storm hydrograph, the armour layer is at its most stable, transport rates are low and comprise largely fine material (sand) derived from winnowing at the armoured surface and/or erosion at the channel margins. This pattern is slightly complicated at increased flows. Transport rates for the rising and falling limbs may differ by an order of magnitude for the same stream discharge. Bedload during the rising limb is comprised of sand to coarse gravel ranges whilst the recession limb carries predominantly fine gravel ranges. This corresponds with the disturbance and restabilisation of the armour layer which regulates the availability of coarse gravel but also finer sand to gravel ranges.

From the foregoing, the assumption that there is an overall relation between discharge and sediment transport (as is accepted for many prediction formulae) would appear to have severe limitations. Indeed, two further relationships ought to be introduced to describe the behaviour of heterogeneous loads: the first, applying to large values of stream power when almost all grain sizes are moving and the effects of heterogeneity are minimal; and, the second, for those stream powers when the effects of the armour layer are most pronounced. Further to this, complications also arise through the assumption that bed material and bed load have broadly the same characteristics (certainly for predictive purposes where sedimentological parameters are realised from streambed material). For much of the time, this is not the case (Maddock, 1976) and, as illustrated above, the armour layer effectively cripples the capacity of flows less than capable of disturbing the whole streambed. Klingeman and Emmett (opp. cit.) have also described the apparent dynamic nature of the armour characteristics, suggesting that its composition is intimately related to the time available for restabilisation between runoff events and the extremes of discharge experienced. Even in cases where armouring is not a direct problem, gravel beds are prone to 'underloose' sediment due to the density of packing and imbrication of the grains. Thus, much higher values of shear-stress are required to initiate motion. Most formulae, on the other hand, presume that the bed will 'overloose' material. (Church and Gilbert, 1975).

The supply of transportable sediment within the gravel-bed channel is rarely uniformly distributed along its length, but becomes incorporated into storage features in a variety of bar forms (Church and Jones, 1982). These features tend to repeat along the channel length in an orderly sequence. Thus, it is possible to identify recurring zones of sedimentation and less stable transport reaches. This implies an inconsistent relationship between bedload transport and discharge during transient runoff events (Klingeman and Emmett, 1982). As a result, the hysteresis characeristics of sediment transport can differ markedly when sampled at points immediately downstream or upstream of a storage area (Figure 2.2; c.f. Meade et al., 1981).

The development of operating procedures to quantify sediment transport rates in gravel-bed rivers has been hampered by the problems described above, and very little progress has been made (Pickup, 1981). The most frequently used formulae include : Meyer, Peter and Muller (1948), Einstein (1942,50) and the Bagnold (1973,77) approaches. The immediate problems with these is their reliance on mean hydraulic and sedimentological conditions whereas we have seen that there may be large spatial and temporal variations. As a result, there are commonly large errors in actual-computed transport rates which at best can estimate within an order of magnitude difference (Kellerhals, 1982). Furthermore, the empirical equations are derived from flume based studies and may be verified on artificially trained channels comprising a narrow range of bed material. Many of the formulae



a) Downstream of storage area.

b) Upstream of storage area.

FIGURE 2.2. HYPOTHETICAL RELATIONS BETWEEN BED-LOAD TRANSPORT AND WATER DISCHARGE WITH REFERENCE TO AREAS OF BED-MATERIAL STORAGE (after Meade et. al., 1981).

verified for gravel-bed reaches are exceedingly sensitive to grain size variation, yet this may be an indeterminate quantity. The problems are well described by Klingeman and Emmett (opp. cit) who suggest "... that even when the empirical relations support each other, they are not well supported by the data."

CHAPTER 3

AN INTRODUCTION TO MINERAL MAGNETIC BEHAVIOUR AND ITS APPLICATION TO ENVIRONMENTAL SYSTEMS

INTRODUCTION

The principles of mineral magnetic behaviour have been described in detail by Tebble and Craik (1969), Cullity (1972) and McElhinny (1973). The application of these principles to the environmental sciences have, in particular, dealt with the palaeomagnetic reconstruction of events such as in the interpretation of continental drift or for sediment dating. Such investigations have relied on the magnetic remanence acquired by sediments in response to past geomagnetic fields; they provide both direction and intensity components of the Natural Remanent Magnetisation (NRM) relative to the earth's magnetic field at the time of deposition.

More recently, a wealth of literature has been compiled which describe the natural magnetic properties of materials (for example, see Dearing (1979), Rummery (1981), Bloemendal (1982) and O'Reilly (1984)). These properties are independent of and unrelated to geomagnetic field but may be used to characterise the mineralogy of a specimen sample. These mineral magnetic properties reflect the iron oxide composition of a sample and may be used to characterise sediment source areas, identify erosion domains and provide an insight into river catchment processes. Thompson et. al. (1980) discuss the application of mineral magnetism to a variety of case studies. These and further examples are presented in Table 3.1..

ATMOSPHERIC		
Oldfield et. al.	1978b	Particulate fallout post - industrial revolution.
Oldfield et. al.	1980	Lake sediment chronology from volcanic ash sequences in the New Guinea highlands, Papua, New Guinea.
Thompson & Oldfield	1978	
O'Garra - Worsely	1982	
LITHOSPHERIC		
Stream borne :		
Oldfield et. al.	1979	Characterisation of sediment type and yield, Jackmoor Brook, G.B.
Walling et. al.	1979	
Oldfield et. al.	1981	Magnetic enhancement of stream bed material.
Arkell et. al.	1981	Magnetic tracing in gravel - bed rivers.
Soils :		
Mullins	1977	Magnetic mineralogy and soil processes,and
Longworth et. al.	1979	processes of secondary formation of magnetic minerals in soils.
Rummery et. al.	1979a	Fire - induced magnetic oxides in soils and lake sediments.
Rummery	1981	" " " "
Lake Sediments :		
Thompson et. al.	1975	Lake sediment core correlation, sedimentation and source erosion rates.
Oldfield	1977	
Bloemendal et. al.	1979	
Thompson & Morton	1979	
Bloemendal	1982	

TABLE 3.1. ENVIRONMENTAL APPLICATIONS OF MAGNETIC MEASUREMENTS.

3.2.1 MAGNETISM

Following Gilbert's (1540-1603) treatise on the earth's magnetic field, the next major development was not to be made until 1825 when, like many other physicists of the time, Oersted began investigating the relationships between magnetism and electricity. Oersted demonstrated that a current flowing through a wire set up a magnetic field around it, and in so doing was able to deflect the needle of a compass held in close proximity to it. Subsequently, in 1832 Faraday showed that the movement of a magnet could induce an electric current. These two principles form the basis of much of electromagnetic theory today.

The magnetic field induced by an electric current may be described by the 'moment' associated with it; that is the turning force experienced at some point near to it. In the case of a current set up by the motion of an electron spinning in orbit around its nucleus, the moment may be described as a product of its atomic radius and charge. However, at this scale two moments can be identified in association with the spinning electron. Firstly : a moment associated with the orbital motion of the electron, the 'Orbital Magnetic Moment'; and secondly, a moment associated with the rotation of the electron about its own axis, the 'Spin Magnetic Moment'. The magnetic moment of an atom as a whole will be the result of all the orbital and spin moments of its electrons (McElhinny, 1973). All minerals exhibit some form of magnetic behaviour as a result of this combined moment.

3.2.2 THE MAGNETIC PROPERTIES OF MATERIALS

The combined effect of all the magnetic moments within a sample, the magnetic moment per unit volume, is termed the Intensity of Magnetisation (or simply magnetisation). In an applied field, the moments rotate in response to the direction of field and provide a net magnetisation in response to the alignment of these moments. This is called the Induced Magnetisation. On removal from the field, depending upon the interaction of these moments (Section 3.2.4.) there may be some relaxation of this alignment and a net magnetisation retained. The net Induced magnetisation is reduced to the (out-of-field) Remanent Magnetisation.

The induced magnetisation in the earth's magnetic field (Ji) is proportional to the applied field:

$$Ji = \chi H$$

At low applied field strengths, χ is a constant of proportionality ie:

$$\chi = Ji/H$$

It is dimensionless and is usually termed the Low Field Apparent Reversible Susceptibility (or Susceptibility).

(Nb. It is given the symbol χ when interpreted for a specific mass, or, K for a specific volume).

3.2.3 VARIATIONS OF Ji WITH H

Most of the magnetic properties of a material can be defined by its response to an applied magnetic field. From Figure 3.1. we can define the initial curve of Ji/H which specifies the susceptibility of a material at point 'a'. At this point, all responses to the applied field are reversible once the sample is removed from that field.

Beyond the point 'b', any further increase in H (the applied field) will produce an increase in Ji; this increase is non-linear and irreversible. At some point, any further increases in H will have no further effect on Ji. This corresponds to the point at which the net magnetic moments of the material are in complete rotation

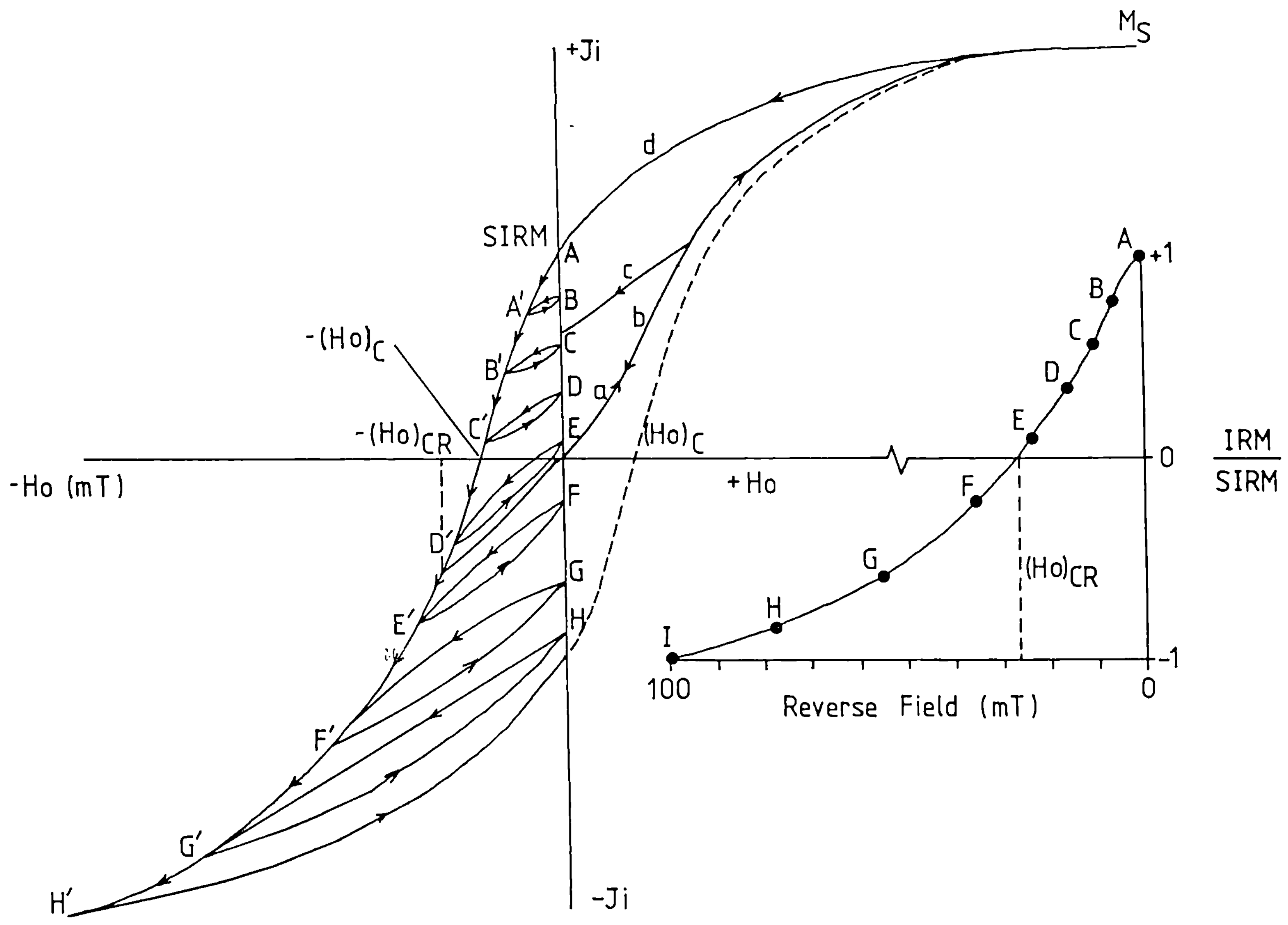


FIGURE 3.1. THE HYSTERESIS LOOP.

in response to the applied field (the nature of this rotation is discussed in detail in section 3.2.4.). The field strength at which this occurs, is called the Saturating Field (H_{SAT}) and the maximum magnetisation induced, the Saturation Magnetisation (M_S). On removal from the field, the material may retain a remanent magnetisation in response to saturation. This is termed the Saturation Remanent Magnetisation, or the Saturation Isothermal Remanent Magnetisation (SIRM) when all measures are carried out at standard room temperature (20°C).

If the material is then placed in a field opposed to the direction of H_{SAT}, the residual magnetism is opposed and may be successively reduced by increasing the reversed field H. The remanent magnetism is a product of the forward saturation remanence minus the reverse magnetisation. On removal from the field, the net moment of the remanent magnetisation may still be in the forward direction (as shown by the curves A'-B, B'-C etc. in Figure 3.1.). Eventually, by increasing the reverse field, a point is reached when the net moments in-field cancel each other out and the net in-field magnetisation is zero. The reverse field (H) at this point is termed the coercive force ($(Ho)_C$).

Continued increases in reverse field will follow the curve until at point D' the net out-of-field magnetisation relaxes to the origin. This point defines the coercivity of SIRM ($(Ho)_{CR}$); it is the reverse field required to reduce the remanent magnetisation from saturation to zero.

Further increases in -H will continue to reduce Ji, until at point H' the material reaches the saturation magnetisation (M_S) in the reversed direction. Out of field, the saturation remanent magnetisation in this reversed direction (-SIRM) equals that measured at point A but has the opposite sign. If the field is now normalised (positive) the curve will follow the sequence denoted by the dotted

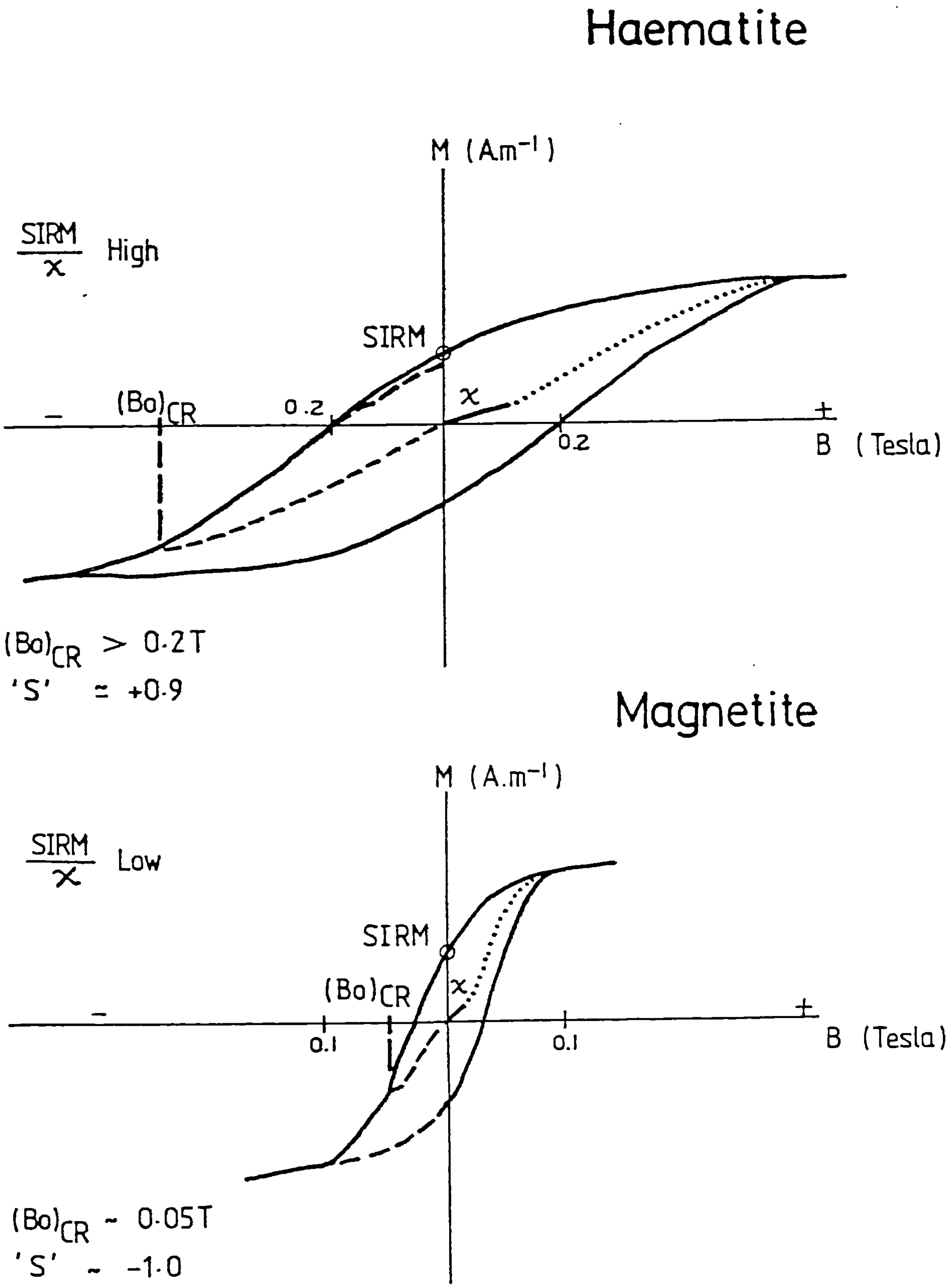


FIGURE 3.2. HYSTERESIS PLOTS FOR MAGNETIC CRYSTALS WITH 'HARD' (ABOVE) AND 'SOFT' (BELOW) ISOTHERMAL REMANENT MAGNETIZATION.

line, again until M_S is reached. This curve defines a geometrically symmetrical shape and provides a repeatable cycle which is called the Hysteresis Loop.

The shape of the hysteresis loop is characteristic of the mineral assemblages (section 3.2.4 and 3.3) and magnetic grain types (section 3.2.4.) contained within a material. At this point, we can usefully define two broad types of hysteresis associated with mineral types (Figure 3.2.): the 'soft' curve associated with magnetite and 'hard' curve associated with hematite. The shape of the hysteresis loop may be simply defined by comparing two discriminating back fields. In recent studies, this has been based on the calculation of a parameter 'S' ($IRM_{O.IT}$/SIRM) which is shown in Figure 3.2. This parameter and others are dealt with in detail in the following sections.

3.2.4 DIAMAGNETISM AND PARAMAGNETISM

Some materials possess no permanent atomic moment; such is the case of two electrons in the same orbit spinning in opposite directions, as in an hydrogen molecule. Such materials are said to be Diamagnetic. When placed in a magnetic field, the electrons rotate so as to oppose the applied field (Lenz's Law). The induced magnetization is weakly negative and quickly lost on removal from the field. Thus, the susceptibility of a diamagnetic material is negative but only small, of the order of $-0.9 *10^{-8}$ m^3Kg^{-1} (for water). By definition, all minerals have a diamagnetic component but in many this is masked by other magnetic phenomena which are discussed below.

Footnote

In keeping with the terminology used by the Liverpool group, $(Ho)_C$ and (Ho)CR are herein termed $(Bo)_C$ and (Bo)CR. B here should not be confused with the measure of Total Induced Remnance (or Flux Density) normally given the symbol B, and measured in Wm^{-2} or T.

In contrast, paramagnetic minerals have an overall dipole moment in the direction of the applied field. However, the thermal motions of the atoms prevent complete alignment of each of the moments so that the overall magnetisation is weakly positive (c.f. Table 3.2). The moment alignment is temperature dependent and is reduced towards higher temperatures to increase the diamagnetic component.

3.2.5 FERRO-, ANITFERRO-, AND FERRI-MAGNETISM.

Unlike diamagnetic and paramagnetic materials, Ferromagnetic substances exhibit a strong positively aligned moment in response to an applied field. Typical ferromagnetic materials such as iron, cobalt and nickel are characterised by self-saturation or spontaneous magnetisation and may have a permanent magnetic dipole moment even in the absence of an applied field (McElhinny, opp.cit.).

The atomic structure of these materials is such that the orbital radii of the inner electrons is very small as compared with the valence electrons which are able to interact with one another to order the spin moments into alignment on the crystal lattice. At absolute zero, all the moments are aligned parallel; increasing temperature reduces the ordering, until at the Curie Temperature (0°C) and above, the ordering is completely destroyed and the system is paramagnetic (Tebble & Craik, 1969).

Antiferromagnetic and ferrimagnetic materials are characterised by subdivided crystal lattices. Two subdivisions (A&B) each exhibit alignment of magnetic moments but they operate antiparallel to one another. If the net moment is zero, the substance is termed antiferromagnetic. Ferrimagnetism occurs when the result of the antiparallel moments are unequal, causing a net spontaneous magnetism. Alternatively, if the moments of the two sub-lattices, though equal, are not exactly parallel, a weak spontaneous magnetism results and the substance is said to be called antiferromagnetic.

3.2.6 DOMAIN FORMATION

Ferromagnetic materials exhibit a strong interaction between neighbouring moments to result in alignment even in the absence of an applied field. However, this fails to explain the fact that iron may be obtained naturally in both the magnetised and unmagnetised conditions. In the demagnetised state, a ferromagnet is divided as a result of the interaction of the spin moments into small regions of very strong magnetisation, called domains.

In an applied field, an internal field is set up so as to oppose magnetisation (Lenz's Law). This field possesses a certain magnetostatic energy, or, energy of demagnetisation. In a saturating field, the magnetostatic energy will be high and, depending upon the size of the ferromagnetic grain, the system will subdivide into domains until it assumes a state of lowest total energy. At each subdivision a boundary, or domain wall, must be formed between opposing domains; this continues until the energy required for the formation of another boundary wall is greater than the consequent reduction of magnetostatic energy. Removal from the saturating field allows the domains to relax into 'easy' directions (positions at which the system requires the least energy to maintain), there is a loss in total moment (section 3.2.3), which on Figure 3.1 is indicated by the decline from M_S to M_{RS}.

The magnetic behaviour of ferrimagnetic and antiferomagnetic minerals is strongly dependent upon variations of their constituent crystal, or grain, size. For small grains, no domain wall can be set up; it is energetically disadvantageous to do so. Instead, the grain forms its own domain and is said to be a Stable Single Domain grain (SSD). At a critical size (>0.05-0.1μm) the grain is able to subdivide further and form multidomain grains (MD). The magnetic behaviour of SSD and MD grains differ markedly.

We have seen that the magnetization of a specimen can be thought of in terms of a rotation and alignment of the orbital and spin moments of its electrons. To bring about a reduction in the mean magnetization by introducing overall non-alignment (below the Curie point) would require magnetic fields of the order of 10^6 Oe (100 Tesla). The effect of the growth of domains (or Nucleation), in subdividing regions of magnetisation into one direction or another is to reduce the difficulty of magnetisation of that specimen (or reduce the energy of demagnetisation). For an SSD grain the energy of demagnetisation will be largely dependent upon rotation; it is only one domain. The loss between M_S-M_{RS} (Figure 3.1.) is minimal. Consequently, both the SIRM and $(Bo)_{CR}$ will be high. In MD grains, the energy of demagnetisation is reduced as a result of the contribution of domain formation, resulting in a much greater loss between M_S-M_{RS}; SIRM is reduced and $(Bo)_{CR}$ much lower in comparison. χ varies only slightly between the MD and SSD grain size and as a result the SIRM/χ ratio may be used as an indication of grain size. For SSD grains SIRM/χ reaches a maximum, and is much reduced for MD grains.

Below the SSD limit (0.03μm), the grains are so small that they undergo frequent spontaneous reversals of the direction of magnetisation as a result the random thermal excitation of the molecules. Their behaviour in a magnetic field is similar to that of a paramagnetic material; they cannot maintain a remanence at room temperature, and

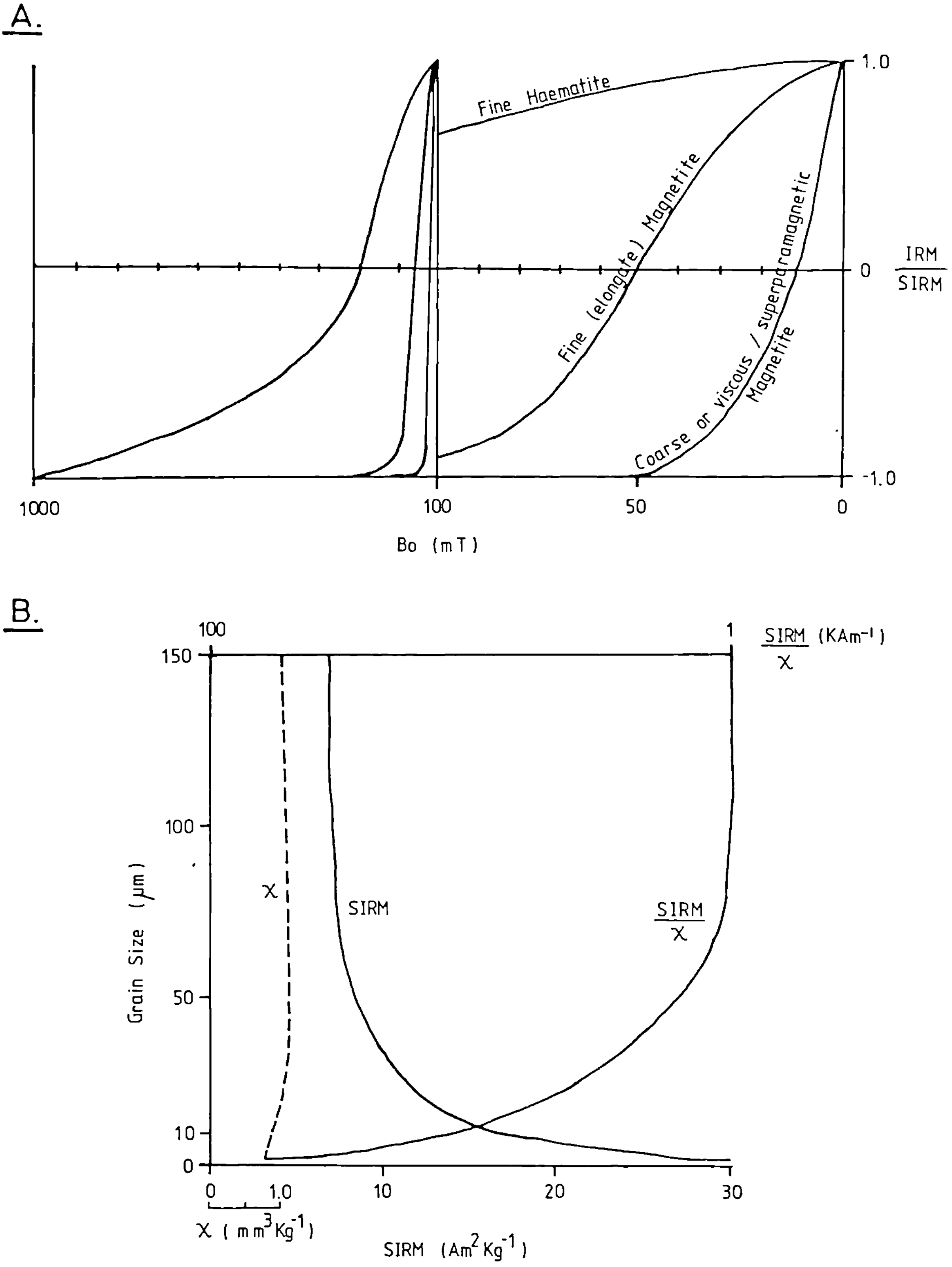


FIGURE 3.3. VARIATIONS IN COERCIVITY OF SIRM FOR CRYSTALS OF MAGNETITE AND HEMATITE (A) AND VARIATIONS IN MAGNETIC BEHAVIOUR WITH MINERAL GRAIN SIZE (B).

(cf. RUMMERY, 1981).

any moment induced by a saturating field is rapidly lost. However, the atomic moments are larger than for normal paramagnetic substances and therefore the immediate alignment in a saturating field is much stronger. This behaviour is termed Superparamagnetism (SP) , and is characterised by a high χ (disproportionately so by volume) and zero SIRM. Stacey and Banerjee (1974) suggest that SP assemblages may have a much greater χ than an assemblage of the same concentration of normal ferrimagnetic grains.

Grain shape (anisotropy) will also effect remanence related properties. Elongate grains, such as Cobalt crystals, have much higher coercivities than square, than do spherical grains. Such effects clearly bear importance to the interpretation of sample measurements especially where there may be a mix of grain size and shapes.

Whilst the principles of grain size variation effects are applicable to all magnetic mineral assemblages, the broad difference in hysteresis characteristics may be used to further classify the response of antiferromagnetic and ferrimagnetic minerals. The characteristics of various sizes of the minerals magnetite and hematite from summaries in Rummery (1981) are shown in Figure 3.3.. For Hematite, the SIRM/χ ratio is greater than 200 KAm^{-1} and the coercivity greater than 0.2T; the remanence is said to be relatively 'hard'. Magnetite, on the other hand, has SIRM/χ ratios between 1.5-50 KAm^{-1} and coercivity values less than 0.05 MT; in this case, the remanence is said to be 'soft'.

The ratio of soft to hard minerals in a sample may be further determined by calculation of the 'S' ratio, the degree of saturation in an applied field of 0.1% following saturation. For Hematite S +0.9, whilst for Magnetite, S=-1.0.

3.3. MAGNETIC MINERALS - ENVIRONMENTAL SYSTEMS

The analysis of magnetic mineral assemblages in lake sediments (c.f. Table 3.1) allows cross-corralation of sediment cores and estimation of sedimentation - erosion rates on a whole catchment basis (Oldfield, 1977). Further to this, Rummery et. al. (1979) and Oldfield et. al. (1979) illustrate the persistence of fire-induced magnetic minerals in soils and their use in reconstructions of fire history and sediment erosion rates. The mechanisms of the natural magnetic enhancement of soil minerals was described by Le Borgne (1955). These give rise to variations which may be diagnostic of source type. On this basis, Dearing (1979) has shown the potential of these techniques in establishing sediment source linkages with lake sediments; further studies by Walling et. al. (1979)and Oldfield et. al. (1979) identify erosion source areas for fluvial suspended sediment load.

A bulk soil sample will normally be composed of a range of mineral components and, thus, its magnetic properties will reflect their combined response to an applied field. The dominant soil magnetic minerals are normally composed of the ferrimagnetic titanomagnetites (or magnetites) in the solid solution series magnetite (Fe_3O_4) - ulvöspinel (Fe_2TiO4); and, the titanomaghemites, the solid solution series magnetite - maghemite (Fe_2O_3). In certain instances, where these minerals are present only in very low concentrations, antiferromagnetic minerals such as geothite, lepidocrocite or the hematite-ilmenite solid solution series, diamagnetic or paramagnetic components may make a significant contribution to the overall magnetic behaviour. However the magnetite content of a soil sample need be only 0.5 - 1% of the hematite content for the two minerals to contribute equally to measures of susceptibility. A range of typical soil magnetic minerals has been discussed by Mullins (1977) and Stacey and Banerjee (1974). These are summarised in Table 3.2..

MINERAL	FORMULA	MAGNETIC PROPERTY	TYPICAL χ ($\mu m^3.kg^{-1}$)	CRYSTAL SYSTEM	
Hematite	αFe_2O_3	Antiferromagnetic	0.27 - 0.63	Hexagonal	Contained in most soil profiles. Typically found in well drained, temperate soils with drier, highly oxygenating conditions.
Goethite	α Fe.O.OH	"	0.125 - 1.26	Orthorhombic	" " " "
Lepidocrocite	γ Fe.O.OH	"	0.5 - 0.75	Hexagonal	Restricted to gleying soils, occurring as bright orange mottles.
Ilmenite	$FeTiO_3$	"	1.7	Hexagonal	
Magnetite	Fe_3O_4	Ferrimagnetic	390 - 1000	Cubic spinel	Occurs both as a primary mineral (esp. derived from basic igneous rocks) and as a secondary Fe oxide (see text).
Maghemite	γ Fe_2O_3	"	410 - 440	Cubic spinel	Secondary soil mineral especially common in highly weathered soils of tropical / sub-tropical nature. May occur as concretions or finely dispersed throughout the profile.
Pyrrhotite	Fe_7S_8	"	53	Hexagonal	

DIAMAGNETIC MINERALS	χ
Quartz	-0.0058
Orthoclase	-0.0042
Water	-0.0090

PARAMAGNETIC MINERALS	χ
Vermiculite	0.152
Montmorrillonite	0.027
Biotite	0.15 - 0.65
Amphiboles & Pyroxenes	0.04 - 0.94
Dolomite	0.011

TABLE 3.2. PROPERTIES OF SOME COMMON SOIL MAGNETIC MINERALS (after Mullins, 1977)

The magnetic mineral assemblages found within a soil may be derived in a number of ways. Primary ferri- and anti-ferrimagnetic minerals are present within the soil matrix essentially as unweathered grains of the parent bedrock. These may be diagenetically altered to secondary minerals by the action of pedogenic processes or by the action of surface fires. Magnetite may occur both as a primary mineral, when derived from a basic igneous bedrock, and as a secondary mineral as a result of the diagenesis of primary mineral sources. In many soils, the abundance of maghemite is much greater than would be expected from the diagenesis of available magnetite (Taylor and Schwertmann, 1974). In such instances, authigenic formation of maghemite (and possibly magnetite aswell) is considered to be of importance. In addition to these sources, a contemporary input of 'new' secondary minerals may be attributed to anthropogenic processes in areas of heavy metal fallout as a result of industrial operations (Oldfield et. al., 1978). These processes are dealt with in more detail below.

Iron released from the bedrock by weathering processes may be transformed to chemically metastable magnetic oxides in the mechanisms schematised in Figure 3.4..

The primary inputs may be contributed to by anthropogenic inputs in the form of atmospheric aerosols, or particulate fallout as a result of fossil fuel combustion or, indeed naturally, by volcanic ash falls (Oldfield et. al., 1980).

The weathering input of magnetic minerals is determined primarily by the characteristics of the underlying lithology. Iron in igneous bedrock is present largely in the reduced ($Fe^{2}+$) state, as silicates, and tends to be converted to the oxidised ($Fe^{3}+$) state by hydrolytic-oxidation reactions. Under high ambient temperatures, together with good aeration, high pH and rapid decomposition of organic matter,

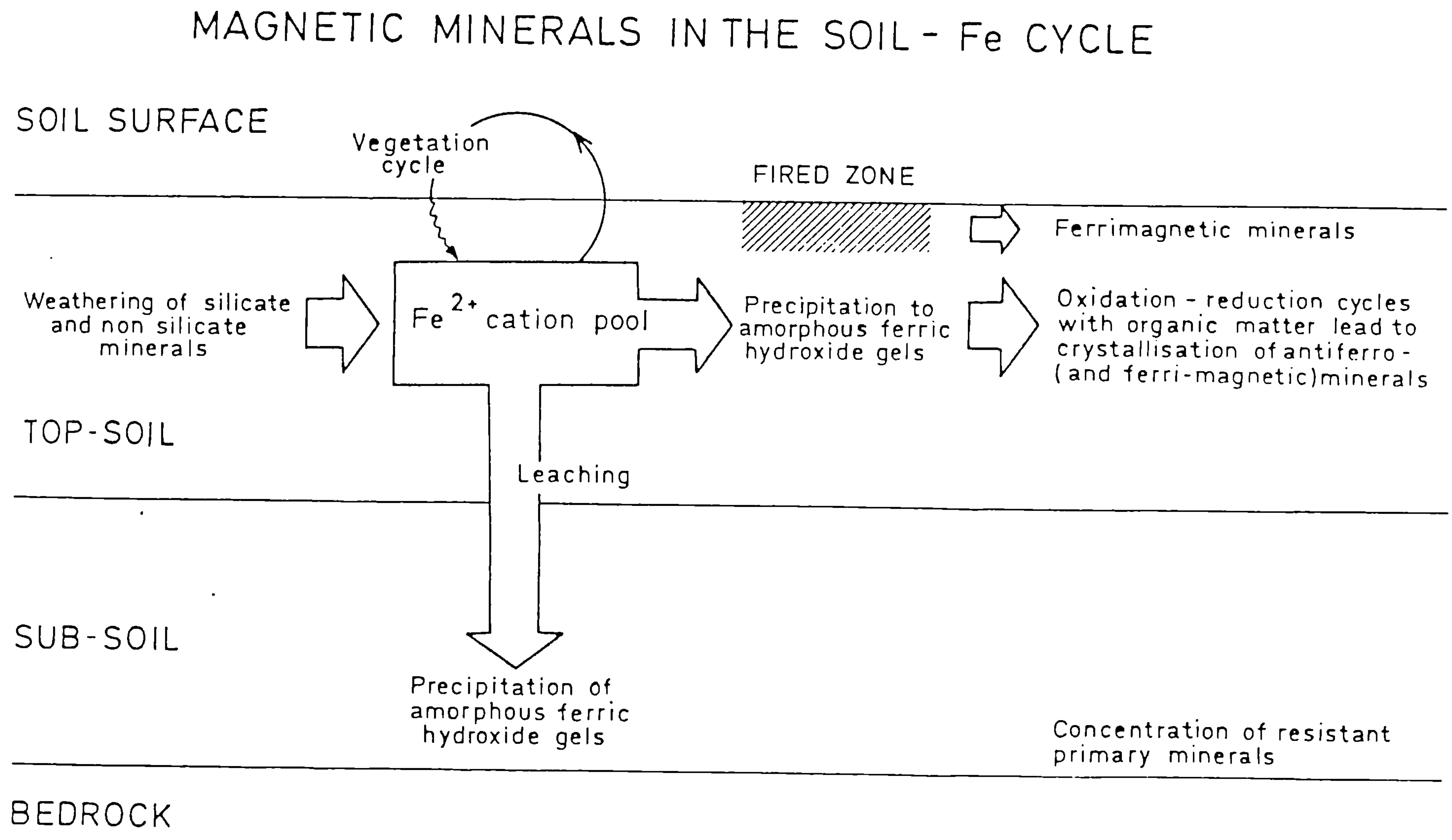


FIGURE 3.4. PATHWAYS OF IRON IN THE SOIL SYSTEM (AFTER DEARING (1979)).

the iron will be converted to hematite (Schwertman & Taylor, 1976) though more frequently to goethite.

Over sedimentary and metamorphic substrate, the mineralogy developed will vary according to the initial concentrations and forms in the bedrock. Many such lithologies are devoid of primary ferrimagnetic minerals and weathering is restricted to the breakdown of iron present in cements, as fine granular inclusions or in clay minerals. Such minerals are fairly easily mobilised following which they will be subject to processes operative within the soil system and conversion to secondary products. Over Triassic Sandstone (c.f. Walling et.al., opp. cit.) typically rich in hematite, long periods of pedogenesis in temperate climates have led to the formation of goethite or more locally (see later) the formation of lepidocrocite.

Ferrous compounds released as a result of weathering processes are rapidly precipitated as *Ferric* (Fe^{3+}) *hydroxide*, an amorphous gel, when mixed with oxygenated soil water. The ultimate fate of the gel is related to the conditions and processes operative within the soil profile. It may precipitate as a thin coating on clay sized particles, or in clay-humus colloids. Alternatively, under podzolization it may be precipitated by effective leaching into a thin concentrated layer in the subsoil which is recognised as the iron pan. Immobilisation occurs as the gel is slowly crystallised during precipitation to anti-ferromagnetic hydrated forms of iron such as Goethite, Lepidocrocite or Hematite.

Secondary enhancement of these minerals, the process of converting paramagnetic and antiferromagnetic forms of iron to strongly ferimagnetic forms (typically Magnetite and Maghemite), has been attributed to a number of processes. Soil samples may typically show values of susceptibility of an order of magnitude greater than the associated bedrock (Longworth et. al., 1979) and frequently greater enhancement towards or at the surface. Under normal pedogenic conditions, this is the result of alternate oxidation and reduction cycles, Le Borgne's (1955) fermentation process. Such processes are fairly ill-defined and still little understood. Vadyunina & Smirnov (1976) demonstrate the variations in iron compounds as a result of repeated wetting and drying associated with seasonal fluctuations in soil water and illustrate the conversion of reduced iron oxides to magnetite in the first instance with re-oxidation to maghemite in subsequent warm, dry conditions.

Bacterial and microbial activity may also be important in the processes of enhancement in creating the conditions essential for the reduction and solution of many forms of iron (Schwertmann and Taylor, opp. cit.). In addition to this, Blakemore (1975) and others have shown that magnetotactic bacteria are able to synthesise magnetite which can then be released at the time of death into surrounding sediments. However, soil bacterial processes may also be effective in reducing the ferrimagnetic content of soils. Gleying, especially, is associated with prevailing chemically reducing conditions to which both primary and secondary iron minerals are prone to solution, and under which the crystallisation of secondary spinel crystals is inhibited. The bacteria Clostridium sp. prevalent under such conditions, have been shown to be effective in the degradation of crystalline iron oxides to reduced, non-crystalline forms.

The main mechanism of enhancement in many soils is that of burning (Mullins, 1977; Longworth et. al., opp. cit.; Rummery et. al., 1979a; Rummery, 1981). The type of magnetic oxides produced in such cases may vary from site to site according to local conditions and the nature of the fire. At best we can suggest that some combination of magnetite and maghemite will be formed, and under prolonged conditions of burning this may be converted to concentrations of hematite. Locally, gleyed soils characterised by higher concentrations of the mineral Lepidocrocite will upon burning dehydrate to maghemite (Mullins, opp. cit.). Oldfield et. al. (1981) and Rummery (1981) discuss the application of these principles to techniques for studying processes of sediment erosion and transport (Section 3.5.).

3.4. TECHNIQUES & METHODOLOGY FOR THE DETERMINATION OF MAGNETIC PARAMETERS

3.4.1 THE MEASUREMENT OF SUSCEPTIBILITY (χ)

The range of instrumentation available for use with reference to the work herein is illustrated in Plate 3.1.. The measurement of susceptibility requires a very low magnetic field generated by a coil of wire carrying a small electric current; in the case of the instruments used here, the field is of the order of approximately 1 Oe or $1*10^{-4}$ Tesla. The control box (Plate 3.1a) supplies a stable oscillator frequency (1KHz) to the coil which, when a sample is placed within the sphere of influence of its magnetic field, changes frequency. This change of frequency is interpreted by microelectronic circuitry within the control box and is displayed as units of susceptibility.

3.4.2 MEASUREMENT OF SIRM AND RELATED PARAMETERS

As opposed to susceptibility, SIRM etc. are measures of remanence measured after removal from an applied field (c.f. section 3.2.3). Variations in H are achieved by placing the sample to be measured within the sphere of influence of a magnetic field at room temperature. A field range of 0 - 1 Tesla may be required to achieve the full cycle

PLATE 3.1. FIELD INSTRUMENTATION FOR MEASURING MAGNETIC SUSCEPTIBILITY SHOWING CONTROL UNIT AND DIGITAL OUTPUT, SEARCH LOOP AND FERRITE PROBE.

of hysteresis including saturation.

Remanence may be measured with a variety of instrumentation; the Liverpool research group relies upon two types:

i) the Parastatic Magnetometer; and,

ii) the Minispin 'Fluxgate' magnetometer.

A general description of the internal workings of each is presented in McElhinny (opp. cit.)

A remanence magnetisation may be induced in a sample by subjecting it to the field generated by either of the following:

1) the pole pieces of a strong electromagnet; fields varying between 0 - 1.0 Tesla; or,

2) the field generated by a Pulse magnetometer; fields varying between 0 - 0.3 Tesla (now up to 0.8T).

Many of the measurements described herein, have been carried out using the latter instrumentation (c.f. Chapter 8) and have, as a result, not been subjected to a full field of 1.0 Tesla. Previous geophysical analyses have been based on a field of 1.0 Tesla to achieve saturation magnetisation within a sample (where this field has been used in the text, the saturated remanence magnetisation induced is termed the SIRM). For many environmental materials, however, saturation magnetisation may be achieved at much lower fields and may be well within the range 0 - 0.3 Tesla provided by the pulse magnetometer. Nevertheless, where a saturating field of 0.3T has been used in the test, the maximum remanent magnetism induced in a sample is termed the $IRM_{0.3}$ (or IRM_{3000}). This is obviously crucial when comparing such parameters as SIRM/χ or 'S' ($IRM_{0.1}$/SIRM) which become $IRM_{0.3}/\chi$ and $IRM_{0.1}/IRM_{0.3}$ respectively.

Measurements of remanence hysteresis are made on samples which have been oven dried (80°) and packed with polyurethane foam (to minimise disorientation) into 10cc containers. The samples may then be placed in the strongest field available (direction marked by an arrow on sample container) to produce either the SIRM or $IRM_{0.3}$. The induced remanence is measured by spinning the sample in the magnetometer. Successive measures of the IRM 'grown' in increasing reverse fields are then made in the same way until the maximum reverse IRM is reached. The results are plotted as points on a coercivity curve which expresses the successive values of IRM in increasing reverse fields divided by the initial 'forward' IRM ($IRM_{0.3}$ or SIRM). An example of such a plot is shown in Figure 3.1. and illustrates the ease at which such parameters as $(Bo)_{CR}$ and 'S' (or $IRM_{0.1}/IRM_{0.3}$) may be quantified.

3.5. ARTIFICIAL MAGNETIC ENHANCEMENT : THE APPLICATION TO SEDIMENT TRACING METHODS

3.5.1 INTRODUCTION

The work discussed in this section is based upon preliminary research carried out by Oldfield et al. (1981) and Rummery (1981). This work described the basic principle of the enhancement of magnetic susceptibility in sediments by heating processes under laboratory conditions. The alteration of the natural mineralogy during heating produces a characteristic magnetisation that may be used as an effective tracer for all sediment sizes.

The scope of the work reported herein is concerned more with the practical and hydrological application of the magnetic technique and is thus based upon the heating procedures and models described by Oldfield et. al (opp. cit). Subsequent developments of the method have been made in the light of field trials by Rummery et.al. (1979b) and in the lab., to extend the technique from dealing with mainly small volumes (20g crucibles) of sediments to dealing with the practical amounts required for tracing (>100Kg).

3.5.2 THE NATURAL MAGNETIC AND GEOCHEMICAL CHARACTERISTICS OF THE SILURIAN SEDIMENT OF THE PLYNLIMON RANGE

The samples used for this preliminary analysis were collected from the upland catchment of the Rivers Wye and Severn. For the most part, this bed material is made up of fragments of Welsh Silurian shales and mudstones (c.f. Chapter 4) derived from coarse gravel colluvial deposits and bedrock, and thus consists of predominantly primary mineral assemblages.

The natural magnetic characteristics and geochemistry of the samples are shown in Tables 3.3 and 3.4. The samples were characteristically iron rich, containing approximately 6% iron, but only weakly magnetic. Values for specific susceptibility ranged between 0.07 to 0.144 $*10^{-6}$ $m^3.kg^{-1}$.

For the most part, the bed-material is made up of paramagnetic forms of iron as illustrated by the low values for specific susceptibility and SIRM and the low values of SIRM/χ ranging 1-5 $K.Am^{-1}$. However, as bed-material particle size decreases beyond 2.8mm, there are noticeable increases in SIRM but more importantly the coercivity curves indicate a much softer remanence (ie, decreasing $(Bo)_{CR}$, c.f. Table 3.3. and Figure 3.5.).

This suggests that the finer shale sediment size ranges are slightly enriched in primary magnetite as compared to the coarser clasts. One possible reason for this lies in the structure and geochemistry of shale deposits which is such that, during diagenesis, fine clay (iron-rich) minerals migrate to the compacting bed surface (Potter, Maynard and Pryor, 1980). Subsequent erosion of the shale bedrock is likely to be associated with the most readily comminuted material ie. that close to the cleavage planes.

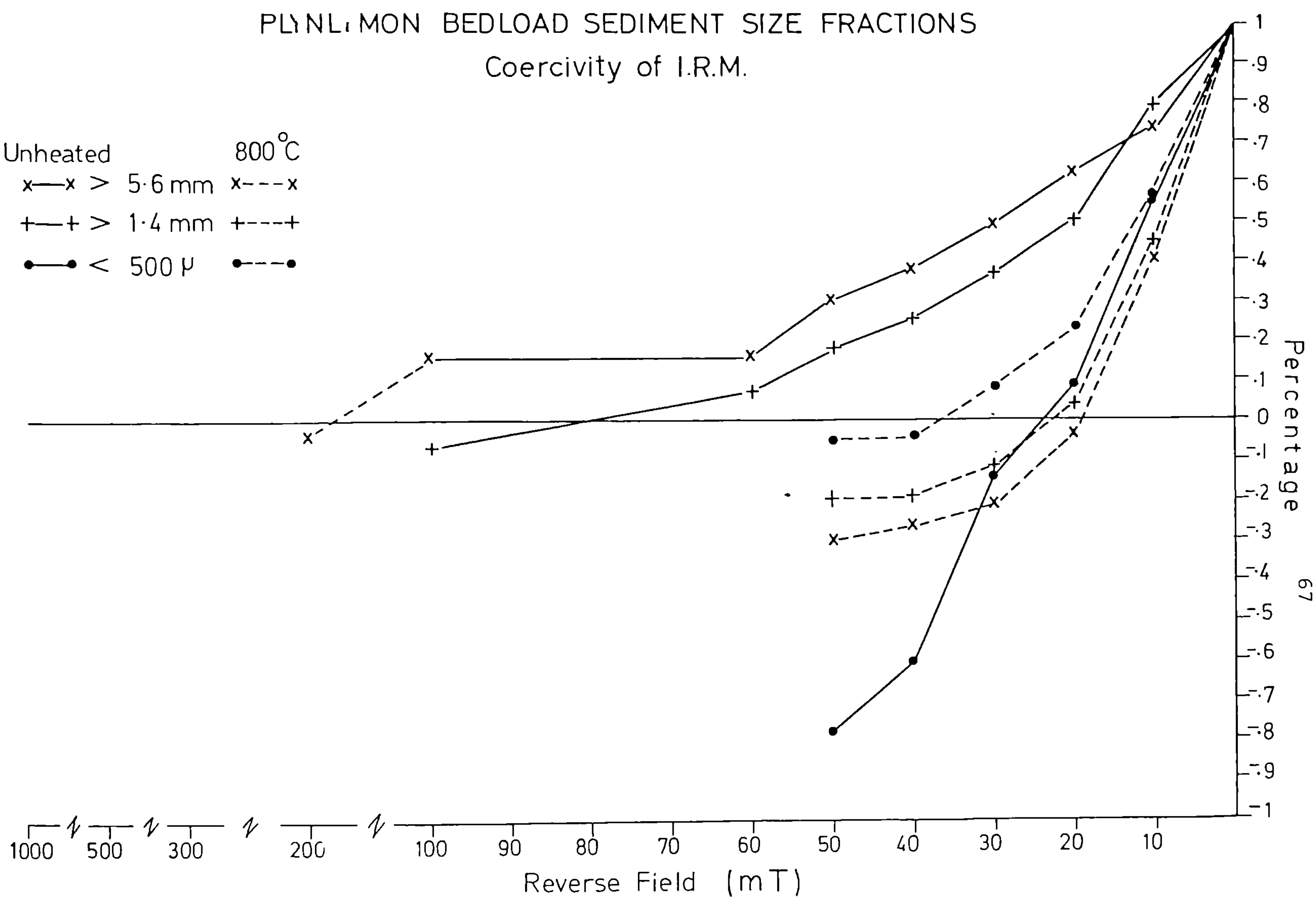


FIGURE 3.5. SUPERPARAMAGNETIC ENHANCEMENT : VARIATIONS IN $(Bo)_{CR}$ WITH PARTICLE SIZE FOR HEATED / UNHEATED BED-MATERIAL SAMPLES.

SIZE RANGE (mm)	χ ($\times 10^{-6} m^3.Kg^{-1}$)	SIRM ($mAm^2.Kg^{-1}$)	SIRM/χ ($KA.m^{-1}$)	$(B_o)_{CR}$ (mT)	'S'
5.6	0.12	0.132	1.09	29.5	-0.51
2.8 - 5.5	0.12	0.173	1.44	10.5	-0.90
1.4 - 2.7	0.12	0.250	2.08	15.5	-0.92
0.71 - 1.3	0.13	0.340	2.46	15.6	-0.90
0.355 - 0.70	0.15	0.495	3.30	15.5	-0.90
0.180 - 0.350	0.11	0.420	3.82	17.5	-0.83
0.090 - 0.175	0.11	0.450	4.10	19.5	-0.81
0.063 - 0.089	0.09	0.441	4.90	22.5	-0.77
0.063	0.09	0.436	4.84	23.5	-0.72

TABLE 3.3. NATURAL MAGNETIC CHARACTERISTICS OF THE WELSH SILURIAN MUDSTONES OF PLYNLIMON.

MINERAL	% CONTENT
SiO_2	65.7
Al_2O_3	19.9
Fe_2O_3	6.8
MgO	1.2
CaO	0.074
Na_2O	0.91
K_2O	3.26
TiO_2	1.14
MnO	0.021
P_2O_5	0.079

10% loss on ignition.

TABLE 3.4. X - RAY FLUORESCENCE OF WELSH SILURIAN MUDSTONES (q.v. Oldfield et. al., 1981).

3.5.3 HEATING PROCEDURES TO PROVIDE OPTIMUM MAGNETIC ENHANCEMENT

The most rapidly measureable magnetic parameter is the specific susceptibility. Consequently, the prime aim in any heating procedure adopted is the maximum enhancement of the magnetic susceptibility within a given sediment size range. This is, of course, offset by the scale, cost and time involved in the operation for the treatment of bulk samples for tracing. Detailed magnetic measurements were carried out by Oldfield et al (opp. cit.) in a systematic study of the enhancement processes with a view to define the controlling variables and to model the changes in mineralogy occurring through different treatments for individual particle size ranges. The variables identified for the optimal recipe are illustrated in Table 3.5.. Magnetic susceptibility, for the Plynlimon sediments, continues to be enhanced to temperatures up to c.950°C (Figure 3.6.); beyond 1000°C the samples undergo partial melting and changes in bulk density and physical characteristics occur. In brief, the recipe identified by Oldfield et al. involves rapid heating of the sediment, in a preheated oven, to 900°C for a period of one to two hours. The material is then spread thinly onto asbestos sheets to facilitate rapid cooling. The effect of changing any one of these variables is shown in the model depicted in Figure 3.7.

Essentially, optimal enhancement of magnetic susceptibility is concerned with the conversion of the available iron minerals into a superparamagnetic (SP) form of magnetite. Within any one treatment, however, there is likely to be a combination of hematite, SP and, SD magnetite. In terms of enhancement, the conversion to hematite (OVER-CONVERSION by prolonged heating) is disadvantageous because the overall enhancement of susceptibility is low. However, under the 'optimal' conditions, each clast is overconverted on its exterior producing a pink colouration (Munsell 10 R 5/8 as opposed to the natural colouration of 9.5YR 4/0). This may be used to the operators advantage in providing

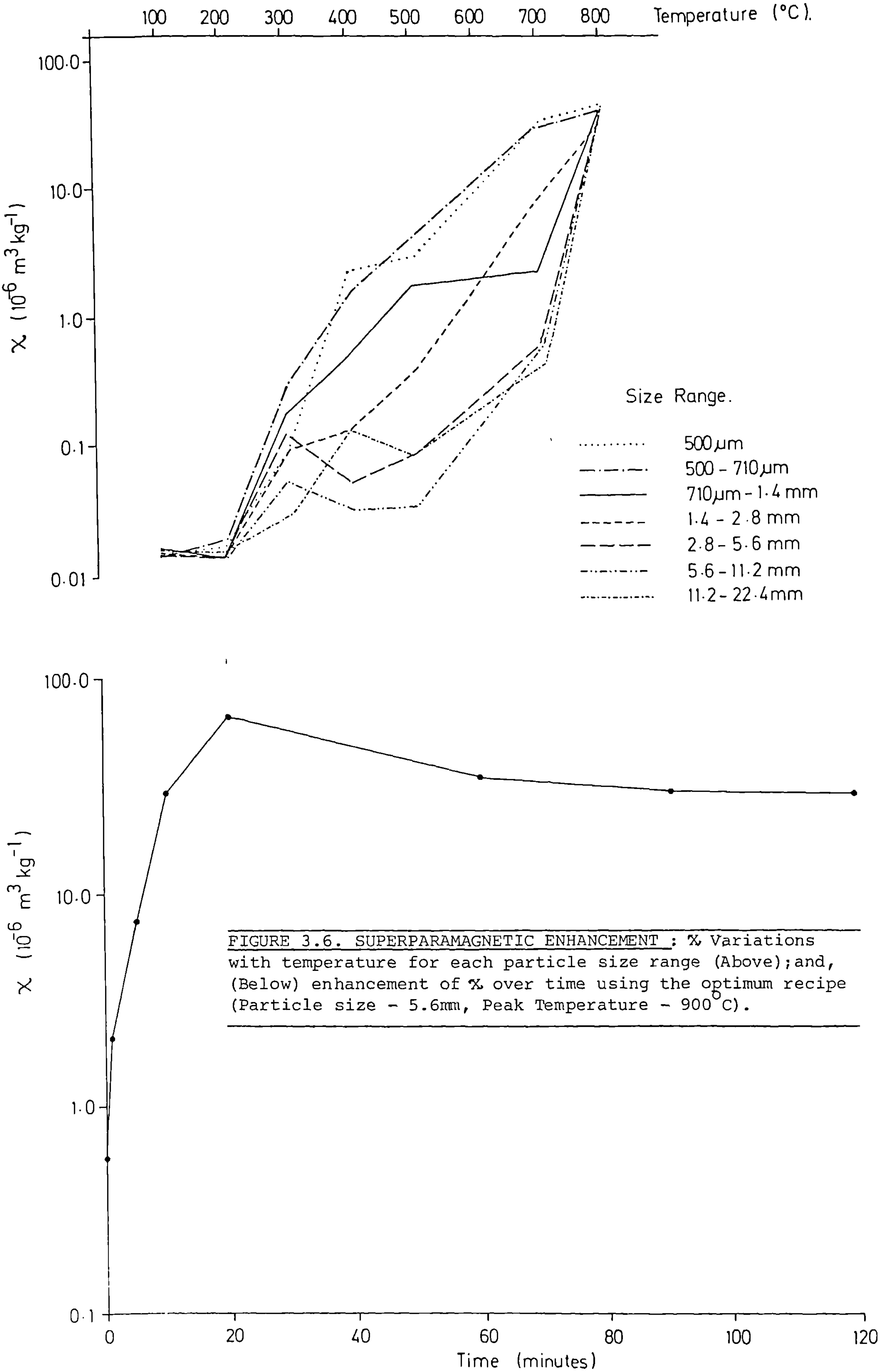


FIGURE 3.6. SUPERPARAMAGNETIC ENHANCEMENT : χ Variations with temperature for each particle size range (Above); and, (Below) enhancement of χ over time using the optimum recipe (Particle size - 5.6mm, Peak Temperature - 900°C).

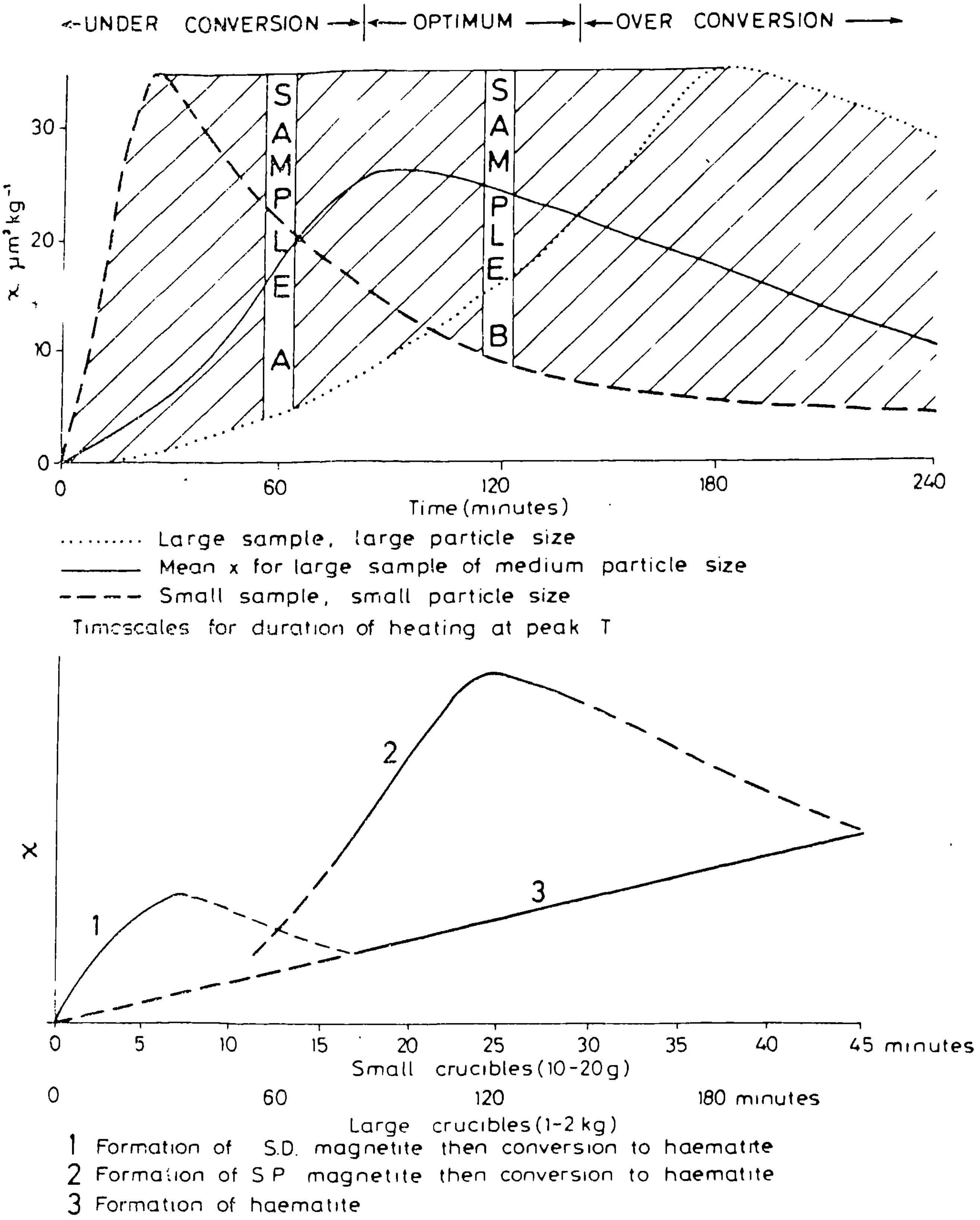


FIGURE 3.7. Schematic model of χ enhancement processes (q.v. Oldfield et. al., 1981). Treatment at 900°C, showing the effects of sample volume and particle size (above) and inferred magnetic mineralogy changes (below) against time. The mean χ enhancement is the average achieved for a whole sample, made up of material at the crucible surface which may be overconverted to material at its centre which may be underconverted. (This principle also applies to coarser size ranges). In practise, the optimum range is shown by Sample B (Time 120 mins.) for small-medium size ranges (to 11.2mm) but for coarser material would be biased more towards the right of the graph and whilst the material may be overconverted, this is more than made up for by its physical size (see text).

A) OPTIMUM FURNACE HEATING PROCEDURES. (q.v. Oldfield et. al.,1981)

i) For optimum enhancement of χ depending on particle & crucible size.

ii) Combination of variables used to produce 100-200 x enhancement for field trial material.

	Variable	(i)	(ii)
A.	Peak Temperature	800 - 950 °C	900 °C
B.	Insertion Temperature	700 - 900 °C	850 - 900 °C
C.	Length of time at peak temperature	20 - 120 min.	100 - 120 min.
D.	Rate of cooling	Rapidly in air or water	On asbestos trays in air @ room temp.
E.	Heating atmosphere	Air	Air
F.	Sample size	0.02-2.0 Kg	1.0 - 2.0 Kg

B) VARIATIONS IN ENHANCED χ BY CONTROL VARIABLES.

A.	Peak Temperature	Optimum 900 °C. Partial melting occurs at 1000+ °C changing Bulk Density.
B.	Rate of Heating	Gradual - weak enhancement. Fast - maximum enhancement.
C.	Time	Optimum - Two Hours. Prolonged heating decreases enhanced χ.
D.	Rate of Cooling	Slow - Decreases maximum χ. Rapid - Maintains maximum χ.
E.	Atmosphere	Soil studies indicate that a reducing atmosphere may improve the enhancemed χ considerably.
F.	Particle Size	Maximum uniform enhancement is achieved for finer size ranges (c.f. Figure 3.7.).

TABLE 3.5. OPTIMUM HEATING PROCEDURES AND THEIR EFFECTS ON PLYNLIMON BED-MATERIAL SEDIMENTS.

a primary identification of the tracer material.

The effects of the optimal enhancement recipe at 900°C are particle size, volume and time dependent (Figure 3.7.A.). For small volumes of material (<20g), maximum enhancement can be attained within 30 minutes. For large quantities of sediment, these are first sieved into whole phi size ranges, then split down into 2Kg batches for treatment. Each 2Kg batch of sediment is treated in a covered mullerite crucible (Oldfield et al.. 1981. used Sillimanite but this was changed to Mullerite because of its resistance to thermal stress and thus longer wear). Splitting the sediment loads in this fashion enables the treatment time to be kept to a minimum. For size ranges up to 32mm, the optimal treatment time was identified at 2 hours (Figure 3.7.A). In terms of the model, this means that the majority of the sample will lie either side of the optimal SP conversion stage. Prolonged heating beyond this point converts all the available iron to hematite and results in decreased susceptibility.

For sediment sizes greater than 45mm, treatment in the above manner tended to produce an extremely violent "exfoliation" reaction during the expansion and contraction of heating and cooling, resulting in a mass of splintered rock fragments. To reduce the thermal stress and maintain particle form, much slower rates of heating and cooling were used. The best results were attained by furnacing 15-20kg of coarse material on an open mullerite tray, raising the operating temperature to 900°C from cold. The temperature was maintained for 4-6 hours then the furnace was opened to cool to room temperature (1-2 days). As predicted by the model, this produced a high concentration of hematite but also a high concentration of SD in addition to SP magnetite (largely in the core and along the cleavage of individual clasts). The overall enhancement of magnetic susceptibility in this

SIZE RANGE (mm)	NATURAL χ	ENHANCED χ	ENHANCEMENT
	($x10^{-6} m^3.Kg^{-1}$)		
1.4 - 2.7	0.144	16.3 ± 0.7	113
2.8 - 5.5	0.143	21.3 ± 1.5	149
5.6 - 11.1	0.122	11.0 ± 0.6	90
11.2 - 22.3	0.112	12.1 ± 2.1	108
22.4 - 44.5	0.098	11.8 ± 3.0	120

TABLE 3.6. ENHANCEMENT OF MAGNETIC SUSCEPTIBILITY (χ) OF SILURIAN BED-MATERIAL SEDIMENTS OF PLYNLIMON.

instance was promoted as a result of the naturally sulphur-rich shales providing their own self reducing environment, the importance of which has been discussed by Oldfield et. al. (opp. cit.) and Mullins (1977; c.f. section 3.3.). As a result, whilst the overall enhancement was lower than compared to the finer size ranges, this is more than made up for by the size of material making it readily detectable.

All the tracer material used for experiments described herein was treated in this way providing an enhancement factor of up to 150. (see Table 3.6.). For material finer than 5.6mm (-2.50), the enhancement was almost uniform enabling predictions of concentrations of tracer "fines" to be made on measurements from subsamples (see Chapter 5). The range of enhancement achieved for coarser material meant that this was impractical and, where possible, tracer clasts need to be identified, sized and recorded.

3.5.4 FIELD MEASUREMENT OF MAGNETIC SUSCEPTIBILITY

The technical details for the measurement of magnetic susceptibility have been described in section 3.4.. These instruments have been used to monitor the presence/movement of magnetically enhanced tracer material in three ways.

Variations in surface susceptibility (ie. presence and absence of tracer material) were measured using a submersible search coil. Much like a metal detector, this instrument responds to changes in magnetic flux around the coil sensor as a result of changes in the volume of magnetic material present. Because the loop is open, and therefore measures variable volumes, its readings are dimensionless and provide a semi-quantitative assessment of the presence and volume of tracer at a point.

The weight or volume of material, at this early stage of development (c.f. Chapter 9), were determined in two ways. Material coarser than 4mm could be identified readily on the basis of its pink colouration (see section 3.5.3) or surface susceptibility, and subsequently confirmed by susceptibility sensing using a hand-held ferrite probe (plate 3.1). Any material separated in this manner could then be assessed in terms of size, weight and shape. Tracer material finer than 4mm was identified by measuring the susceptibility of subsamples of collected material, either from bedload traps or from the channel storage areas. As indicated in 3.5.3, the finer clasts have a much more uniform enhancement and enable accurate calibration curves to be drawn of mass susceptibility* versus percentage concentration in known weights of unenhanced bed material. This enables a rapid assessment of fine tracer concentration but of course requires either the trapping of material, or the disturbance of the channel bed to collect samples.

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* Mass Susceptibility : χ per unit weight measured within an enclosed loop sensor - units $m^3.kg^{-1}$.

CHAPTER 4

4. THE STUDY AREA

4.1. INTRODUCTION:

This research was carried out within the headwater catchments of the Rivers Wye and Severn in mid-Wales. The upland catchments of these two rivers have been monitored by the Institute of Hydrology since 1968, to provide comparative data regarding the hydrological processes of forest (Severn) and grassland (Wye) catchments. Two main initial objectives were defined : to investigate whether the annual yields differed between the catchments; and, to investigate the effect of afforestation on the timing and magnitude of storm flows. (Newson, 1979).

Since 1971, the implications of changing land use on the hydrogical system have been further investigated to include the effects on water quality, erosional and depositional processes. Upland sediment source areas have been identified and the dynamics of sediment transport studied (Newson,1980 a&b) More recently, this type of study has been extended to investigate both the upland source areas for coarse fluvial sediments and the middle, piedmont reaches of storage and throughput (MAFF/IOH Project 73: Erosion and Deposition by Rivers). However, the Institute of Hydrology's Plynlimon Experiments have been concerned largely with monitoring sediment dynamics over the long term with special attention paid to erosion source areas, they have never investigated the routing of sediments through reaches or studied individual sequences of transporting events. This, coupled with the intensive network of instrumentation built up by the Institute of Hydrology at Plynlimon provides an excellent basis for the development of new techniques for application in fluvial geomorphology. Further to this, the natural magnetic characteristics

QUATERNARY	SUPERFICIAL DEPOSITS	Recent & Pleistocene	Alluvium Peat Colluvium, fluvioglacial terrace gravels Head, Scree Till
LOWER PALAEOZOIC	SOLID FORMATIONS		
	SILURIAN	Valentian	FRONGOCH FORMATION Shales & mudstones with thinly banded sandstones and siltstones.
			GWESTYN FORMATION Shales & mudstones with thinly banded siltstones and impure limestone beds.
	UPPER ORDOVICIAN	Van Formation	UPPER Soft blue shales & bronze weathered mudstones. LOWER Massive gritstones, mudstones and conglomerates.

TABLE 4.1. GEOLOGICAL FORMATIONS OF THE PLYNLIMON CATCHMENTS AND STUDY AREA.

Sources : Rudeforth (1970)
Newson (1976)

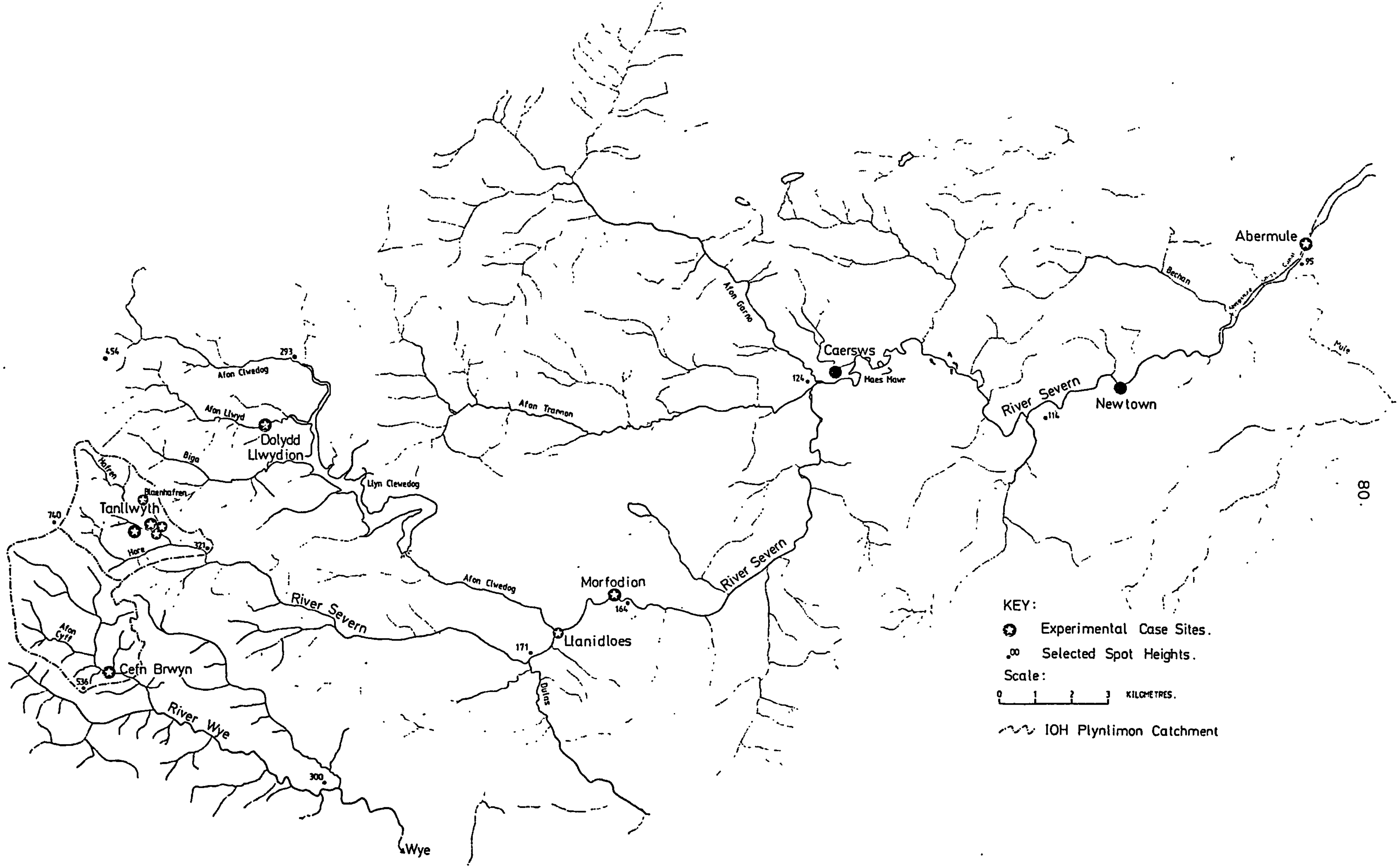


FIGURE 4.1. STUDY AREA: Upland reaches of the River's Severn and Wye.

of the shale bedload compare favourably with those of the lithologies studied by Rummery (1981) thus readily lending itself to the development of the magnetic technique. The area of study for this work (Figure 4.1.) encompasses five case studies based in the upland reaches of the River Wye and the River Severn downstream as far as the Severn Trent Water Authority gauging station at Abermule, near Newtown (GRID REF: SO157945).

4.2. SOLID GEOLOGY

The solid geology of the study area consists almost entirely of shales, mudstones and fine sandstones laid down during the Lower Palaeozoic (Table 4.1). Non-calcareous lithologies from the Upper Ordovician make up the crest of Plynlimon, whilst the surrounding uplands are underlain by Silurian deposits. The main structural trends are aligned approximately NE/SW and these are picked out by many of the tributaries to the west of Plynimon (GRID REF: SN830898) and have a major topographic effect on the upper slope of this area causing a characteristic ridge and step profile. The major physiographic outlines of the area are associated with the Tertiary and Quaternary periods and will be discussed in the following sections.

4.3. DRAINAGE AND RELIEF

The physiography of the catchment prescribed by the boundaries of the IOH Plynlimon research catchment have been described by Newson (1976), and for much of Wales, by Brown (1957, 1960).

The rivers Wye and Severn begin their courses on the eastern slopes of Plynlimon (Grid. Ref SN830898) which lies at the most westerly extreme of the study area. Plynlimon, which rises to a height of 2468 ft.

	SURFACE	HEIGHT ABOVE O.D.		CHARACTERISTICS	
		(metres)	(feet)		
A	MONADNOC	610	2000	Main structural surface	Summit plain of Wales
B	HIGH PLATEAUX	518 - 579	1700 - 1900	Middle Tertiary (Miocene) uplift.	Dominated by convex slopes, gently rolling hills and bog filled hollows.
C	MIDDLE PENEPLAIN	366 - 457	1200 - 1500	Negative base level change.	More dissected, well preserved.
D	LOW PENEPLAIN	244 - 366	800 - 1200	Negative base level change and adaptation to Palaeozoic structure.	Gently sloping, rolling, quite dissected.
E	COASTAL PLATEAUX	122 - 214	400 - 700	"	Present cycle relating to Pliocene-Pliestocene sea-level changes and accelerated incision and river capture.
F	COASTAL FLATS	8	25		

TABLE 4.2. THE PHYSIOGNOMY OF WALES : EROSION SURFACES (after Brown (1960) and Rodda (1970)).

(700m), is carved from a thick mass of Upper Bala mudstone, grits and shales. In form, Plynlimon stands as a true textbook monadnoc, its concave slopes rising to form the main Plynlimon dome. As opposed to much of the surrounding region, Plynlimon is the only feature which can be attributed a definite structural origin; the dome is a complex uplift, skewed slightly eastwards of the structural apex of a number of north-south anticlinal axes. The main axis lies in line with the eastward curve of the Jurassic/Cretaceous outcrops of England and the pre-glacial drainage divide from Liverpool to Bristol. In comparison, the surrounding landscape is composed of a series of plateaux; peneplain erosion surfaces which have formed in response to the regression of late Tertiary sea levels.

The surface of Plynlimon, in common with a number of other features in Wales, for example, Cader Idris and Berwyn in the north which rise to an altitude of 1700 to 2000 feet (520-610 metres), provide the oldest erosion surface in Brown's thesis (Table 4.2). Brown sees this as a warped summit plain whose form envelopes the whole of the present day landscape.

To the east of Plynlimon, as over the whole of Wales today, the landscape between 600-2000 feet exhibits clear evidence of the evolution of three peneplain surfaces (Brown, 1957, 1960; Rodda, 1970). These are evident although not traceable throughout as extensive features due to past weathering and erosion.

Brown sees these surfaces as stages of landscape development as opposed to being structural in origin although structure has occasional influence in causing a break in form through the outcropping of more resistent gritstone and conglomerate lithologies within the weaker shale strata. The peneplains are strongly interdigitate, both the Severn and Wye exhibit low and mid peneplain surfaces traceable well into their upper tributaries.

Each of the peneplain surfaces is seen as a sub-aerial erosion response to middle to late Tertiary negative base level changes, each intimately related to the drainage pattern development: Brown's interpretation of the original drainage pattern of Wales suggests a radial pattern flowing off the slopes of Plynlimon. At this time, both the Severn and the Wye flowed towards the south-east and Brown suggests that the Severn and Clywedog were infact tributaries of the Wye at this stage. Subsequent erosion produced progressively better adjustment of drainage pattern to structure. However, the magnitude of Pliocene - Pleistocene sea level changes created conditions of such negative movement of base level that accelerated incision provided considerable river capture and diversion. During this time, the Severn and Clywedog were diverted and began to flow eastwards.

4.4. THE PLIESTOCENE LEGACY

The Pleistocene chronology of Wales, like most of Great Britain is still subject to controversy. Substantial reviews are provided by Bowen (1973, 1974) and Lewis (1971). A major legacy of the Pleistocene has been the deposition of coarse superficial drift which has since been subjected to post-glacial weathering processes and Holocene fluvial modifications.

Plynlimon was originally conceived as a centre of ice dispersion, from which radiate remnant trough-shape valleys typical of glaciated regions. However, deposits of indisputable glacial origin are rare: many described as till or boulder clay deposits make up the slopes of concave valley sides and are now thought more likely to be depositional sequences of scree or drift. Arguments central to this are presented by Watson (1970, 1971) who has interpreted the stony-clay character of these deposits as soliflunctional head in origin. On the other hand, Bowen (1974) suggests that the blue head/till deposit commonly found in river bank sequences exhibits sedimentological and morphological characteristics which support a glacial origin but that such a Devensian ice coverage would be masked by the overwhelming periglacial inheritance. Rounded pebbles found within these deposits are, frequently, deeply striated; genetically similar to fluvioglacial outwash material, though the lack of preferred orientation and bedding within drift terraces suggest that these may have been redeposited through solifluction from upslope. These inferences are supported by Vincent (1976) and Bowen (opp. cit). Positive evidence for a late Devensian glaciation of South Wales is discussed by Archer (1968) and Bowen (1970), for the mid-Usk by Williams (1968) and Lewis (1971), and for the Welsh borderland by Bowen (1973).

Newson's review (1976) suggests a Pleistocene area of shallow valley glaciers through the Devensian surrounded by low level, active periglacial and ice margin features which later ornamented a fundamentally glaciated landscape. As a consequence, the Tertiary surfaces discussed in section 2.2, are still intact; the main valleys of the upper Severn and Wye do not show the typical trough shape as expected of extensively glaciated areas. Northward facing slopes maintain a terrace cover of stoney clay till, though are significantly disturbed as a result of late Pleistocene solifluction processes. Elsewhere, solifluction has removed the drift mantle from all but the lower concave valley sides leaving a stratified scree cover of gravel sized fragments of shale and mudstone which is frequently gullied or spreads as fans across the main valley floors. Where gritsones go to form the matrix of these deposits, the texture is crudely blocky and angular.

To the north-east, where the catchment descends to the piedmont valley of the Severn, the terrain maintains the characteristic Tertiary erosion surfaces though the basal valley slopes contain extensive terrace deposits. These are particularly marked along the south-west facing slopes of the area around Caersws (Grid. Ref. SO 033918). Elsewhere, the Pleistocene has left a hummocky, drumlinoid terrain within a central 'till' basin. The main Severn valley is formed on extensive river drift interrupted in areas by glaciofluvial gravel deposits, bordered by marginal kame terraces.

The nature and distribution of these deposits may be used in the reconstruction of palaeohydrologic environments of mid-Wales (See Lewin, 1977; Gregory, 1979). The Devensian environment, whether glacial or periglacial provides the scene for considerable redistribution of coarse sediment loads. Periglacial and/or paraglacial processes provided high local supplies of material, as seen above. Periglacial hydrology is typified by high seasonal fluctuations of discharge with flow rates far in excess of contemporary temperate regimes. All Welsh rivers were, at sometime during the Devensian, subjected to periglacial regimes. Furthermore, flows generated from paraglacial sources would be augmented by outwash or seasonal and diurnal fluctuations in meltwater, from this, it may be concluded that Devensian fluvial geomorphology consisted of high energy, braided regimes with flows (and sediment transport rates) far in excess of those achieved at present. In contrast, Holocene climatic amelioration suggests a hydrologic regime similar to that of today with the exception of short-term changes in flow and sediment transport rates produced by changes in vegetation (blanket peat development <u>c</u> 5000 BP.) and the Neolithic activities of man. As a consequence the contemporary channel, with a much reduced seasonality of flow and less significant slope input of material, is engaged in sorting and reworking the Quaternary sediment load through the system. Hence the routing and storage of coarse sediment is of greater interest than the continuous throughput of sediment loads.

4.5. THE POST-GLACIAL PERIOD: SOIL AND VEGETATION DEVELOPMENT

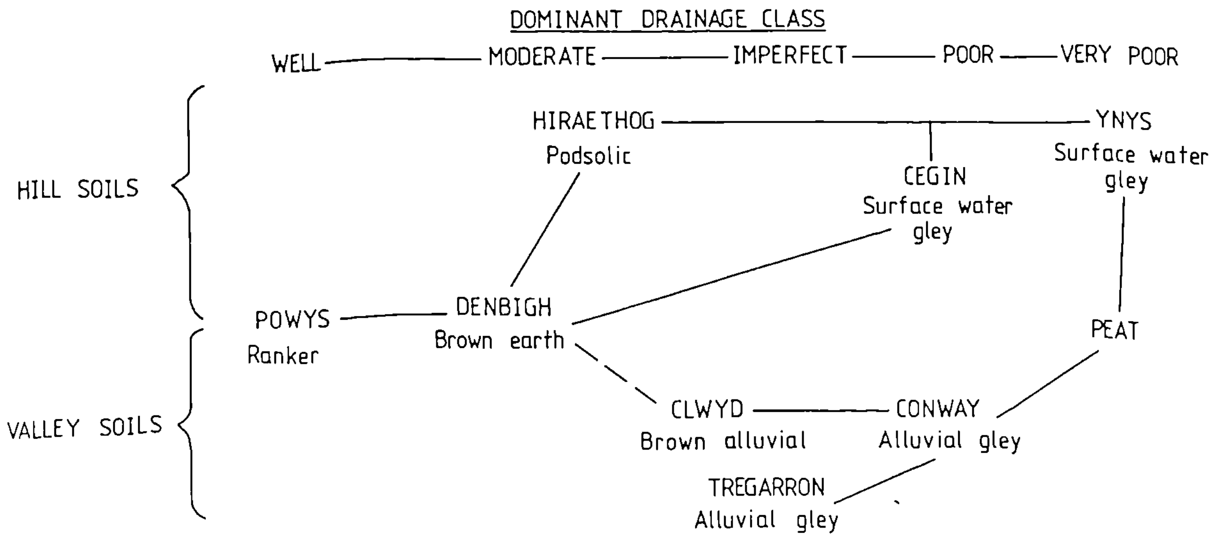
The Post-glacial, present interglacial, period provides the background for the development of the soils and vegetation of the region together with the accentuation of alluvial deposits in the piedmont channel reaches. Like much of upland, oceanic North-west Europe, Plynlimon exhibits areas of blanket peat; particularly extensive over plateaux and gently shelving depressions, and shallower or absent on the steeper slopes. These deposits provide a chronology spanning the 11,000 years B.P. described by Moore (1968,1970,1973,1977), Taylor (1973) and Smith (1970).

Deteriorating edaphic conditions c 5000 B.P. provide the basis for initiating peat blanket cover; the formation of a raw, mor humus over developing peaty podsols and accompanied gleying caused through the decay of the A2 horizon. This development has been accelerated by successive interference; not least due to further climatic deterioration c 2500 B.P. and as a consequence of Neolithic forest clearance and grazing producing substantial rises in watertable levels and increasing water availability. This has provided a peat soil system dominated by the Caron, Ynys, and Hireathog series. Elsewhere, the activity of post-glacial weathering processes has served to produce a finer, cohesive soil mantle over coarse Pleistocene superficial deposits.

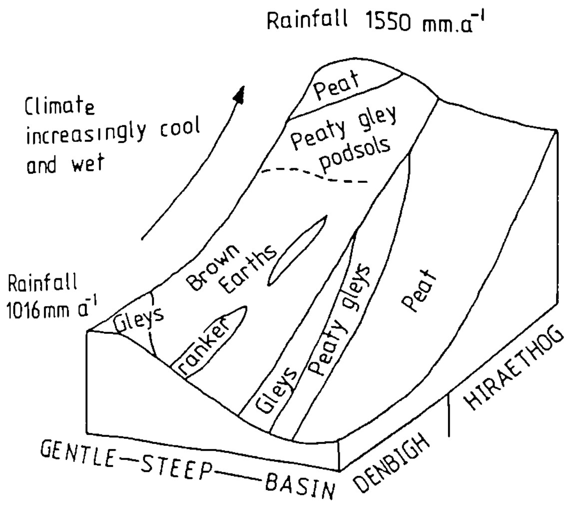
The main soil groups of the region are discussed in the Soil Survey Record (No 28) and by Rudeforth (1970). These are described schematically in figure 4.2.

The weathering of Lower Paleaozoic bedrock provides most of the

A. DRAINAGE AND SOIL TYPE



B. TOPOGRAPHY AND SOIL TYPE



SERIES	TYPE	LITHOLOGY
Powys	Ranker	Silt-clay loam over Lower Palaeozic mudstones & sandstone
Denbigh	Brown earth	Silt-clay loam over drift
Hiraethog	Peaty gley podsol	" " " " "
Cegin	Gley	" " " " '
Ynys	Peaty gley	" " " " "
Peat	-	-
Conway	Gley	Silt-clay loam over alluvium from lower palaeozoic rocks
Clwyd	Gley brown earth	" " " '
Tregarron	Gley	" " " ' (with peat layers)

FIGURE 4.2. THE RELATIONSHIP BETWEEN THE MAJOR SOIL ASSOCIATIONS OF THE REGION TO DRAINAGE (A) AND TOPOGRAPHY (B). (AFTER RUDEFORTH (1970) AND SOIL SURVEY RECORD N$^{\underline{O}}$ 28 (1975))

minerals found within the soil profile; principally mica, quartz, orthoclase, limonite, ilmenite, magnetite and pyrites with inclusions of the clay minerals illite and chlorite. Soil texture depends mainly upon the degree of weathering of the shale bedrock and for the most part will be found as medium to fine silt. Where, as in many instances, the soil has developed over superficial deposits the main influences on the character of the profile are in texture and permeability.

The most important determinant of soil type and development to the area is moisture regime, particularly its downslope transfer as the underlying lithologies are only slowly permeable. On convex and upper plane slopes, moisture saturates the colluvium for short periods and oxidative weathering allows the accumulation of Ferric hydroxide. As the colluvium thickens downslope, the soil profile changes from rapidly transmitting water to periodically saturated or saturated within hollows. Such conditions set up a reducing environment which inhibits chemical and bacterialogical formation of hydrated ferric oxide or hydroxide. Instead, the profile is characterised by an olive-grey colour with occasional ocherous stains where macropores permit rapid by-pass flow of soil water and consequent re-oxidation of minerals.

Regional variation in moisture through precipitation and evapotranspiration also provide distinct soil variations, Newson (1981) discusses the variation in spatial intensity of precipitation between the upland and piedmont areas of the River Severn. Cool, wet upland soils are if not blanket peat, podsolised soils as a result of the inability to break down organic matter in the mineral soil under the prevailing conditions. The extent of podsolisation is closely related to the degree of slope; steeper slopes provide a much more rapid transfer of soil water, leachingis reduced, and thus podsolisation is often less well developed.

The soils of the piedmont areas have developed over extensive drift deposits of till, gravel terraces and glaciofluvial gravel deposits. The main valley soil associations are derived from alluvium over coarse gravels. The reduced rainfall and generally warmer climatic conditions, as compared to the uplands, stimulate the activity of soil micro-organisms and this, together with a slower rate of leaching, allows adequate mixing of the mineral soil. For the most part, this results in an even distribution of ferric hydroxides down through the soil profile, producing a yellowish-brown Brown-earth soil type. These may be distinguished by their ability to incorporate surface organic matter and drainage class, related to slope and substrate type. On the hillslopes bordering the piedmont floodplain the soil drainage may be impeded by the occurrence of an impermeable till fairly close to the surface, which has resulted in the development of a surface-water gley soil (cambic stagnogley). The extent of these soils reflects the distribution of the impermeable till over gentle to moderate valley side slopes. Elsewhere, on the steeper, well drained slopes, the soil units are predominantly brown-earths. This association with drainage is repeated for some soil associations in the valley floor areas where soils have developed over more readily drained raised gravel terraces associated with Quaternary glacifluvial deposits on the valley floor. However, over much of the wide alluvial floodplain, high water tables and periodic flooding have resulted in gleying at depth producing an alluvial ground-water gley soil.

The geomorphological significance of these soil types and their associations with Quaternary deposits are discussed more fully in section 4.7. At this stage, it is apparent that in the uplands, once the relatively inactive organic surface is breached, rivers may provide themselves with a variety of sediments as a result of their own incision (see Lewin, 1981). The coarse nature of the upland deposits, however, means that the upland sediment system is dominated by bed-material loads.

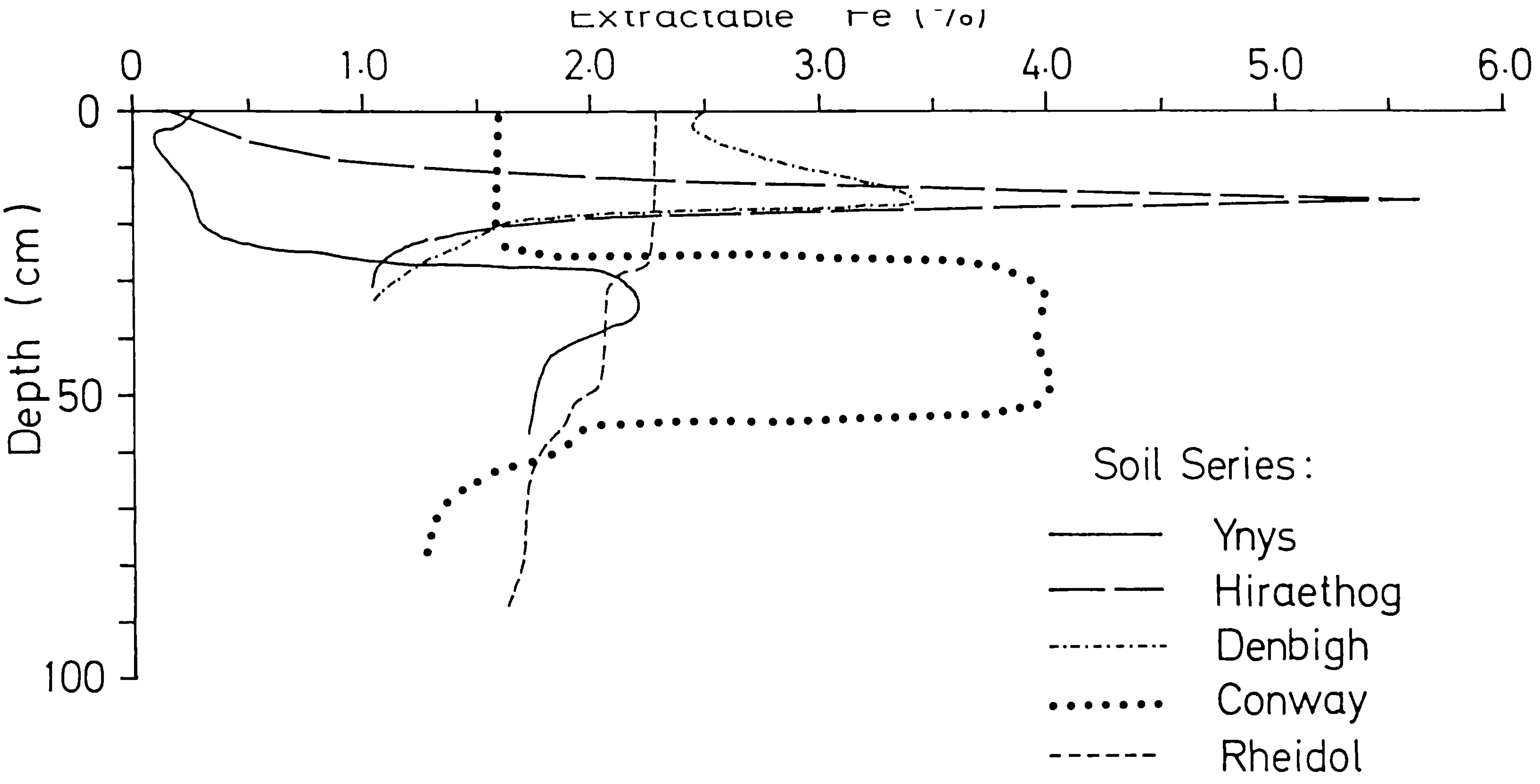


FIGURE 4.3. TOTAL PERCENTAGE OF EXTRACTABLE Fe BY DEB'S (1950) METHOD FOR THE MAJOR SOIL ASSOCIATIONS. (see text)

By contrast, the piedmont system is engaged mainly in the reworking of deposits derived from upland flooding. Coarse sediment material moves only slowly and infrequently through the piedmont channel and in doing so is incorporated into meso-scale bedforms as channel bars. Local deposition and channel bed aggradation may force bank erosion of the valley deposits and provide increased loads of suspended material (as compared to the uplands), derived from the fine, valley alluvial soil associations.

Some broad inferences as to the nature of the magnetic mineral characteristics of these soil associations and thus of potential erosion source areas, may be made with reference to Chapter 3 and to the soil forming processes described above. The total percentage of extractable iron (by Deb's (1950) method) in the profiles of some common soil types is shown in Figure 4.3. (Soil Survey, opp. cit). Whilst this is not an indicator of magnetic characteristic, it does serve to highlight the broad differences between soil types. The upland soil associations, Hiraethog and Ynys, have a very low iron content in the upper horizons due to the organic nature of the soils. The peak iron concentration occurs at depth, associated with elluviated (A&B) and iron pan (Bfe) horizons in the Hiraethog soils, and with the Bg horizon of the Ynys soil associated with periodic saturation causing reduction and segregation of iron in the profile. In each case, the peak iron content is associated with the reoxidation and deposition of hydrated ferric hydroxide and ferric oxide. The exact mineralogy is not determined by the Soil Survey, although they do suggest that stagnogley soil associations contain some lepidocrocite (FeO.OH). The magnetic behaviour of these minerals is antiferromagnetic and will be characterised by very low mass susceptibility (of the order of 0.1 to 1.25 $\mu m^3.kg^{-1}$) and high coercivity (see figure 3.2.).

By contrast, the piedmont soil associations have a much higher surface horizon content of iron, primarily because of the overall inorganic profile but also due to the lack of leaching. Ferric hydroxide is found throughout the soil profile but towards the surface insitu weathering and increased microbial activity (as compared to the uplands) combine to produce a much higher iron and sesqui-oxide content. As a result, much more enhanced magnetic characteristics should be expected in the piedmont soils, increasing towards the surface in association with the development of a distinct topsoil horizon. In such instances, the mineral magnetic properties would be of a much "softer" nature than that of the upland soils with generally lower 'S' and B_O(CR) values and increased surface susceptibility values. However, where the profiles exhibit gleying a hard remanence (high B_O(CR) and 'S') and reduced susceptibility should be expected. Field and laboratory study of these characteristics is described in Chapter 8 .

4.6. CONTEMPORARY VEGETATION AND LAND USE

In common with the majority of mainland Britain, the vegetation of the research area has been considerably modified by both man and animals. Cultivation has modified much of the piedmont and lower valley landscapes, whilst grazing and more recently, afforestation have altered the upland plant communities.

In the uplands around Plynlimon, the remnant semi-natural vegetation pattern shows clear interrelationships between soil type and soil microclimate as described earlier in this chapter. They may be simply classified into habitats based upon drainage class and altitude with their associated dependance upon slope, aspect and soil profile thickness. Three major upland plant communities may be identified (see Newson, 1976):

i) Semi-natural Grassland: which covers the largest area of the

Wye catchment and is made up of Festuca-Nardus Nardus-Festuca plant associations on long, well drained slopes with podsolised soils. This group also contains a number of arctic alpine species.

ii) Mire : which is typically located in valley bottoms and are thus mesotrophic (ie they receive both rainfall and throughflow inputs of water and nutrients). The community is dominated by Juncus sp. and Eriophorum sp. and may be found in association with the main stream heads of the catchments.

iii) Heathland : this community is dominated by Eriophorum sp. associated with Vaccinium , Calluna Vulgaris or Nardus sp.

These communities cover much of the upland Wye catchment which is itself part of a 20.24 km^2 area of farmland which is being developed for mountain sheep grazing.

The upland Severn catchment, has been used as a major forestry plantation since it was purchased by the state in the 1930's. The main species growing is Sitka spruce (Picea Sitchensis) which has been planted at various times with Norway Spruce (Picea Abies), Japanese Larch (Larix Kaempferi), Scots Pine (Pinus Sylvestris) and Lodgepole Pine (Pinus Contorta). Since the 1940's, the plantation has been pre-ploughed by tractor and, more recently, intensively drained by open ditching. The geomorphological implications of these operations is discussed further in section 4.7.

To the east, the piedmont Severn valley is used predominantly for farmland, most of which is under permanent pasture cover for beef, dairying and sheep production. Some larger farm units produce arable crops such as wheat, barley and winter rye for use as early spring grazing. There are few commercial forest or woodland areas in the Severn valley; some farm owned units are managed at the flanks of the valley floor, producing Beech (Fagus sylvatica), larch (Larix decidua) and Hazel (Corylus avellana).

4.7. CONTEMPORARY FLUVIAL GEOMORPHOLOGY

4.7.1 INTRODUCTION

The contemporary fluvial systems of the Wye and Severn, within the study area, may be simply characterised by their form and process into mountain and Piedmont streams (Newson, 1981). In section 4.3. it was shown that the morphology of the river channels is intimately related to stages of Tertiary peneplanation though more significantly to river capture during the Pleistocene. The distinction may be simply illustrated by Figure 4.4. which shows the two main slope changes in long profile for the Wye and Severn. The mountain stream sections are characterised by steep channel gradients together with steep valley-side slopes (section 4.3.). Piedmont streams, although developed on a smaller scale in the U.K. as compared with the traditionally accepted piedmont channels of the USA, show a noticeable reduction in channel gradient and, for the most part, flow through a wide floodplain.

The upland channels of the Wye and Severn fall into two further categories:

i) forest drainage ditches:with banks up to 2 metres high in peat and colluvium with a bed of fine to coarse gravel overlaying the shale bedrock; and,
ii) the main upland sub-catchment tributaries of the Tanllwyth, Cyff, Gwy, Nant Iago, Hafren and Hore.

The natural channels combine a mix of rock controlled reaches providing an irregular planform in which the rock itself also comprises the bed. Loose, eroded bed-material is stored within the channel as fine to coarse gravel shoals. Sediment movement in these reaches occurs in response to rapid and extreme upland flooding. As the channel develops

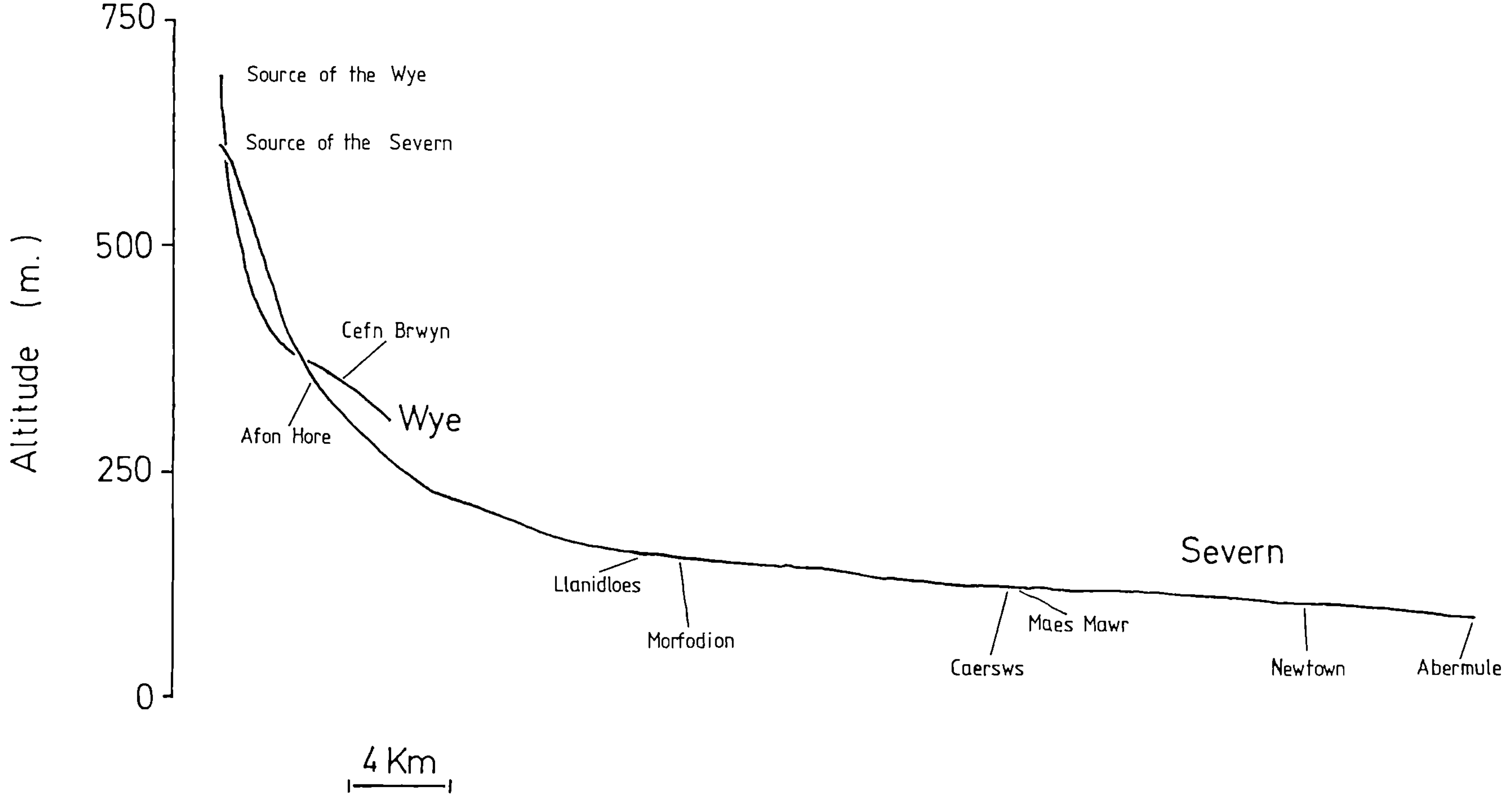


FIGURE 4.4. CHANNEL LONG - PROFILES FOR THE STUDY REACHES ILLUSTRATING THE POSITIONS OF THE STUDY REACHES AND AREAS OF PREVIOUS RESEARCH DISCUSSED IN THE TEXT.

downstream, bedrock may still exert considerable control but the reaches are more typically characterised by a wide shallow cross-section in a meandering planform which may be locally divided by gravel bar deposits. In some locations, the meanders have become confined as they impinge upon large scale fluvio-glacial deposits, whereupon the development of large bluff faces occurs (Lewin, et. al., 1974).

The piedmont channels of the Wye and Severn are actively meandering through a floodplain of fine alluvial sands, underlain by coarse colluvial sediment, much of which goes to make up the channel bed. Diagonal and point bars are the most common feature of the storage of coarse gravel in the piedmont reaches. The channel is confined within steep to vertical banks of the composite floodplain sediments which provide a rapidly eroding cliff face some 1-2 metres in height. The hydrological processes in the piedmont zone are intimately related to phases of upland flooding, though for the Severn discharges are a combined effect with the regulation operations of the Clywedog Rese oir. The Reservoir, completed in 1968, is used for flow augmentation for the major abstractors enroute to Bristol, and to provide flood mitigation for those areas within the upper reaches of the Severn. The reservoir maintains a statutory minimum flow (measured at Bewdley, 160 km downstream), slightly higher than the natural flow at times of drought to provide adequate water for supply purposes. Natural, or unhindered flows are still provided by the upland tributaries of the Severn and Afon Dulas.

4.7.2 HYDROLOGY

The westerly position of the catchments in Great Britain provides them with seasonally affected rainstorms of long-durations particularly during the winter months of November to January which provide 40% of the total annual rainfall. However, in terms of geomorphological effectiveness, infrequent convective downpours during the summer months

	Plynlimon Catchments		Piedmont River Severn	
	Severn	Wye	Trefeglwys	Newtown
Relief (m)	409	399	146	98
Mean Annual Rainfall (mm)	2230	2349	1125	923
Mean Annual Runoff (mm)	1398	1964	c 1200 *	
Drainage Area (Km^2)	8.92	10.82	187 *	
Approx. Stream Slope ($m.Km^{-1}$)	80.6	71.4	2 - 10	2.0
Network Power	92	69	-	-

Sources :

Clark & McCullogh	1979
Gregory	1979
Newson	1981
NERC Flood Studies Report	1975
Soil Survey Record 28	1975

* Severn Trent Water Authority gauging station at Dolwen (GRID REF. SO997852), period 1980-1981.

TABLE 4.3. PHYSIOGRAPHIC CHARACTERISTICS OF THE STUDY AREA.

may be of considerable significance (Newson, 1980a). The brunt of these episodic downpours is restricted largely to the uplands due to their westerly position; further eastwards, precipitation on the piedmont zones is significantly reduced (approximately 40% of that falling in the uplands, c.f. Table 4.3.).

Upland Runoff in response to these episodic outbursts, although dependent upon antecedent conditions aswell, is equally rapid. Highly permeable and shallow slope deposits, coupled with the affect of saturated heaths and mires in the uplands, respond rapidly to infiltrating stormwater (Knapp, 1974) providing a form of 'physiographic aggravation' to upland flooding (Newson, 1976). The Flow ranges in the Plynlimon catchments are frequently in excess of three orders of magnitude, with flood peaks reached in under one hour. These flows may, to some extent, be augmented by pipeflows caused through the development of large macropore systems in the peaty soils (Gilman and Newson, 1980).

The widespread deposits of ombrogenous peat above 300 metres plays a vital role in water transfer/supply in the uplands; peat is able to store water then release it slowly during drier summer conditions thus maintaining streamflow. Detailed hydrological properties of the peats are difficult to define. They are dependent upon the degree of decomposition (Wilcock, 1979) aswell as topographic factors (Taylor & Tucker, 1973). Highly fibrous, relatively undecomposed peats have higher hydraulic conductivities, but at the same time may retain much more water at saturation than a more decomposed deposit. The peat deposits on the Plynlimon range are dominated by the Caron and Hiraethog series (c.f. section 4.5.). The former comprises peat/peaty gley deposits, generally in excess of 40 cms depth, capping the interfluves, whilst the Hiraethog is a complex of peaty podzols covering gently sloping areas (10-15° slopes). Whilst both are classed as poorly draining deposits, the latter has been associated with the development of extensive soil pipe networks (Gilman & Newson, opp. cit).

The recent rapid expansion of upland land-use, for forestry, improved pasturing or water regulation, has necessitated developments in techniques to improve the drainage of the naturally boggy terrain. For the Plynlimon catchments, this has been attempted by deep ploughing of open drainage ditches. The consequences of draining in this fashion are dependent upon topography and stage of land-use development. As some 20% of the land surface, in the first instance, is covered by drains, direct storm runoff is increased and the timing to peak discharge is reduced (Binns, 1979). For a forested catchment, as the stand grows, the closing canopy reduces this effect and this together with an increased evapotranspiration rate gradually reduces the enhanced flashy regime. With a mature stand, the increase in evapotranspiration becomes considerable; Clarke and McCullogh (1979) quote evapotranspiration rates for the forested Severn Catchment at 21-28% higher than those for the adjacent pastured Wye. These results are in accordance with Laws (1956) hypotheses and suggest an annual loss in runoff of some 12% of potential available water.

In comparing the hydrographs for the two catchments, there appears to be little difference in storm hydrograph dimensions between the forested and ditched Severn as compared to the unditched, pastured catchment of the Wye (Newson, 1980a,b). However, the longterm response for water supply purposes clearly questions the role of afforestation in catchment headwaters.

With regard to the flow regime of piedmont channels, we are mainly concerned here with the River Severn extending downstream to the Abermule gauging station (Grid Ref. SO157945). Whilst stormflows are still highly dependent upon upland flooding, the effect of the Clywedog operations is in reducing the sudden rise in stream stage during high flows and also to reduce the extremes of high flows and augment those at lower flows. Nevertheless, the full potential of the Clywedog reservoir has yet to be realised (Hey, 1984). However, any proposed increases in flow regulation could have significant impact on the piedmont channel; there is already evidence of erosion of the channel downstream from the outfall to Llanidloes Bridge (Gr.Ref. SN953844) and considerable deposition below this point to Morfodion (G.ref: SN 974857). Considerable changes in channel geometry occur downstream of this point associated with stable and unstable channel regimes. In constructing flow duration curves for the River Severn, based upon water demand predictions to 1999, Hey suggests the long term effects of regulation will be unnatural amounts of erosion and deposition because the frequency of flows above the threshold for bedload transport will be increased. Sediment tracer studies carried out by Hey under controlled releases from Clywedog indicated that the threshold discharge for bedload transport occurred at approximately 12 cumecs (or 1000 Ml d^{-1}). This figure is only slightly less than that experimentally derived by Thorne and Lewin (1979) of 17 cumecs for sites in the region of Maes Mawr and Caersws on the River Severn (c.f. Figure 4.1.).

4.7.3 GEOMORPHOLOGY

The division of mountain and piedmont streams, which were discussed at the beginning of this section, may be taken one step further by analysing the changes in geomorphic processes. The intensity of rainstorms over the upland catchments, suggest that these are the areas prone to high rates of erosion and transport of material. Indeed, Harvey (1974) suggests that erosion in upland Britain operates at a greater pace than generally realised; in mid-Wales a typical range of depths of erosion spans 0 - 7.5 cm.a^{-1}. (Slaymaker, 1972). However, from the preceeding discussions, it can be seen that the most likely sources of material are frequently concealed beneath a cohesive mantle of peat or podzolic deposits. Beneath these, the Quaternary has provided a wealth of coarse, heterogenous material providing the potential for a range of sediment loads, as either bed or suspended sediment. The efficiency of the peaty mantle, means that 'natural' erosion rates are low; for example:-

SITE	ANNUAL YIELD* $(m^3.Km^{-1})$	RESEARCHERS
AFON CYFF	2.5	Newson (1981)
MAESNANT	1.1	Lewin et. al.(1974)
NANT IAGO	1.2	Lewin & Wolfenden (1978)

* (Annual Yield of Gravel Bedload)

Lewin et.al.(1974) monitored sediment yields for the Maesnant catchment on the western slopes of Plynlimon. They were able to identify three main forms of bank erosion, the dominant source in each case being pre-weathered soliflual material:-

1) Basic, eroding channel banks up to a metre in height, cut randomly in local material.
2) Large bluff faces up to 4.8 metres high by 10 metres in length, being actively undercut by stream action.
3) Crescentic slips (sheep hollows) which may supply material at extreme flows.

Even on less stable nineteenth century mine spoil heaps, Lewin and Wolfenden (1978) recorded sediment yields as low as $1.2m^3km^{-1}a^{-1}$.

By contrast, erosion rates recorded in catchments with considerable drainage ditching operations ranged between 2.0 to 308 $m^3km^{-1}a^{-1}$. Newson (1980b)suggests that where the cohesive surface layers of peat are broken, such as in the case of open ditching for forest drainage, increased bedload yields are a result of increased erosional source areas. The Institute of Hydrology (Plynlimon) has been monitoring sediment yields from the Afon Cyff (Pastured) and the Tanllwyth (Forested; ditched) since 1973 to isolate such patterns. The results from these observations are discussed in detail in Chapter 5. The lack of any observable difference in hydrograph dimensions between the two catchments, but a much higher drainage density, as a result of ditching, in the Tanllwyth suggests a much larger area for potential bedload erosion. As opposed to the results discussed by Binns (1979), Newson's data indicates that ditching may continue to impact well into maturity of the forest stand, although the periodsof activity are difficult to predict.

Bedload yields are not completely event-dependent; even with events of similar magnitude, yields may vary considerably. Antecedent conditions appear to play a vital role in providing transportable material (c.f. Chapter 2.). The typical upland substrate of peat over less cohesive colluvium may be susceptible to increased erosion because of the response of the large intergranular voids to variations in soil moisture. Gilman and Newson (1980) identified periods of dry and wet extremes vital in producing peat cracking and providing macropore structural development for piping. The seasonal effects of frost in structural heaving, followed by rain and melting sequences, in providing available sediment was identified by Lewin et.al.(opp. cit). Similar processes have been identified by Ball and Goodier (1970) operating on the hillslopes of Snowdonia, and by Harvey (1974, 1977) in the Howgill Fells, Westmorland. Harvey suggests that the frequency of events for slope production of coarse material is high, whilst that of floods competent to shift the available load is very low (of the order of one per annum). The efficacy of a flood event is intimately related to antecedent conditions (both of moisture and event frequency) aswell as event magnitude. Newson (1980a) has illustrated that periods of high antecedent moisture conditions followed by intense rainfall are associated with considerable slope failure and supply of sediment to the river channel. Conversely, short, intense storms during relatively low antecedent moisture conditions are associated with major impact on the river channel sediments. Such events provide the main stimuli to bank undercutting and failure and the redistribution of channel bed material producing significant channel change. Similar patterns have been observed in the Afon Crewi, west Wales (Blacknell, 1981) where bank failure was enhanced by the combined action of frost heave in an already wetted soil and increased hydraulic stress applied during subsequent floods.

We have seen that the hydrological processes of the piedmont zone are dependent upon the combined effects of upland flooding together with flow regulation by the Clywedog resevoir. Piedmont geomorphology in response to this flow regime, involves the reworking of previously derived upland sediments together with the increased supply of coarse material as a result of contemporary upland land-use practises. The coarse sediment load moves slowly and infrequently through these reaches, and for the most part is stored as channel bars of an intermediate scale. Observations regarding the dynamics of these features are few but suggest that they are stable throughout the majority of high flows (Church and Jones, 1982). As a consequence, the sediment transport system of the piedmont channel is characterised by a higher ratio of suspended to bedload movement (McManus and Al Ansari, 1975).

Thorne and Lewin (1979) were able to identify a number of thresholds of scale of sediment movement and channel change for the piedmont Severn using painted tracer pebbles. The threshold for incipient motion of the bed sediment was in the region of 17 cumecs. At these flows, transport is concerned mainly with the redistribution of loose available material, accumulated as a result of bank failure and winnowed from an essentially armoured bed. Major planform changes occur at discharges in excess of 65 cumecs ie. greater than bankfull discharge. Such extreme events are very rare (Lewin, 1978), but may be associated with rapid bank recession, cut offs and flood chutes. Intermediate discharges (between

12 - 70 cumecs) are associated with an orderly development of the meandering planform. Bedload transport at these discharges is spatially and temporally variable and occurs mainly in response to hydraulic conditions at meander inflexion points. This is particularly concerned with the development of secondary circulation cells which produce increased shear stress at the outer bank of the meander and are effective in transporting loose debris accumulated by bank slope failure and increasing bank erosion. Basal scour and the disturbance of the major sediment storage features is limited to the extreme events; for the majority of flows, sediment is derived from the channel banks. Increased shear stress at the channel banks may also be created as a result of flow divergence around the more stable gravel bars.

Bank erosion on the Severn is exacerbated by their composite profile of fine alluvial sands overlaying coarse gravel colluvium. The latter are highly susceptible to erosion causing severe undercutting of the bank sections and large overhanging blocks which are liable to failure and collapse. The latter process is derived through mechanical rather than fluvial forces (Thorne, 1982). Gravitational strain of the overhanging block is aided by weakening along ped structures. Cycles of wetting and drying cause swelling and shrinkage of the interped fissures and the development of larger desiccation cracks. Consequently, the added strain through the weight of draining water following flood recession or high antecedent precipitation conditions may aid failure. Rates of bank erosion as a result of fluvial undermining and mechanical failure range between 300 and 600 $mm.a^{-1}$. This compares with a much lower rate associated with weathering alone, of 20-200 $mm.a^{-1}$. (Thorne and Lewin, opp. cit).

Bank erosional processes may be increased by the lack of available fine sediment material for transport. Finer sediments are noticeably rare on rivers draining the palaeozoic shales of mid-Wales (Lewin, 1978). In addition, one of the effects of flow regulation is in reducing the supply of finer material from upstream (Grimshaw and Lewin, 1980). Further to this, there is evidence to suggest that flow regulation by the Clewedog may exacerbate these effects even though the regulation potential has yet to be fully realised (Hey, 1984). Erosion at the outflow area has caused significant deposition downstream of Llanidloes (see earlier) and areas of unstable planform are being rapidly eroded by intermediate, increased regulated flows and by flow divergence around coarse bed material features. The lack of available fine material for transport may also be associated with channel armouring which is suggested to be a corollary of flow regulation (Livesey, 1965). As a consequence, regulation may significantly reduce the ability of a river to move its bed materials (Kellerhals, 1982) and the available stream power is effective in increased bank erosion.

Analyses of the historic and contemporary development of channel planform of river reaches in mid-Wales show varying rates of activity. In a study of 100 randomly selected reaches, Lewin et. al.(1977) showed that 25% of those reaches were actively meandering at rates of between 0.1 to 5.5% of the channel area per year. However, 75% showed no measurable change for periods up to 78 years. Channel planform changes in those active reaches are related to the stability, accumulation and development of gravel bar assemblages. In many instances, barforms may be stable with respect to planform evolution; bed sediment transport is little more than a local flux and flow divergence or constriction, as a result of coarse sediment storage, increases channel lateral migration and cut-bank recession producing high suspended sediment loads (q.v. Lewin, 1981). By comparison, Lewin (1978) describes evidence for the River Rheidol of bars moving through meander reaches atleast under extreme flows. In such instances, the bars may not be stably located with respect to planform (Lewin, 1981) and the passage of the bar (or bars) causes local channel deformation enroute downstream. The River Rheidol in particular, is characterised by an high rate of lateral mobility (Lewin, 1981) and whilst floodplain sediments illustrate an accumulation of coarse gravels in relic lateral and medial bar forms, the vertical changes in the short term (100 years) are not nearly as pronounced as lateral movements. As Lewin points out, however, "We do not yet have nearly enough information about these activity patterns and rates.." . This investigation was initiated to provide a technique to monitor gravel bed sediment transport and to identify the processes occurring at various stages within the contemporary system.

4.8. PURPOSE OF THE PRESENT INVESTIGATION

The main purpose of this investigation was to develop a procedure for magnetically tagging stream bed material and to test its efficacy within the gravel bed reaches of mid-Wales. The practical assessment of bedload transport is fraught with problems, particularly so in heterogeneous bed material. Of the available techniques, many have been developed for a narrow range of particle sizes. Magnetic enhancement is applicable to all grain sizes for the Plynlimon shales (Chapter 3) and is a non-destructive and environmentally harmless tracing system.

Five case studies have been carried out, the aims of each is dealt with in detail individually (Chapters 5-8 inclusive). The broader aims of these studies are complementary to those undertaken by IOH in establishing the MAFF Project 73 and are concerned with the understanding of the dynamics of sediment transport in gravel bed channels. Traces were initiated at several stages within the system described above:-

1) In the uplands : investigating the bedload dynamics in the rapidly eroding drainage systems and the sediment throughput in larger upland channels.;

and, 2) In the piedmont zone : to investigate the effect of channel storage elements under regulated flows.

Using the principles outlined by Walling et. al. (1979) the natural magnetic characteristics of suspended sediments were also analysed to identify source contributing areas and to isolate any stormflow variations.

CHAPTER 5

5. TRACER EXPERIMENTS IN ERODING FOREST DRAINAGE CHANNELS: THE SOURCE AREAS

5.1 INTRODUCTION

This chapter is concerned with the initial application of magnetic tracing under 'controlled' channel conditions. Tracing has been carried out within forest drainage channels; ditches cut into the upland peaty deposits of the Plynlimon range in an attempt to drain the naturally high water content. Bedload transport in these ditch systems is constantly monitored by "pit-slot" type bedload traps (c.f. Chapter 2). In this case, tracer movement within the channel may also be monitored by capture within the bedload traps.

Studies over the past ten years (Newson 1979, 1980b;Robinson, 1980, Robinson and Blyth, 1982) have questioned the efficacy of ditching practises in environmental management of the uplands, and suggest that (at least during the first five years of operation) these may do more harm than good by initiating erosion and supplying large volumes of coarse sediment to the main river channels.

This chapter describes the application of the magnetic tracing technique to the study of sediment transport within two such systems. In addition to this, the experiments were used as a pilot study of both tracer and instrumentation to aid the development of experimental and instrumental design and survey procedures for subsequent tracers.

5.2. DRAINAGE DITCHING AND INCREASED SEDIMENT YIELDS FROM SMALL CATCHMENTS

The natural high water storage capacity of British upland soils has been a major handicap to any land development, and not surprisingly, along with the expansion of forestry, agriculture and water resources into the uplands there has been a need for efficient drainage practices. Over much of Britain, this has been attempted by open ditching. Under forest and moorland, such a practise has served to increase erosion and consequent sedimentation.

The present semi-natural surface of much of the British uplands provides a cohesive matrix of pasture vegetation over organic soil horizons which are seldom breached by running water. However, if the surface is broken artificially, for example by ditching prior to afforestation, the underlying, much less cohesive glacial and periglacial sediments are exposed and erosion can be rapid.

In an attempt to assess the impact of these practises, the Institute of Hydrology (IOH) has, since 1973, continuously monitored sediment yield from two sub-catchments within the Plynlimon Experimental Catchment; the Cyff ($3.13Km^2$, pasture), and the Tanllwyth (0.89 Km^2, forested) (c.f. Figure 4.1.). Since the major sediment movement in these channels is in the gravel and coarser size ranges, up to boulders weighing 20-100 Kg each, bed-load has been the dominant interest. Suspended sediment concentration rarely exceeds 100 $mg.l^{-1}$ (see Chapter 8). Concrete lined bedload traps (capacity $10.5m^3$) have been operated by the Institute of Hydrology on both the Cyff and Tanllwyth to determine bed-load yields.

Painter et al (1974) showed that from an initial comparison of yields, the greater erosional activity of the forested Tanllwyth against the Cyff was apparent, and implied that the increased bed-load yield was derived from increased source area erosion associated with forest drainage ditches. In order to confirm this impression, IOH undertook a detailed erosion survey of individual ditch systems within the Tanllwyth equipped with small weirs and bed-load traps. These together with further reconnaisance studies using temporary traps installed in other ditched areas of mid-Wales are presented by Newson (1980b) The results may be summarised thus (Newson, Pers. Comm):-

1) In eight years of sediment trapping, the yield of the forested Tanllwyth is approximately five times that of the pastured Cyff.

2) Physiographic characteristics and delivery ratios do not explain the difference; this is probably due to the use of open ditches in the Tanllwyth to prepare the ground for planting trees.

3) Yields of sediment from drainage ditches in the Tanllwyth vary from 3.5 to 20.5 times those of the pastured Cyff. In places the cross-sectional area of the ditches has increased by more than 100% since construction in 1950; current yields suggest that erosion has since slowed as compared to the initial years (see Robinson 1980).

4) Elsewhere in mid-Wales, drainage ditching areas are equally well eroded, even on pasture where ditching is a cheap form of drainage for hill farmers.

5) Ditch design and siting could be improved to avoid erosion. Two common factors identified by Newson (1980b)are gradients in excess of 2° and the rupture of soil mineral layers beneath the peat surface during excavation.

Following the recommendations of the Forestry Commission (1979), the expansion of upland afforestation seems likely to continue, as does the improvement of land drainage for hill farming. Similarly, the technique of open ditching, which provides the cheapest and quickest option for upland drainage, is also likely to remain in practice.

Further studies of erosion and sediment transport in upland ditch systems have been carried out to assess the downstream impact of any increased sediment yield and to provide alternatives to drainage design.

5.3. BED-LOAD MOVEMENT IN ERODING DITCHES

From February 1976 to February 1981, three traps were operated on two individual ditch systems in the Tanllwyth area mentioned above, to identify the processes of erosion, storage and sediment transport in relation to individual flood characteristics. During the latter period, continuing observations were complemented by magnetic tagging

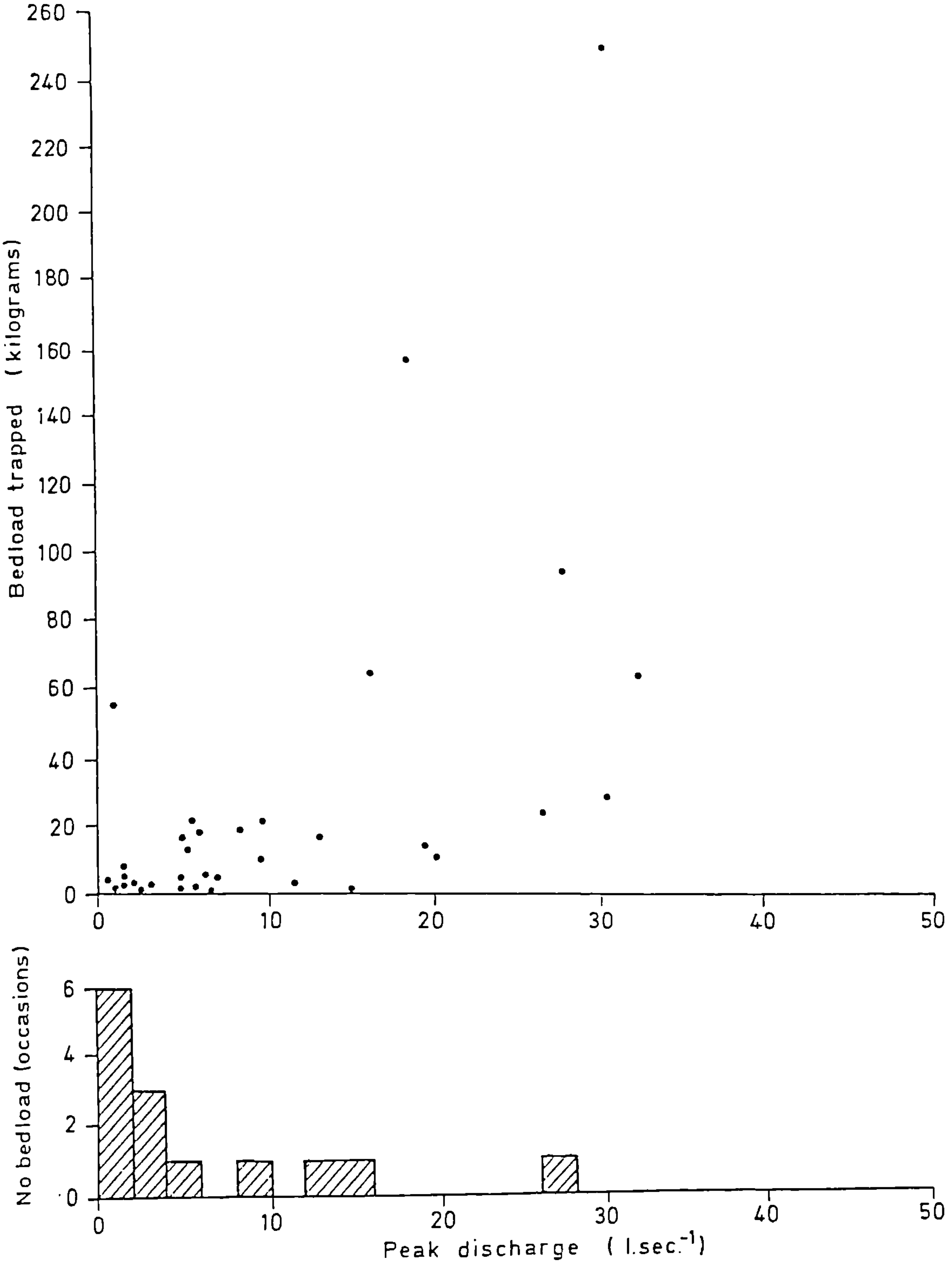


FIGURE 5.1. RELATIONSHIP BETWEEN PEAK DISCHARGE AND WEIGHT OF BED-LOAD TRAPPED FOR THE DITCH LTAN6'A, 1976 - 1981 (ABOVE); AND, THE NUMBER OF OCCASIONS FOR WHICH NO SEDIMENT WAS TRAPPED RELATED TO FLOW CLASS (BELOW).

of sediment within each ditch.

The relationship between the instantaneous maximum flow (measured at a small weir downstream from each trap) and bedload yield for each event during the period 1976 to 1981 are shown in Figure 5.1. Whilst a general trend towards a positive relationship appears, the deviation in the data shows the wide variation in the work achieved by different flood events, and in particular, the lack of any bed-load movement even during moderate flows. This implied that for a range of flows, bed material may become supply-limited. Indeed data from the main traps operated on the Cyff and Tanllwyth both support this, illustrating supply and transport-limited phases of channel activity (Newson 1980a).

These results should not be unexpected; hysteresis and supply-limitation phenomena are commonly observed with suspended load (Walling, 1974; Heidel, 1956) and Dissolved load (Hem, 1970; Hendrickson & Krieger, 1960), and it seems reasonable to expect similar phenomena with bed-load transport. However, whilst similar phenomena have been identified elsewhere (Emmett, 1976; Andrews, 1979) very little account is taken in conventional calculations of bed-load yield. As Emmett points out:

> 'At high values of streamflow .... bed-load transport rates are correlative with a predictable proportion of stream power expenditure..'

..but for a range of floods and even within a flood, predictions based upon bed-load equations can become unreliable. This may be particularly so in cases of short term variations in stream discharge (Chapter 4). The data would suggest that estimations of bed-load yield should take into account the supply of material to the channel and controls on its availability, once in the channel, for transport (Gomez 1979). Under these controls, the rate of sediment transport for low to moderate flows will be dependent not only upon the momentary discharge, but also on the recent history of the channel (Bogen, 1980). This may include seasonal changes in source-generating areas (Nanson, 1974), the interdependency of flood events (Leopold & Emmett, 1976; McGreal & Gardiner, 1977) and related changes in channel storage of material (c.f. Chapter 2).

Since sediment supply is difficult to measure continuously, the answers to these problems would appear to lie in the analysis of multivariate time series, serial autocorrelation techniques, or in the use of tracing techniques to investigate transport mechanisms in isolation from supply mechanisms.

5.4. MAGNETIC TRACING WITHIN DITCH SYSTEMS

Methodology:

Two experimental lengths of eroding forest ditch in the Tanllwyth system were chosen for study (Figure 5.2.):

1) LTan6'A : An extremely high energy system, geometrically simple and similar to a laboratory flume; the trace was carried out in 15 metres of ditch upstream from a bed-load trap;

and,

2) LTan : A low energy system, sinuous in plan and profile and eroded in places to rough outcrops of bedrock within the upper 30 metres of the experimental reach. The total trace distance to bedload trap was 60 metres.

The difference in energy between the systems, an assessment based on previous field evidence, was used to study the efficiency of the technique in tracing two particle size ranges : in the LTan, finer trace particle sizes between 1.4 - 11.1 mm; and, in the LTan6'A, coarser material between 5.6 - 44.5mm. The particle size distribution curve for the total size range of bed-load trapped for the period February 1976 to

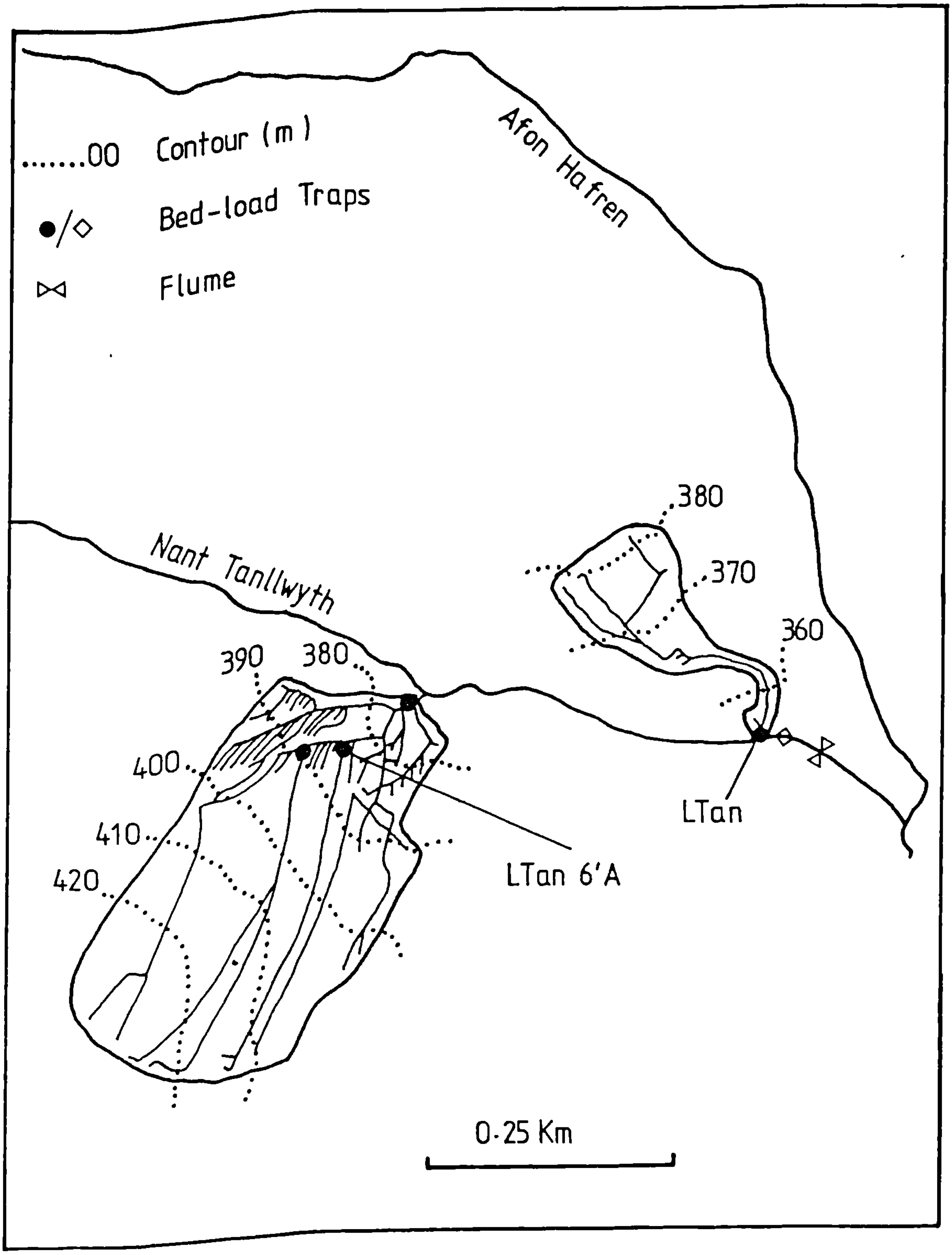


FIGURE 5.2. LOCATION OF THE DRAINAGE DITCHES LTAN AND LTAN6'A.

(c.f. Figure 4.1.)

SIZE RANGE	LTAN			LTAN6'A		
(mm)	(g)	(%)	Tracer Wt. (g)	(g)	(%)	Tracer Wt. (g)
1.4 - 2.7	300	5.8	300 *	700	9.1	
2.8 - 5.5	600	11.5	600 *	1050	13.7	
5.6 - 11.1	900	17.7	900 *	1750	22.7	1750 *
11.2 - 22.3	800	15.1		1400	18.2	1400 *
22.4 - 44.5	800	15.4		1400	18.2	1400 *
45.0 +	1800	34.6		1400	18.2	

TABLE 5.1. FOREST DRAINAGE DITCHES : PARTICLE SIZE CHARACTERISTICS OF THE TWO SHOALS REPLACED SHOWING THE WEIGHT OF TRACER MATERIAL PER SIZE RANGE USED IN EACH TRACER EXPERIMENT.

1980 is shown in Figure 5.3..

For each ditch, a substantial shoal was chosen as the emplacement site for tracer material. Gravel removed from the shoal was dry sieved for size then the specific size ranges to be monitored experimentally were enhanced and replaced within the sample. These data are summarised in Table 5.1.. The shoals were then reconstructed at their original positions and left to stabilise under the low summer flows of 1980.

Together with topographic survey, a background survey of surface susceptibility within each ditch was carried out with the search coil at a series of fixed points of 1 metre intervals between shoal and trap. Following each major flow period subsequent to the shoal emplacement, the surface susceptibility surveys were repeated at each fixed point and also where appropriate inbetween (ie. at points of shoal build-up between rock outcrops) to assess any transport of the tracer material. The bed-load traps were emptied on each occasion and analysed for the tracer material. Analysis was undertaken in two ways in accordance with the size range of tracer used. For material coarser than 4mm, identification by colour or surface susceptibility was subsequently confirmed by susceptibility sensing using the hand-held Ferrite Probe described earlier (Chapter 3.). Material separated in this manner was subject to further analysis of weight, size and shape. As a check on the accuracy of this separation procedure out in the field, for each trapped load a subsample (approximately 30 kg) was taken back to the laboratory, dry seived and checked for any remaining magnetic clasts. Any additional magnetic material found in the subsample was analysed for size and weight and calculated as a percentage of the subsample weight and of the trapped weight as a whole. The latter value was calculated as the additional weight of tracer material missed in the

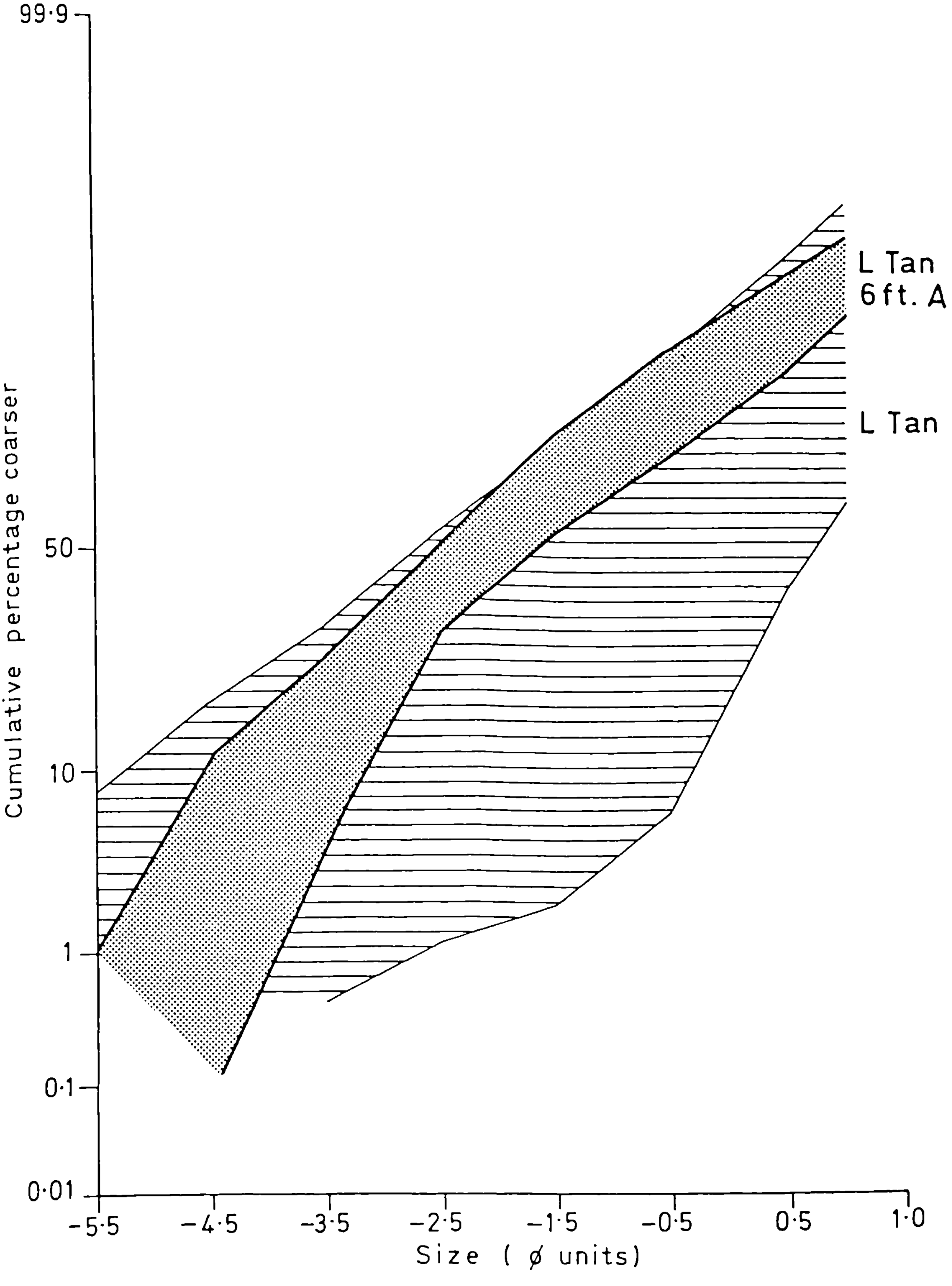


FIGURE 5.3. PARTICLE SIZE DISTRIBUTION CURVES FOR THE TWO STUDY SITES INDICATING THE RANGE OF MATERIAL BEING TRAPPED OVER THE STUDY PERIOD, PARTICULARLY THOSE OCCASIONS WHEN ONLY FINE MATERIAL WAS TRANSPORTED.

field analyses, and was used to correct the total trapped weight of tracer following each period of high flows. For each flood-load analysis, the average error in recovery was approximately 2.4% the recovery in the field approximately 93 - 97%.

For finer material than above, the 250 ml bottle sensor was used to measure the mass susceptibilities of dry sieved samples in each tracer range. By comparing the mass susceptibility vs. percentage concentration of tracer for each size range used in the trace, the total weight of tracer was estimated.

The ditches were monitored up to June 1981 covering seven sediment effective flood events following which attention was turned to the major traces discussed in Chapters 6-8.

5.5. RESULTS AND DISCUSSION

A summary of the data collected for each trace is presented in Figures 5.4 and 5.5.. These show the surface susceptibility change for each survey related to ditch topographic profile, together with cumulative plots of the volume of tracer trapped in each size range plotted against the storm hydrograph for the study period.

The cumulative plots of trapped magnetised material exhibit remarkably similar features though at different scales; the range of data for the LTan ditch was 0-200g whilst for the LTan6'A, 0-2500g. Trapped weights of tracer material for the first high flow event (7.10.80) were 82g and 2228g respectively. For both ditches, the graphs show the influence of three main periods of high flow early on in the experiment. More moderate events between and following these spates exhibited much lower recovery rates. These relate to individual characteristics of each ditch trace.

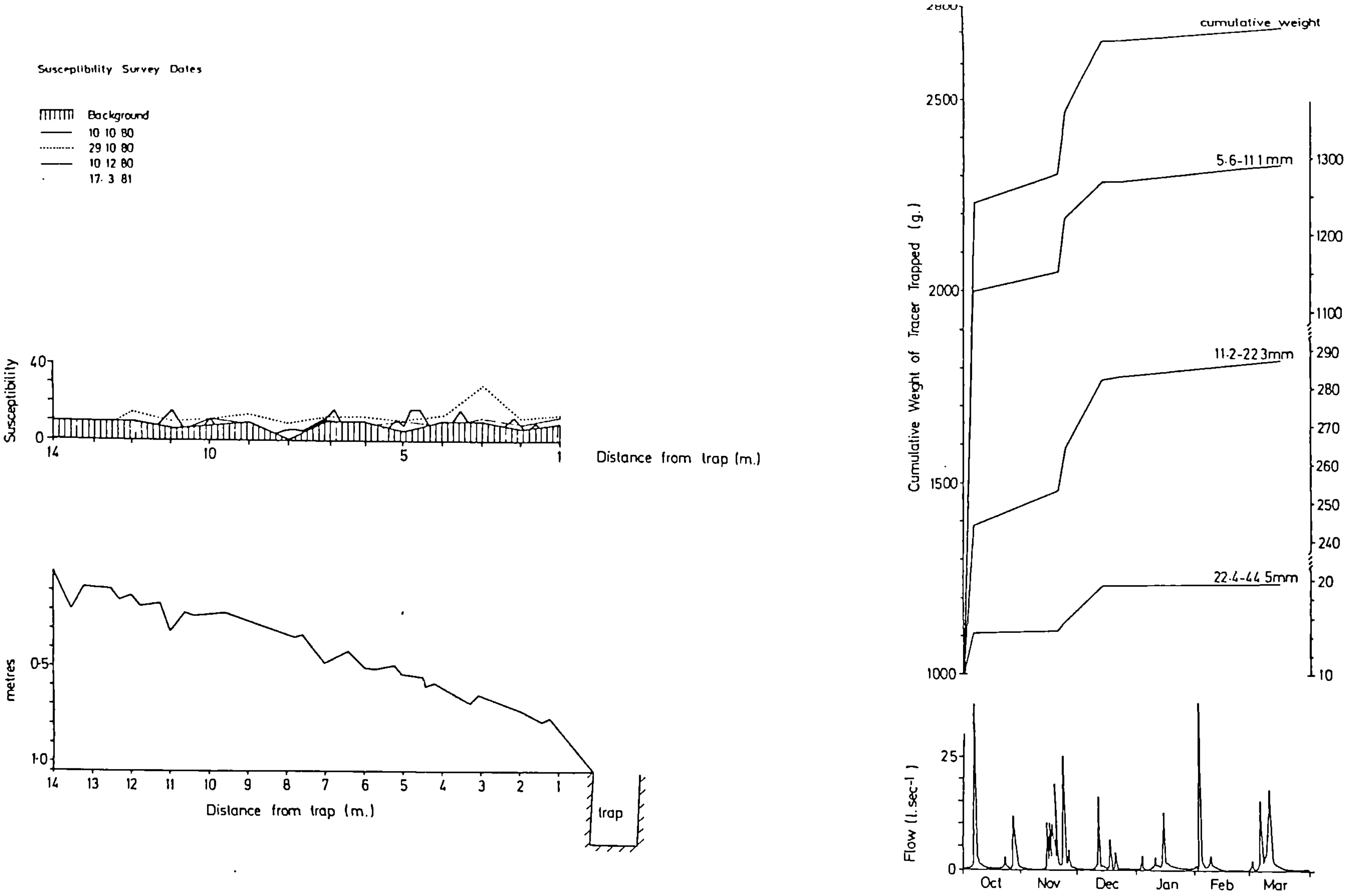


FIGURE 5.4. TRACER MOVEMENT AND RECOVERY FOR THE DRAINAGE DITCH LTAN6'A. FIXED POINT SURFACE SUSCEPTIBILITY SURVEYS (top left) ARE PLOTTED ABOVE THE LONGTITUDINAL PROFILE OF THE REACH AND (right) HYDROGRAPH - TRACER YIELD RELATIONSHIPS FOR THE STUDY PERIOD.

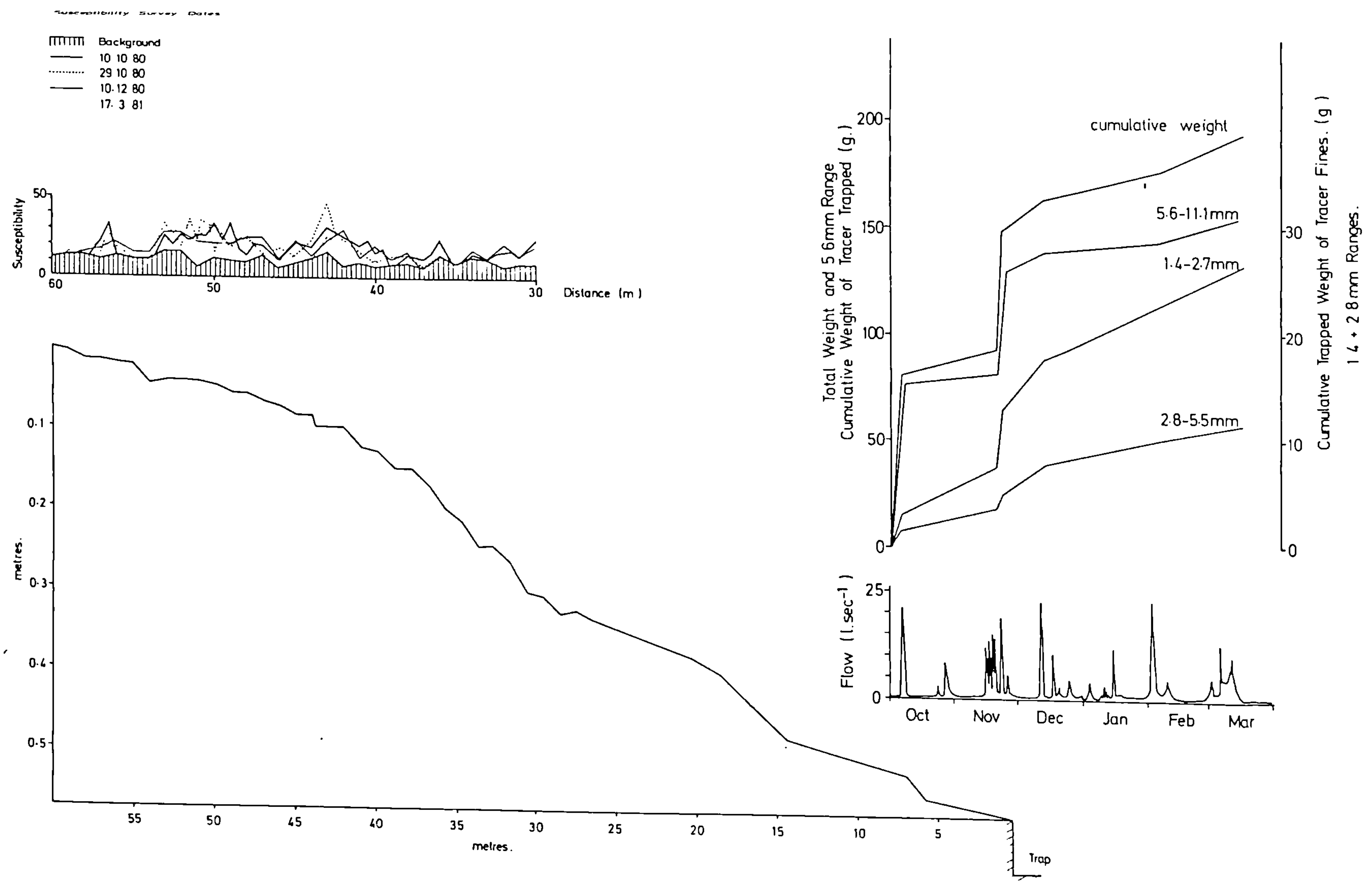


FIGURE 5.5. TRACER MOVEMENT AND RECOVERY FOR THE DRAINAGE DITCH LTAN. (see legend Figure 5.4. for details)

SIZE RANGE (mm)	TOTAL WEIGHT EMPLACED (g)	TOTAL WEIGHT RECOVERED (g)	% RECOVERY
LTAN6'A			
5.6 - 11.1	1750	1241	71
11.2 - 22.3	1400	992	71
22.4 - 44.5	1400	634	45
Total	4550	2867	63
LTAN			
1.4 - 2.7	300	See Text.	
2.8 - 5.5	600		
5.6 - 11.1	900		

TABLE 5.2. TOTAL RECOVERY OF TRACER TO 1.6.81 AGAINST THE ORIGINAL WEIGHTS PER SIZE RANGE ON EMPLACEMENT.

Data for the LTan6'A experiment indicated that the ditch was clear of tracer by the survey of March 1981. Whilst the cumulative weight of tracer trapped had levelled out, indicating exhaustion of the tracer supply, surface susceptibility surveys indicated a return to background levels. On this basis estimates of total recovery of tracer were calculated (Table 5.2.). The total tracer yield from the separation procedure described above, was 63% of the original weight emplaced; 40% of this was derived in the first and largest flood during the study period (peak discharge 37 $l.sec^{-1}$).

Surface susceptibility surveys illustrated the different sediment storage characteristics of the ditches, which also go to explain the variation in tracer delivery rates. These differences were largely due to the topographic variations between the channels as identified earlier. For LTan6'A, the steep, smooth u-shaped profile throughout, offered little opportunity for discrete shoaling, and the tracer material identified by surface susceptibility measurements appeared spread thinly throughout the channel length, particular concentrations lying in the lea of coarse clasts.

For the LTan trace (Figure 5.5), the tendancy towards shoaling was illustrated by the surface susceptibility surveys and topographic profiles, which from the first post-emplacment survey showed high concentrations of tracer in areas of reduced gradient or in association with bedrock obstructions in the channel. Field observations noted the build-up of distinct shoal units downstream from the emplacement zone. Figure 5.5 illustrates this, showing from the first survey, a build-up of tracer into three main units with a peak in susceptibility at approximately 43 metres upstream from the trap. Whilst subsequent floods altered the surface susceptibility readings slightly, perhaps indicating some local exchange between shoals, there appeared very little further downstream transport during the study period.

A mathematical assessment of the extent of tracer movement may be made bearing in mind the following limitations of the technique in its present form. The design of the search loop is such that the data provided by surveys of bed-surface susceptibility are only semi-quantitative estimates of tracer concentration (c.f. Chapter 3); susceptibility readings may be affected by concentration, size and position of the magnetic material in relation to the search loop. The instrument readings do not provide any distinction between particle sizes. Nevertheless, the results as shown in Figure 5.5. present a rapid and useful assessment of the tracer position following storm flows. On the assumption that the tracer has been well mixed with the bed sediment upon movement, these data provide an indication of the tracer activity per unit area of streambed. In this form, the results lend themselves readily to the spatial integration technique (Crickmore, 1967) for describing bedload transport. The integration of the observed spatial and temporal changes in tracer activity (susceptibility as shown in Figure 5.5.) define a "centroid position" of the tracer sediment at successive sampling times enabling the velocity and discharge of the bedload to be calculated. This technique has been successfully employed for a number of studies; see for example, Rathbun and Kennedy (1978) and Thomsen (1980).

The centroid position (P_t) of the tracer mass, at time T, may be calculated by the formula:

5.5.1. $$P_t = \left(\frac{\int_0^\infty \text{NadxX}}{\int_0 \text{Nadx}} \right)$$

and, its velocity (U_t) between successive observations at times T1 and T2, by:

5.5.2. $$U_t = \left[\left(\frac{\int_0^\infty \text{NadxX}}{\int_0^\infty \text{Nadx}} \right) T2 - \left(\frac{\int_0^\infty \text{NadxX}}{\int_0^\infty \text{Nadx}} \right) T1 \right] \cdot \left(\frac{1}{T2-T1} \right)$$

where:

DATE OF OBSERVATION	CENTROID POSITION (m)	DISTANCE DOWNSTREAM (m)	PREVIOUS PEAK DISCHARGE ($1.sec^{-1}$)
10.10.80	15.17	15.17	18.0
29.10.80	14.60	-0.57	7.5
10.12.80	14.98	-0.19	16.0
17. 3.81	16.93	1.76	19.5

TABLE 5.3. LTAN : DISTANCE OF TRACER MATERIAL TRANSPORT PER FLOOD CALCULATED USING THE SPATIAL INTEGRATION METHOD (Crickmore, 1967), BASED ON SURFACE SUSCEPTIBILITY VARIATIONS.

Na is the activity of tracer (susceptibility) per unit surface area; and; X is the distance from the origin.

Using these formulae, a series of positions of the centroid of the tracer mass were defined (Table 5.3.). These figures indicate two periods of tracer movement, associated with flows $\geqslant$ 18 l.sec^{-1}, the largest distance of tracer movement occurring during the first high flow event of the winter of 1980 (dated 10.10.80; previous peak discharge 18 l.sec^{-1}).

The data for subsequent events indicates very little movement of the tracer with the calculated centroid position wavering between 14.6 to 15.17 metres. The reason for this apparent negative movement of tracer is largely explained by the constraints of the technique as outlined above; the geometric relationship of the search coil to the tracer concentration on the streambed may produce significant variations in surface susceptibility readings. Further to this, any compaction of the bed sediment and/or burial of tracer, may reduce the surface susceptibility. Despite these variations, the data indicate a static position of the tracer centroid at 14.9 ± 0.3 metres (downstream from the emplacement site) for the period following 10.10.80 to February, 1981. This supports the inferences drawn from the susceptibility surveys (Figure 5.4.) that, subsequent to the initial phase of transport, the tracer was static even at flows equal to that of the first event. These data, together with the observations described above, suggest that subsequent to the initial phase of transport associated with the onset of high flows, the tracer sediment became incorporated into channel storage units, typically as small gravel shoals approximately 1.0*0.4*0.1 metres in size. The stability of these shoals would appear to be the controlling factor in determining tracer movement. This may be related to local bed topography, particularly bed-surface slope which for this part of the LTan channel ranges between 0.001 to $0.009m.m^{-1}$. Using a derivation of the Schoklitsch equation (Bathurst, 1984):

5.5.3. $$Qc = 0.15s^{-1.12}.g^{\frac{1}{2}}.D50^{3/2}$$
(per unit width of channel; $m^2 sec^{-1}$)

where
Qc = critical water discharge for sediment motion;
s = slope;
g = acceleration due to gravity;
D50 = median size of bed material;

critical discharge for these slope angles would need to be in the range of $40+l.sec^{-1}$ to initiate sediment movement. However, the calculated critical discharge for the emplacement site was equally high (local bed slope $0.0074mm^{-1}$). This suggests that either the original emplacement site was unstable, or that the Schoklitsch calculations provide an overestimate of the critical discharge.

A second phase of tracer movement was indicated by the spatial integration calculations shown in Table 5.3.. Despite a previous peak discharge of 19.5 $l.sec^{-1}$, the total distance of movements of the tracer centroid was calculated at only 1.76 metres, as compared to 15.17 metres for the first winter flood event at a peak discharge of 18 $l.sec^{-1}$. Thus, the Schoklitsch calculations would appear to be overestimating the critical discharge required for sediment movement although the relative magnitude of the changes due to slope provide a good indication of transporting and depositing reaches within the LTan system.

During the first post-emplacement survey, measurements of surface susceptibility revealed a number of magnetic clasts coarser than those used for the trace, located between 30 to 50 metres upstream from the bedload trap (c.f. Figure 5.5.). These appeared to be remnant from the initial tracing trials held within this system as reported by Rummery (1981). Further laboratory analyses were undertaken to verify the source of this material. The differentation between the two introduced magnetic materials should be made. Herein, any material attributed to earlier traces by Rummery will be classed as, "exotic".

Further laboratory analyses of mass susceptibility were able to distinguish between the much less magnetic exotic material (χ=6.97 $*10^{-6}$ m^3Kg^{-1}) as opposed to that of the tracer material (χ=12 * 10^{-6} $m^3.Kg^{-1}$) and compare favourably with the data presented by Rummery. The much lower susceptibility of the exotic material provides some explanation as to why it did not show up on the initial background susceptibility surveys for the current tracer experiments (Figures 5.4. and 5.5.). These much lower susceptibilities were probably reduced further to the 'apparent surface' susceptibility as a result of burial within shoals. Despite the presence of coarse exotic tracer material within the LTan channel, no magnetic material coarser than the 5.6-11.1mm size range was found within the trapped load following each high flow period. Estimates of trapping concentrations, used to plot Figure 5.5. have been made on the basis of the characteristics of the introduced tracer material not of the exotic material characteristics (Rummery (1981) suggests that much of his introduced tracer load, in the ranges finer than 2.8mm, was trapped during his period of observations). On this basis, whilst the trapping concentrations for both ditch systems showed a similar response to flooding (at least in the earlier stages), the significantly lower concentrations trapped in the LTan suggested that the emplaced tracer material was held in storage further up-channel from the bedload trap. This pattern was also identified by the susceptibility scans and by the spatial integration method for calculating bedload yield.

Estimates of emplaced tracer recovery rates (Table 5.2.) have not been made for the LTan system because of the possible complications

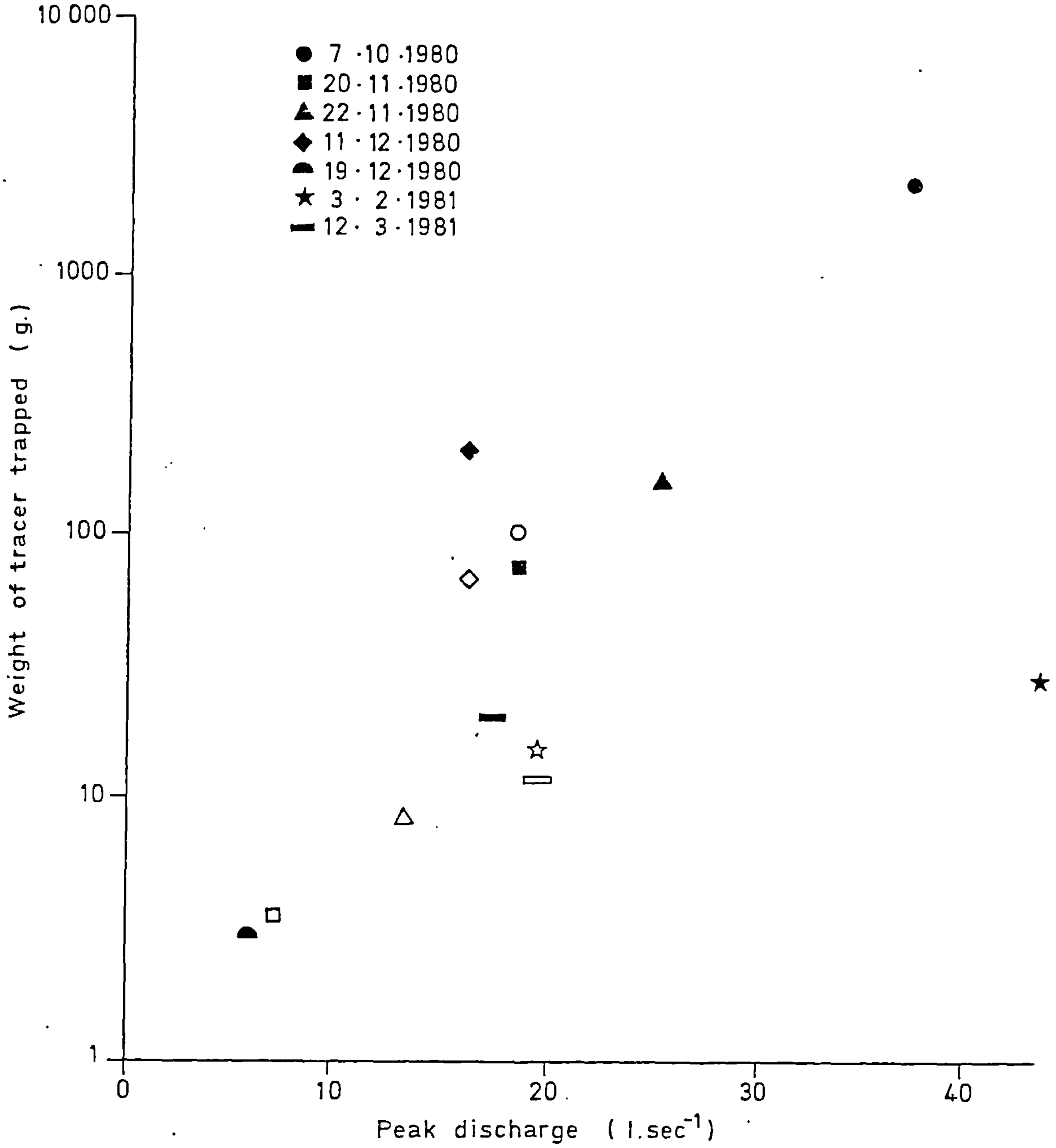


FIGURE 5.6. RELATIONSHIP BETWEEN WEIGHT OF TRACER MATERIAL TRAPPED TO PEAK DISCHARGE FOR EACH STORM EVENT; Open Symbols - LTAN;

Closed Symbols - LTAN6'A.

arising from two sources of magnetic material. As a coarse estimate, if we assume that there is no exotic material within the trapped load, the cumulative trapped weight of magnetic material was 198.57g, providing a recovery percentage of 11% of the total emplaced load. ie. some 89% of the emplaced tracer (at least) was still in the system at the end of the observation period.

As the main aim in these experiments was to examine transport conditions in eroding ditch systems and, if possible, identify critical flows for bed-load transport, the tracer recovery data were plotted against peak discharge for each ditch weir (Figure 5.6.) With a correlation coefficient of 0.51, a weak relationship between tracer yield and peak flow was illustrated; the deviation in trapping concentrations resembled the complexities discussed earlier with reference to Figure 5.1.. There are two immediately apparant reasons for this. Firstly, for the ditch LTan6'A, exhaustion of the tracer supply limits data for the latter observations. From the above however, for the LTan system this was clearly not the case and storage became a dominating factor. Whilst the relationship held true for the broad range of flows, this was interrupted by periods where frequent moderate flows and even larger events (subsequent to the main release of tracer) released low concentrations or no tracer at all.

5.6. SUMMARY

1) These initial applications of the technique illustrate its immediate success on small scale systems. Recovery rates of the tracer, where appropriate, totalled 63%, up to 71% for individual size ranges.
2) Transport - recovery rates were influenced markedly by the initial flooding of the winter season 1980. For LTan6'A, 40% of the total

emplaced tracer load was yielded during this one event. For LTan, this was reflected by the main period of transport within the system, the mean transport distance was approximately 15 metres.

3) For the range of flows studied, variations in peak discharge were not capable of explaining the full variation in tracer/bed-load yield. At extreme values of flow, bed-load rates were correlative but for moderate events a wide deviation in yields was observed.
4) Channel storage of sediment was observed as a controlling factor on the release of tracer. This was exhibited markedly by the LTan system for which tracer concentrations were shown to be static subsequent to the main release.
5) Again, in response to the initial spates of winter 1980, exotic tracer was released from storage. Using its individual magnetic signature, (Oldfield et al. 1981) this was identified as a previous trace load which had been held in storage for two winter seasons. Its lower Mass Susceptibility contributed to its remaining unseen during the background surveys of each ditch.

5.7. CONCLUSIONS

The applications described in this chapter have illustrated a number of points which hold implications for subsequent experiments in downstream channels where sediment supply may be limited to seasonal response.

The variations observed in tracer yield in response to upland flooding suggests that, for the most part, forest drainage channels are supply-limiting. Furthermore, the data described above show that bed-load yields may be more closely related to the storage and throughput characteristics of the eroding channel, rather than being predominantly supply-limited as a result of the dependancy upon the weathering of bank sediments. If, as Robinson (1980, 1982) suggests, the bed-load

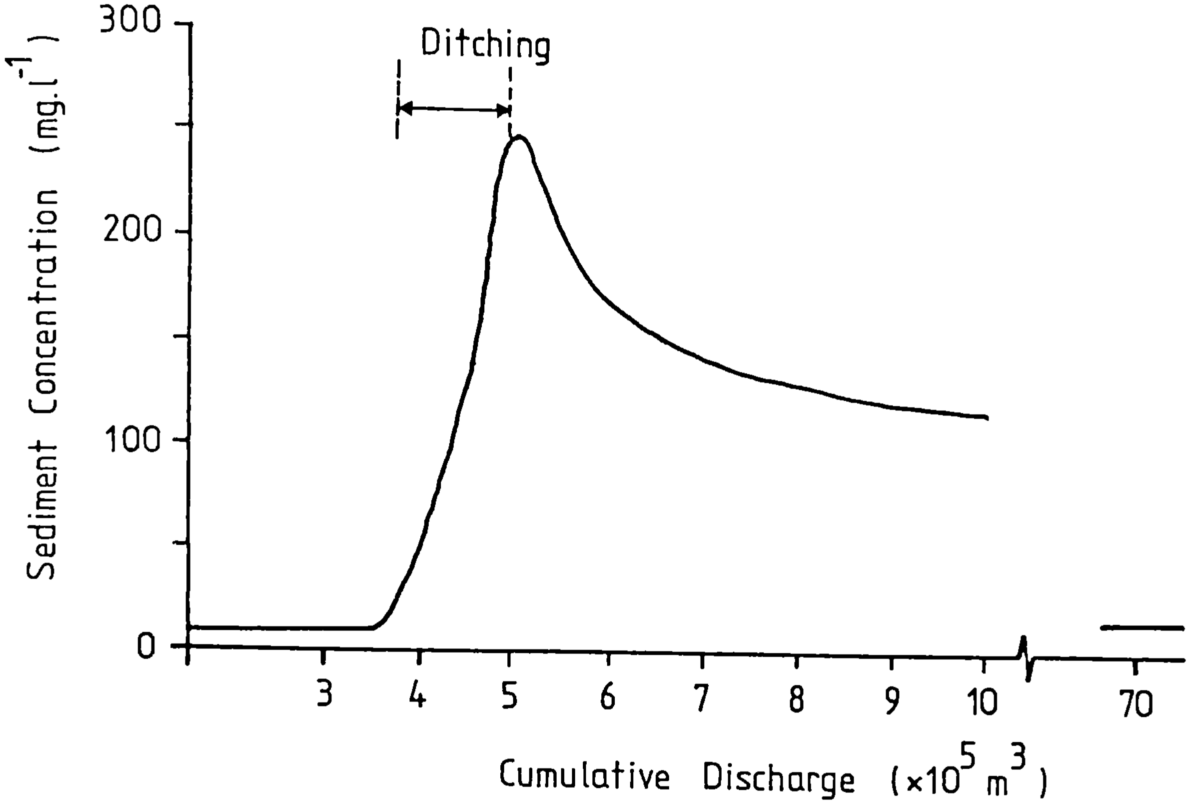


FIGURE 5.7. AVERAGE PATTERN OF SEDIMENT LOSS AS RECORDED FOR THE COALBURN CATCHMENT (AFTER ROBINSON (1980) AND ROBINSON AND BLYTH (1982).

yield from drainage ditch systems declines after an initially high (catastrophic!) yield period (Figure 5.7.), then these results should not be unexpected. However, Robinson's data show that subsequent yields remain at some four times higher than pre-ditching levels and would suggest considerable increases in sediment delivery rates to the piedmont river reaches.

The main phase of tracer transport - release occurred within the first flood of the observation season. For LTan6'A this was the largest discharge event (Peak discharge 37 $lsec^{-1}$), for LTan, one of the largest (Peak discharge 18 $lsec^{-1}$). No successive, individual event yielded as large a volume either in transport or trapping. This may be taken to infer some inherent instability in the trace shoal, as has been recognised in many of the earlier tracer experiments. However, given the length of time the shoal was given to stabilise under low flows (3 months) this seems unlikely. Rather, as identified by Leopold & Emmett (1976) transport rates for the same range of flows were greater for the initial stream rise at the start of the winter season than for any subsequent event. For the LTan system, this was associated with the destruction of the tracer shoal and an immediate build-up of disctrete storage shoals which remained static for much of the remainder of the trace period.

The experiments illustrated the identification of two trace source materials, definable by their clearly different magnetic mass susceptibilities. Whilst these can be distinguished rapidly using other magnetic characteristics field surveys using the Search Coil were limited to measures of susceptibility alone and as a result, could not discriminate between the exotic or the introduced trace load. Such surveys require back-up laboratory treatment and a necessary removal of tracer from the system.

The removal of any tracer material, or indeed any bed-material, was avoided for two reasons. Firstly, any such removal of the bed sediment requires some disturbance of the bed structure and subsequent weakening providing for easier erosion. Secondly, any removal of material complicates the recovery rates when using such small emplace loads (LTan 1800g).

The application of the magnetic tracing technique to small drainage ditch systems, has illustrated that even at this scale channel storage may be of considerable importance to the transport systems of gravel bed streams. The consequence of previous studies by Rummery (field trials carried out in winter 1978/9) has indicated how this element of storage has been largely underestimated in the past. The lack of tracer movement was also reflected in the total throughput of sediment in the ditch systems; bedload trapping in both drainage ditches was reduced (Newson, pers. comm.). This suggests that bed material may be stored as bedforms for considerable periods of time.

The irregularity of coarse sediment transport in these systems has made the prediction of critical flows extremely difficult. Transport of tracer material occurred at flows in excess of 16 $l.sec^{-1}$. However, even at the extremes of the observed flow range during this study, transport rates were not comparable.

CHAPTER 6

SEDIMENT TRACING IN LARGE UPLAND GRAVEL REACHES: CASE STUDIES FROM THE UPLAND TRIBUTARIES OF THE RIVERS WYE AND SEVERN

6.1. INTRODUCTION

The data described in Chapter 5 have illustrated the use of the magnetic tracing technique under semi-controlled conditions; tracer recovered by sediment trapping within a reach could be compared with the initial quantity emplaced. However, it is in much larger gravel bed reaches that sediment tracing will be of most use. At this scale, sediment tracing serves as a valuable alternative to large bedload traps, or as a complement to bed material sampling devices at the stream bed which may frequently lead to considerable innaccuracies in estimations of transport rates as a result of the discontinuous nature of bedload movements.

This chapter describes the application of magnetic tracing to two case study areas on the upland tributaries of the Rivers Wye and Severn. These were thought to be representative of characteristic field problems in monitoring sediment transport processes. Two main objectives were identified: to test the efficacy of the technique within large, gravel-bed reaches; and, to provide observations of gravel bedload transport within upland British rivers.

6.2. METHLODOLOGY AND SITE DESCRIPTIONS

To extend the tracing technique to larger upland reaches, two stream sites were studied: An upland reach of the Wye; and, an upland tributary of the Severn, Afon Llwyd.

Tracing on the Afon Llwyd was carried out downstream from the Dolydd road bridge (Grid Reference: SN874906; Figure 6.1.). The area

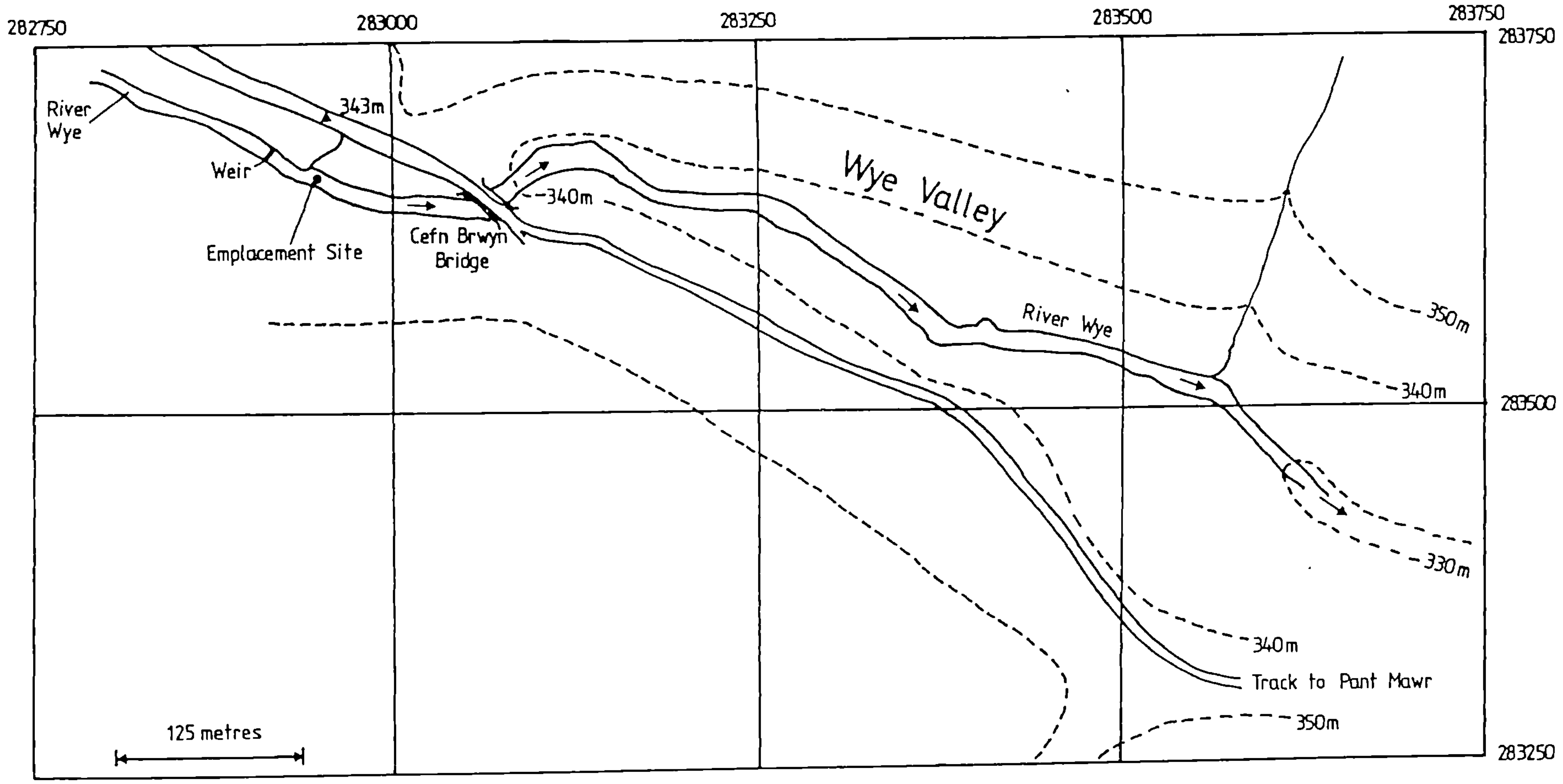


FIGURE 6.1. CEFN BRWYN (RIVER WYE CATCHMENT) : POSITION OF STUDY REACH.

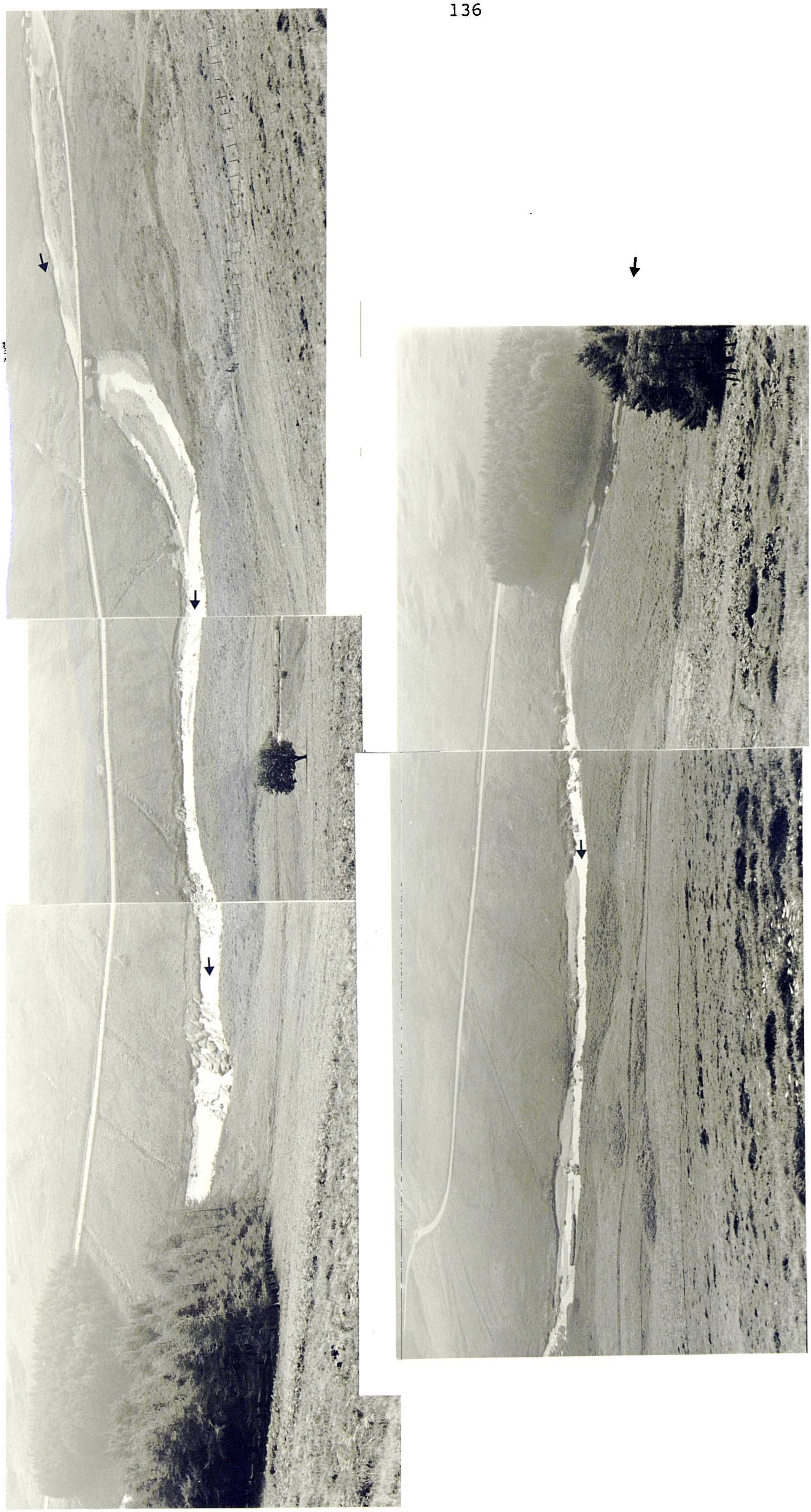


PLATE 6.1. CEFN BRWYN (RIVER WYE CATCHMENT) : STUDY REACH.

PLATE 6.2. (ABOVE) CEFN BRWYN : EMPLACEMENT SITE AND SURVEY GRID,

(BELOW) DOLYDD, AFON LLWYD : BRIDGE SHOAL STUDY AREA.

has, in the past, been frequently used as a site of gravel excavation for use as farm road aggregate. The excavated trenches have been monitored by IOH as ad hoc bedload traps since August 1979. Their observations have suggested that the shoals immediately downstream of the Dolydd bridge are very mobile features allowing both throughput and storage of sediment. The main aims of tracing here were to monitor sediment movement from bridge pier structures and to observe sediment routing through downstream channel shoals as a result of both natural processes and of gravel excavations within the main shoal.

Tracing experiments on the River Wye were carried out downstream of Cefn Brwyn (Grid Reference : SN829837; Figure 6.2). Situated immediately downstream of the IOH Compound Crump Weir, this reach provided an opportunity to monitor loss and subsequent downstream movement of tracer sediment without the additional complication usually found in natural channels, of bedload supply from upstream (it being trapped in the substantial weir pool). The initial 130 metres of stream reach is comprised of a wide, shallow channel, the bed-surface of which is composed of a lag of coarse gravel and exposures of solid bedrock. At approximately 130 metres downstream, the channel meanders and has associated with it a series of lateral and medial bars (Plate 6.1.). Downstream from this point (approximately 200 metres) channel reaches are predominantly rock controlled with outcrops of solid rock bluffs. Bed material storage areas are widely spaced, again in response to channel bends or areas of slack water between bare rock rapids.

Tracer Emplacement

In both cases, a detailed survey was carried out prior to tracing, to assess variations in surface and subsurface bed material size ranges. Particle size distribution curves for the two sites are shown in Figures

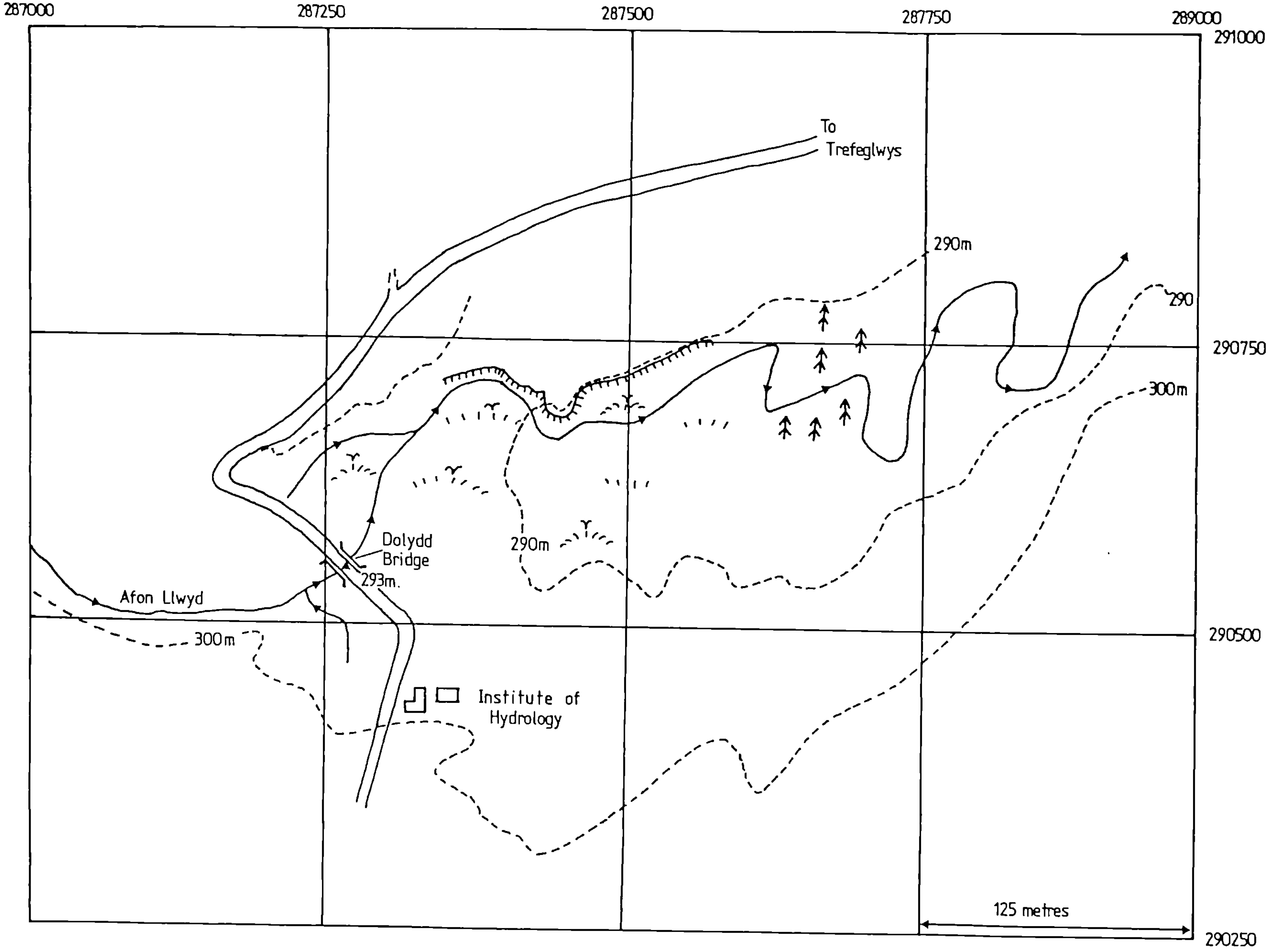


FIGURE 6.2. DOLYDD (AFON LLWYD) : POSITION OF STUDY REACH.

SIZE RANGE	DOLYDD		CEFN BRWYN	
(mm)	%	Weight (Kg)	%	Weight (Kg)
90 +	5	6.6	5	6.6
45.0 - 89.0	42	55.4	27	35.4
22.4 - 44.5	20	26.4	29	38.0
11.2 - 22.3	12	15.8	29	38.0
5.6 - 11.1	8	10.6	6	7.9
2.8 - 5.5	5	6.6	4	5.2
1.4 - 2.7	4	5.3		
1.4	4	5.3		
TOTAL	100	132.0	100	131.0
DATE EMPLACED	3.6.80		8.5.80	

TABLE 6.1. PARTICLE SIZE CHARACTERISTICS OF TRACER MATERIAL EMPLACED FOR CEFN BRWYN (RIVER WYE) AND DOLYDD (AFON LLWYD) TRACES.

6.3. and 6.4.. To insure that, as far as possible, the tracer behaved in the same manner as the local sediment, material was extracted from an upstream shoal prior to tracing, enhanced in the manner described in Chapter 3, and emplaced in the size ranges shown in Table 6.1. In both cases, tracer material was 'seeded' onto the bed surface at the emplacement sites shown in Figures 6.5. and 6.6.. In the case of the Dolydd site, this required varying particle size ranges in response to bed material size variations on the shoal surfaces beneath the Dolydd bridge. For the Cefn Brwyn site, pegs were installed on the river banks delineating a grid extending 10 metres downstream (plate 6.2.). Tracer material was seeded over a 4 x 4 metre area in the upstream portion of this grid, approximately central to the channel (c.f. Figure 6.5.). At the intersection of each tramline on the emplacement grid a 90mm yellow painted pebble was placed on the bed and was used as an indicator of tracer movement.

Emplacement was carried out in the early summer of 1980 (Cefn Brwyn: 8.5.80; Dolydd: 3.6.80), so that at each site a period of settlement to low flows was achieved before the onset of winter flows.

Methods of Survey

The movement of tracer material was assessed following each spate of high flows. A three tier system of survey was used at both sites:

1) A detailed survey of bed surface susceptibility was carried out at the emplacement site. At Cefn Brwyn, this was made by taking a sequence of ten randomly spaced readings within the 1x2 metre squares defined by the emplacement grid. An early indicator of sediment transport was given by the movement of the coarse painted clasts. For the Dolydd trace, the emplacement site was monitored along fixed transect lines between the bridge piers. A series of eight transects were continuously monitored for changes in bed-surface susceptibility by monitoring changes

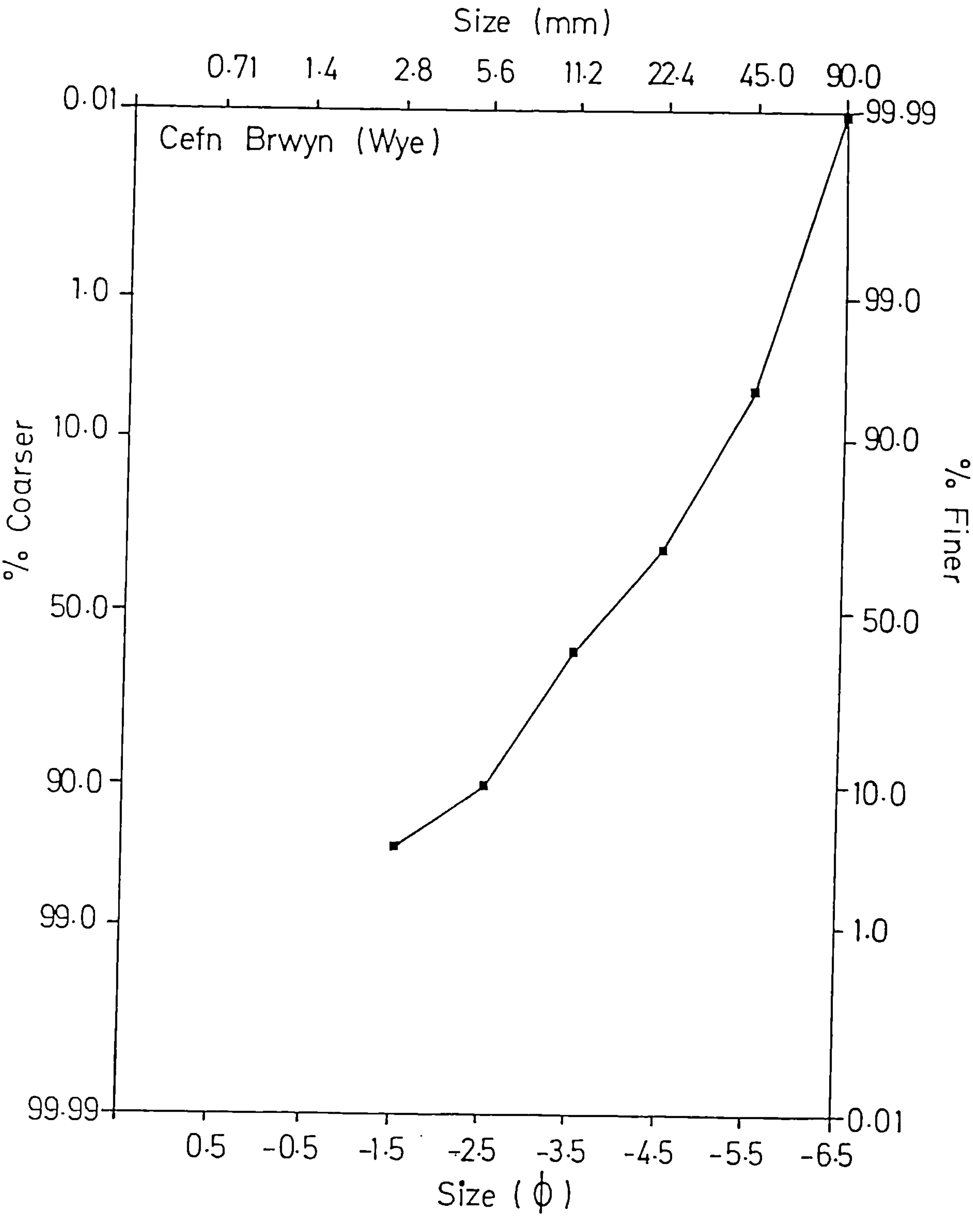


FIGURE 6.3. CEFN BRWYN : PARTICLE SIZE DISTRIBUTION CURVE OF EMPLACED TRACER MATERIAL.

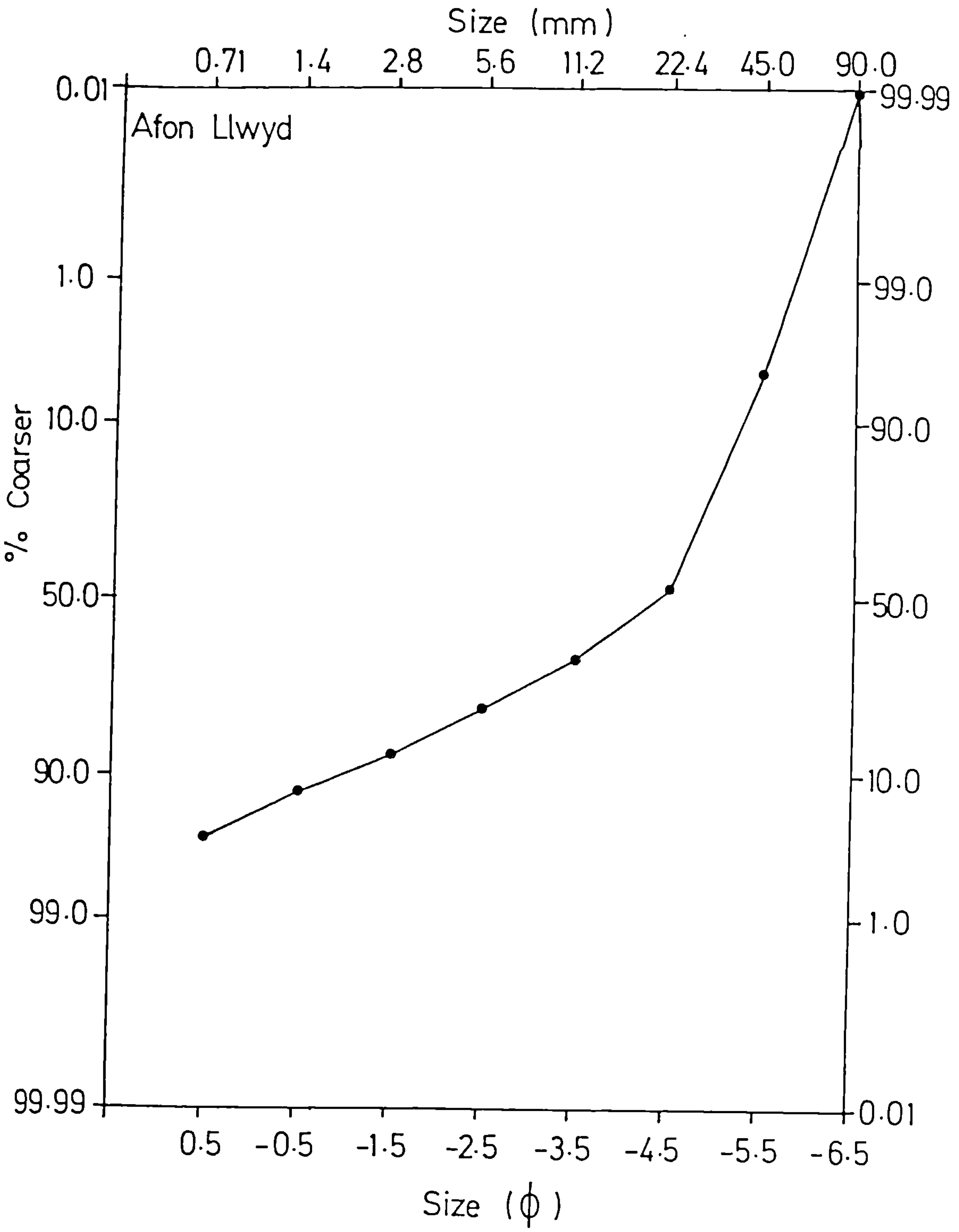


FIGURE 6.4. DOLYDD : PARTICLE SIZE DISTRIBUTION CURVE OF EMPLACED TRACER MATERIAL.

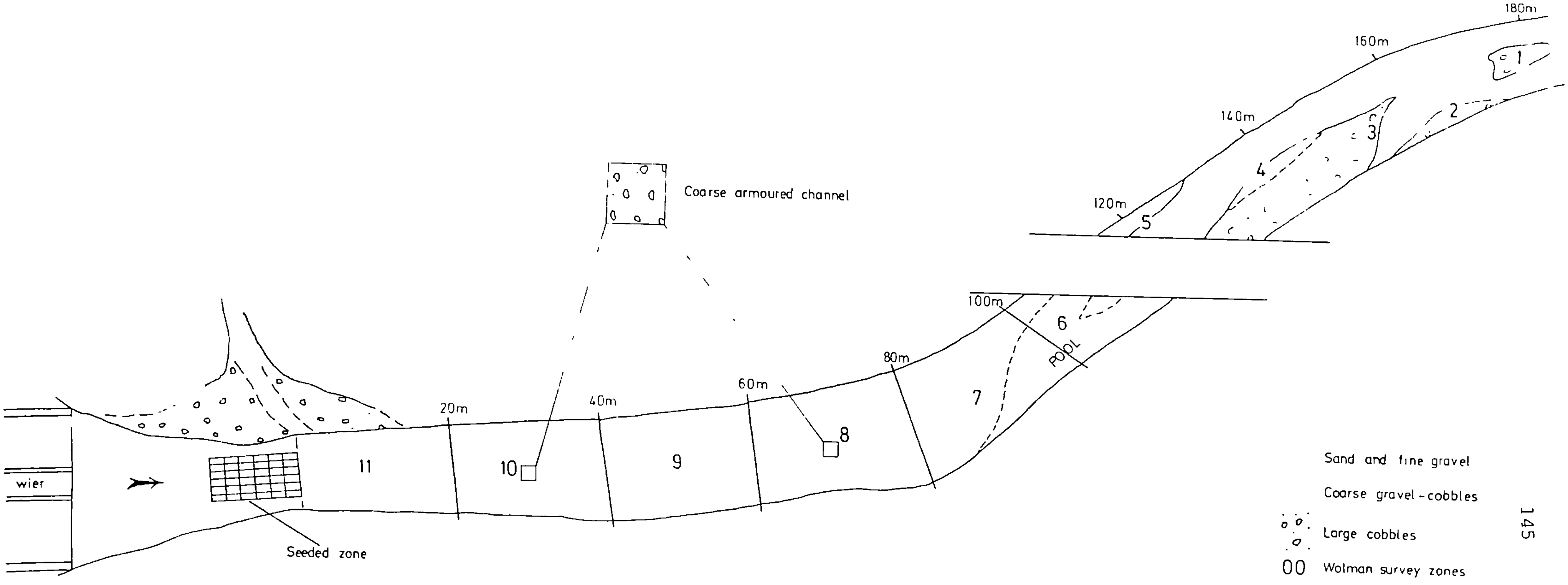
at each 20 cm interval.

2) At the initial phase of surveying, the extent of tracer movement within the immediate 100 metres downstream was assessed by using transect surveys across the channel. These were spaced at 20 metre intervals downstream, with surface susceptibility readings being taken at 0.5 metre intervals along each transect.

3) For most of the study period, longer distant tracer movement was assessed by monitoring channel segments or individual bedforms using a "Wolman 100" approach. This was a hybrid approach of that used by Wolman (1954) and later developed by Leopold (1970), and involved the collection of 50-100 susceptibility readings on each significant bedform or section of channel thalweg. In the case of the Cefn Brwyn trace, where a substantial shallow gravel reach with its obvious bedforms had to be surveyed, a number of 20 metre channel segments were identified (Figure 6.5.). Within each 20 metre reach, the Wolman approach was carried out on a random-stratified basis, taking 20 readings along each of five transects parallel to flow.

Surface susceptibility variations were monitored using the field susceptibility search loop, and subsequent identification of individual magnetic clasts using a ferrite probe (c.f. Chapter 3).

Identification of individual clasts was attempted whenever the flow conditions permitted. In such instances, at each point of high surface susceptibility (as compared with a normal reading of 16-20 units), the channel bed was studied for the bright pink colouration of the tracer and where found, was carefully removed, sized and replaced at its original position within the sediment matrix.



CEFN BRWYN : River Wye Study Area.

FIGURE 6.5. CEFN BRWYN : SKETCH PLAN OF EMPLACEMENT AREA AND POSITIONS OF SURVEY ZONES.

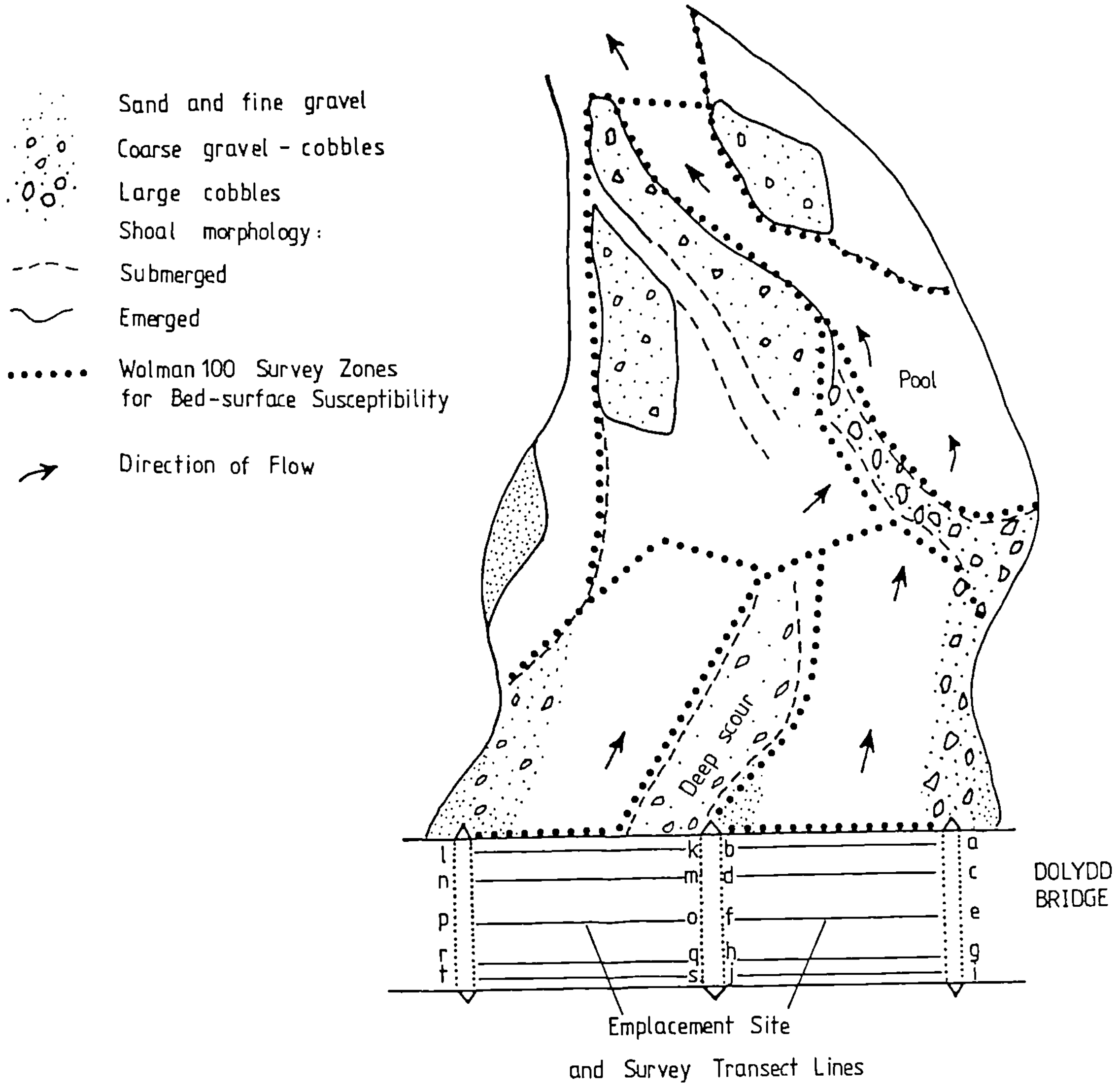


FIGURE 6.6. DOLYDD : SKETCH PLAN OF EMPLACEMENT AREA AND POSITIONS OF SURVEY ZONES.

6.3. CEFN BRWYN AND DOLYDD TRACES

MOVEMENT OF TRACER FROM THE EMPLACEMENT SITES

Periods of tracer movement (loss) from the emplacement zones of both the Dolydd and Cefn Brwyn traces were inferred by changes in the pattern of bed-surface susceptibility at each site. These are summarised in Figures 6.7. and 6.8. for Cefn Brwyn and Dolydd respectively. For the period from emplacement of tracer to October 1980, both sites exhibited a period of stabilization to flows <10 cumecs. These also provided some localised transport of tracer; of the order of 1-1.4 metres as shown by Figure 6.7. (Plot B). The overall small increase in bed surface susceptibility exhibited in Plot B suggests that any localised transport of tracer is concerned with the smaller grain sizes of the emplaced tracer load.

The main period of tracer movement began in response to the first storm event of the winter 1980/81 (see Figures 6.7. and 6.8.; Tables 6.2. and 6.3.). This was preceeded by a period of 4-6 months where the flows were less than 10 cumecs.

For the Cefn Brwyn trace, peak flows of 18.6 cumecs were associated with considerable movement of tracer out of the emplacement zone; bed surface susceptibility levels dropped to an average of 30 units (Figure 6.7., Plot D). Bed surface susceptibility varied little in response to subsequent flows as high as 10 cumecs (28.10.80; Plot E). However, following a period of two storm events (20.11.80, Qmax 17.5 cumecs; 22.11.80 Qmax 16.6 cumecs) measures of mean bed surface susceptibility returned almost to background levels, suggesting that the majority of emplaced tracer load had been transported downstream. Small variations in bed surface susceptibility (Plot F) were associated with the presence of coarse tracer clasts which had been buried at the emplacement site.

FIGURE 6.7. CEFN BRWYN : SURFACE SUSCEPTIBILITY SURVEYS OF THE EMPLACEMENT SITE ILLUSTRATED BY ISOPLOTS, SHOWING TRACER LOSS ASSOCIATED WITH THE MAIN FLOOD EVENTS OBSERVED.

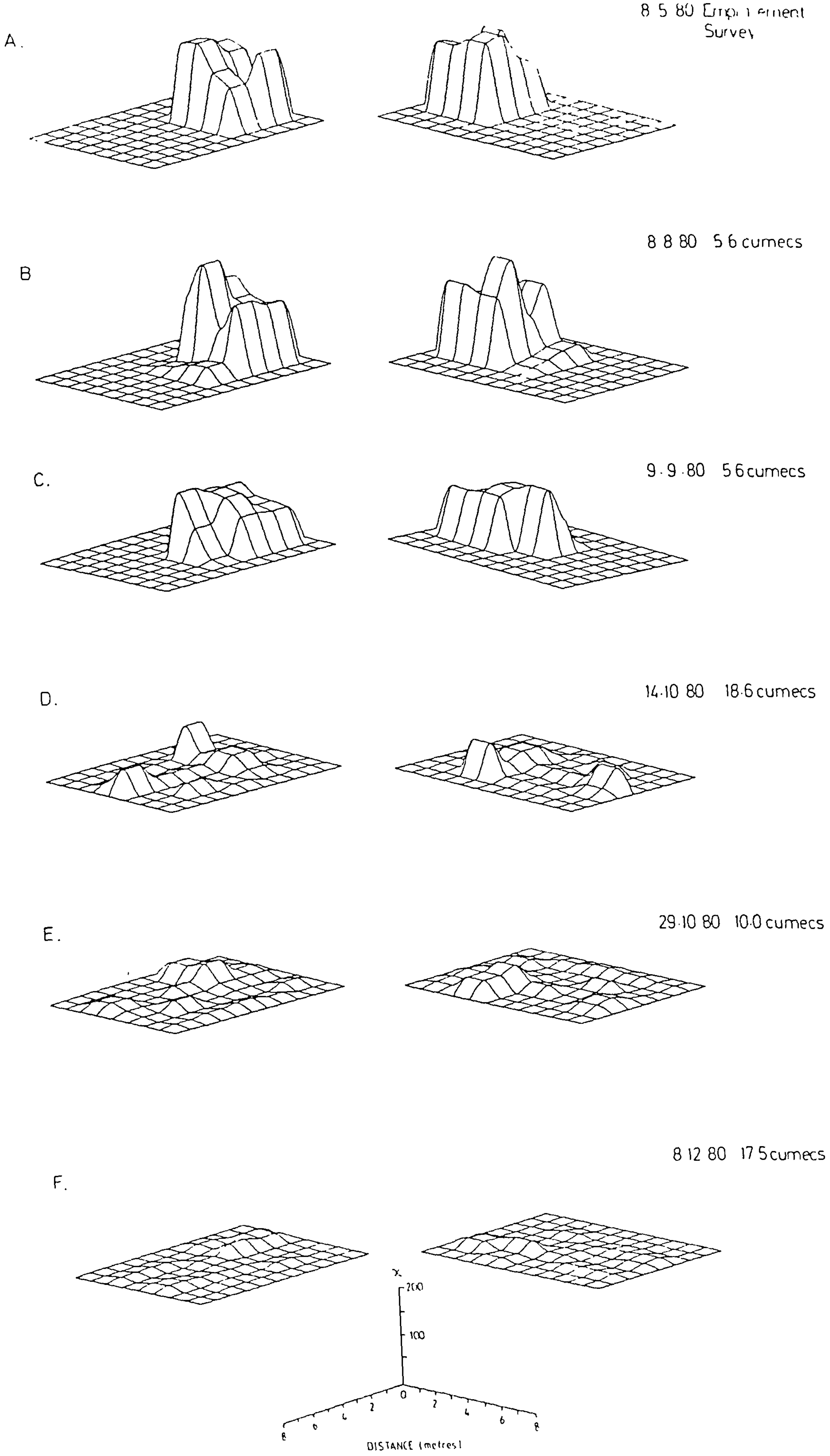
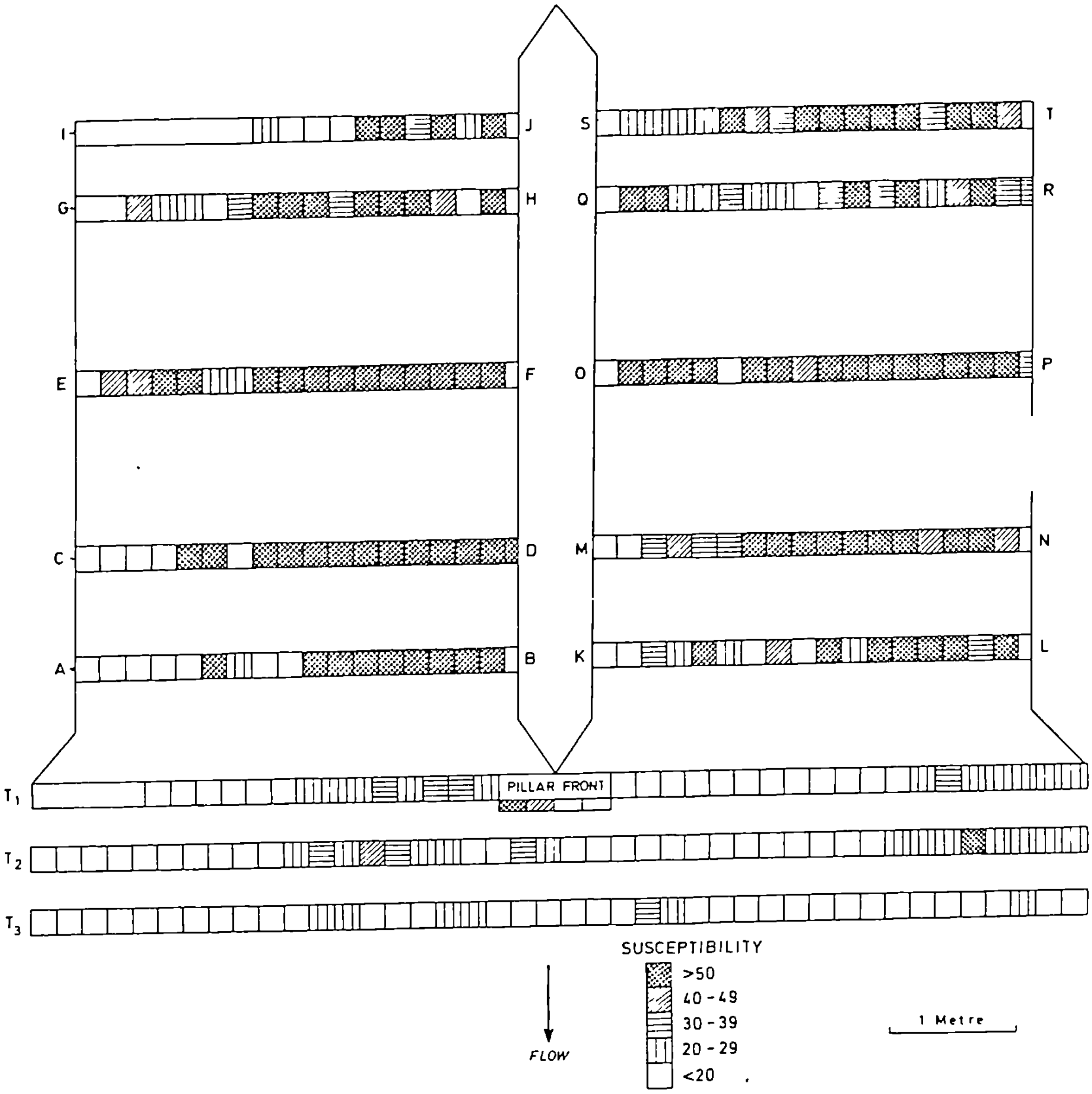
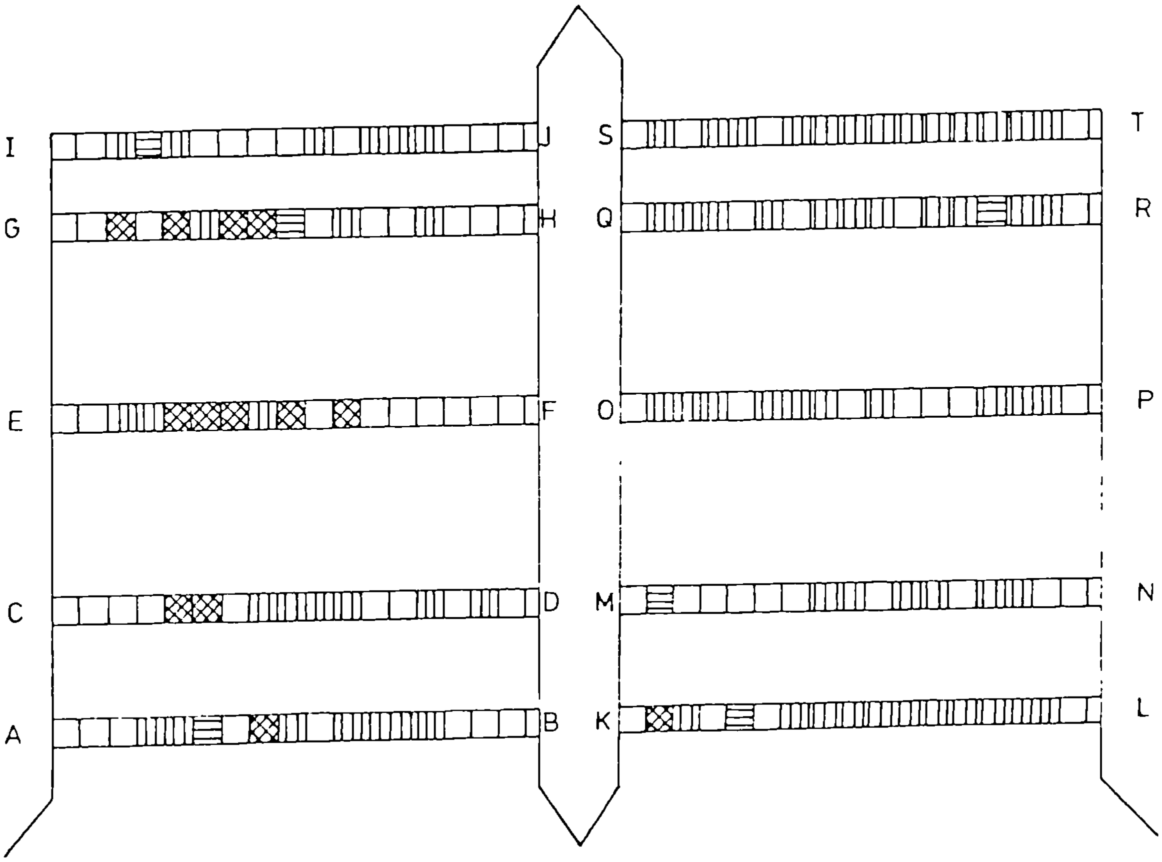

A.
8 5 80 Emplacement Survey
B
8 8 80 5 6 cumecs
C.
9·9·80 5 6cumecs
D.
14·10 80 18·6cumecs
E.
29·10 80 10·0 cumecs
F.
8 12 80 17 5cumecs
%
200
100
0
2
4
6
8
DISTANCE (metres)

FIGURE 6.8. DOLYDD : TRANSECT SURFACE SUSCEPTIBILITY SURVEYS OF THE EMPLACEMENT AREA FOLLOWING A PERIOD OF SETTLEMENT AFTER EMPLACEMENT (ABOVE) AND IMMEDIATELY AFTER THE MAIN FLOOD EVENT FOR THE REACH 6.10.80 (BELOW).


I
J
S
T
G
H
Q
R
E
F
O
P
C
D
M
N
A
B
K
L
PILLAR FRONT
T_1
T_2
T_3
SUSCEPTIBILITY
>50
40 - 49
30 - 39
20 - 29
<20
FLOW
1 Metre


I
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L

Particle size variations between flood events, and burial processes are described in more detail in the following sections.

At the Dolydd site, discharge variations were of a much higher magnitude, of the order of 130 cumecs for the 6.10.80 storm event.† The pattern of bed surface susceptibility changes (Figure 6.8., Plots A-C) indicate a very similar pattern to that observed at Cefn Brwyn. However, scour and transport of tracer material was confined to the true left hand side of the channel at the emplacement site (sections K to T, Figure 6.8.). By contrast bed surface susceptibility surveys over the right hand side of the emplacement zone (sections A-J) indicated a very slow but progressive decline throughout the monitoring period, suggesting very little loss of tracer material and a probable gradual burying of the coarser tracer clasts. These patterns of scour were also illustrated by topographic surveys (Figures 6.9.), showing localised scour at the bridge site and a downstream accumulation of material at the Dolydd shoal.

Scour at bridge crossings is a commonly observed problem normally associated with channel constriction and localised turbulence caused by the positioning of bridge piers. Localised scour depends upon the approach stream velocity, depth, and natural variations in sediment transport rates within the bridge reach (Richards, 1982). No detailed hydraulic observations have been made for this reach, thus, the causes of localised scour in the region of the Dolydd bridge are largely hypothetical.

Whilst the Dolydd channel is, in general, actively meandering, the reach immediately upstream of the bridge is straight and this would

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† The accuracy of flow measures at this site is suspect; unit hydrographs suggest that 'recorded' flows may be 2-4 times larger than predicted. Adjustments from stage charts to a unit hydrograph calibration has not been made because of lack of data at high flows (Newson, Pers. comm.).

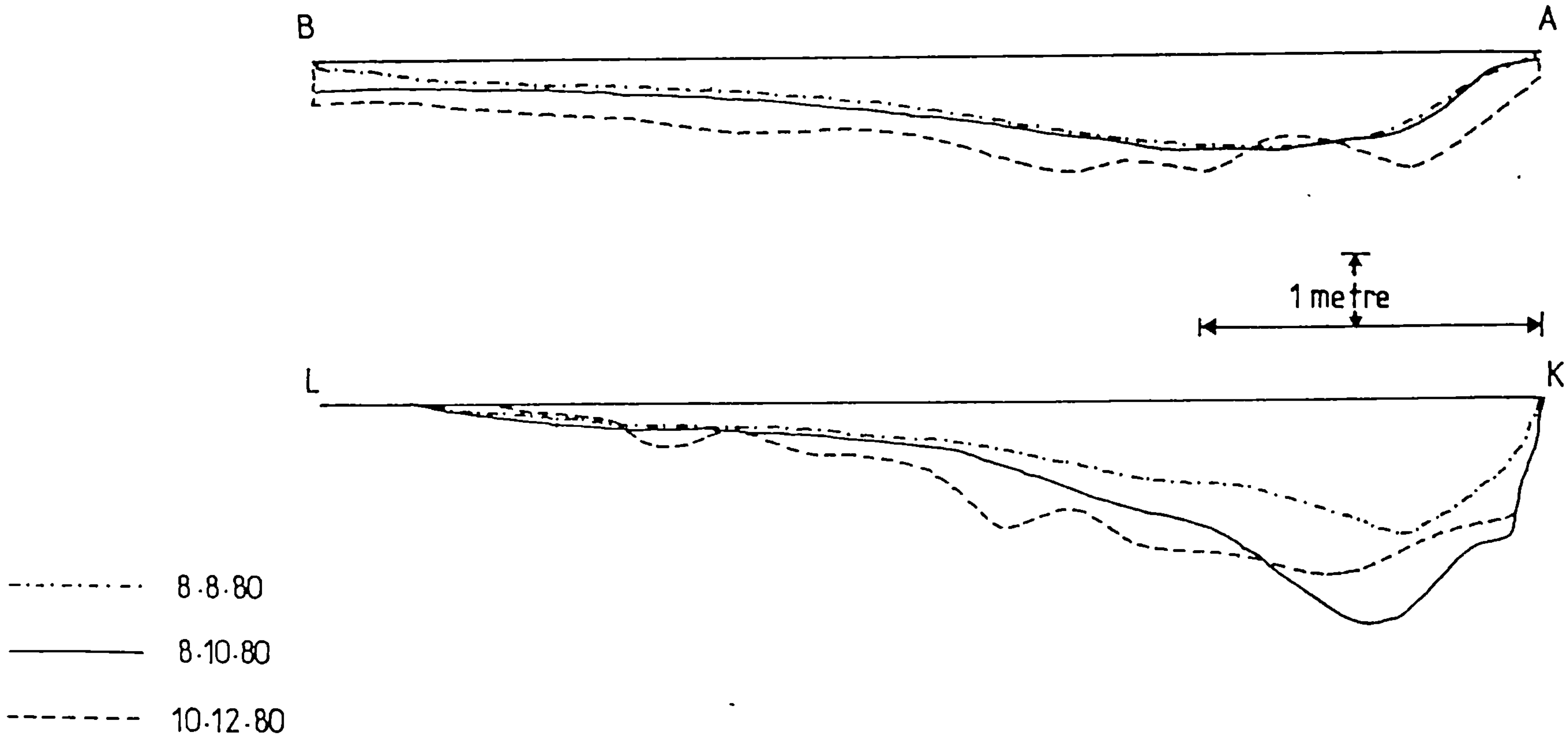


FIGURE 6.9. DOLYDD : TOPOGRAPHIC CHANGES OF THE RIVER BED BENEATH DOLYDD BRIDGE ILLUSTRATING THE DEPTH OF SCOUR (SECTIONS K - T) ASSOCIATED WITH TRACER LOSS FOR THE EVENT 6.10.80.

suggest that rather than as a result of channel induced secondary flows, scour may be associated with the position and shape of the bridge piers relative to the oncoming flow. The bridge is centred to the right of the channel centre line, thus, it is plausible that the main flow velocity (of this straight channel segment) will be constricted by the bridge piers mainly within the cross-sections K to T. Furthermore, this constriction may be exacerbated by the shape of the pier nose to oncoming flows; water hitting the upstream nose of a bridge pier is drawn into a horseshoe vortex which has been shown to cause appreciable scour at pier heads and abutments (Garde and Ranga Raju, 1977). Whilst, for the Dolydd, the effect of pier nose shape is reduced by their triangular form (approximately 70°), data from Garde and Ranga Raju would suggest that there may still be appreciable aggravation of the scour processes.

6.4. DOWNSTREAM PATTERNS OF TRACER MOVEMENT

6.4.1 CEFN BRWYN

The downstream transport of tracer material has been illustrated in two ways: by using the total number of bed-surface susceptibility readings above background per survey zone: and, where tracer clasts were identified, the number and size variations downstream. The results of surveys from 8.5.80 to 12.3.81 are shown in Figure 6.10.

Recovery rates of tracer have been estimated from those surveys where tracer clast size has been identified using an average particle weight per size range, the total number of clasts identified was converted to a weight and estimated as a percentage of the emplaced load. For the five surveys to September 1982, the recovery of identified tracer material has rarely exceeded 5%, ranging from 0.5 to 6.0% for individual surveys. In addition to this, tracer recovery in the field

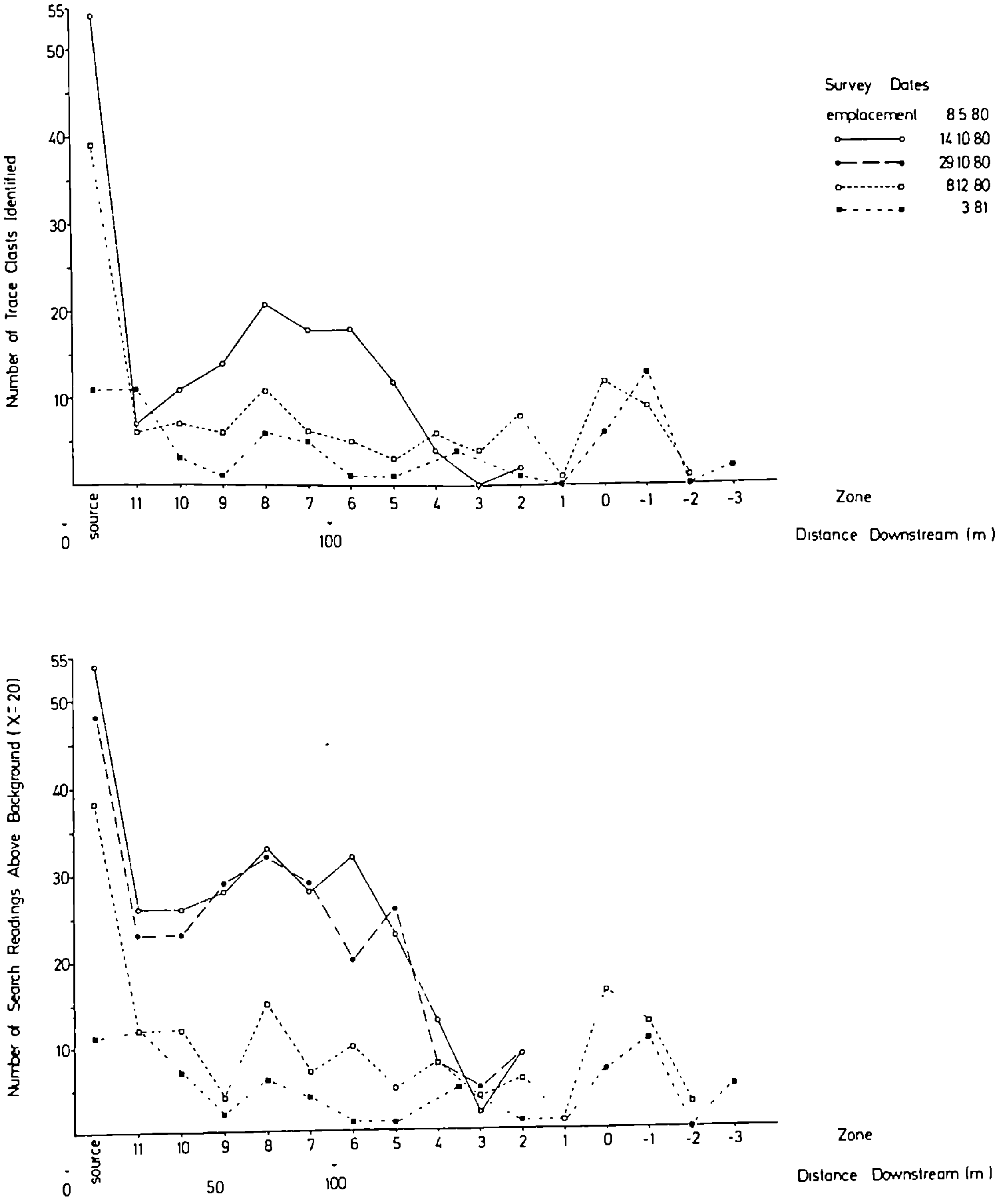


FIGURE 6.10. CEFN BRWYN : THE RELATIONSHIP BETWEEN SURFACE SUSCEPTIBILITY (BELOW) AND IDENTIFIABLE TRACER MATERIAL (ABOVE) PLOTTED AGAINST DISTANCE DOWNSTREAM FROM THE EMPLACEMENT SITE.

has been largely confined to those fractions of bed material coarser than 5.6mm (-2.5∅). The main reason for such low recovery rates is that disturbance of the bed fabric, to identify tracer clast size, was kept to a minimum at all times so that atypical transport rates were not encouraged. Only those clasts at or visibly close to the bed-surface were removed for sizing. Furthermore, the lack of recovery of finer trace clast sizes reflects the low volumes emplaced in the stream bed combined with mixing and dilution processes downstream. Grab samples of bed material of the order of 2-3 Kg showed no evidence of finer tracer material.

Tracer recovery rates may also be affected by the efficiency of detection when using the Wolman style sampling procedure adopted for survey. A sample size of 100-200 readings (depending upon the size of bedform under study) was felt to be the maximum manageable, given the time and labour available to cover a channel length of 1Km.

The relationship between the number of tracer clasts identified to the number of bed surface susceptibility readings above background levels (Figure 6.10.) indicates that more rapid surveys of susceptibility alone may be used to assess tracer movement. The apparently higher recovery rates from the latter method suggests that it may provide a more accurate survey of tracer volume per unit area. Using these data as minimum estimates of the number of tracer clasts recovered at each survey, recovery rates can be increased from 6% to 12+%. However, as emphasised in Chapter 3, the geometric constraints of the search loop mean that individual readings may not be reliably used as an indication of tracer clast size.

The efficacy of the search loop readings in identifying individual tracer clast sizes is illustrated in Figure 6.11. Whilst there was a general trend of increasing tracer clast size with increasing bed

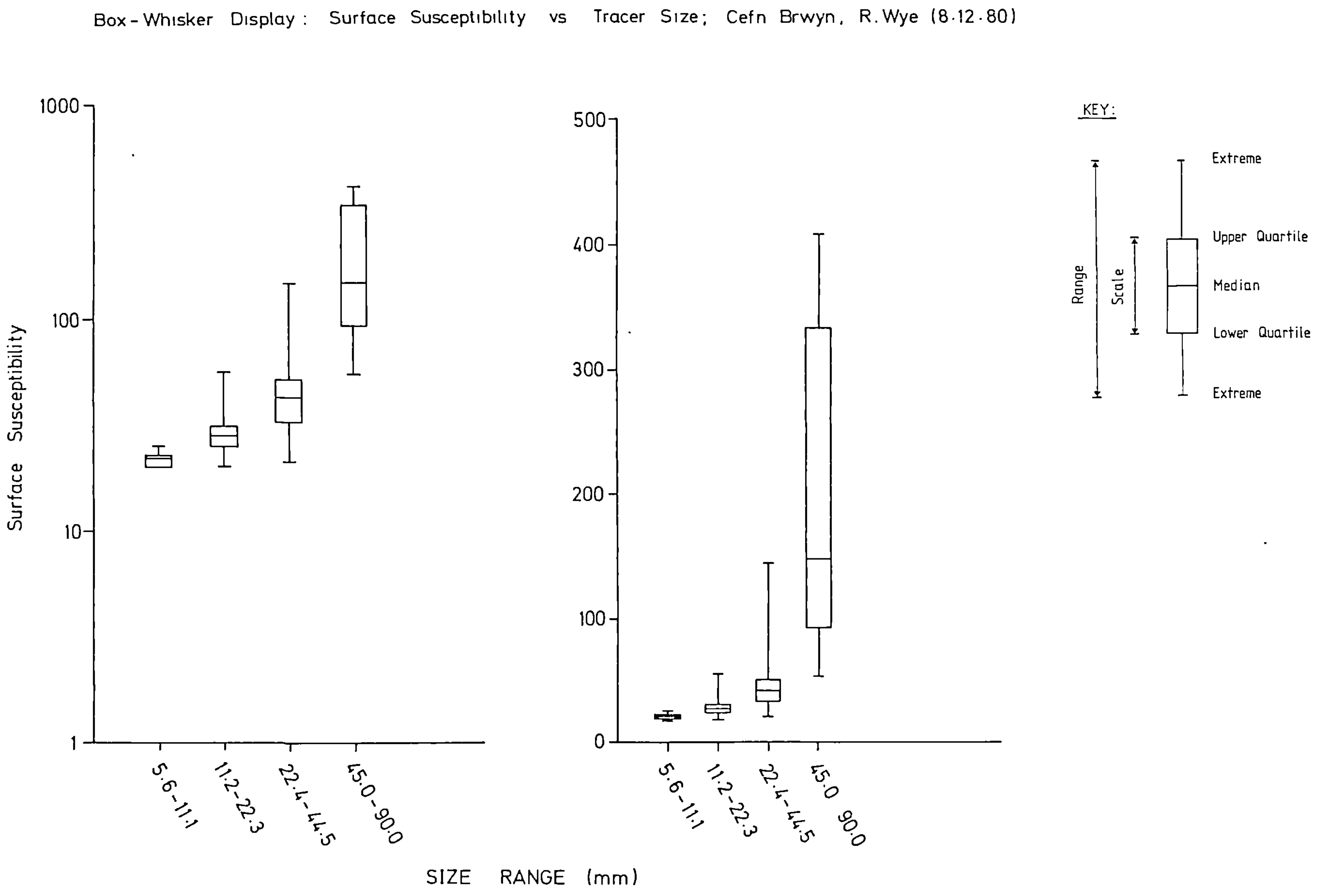


FIGURE 6.11. CEFN BRWYN : VARIATIONS IN BED-SURFACE SUSCEPTIBILITY READINGS RELATED TO TRACER CLAST SIZE (see text)

surface susceptibility, the variability in the data (particularly for coarser clast sizes) was considerable. Susceptibility data generated by the search loop are an average of the area of bed immediately below (approximately 0.035m^2 per reading) together with the effect of any coarse magnetic clasts near to the loop. Susceptibility readings represent a combination in most instances, of natural and enhanced bed material. In many instances, the effect of natural bed material was in providing an overburden of sediment as tracer material became buried. Consequently, for tracer material finer than 11.2mm, bed surface susceptibility data were frequently close to background levels reflecting their positions within the interstices of larger clasts. Recognition, in addition to slightly higher susceptibility readings, was based upon the colour change involved in enhancement. Tracer material coarser than 11.2mm was more readily identifiable using bed surface susceptibility readings; the variation in data shown in Figure 6.11. may be more readily associated with the shape and position of the clast in relation to the search loop; disc shaped clasts on one side may have less effect than a spherical shaped clast of the same size range.

Downstream patterns of tracer movement are illustrated in Figure 6.12.. This shows, for each survey, the cumulative percentage recovery of identified tracer material plotted against distance from the emplacement zone. The pattern of tracer recovery indicated by individual surveys suggests some continuity of tracer transport downstream with a slow progressive increase in the volume of tracer material recovered downstream to the last survey date, 15.9.82. In the first major period of transport, associated with the 6.10.80 storm event, tracer material was found throughout the whole length of the channel thalweg to Cefn Brwyn Bridge; a maximum distance of transport of 160m. Almost 30% of

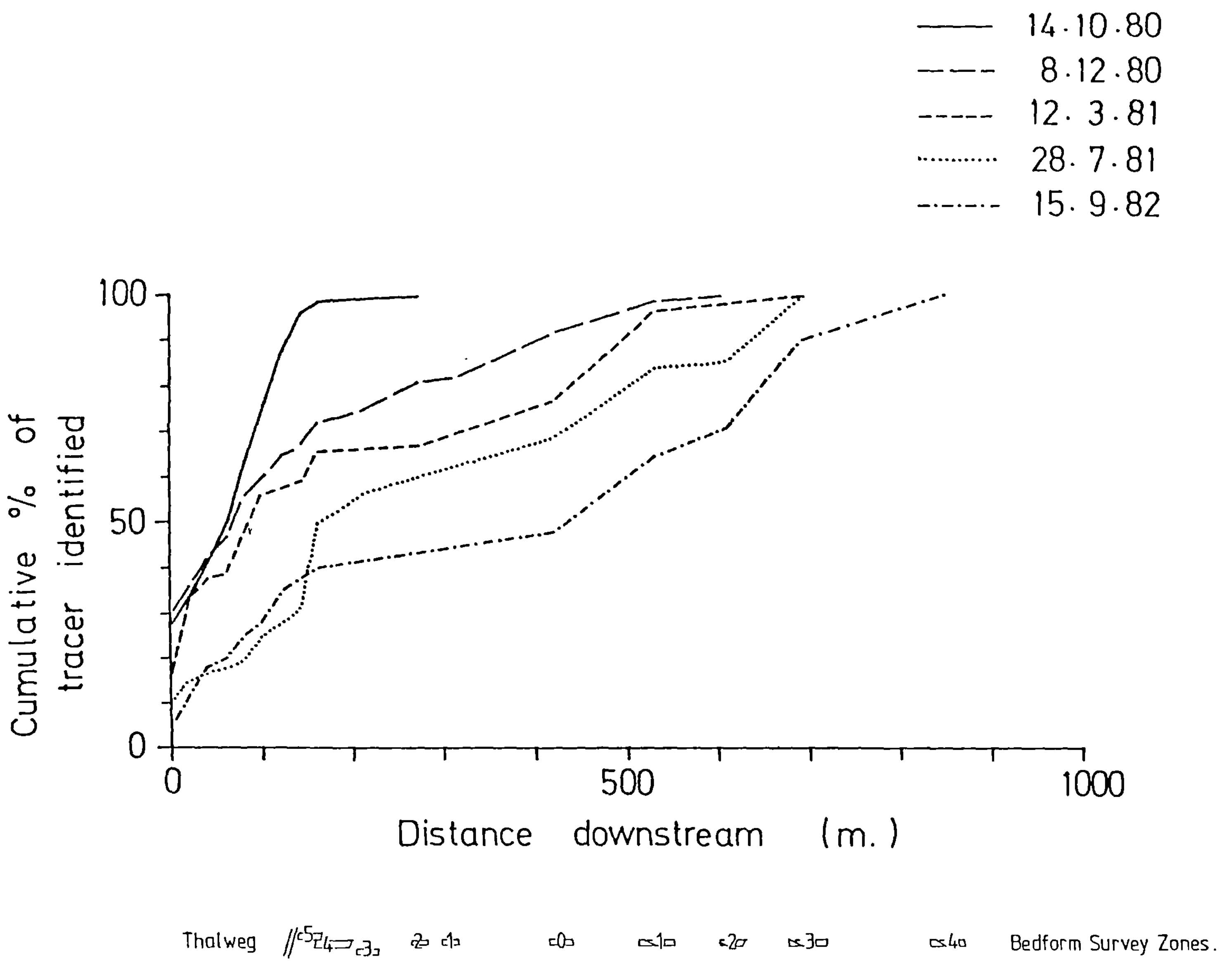


FIGURE 6.12. CEFN BRWYN : CUMULATIVE RECOVERY OF TRACER MATERIAL PLOTTED AGAINST DISTANCE DOWNSTREAM FROM THE EMPLACEMENT SITE.

the tracer recovered was located at the emplacement site, the remaining 70% spread equally over the downstream channel length. Subsequently surveys indicated a gradual reduction in tracer recovery at the emplacement site and an increase, associated with an increasing maximum distance of transport, downstream. However, movement downstream from the main channel thalweg, as shown by the stepped response in Figure 6.12., indicates discrete zones of increased tracer concentration. These were associated with, in the first instance, the group of channel bars immediately downstream of Cefn Brwyn Bridge (Survey zones 1-5), and subsequently with shoal zones under slack water between rock bluffs and rapids and smaller lateral bar features (c.f. Plate 6.1.). If Figure 6.12. is representative of the total body of tracer material, then the data would suggest that, following 23 months of study, at least 30% of the emplaced load was still held upstream of Cefn Brwyn Bridge.

Unlike several previous studies (see Mosley, 1978; Leopold et al, 1964; Church & Jones, 1982), there was no evidence to suggest that tracer deposition occurred in a pattern determined by bedforms. In the first instance this may reflect the length of channel under observation, 1Km. At such a 'local' scale, tracer deposition was in response to meander associated bedforms and bed macro-relief, related to outcrops of solid bedrock.

6.4.2 RELATIONSHIPS BETWEEN TRACER SIZE DISTRIBUTION AND DISTANCE OF TRANSPORT

Figure 6.13. shows the relationship between recovered tracer particle size related to the mean distance moved for each observation period. The data illustrate a negative correlation for those size ranges recovered (2.8 to 64.0mm). However, the main diagram of Figure 6.13. illustrates considerable scatter in the data and suggests that the relationship

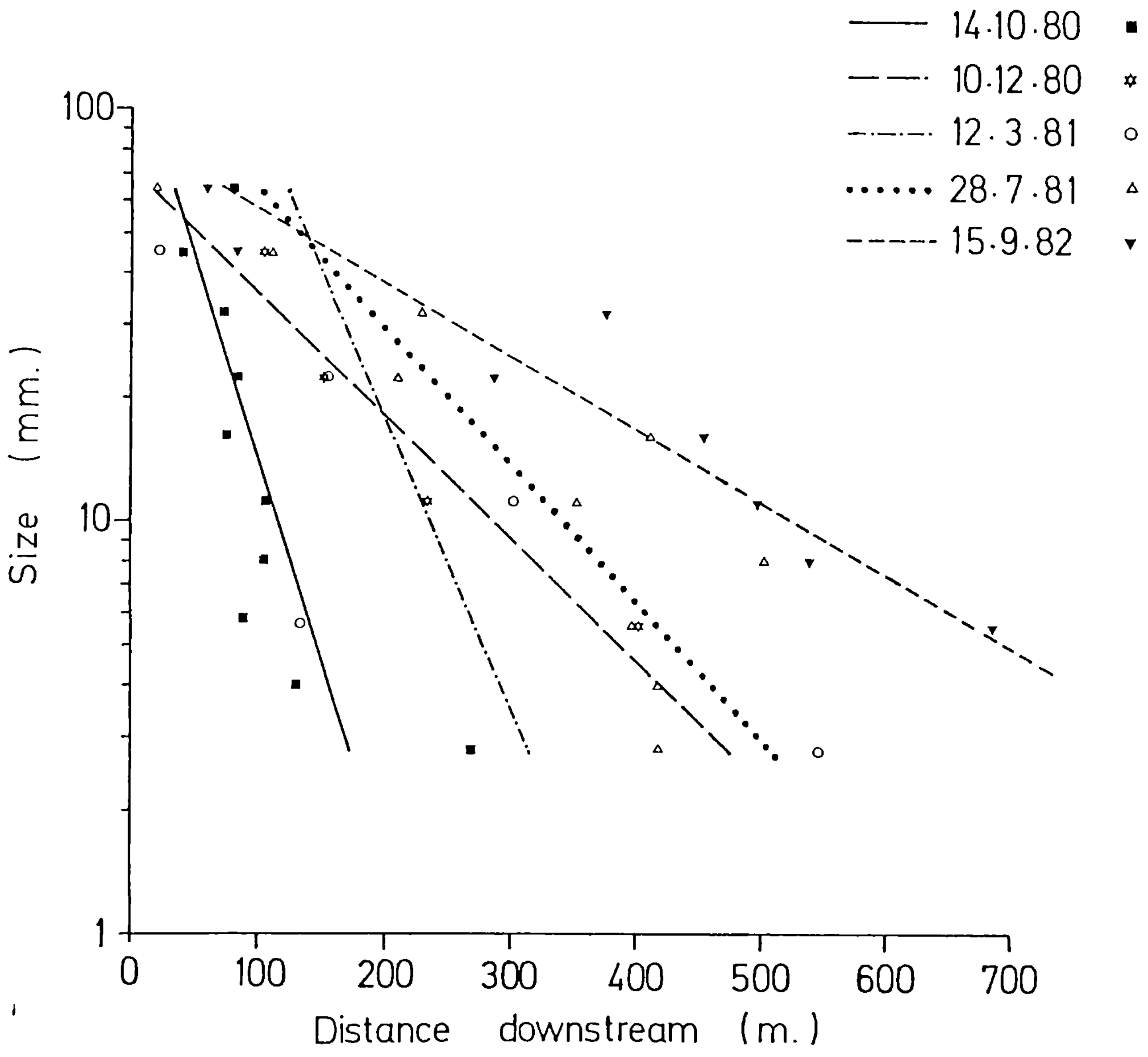


FIGURE 6.13.A. CEFN BRWYN : DOWNSTREAM TRENDS OF RECOVERED TRACER SIZE VARIATIONS - LINEAR TRENDS.

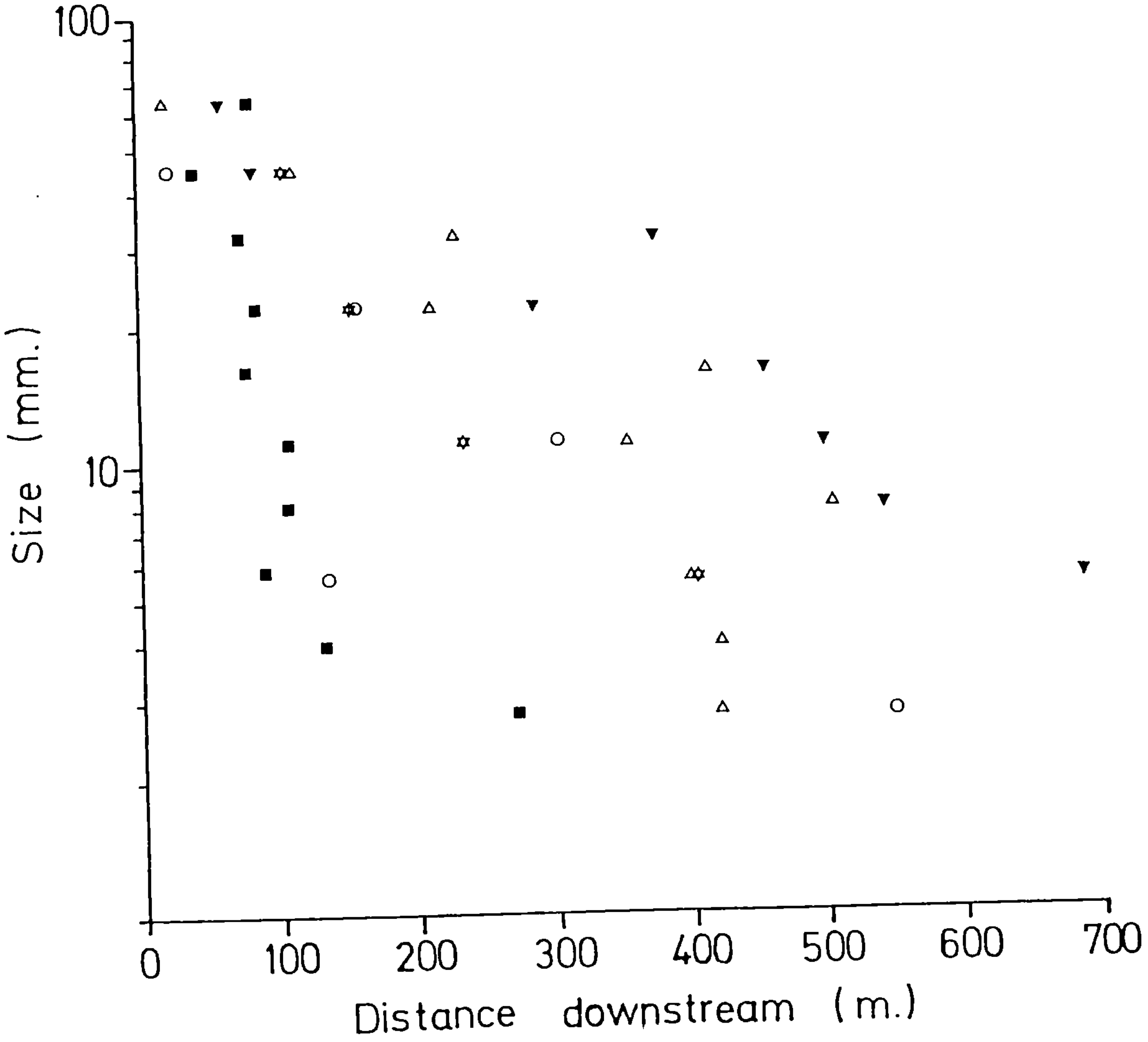


FIGURE 6.13.B. CEFN BRWYN :

DOWNSTREAM TRENDS OF RECOVERED TRACER SIZE VARIATIONS

- RAW DATA.

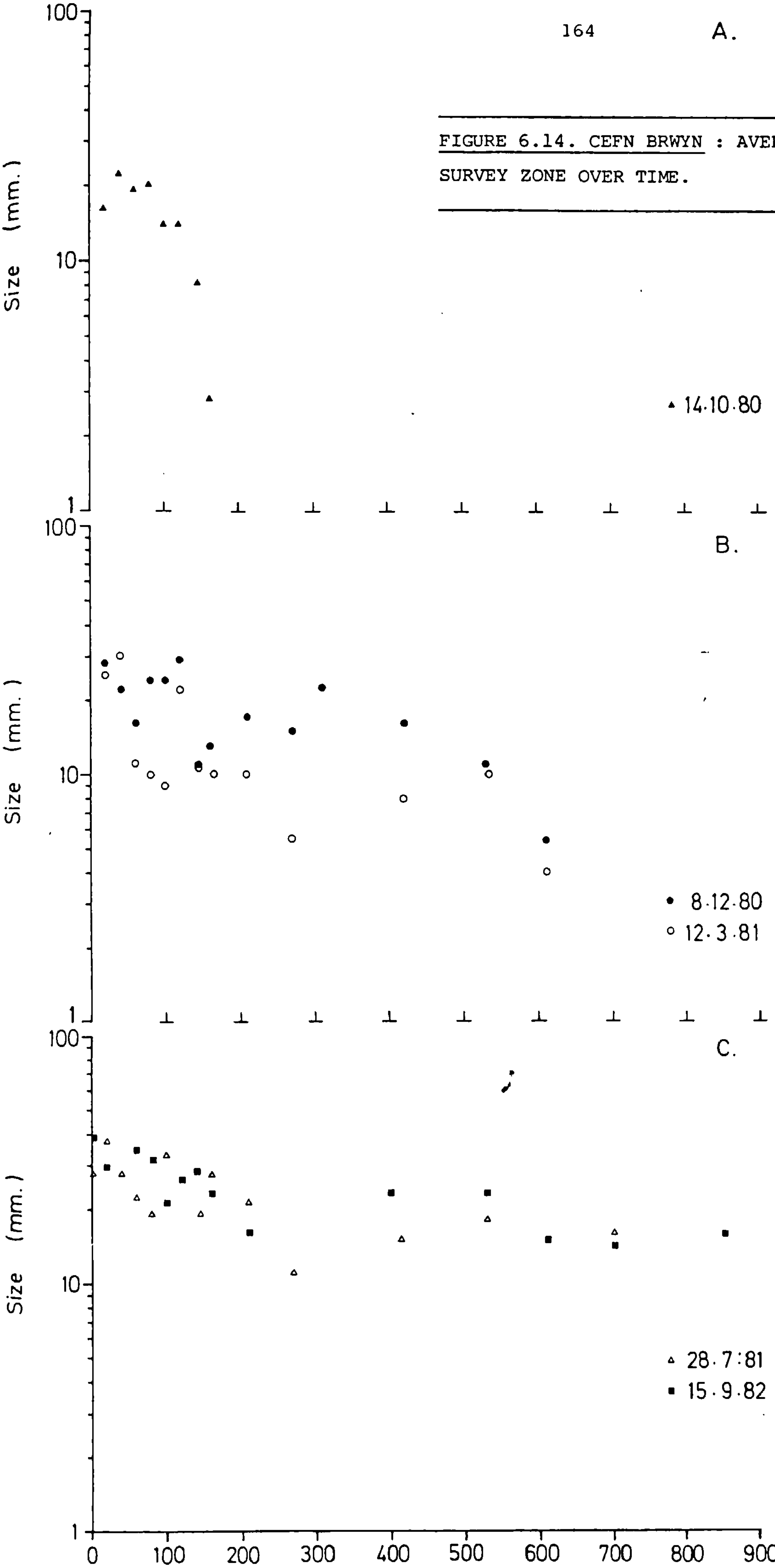
is more complicated than a simple downstream decline in tracer particle size. The variation in trend of the regression lines for successive observations (Figure 6.13., inset) reaffirms the pattern suggested above, of increasing distance of transport of tracer. However, the range of tracer sizes recovered at the upstream and downstream extremes of each survey would appear to remain constant. This suggests that a process of segregation is occurring within the main channel thalweg and that only sizes below a certain threshold are maintained for downstream transport. To test this hypothesis, further analyses of the particle size ranges at each downstream survey zone was carried out (Figure 6.14).

The initial major period of tracer movement (6.10.80) was associated with a clear downstream fining sequence throughout the channel bed to the bridge shoals approximately 160 metres downstream. The coarser clasts appeared to dominate the first 80 metres of this reach. By contrast, subsequent observations illustrated a bimodal tendancy in the recovered tracer sizes which could be differentiated on either side of the Cefn Brwyn Bridge. In most instances, the material downstream of the bridge was finer in size than 22.4mm.

The complexities of the relationship between tracer size and distance of transport are more clearly shown in plots B and C (Figures 6.14). Throughout the study period, a lag of coarse tracer material (>16mm) was evident within the thalweg upstream from the bridge; this illustrated a consistent trend of downstream fining. Similar patterns were also exhibited by the sequence of channel bars immediately downstream of the bridge. However, this trend is disrupted by the apparent coarsening of recovered tracer at sites downstream (survey zones 0 and -1 at approximately 425m and 530m respectively). Downstream from these points there was again a downstream fining tendency.

The differences in size characteristics of tracer recovered upstream and downstream of Cefn Brwyn bridge are illustrated in Figure 6.15.

FIGURE 6.14. CEFN BRWYN : AVERAGE TRACER SIZE PER SURVEY ZONE OVER TIME.



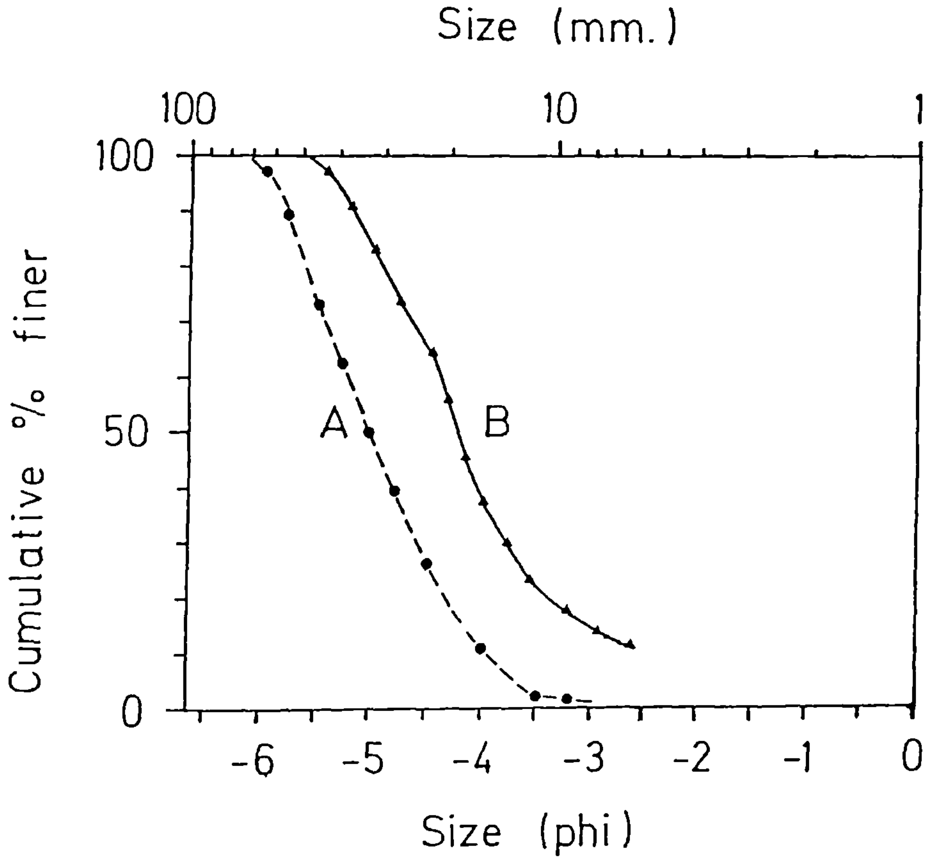


FIGURE 6.15. CEFN BRWYN : TRACER MATERIAL SIZE DISTRIBUTION CURVES FOR MATERIAL RECOVERED IN THE RIFFLE AREA UPSTREAM FROM CEFN BRWYN BRIDGE (A) AND FROM THE SHOALS IMMEDIATELY DOWNSTREAM (B). SURVEY DATE 15.9.82.

Whilst the data described above indicate that all tracer sizes have moved at some time or another, these data suggest that the bulk of the coarse tracer material has remained in the main channel thalweg, upstream from the bridge, moving very slowly if not infrequently downstream. The difference in size of the two tracer populations is 0.75 ∅. The reason for this may be that in straight, rectangular channels neither the depth or secondary flows are sufficient to move the coarse material. Consequently, finer material is removed by storm flows and deposited by a process of sorting downstream.

Comparing the recovered tracer size distribution curves for the whole population at each survey (Figure 6.16) indicated further complexities in the transport processes. Size distribution curves for observations on 8.12.80/28.7.81/15.9.82 plotted consistently at 0.5-1.0 ∅ range coarser than those for the periods 14.10.80 and 12.3.81. Analysis of the hydraulic data in Table 6.2 indicated that these latter observations were associated with surveys following very high peak storm discharges. By contrast, the other observations were associated with surveys following relatively low flood flows, peaking frequently over a period of days. The data would suggest that the higher storm flows of the former were of a sufficient competence that the whole bed fabric was disturbed enabling the release of a much wider range of tracer sizes.

These observations are consistent with those of Scott and Gravlee (1968) and Laronne and Carson (1976) in suggesting that larger floods are able to activate the river bed to a much greater depth and that the post-depositional bed following such less frequent events will be more stable. As a result, those observations of the 8.12.80 survey would be consistent with a stable bed containing a store of tracer material. Similar processes are discussed by Klingeman and Emmett (1982) and Milhous and Klingeman (1973), showing that at high flows

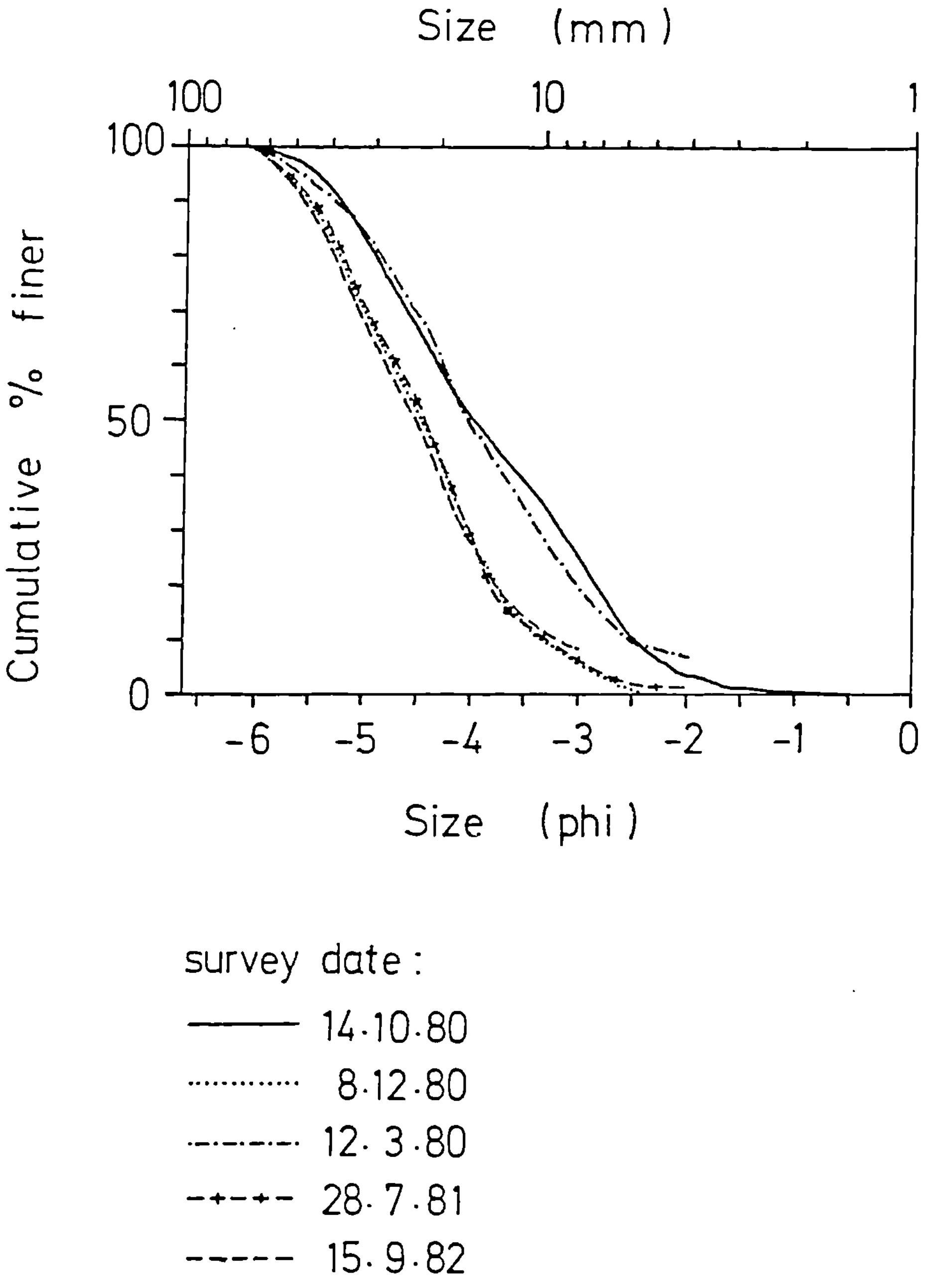


FIGURE 6.16. CEFN BRWYN : TRACER MATERIAL SIZE DISTRIBUTION CURVES FOR MATERIAL RECOVERED AT EACH SURVEY DATE.

a coarse surface layer (armour) will break up to provide a whole range of sediment sizes for transport. The armour layer is effective in controlling the availability of a reservoir of sediment (Chapter 2); its stability is related to the time period between runoff events and the magnitude of individual storms. In the case of Cefn Brwyn, it is likely that the transporting efficiency of flows is limited by the lack of supply of bed material (particularly finer size ranges) from upstream as a result of capture in the weir pool of the IOH compound crump weir. As a result, the gradual selective removal of finer material results in a progressively coarser armoured surface which will be stable at the majority of flows. Similar processes have been described by Little, Campbell and Meyer (1972). The formation of such a surface may also provide more "hiding places" for tracer sediment.

. 5. MOVEMENT OF PAINTED PEBBLES

The movements of painted pebbles during the Cefn Brwyn tracer experiment are summarised in Table 6.2.. The painted pebbles, all 64+mm in size were observed following movement from the intersections of the survey grid over the emplacement site. The data describing mean transport distances confirms those conclusions drawn above ie. that the bulk of the coarse tracer load was still in the main channel thalweg upstream of Cefn Brwyn Bridge at the end of the observation period. The variations in extremes of transport distances (MAX/MIN) illustrate the importance of burial processes affecting coarser tracer material, a combined effect of the segregation of finer material and the gradual armouring of the channel bed.

DATE	14.10.80	8.12.80	12.3.81	28.7.81	15.9.82
Number found	8	8	11	10	4
Minimum Distance (m)	1.5	4.0	2.0	10.0	20.0
Mean Distance (m)	9.25	85.0	42.0	38.0	97.5
Maximum Distance (m)	29.0	420.0	120.0	200.0	120.0

Total Number Emplaced = 12

TABLE 6.2. CEFN BRWYN : DISTANCE OF TRANSPORT FROM EMPLACEMENT SITE AND RECOVERY OF PAINTED TRACER MATERIAL.

6.6 DOLYDD LLWYDION, AFON LLWYD

6.6.1 DOWNSTREAM PATTERNS OF TRACER MOVEMENT

Downstream patterns of tracer movement for the Dolydd reach have been summarised in Figures 6.17 -6.19. These illustrate:

i. cumulative plots (%) of tracer clasts identified at each individual survey following movement from the emplacement size;

ii. the range and number of bed-surface susceptibility readings above a background level; and,

iii from ii, cumulative plots (%) of the number of bed-surface susceptibility readings above background against distance downstream from the emplacement site.

Two main features are apparent from the data. Firstly, in comparison to the Cefn Brwyn trace, the pattern of tracer recovery indicates a substantially high proportion of tracer was transported to, and retained at, the main channel shoal within 80 metres of the emplacement site during the study period. Secondly, in a similar manner to the Cefn Brwyn trace, deposition of tracer beyond the main bridge shoal (Figure 6.17 A to E) occurred in response to channel storage areas, in most instances lateral or medial bars associated with a meandering planform.

The broad relationship between bed-surface susceptibility and tracer clast size established for Cefn Brwyn (Figure 6.11) may be used here for a subjective assessment of size variations from Figure 6.18. Bed surface susceptibility variation decreased with observations downstream, as did the magnitude of susceptibility readings suggesting an overall pattern of downstream fining (see later). However, the range and quantity of bed surface susceptibility readings over the main Dolydd shoal suggested that a wide range of tracer clast sizes was held in storage here. However, the limitations imposed in sampling individual bed features ie. in not disturbing the bed fabric, meant that overall tracer recovery rates

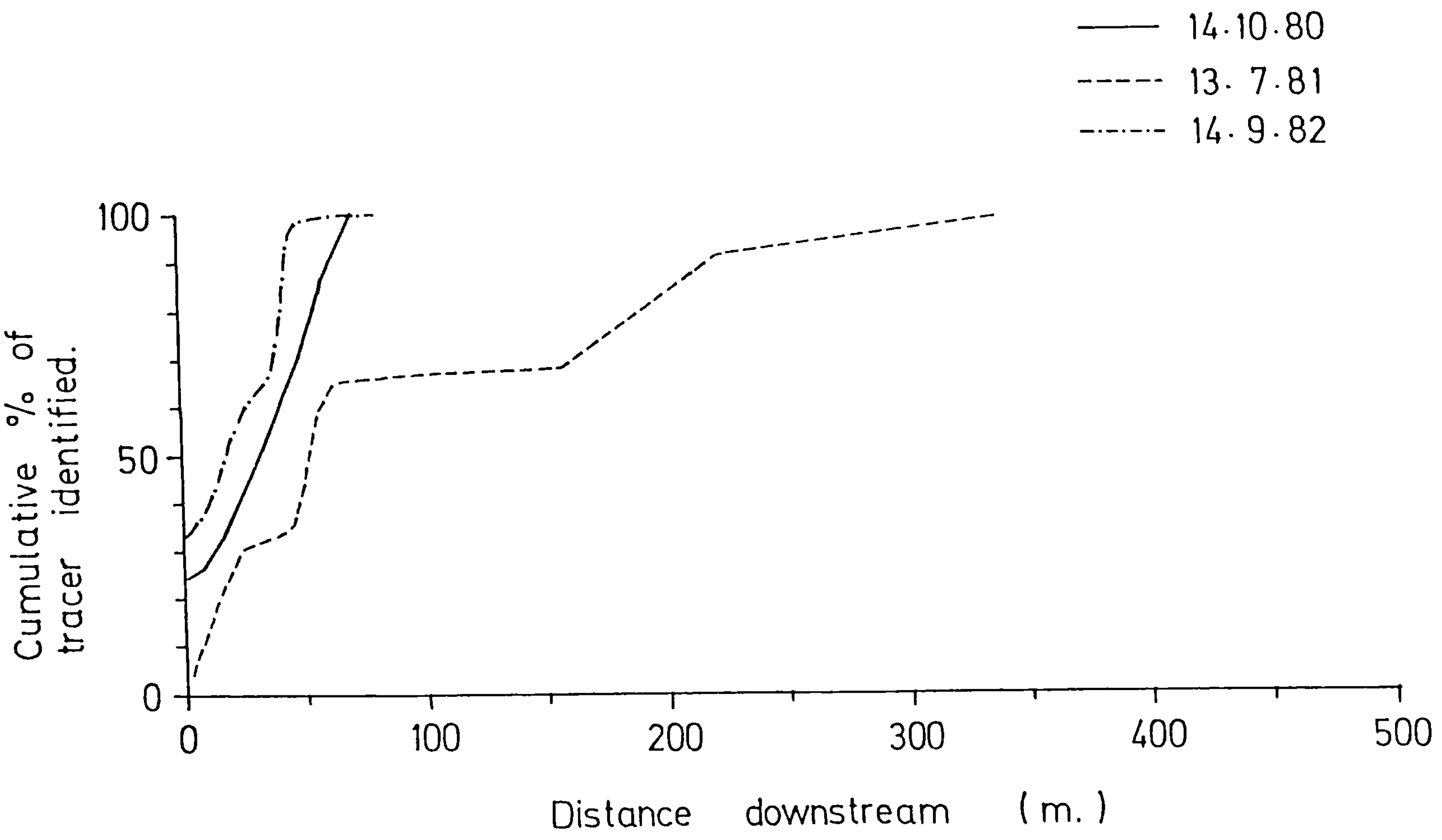


FIGURE 6.17. DOLYDD : CUMULATIVE PERCENT RECOVERY OF TRACER MATERIAL PLOTTED AGAINST DISTANCE DOWNSTREAM FROM THE EMPLACEMENT SITE.

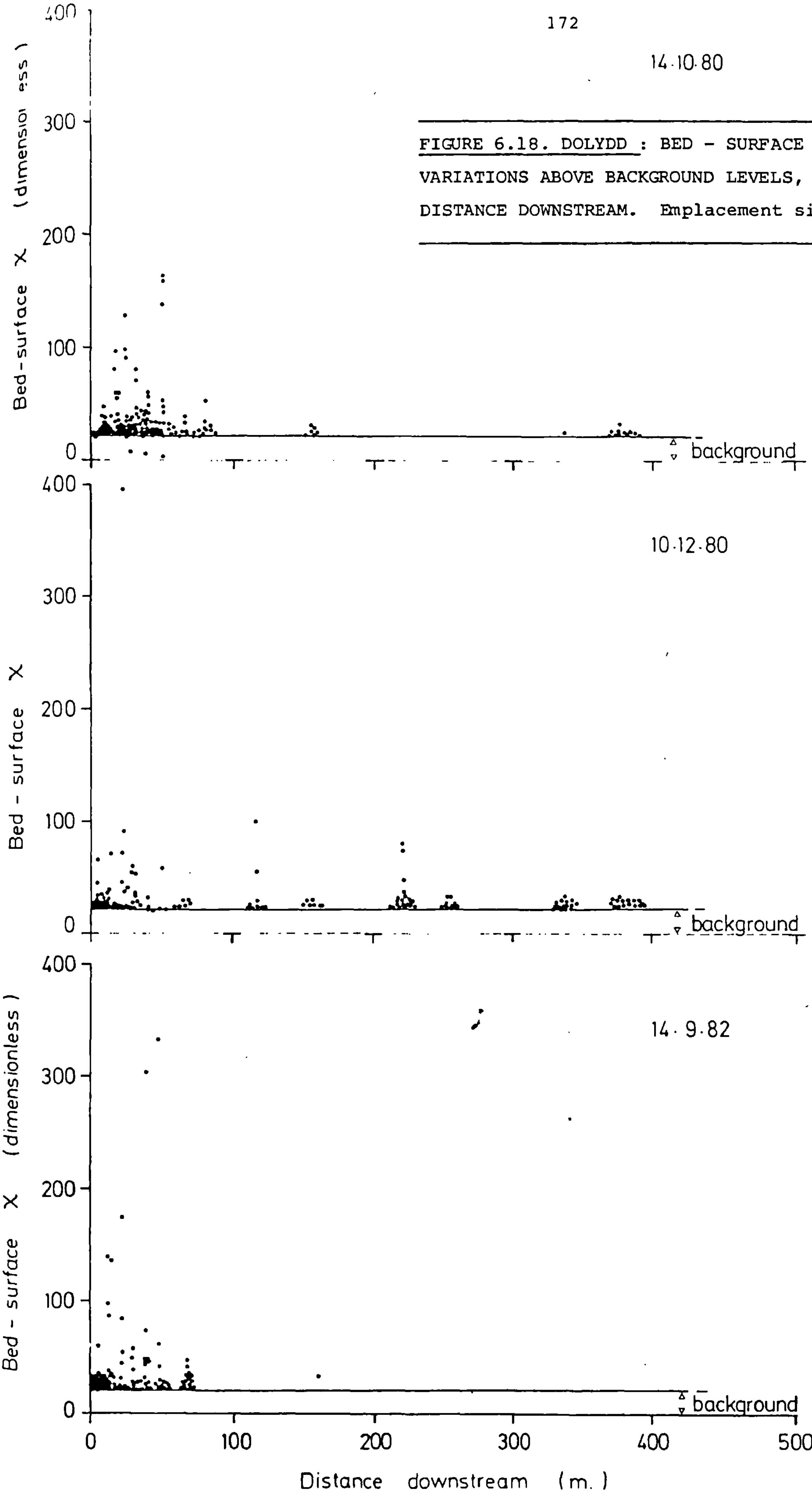


FIGURE 6.18. DOLYDD : BED - SURFACE SUSCEPTIBILITY VARIATIONS ABOVE BACKGROUND LEVELS, PLOTTED AGAINST DISTANCE DOWNSTREAM. Emplacement site not included.

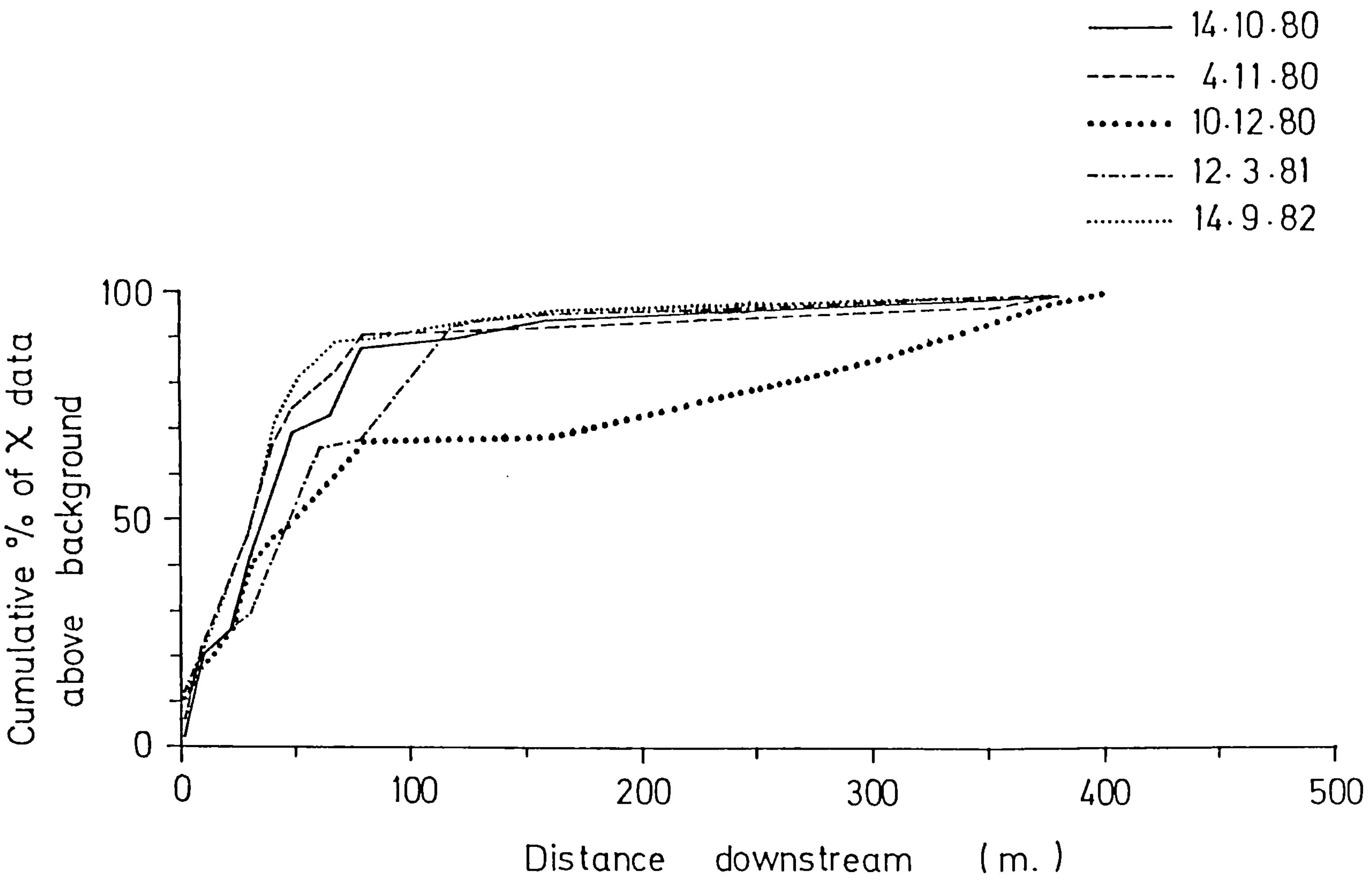


FIGURE 6.19. DOLYDD : CUMULATIVE PERCENT OF BED-SURFACE SUSCEPTIBILITY READINGS ABOVE BACKGROUND LEVELS PLOTTED AGAINST DISTANCE DOWNSTREAM FROM THE EMPLACEMENT SITE.

were very low (<2%). On the basis of the relationship between bed-surface susceptibility and tracer recovery established for Cefn Brwyn (Figure 6.10), the number of high susceptibility readings occurring on the main Dolydd shoal was assumed to be representative of the actual response of tracer to transport processes. As a result, the data from Figures 6.17 and 6.19 suggest that between 70-80% of the tracer load has remained stored within the Dolydd shoal subsequent to the main flood event of 6.10.80.

As indicated above, tracer material was also recovered some distance downstream from the main Dolydd shoal in association with other storage features. However, the overall low recovery rates and gradual loss of this material (as shown in Figures 6.18 and 6.19) suggests that the volume of material transported beyond the Dolydd shoal area was not as large as the 25% suggested by Figure 6.17. Further analysis of size characteristics was carried out to verify this.

THE RELATIONSHIPS BETWEEN TRACER SIZE AND DISTANCE OF TRANSPORT

Downstream variations in recovered tracer material size range are summarised in Figure 6.20, relating mean tracer clast size to distance from the emplacement site. Two features are apparent from the data.

Firstly, an overall trend of downstream fining of recovered tracer material was observed for the study period. Tracer material recovered downstream from the Dolydd shoal ranged in size from 2.8mm to a maximum of 22.4 mm. By contrast, a much wider range of tracer clast size was represented within the fabric of the Dolydd shoal, although this too exhibited a pattern of downstream fining over a much shorter distance. The coarsest size range of tracer clasts were located at the upstream bar head, fining downstream to the bar tail.

Size distribution curves were analysed for the recovered tracer

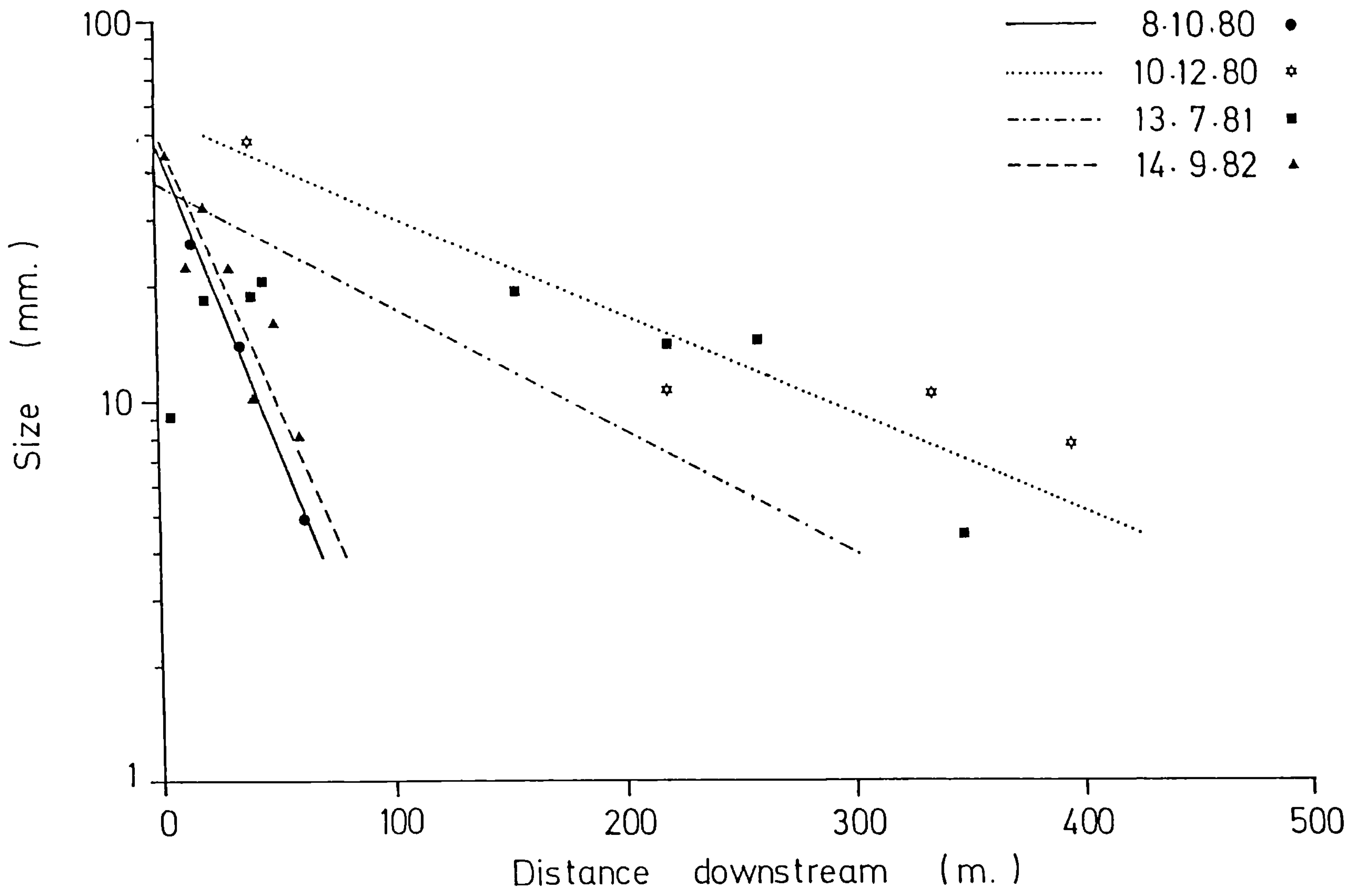


FIGURE 6.20. DOLYDD : DOWNSTREAM TRENDS OF AVERAGE RECOVERED TRACER MATERIAL SIZE.

material and compared to those of the original emplaced tracer material to monitor any changes occurring through successive surveys. (Figure 6.21). These were dominated by the coarser size ranges, probably related to the development of an armour surface over the shoal (c.f. Figure 6.22). Some selective removal of material < 16mm occurred over successive surveys and was transported downstream as indicated by Figures 6.19 and 6.20. This did not appear to be a continuous process as very little tracer material was recovered downstream of the Dolydd shoal subsequent to the 13.7.81 survey. Despite the obviously coarse nature of the recovered tracer material, this was consistently finer overall than the emplaced tracer load. Only the tracer material remaining at the emplacement site exhibited similar characteristics, excepting the lack of finer clast sizes, reflecting the original pattern of scour at the bridge and the stable nature of this part of the Afon Llwyd channel.

More detailed spatial variations of tracer recovery on the Dolydd shoal are illustrated in Figure 6.23. Patterns of recovery related to bed-surface susceptibility indicated the lack of movement at the right bank side of the emplacement site and deposition towards the bar tail. The downstream fining of recovered tracer clasts is evident from mean clast size data, and is broadly repeated in variations of maximum clast size. Upstream coarsening of tracer clast size would appear to be related to accretion at the bar head and scour of finer particle size ranges in the pool sections of the Dolydd bridge. In addition, the main flow lines (shown on Figure 6.22) are associated with coarser tracer deposits than those found at the exposed bar areas, the bar tail and small shoals downstream from the main Dolydd shoal. This suggests, as noted by Bluck (1982), that turbulence at the sediment-water interface together with variations in the surface fabric of the shoal are important controls of clast segregation. Maximum clast size data over the study period,

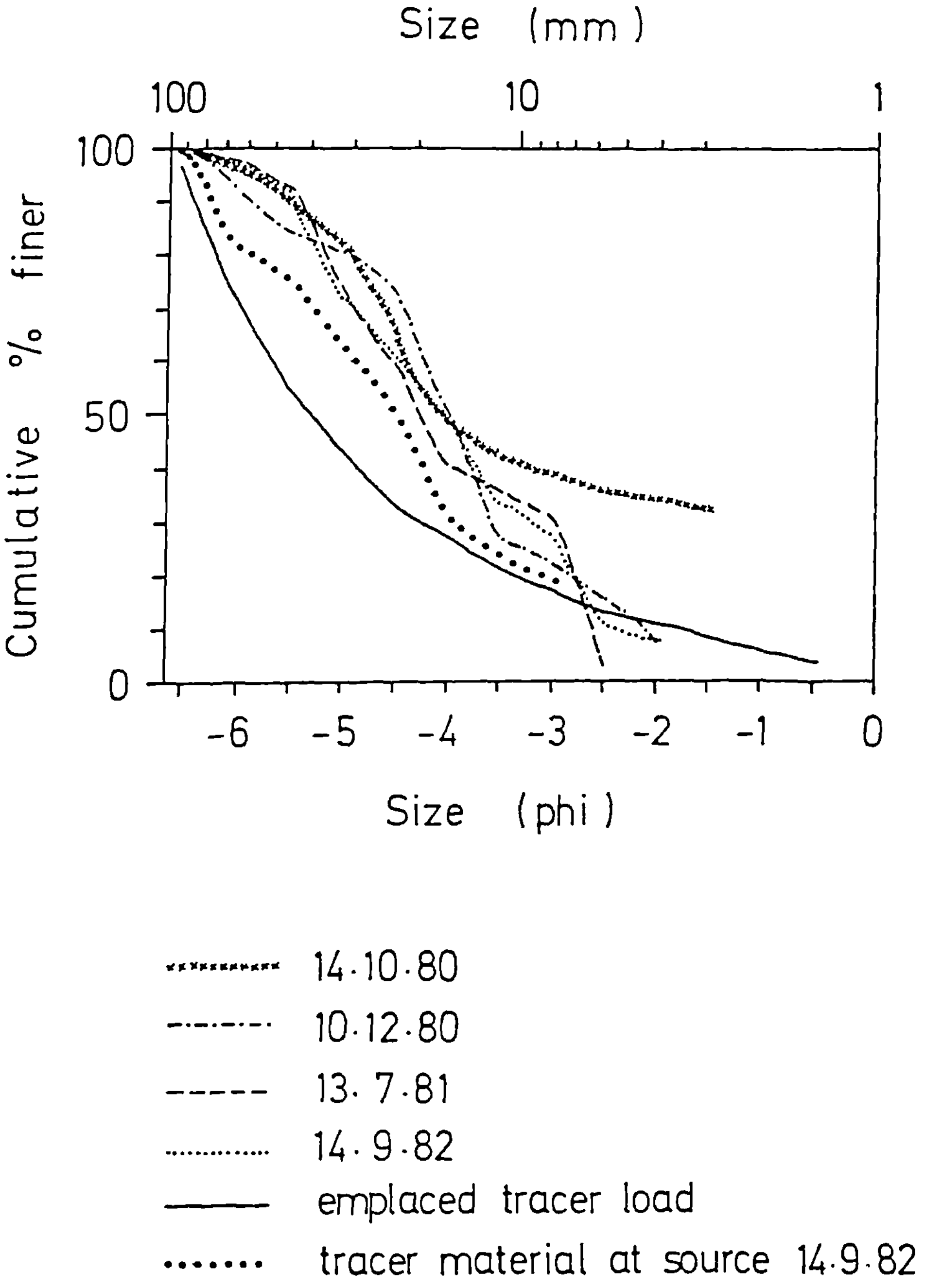


FIGURE 6.21. DOLYDD : RECOVERED TRACER MATERIAL SIZE DISTRIBUTION CURVES FOR EACH SURVEY DATE.

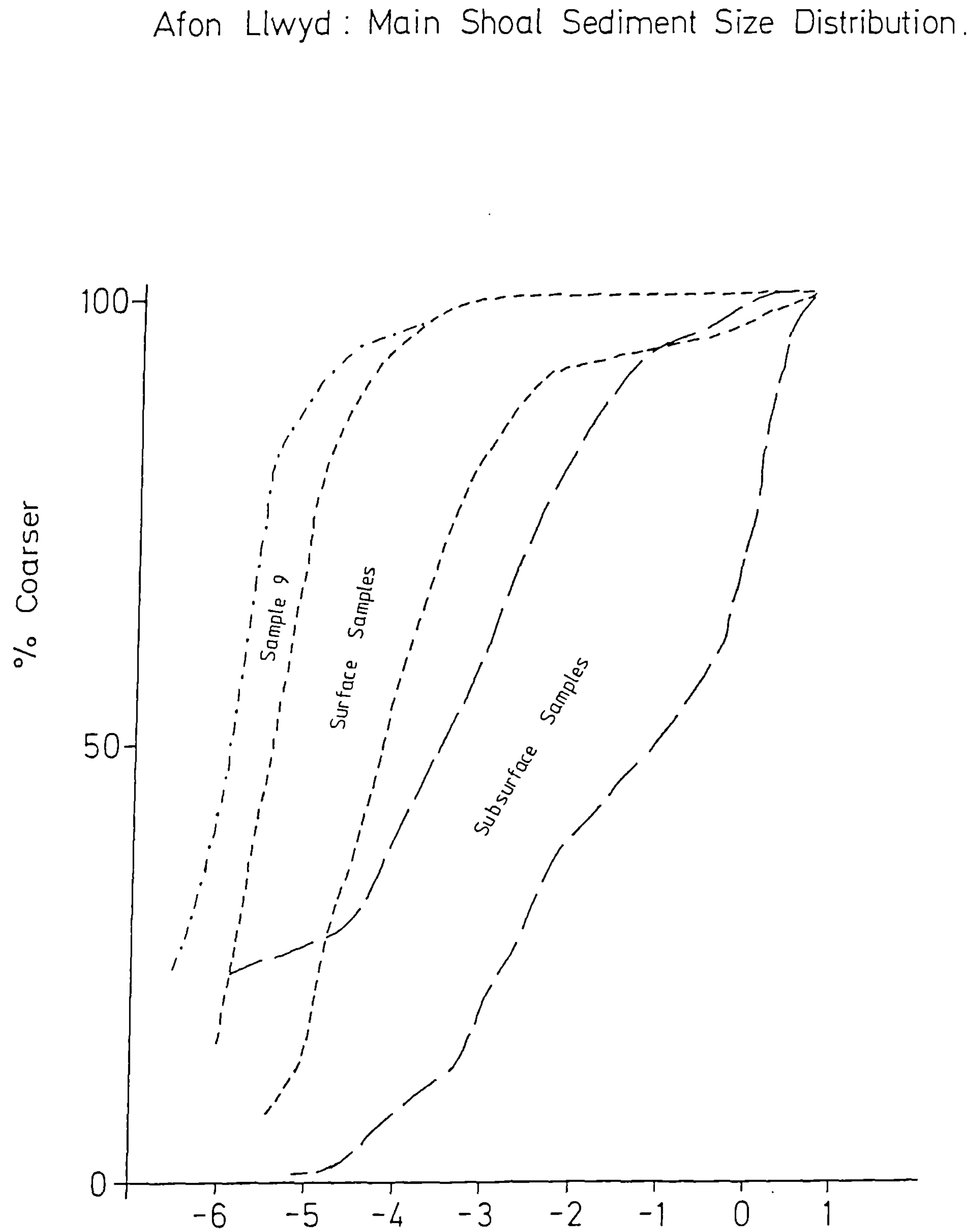


FIGURE 6.22. DOLYDD : MAIN BRIDGE SHOAL; PARTICLE SIZE DISTRIBUTION CURVES FROM AUGER SAMPLES OF BED-MATERIAL ILLUSTRATING THE SHOAL'S COARSE, ARMOURED SURFACE.

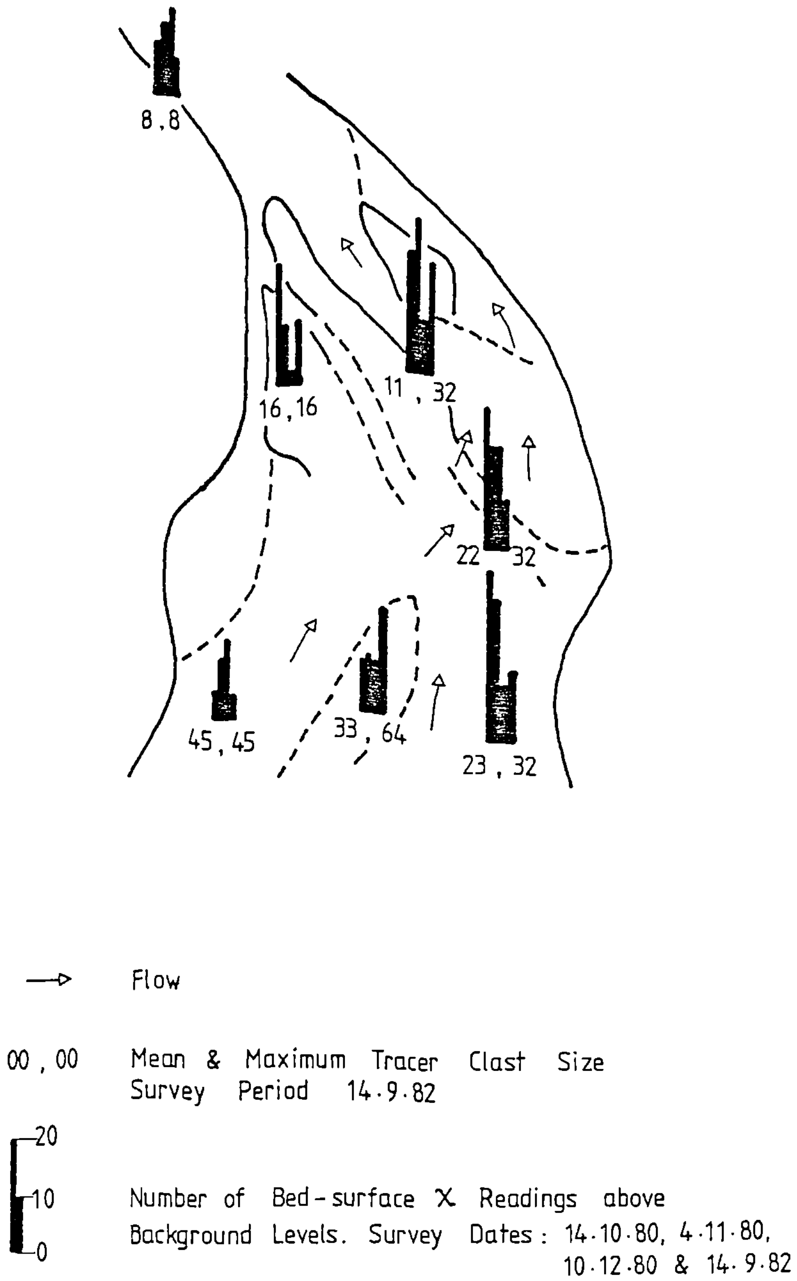


FIGURE 6.23. DOLYDD : SPATIAL AND TEMPORAL TRENDS IN TRACER MATERIAL RECOVERED FROM THE MAIN SHOAL AREA (Figure 6.6.). HISTOGRAMS ILLUSTRATE TEMPORAL CHANGES IN BED-SURFACE SUSCEPTIBILITY.

show a progressive coarsening indicating a slow increase in stability of the armour surface. This was paralleled by a gradual decrease in the recovery of tracer material (as indicated by bed-surface susceptibility data) over the Dolydd shoal reach, suggesting a process of supply limitation by the redevelopment of a stable armour surface.

6.7. CONCLUSIONS

Tracer recovery rates from relatively large reaches of the Upper Severn and Wye tributaries were extremely low, of the order of 1 to 12%. This reflects the processes of sediment movement in gravel bed channels combined with (on hindsight) rather low volumes of tracer material emplaced. Tracing over similar distances, although slightly wider channel reaches, Mosley (1978) used $3m^3$ bulk volume of tracer or approximately 5 Tonnes. Nevertheless, Mosley's recovery rates were equally low, of the order of 5% by volume. Patterns of bed-surface susceptibility,whilst not providing an accurate assessment of size or volume of tracer at a point, gave consistently higher recovery results based on those tracer clasts which remained unseen below the bed surface. From the data collected above, the maximum operable depth of the equipment used was approximately 10cms, depending upon the size and position of the tracer clast related to the depth of overburden of the unenhanced bed material.

In comparison with many other tracer experiments (c.f. Chapter 2), the quantification of transport rates and description of patterns of movement were much more difficult to approach. The derivation of such data is normally based upon a range of flow conditions and a range of sediment transport rates. However, supply limitation of tracer sediment by storage in bedforms or as a result of channel armouring, meant that sediment movement was related to one to two individual events of very high discharge magnitude or as a result of sediment grading by winnowing from an armoured surface. Conventionally accepted calculations of sediment transport - discharge rates related to such approaches as those of Shields or Bagnold (see Chapter 2) could not be made. However, the problems described above are typical characteristics of gravel bed transport processes. More detailed analyses would benefit from the emplacement of measuring equipment in the stream bed as has been described by Ergenzinger and Conrady (1983) and Reid et al. (1984), which would be able to supplement data on the frequency and magnitude of sediment transport with data describing conditions for incipient motion and deposition, and the changes in tracer volume stored in a cross-section at any one point in time.

One major problem with the magnetic technique at this scale of study is the collection and presentation of data to represent tracer movement. Spatial variations in bed-surface susceptibility in longitudinal cross-section provide a useful assessment of tracer movement. However, when presented as a 2 dimensional view of the river channel (Figure 6.24), much of this pattern may be lost. Isolated areas of densely packed susceptibility contours frequently represent the occurrence of only 1 or 2 tracer clasts which form the coarse armoured channel surface. Finer tracer material buried beneath the armour layer are not so easily identified by this approach. When compared with other figures used in this chapter, such an approach appears unrepresentative of the volume of tracer at a point. These problems may be overcome by further developments in instrumentation which might be pursued along the lines of industrial pipe detectors of a manageable size when immersed to the stream bed. Any further design must incorporate a more sensitive search loop capable of penetration beyond the 10cm limit described above. Despite these reservations, the data described above would suggest that the technique still holds considerable potential.

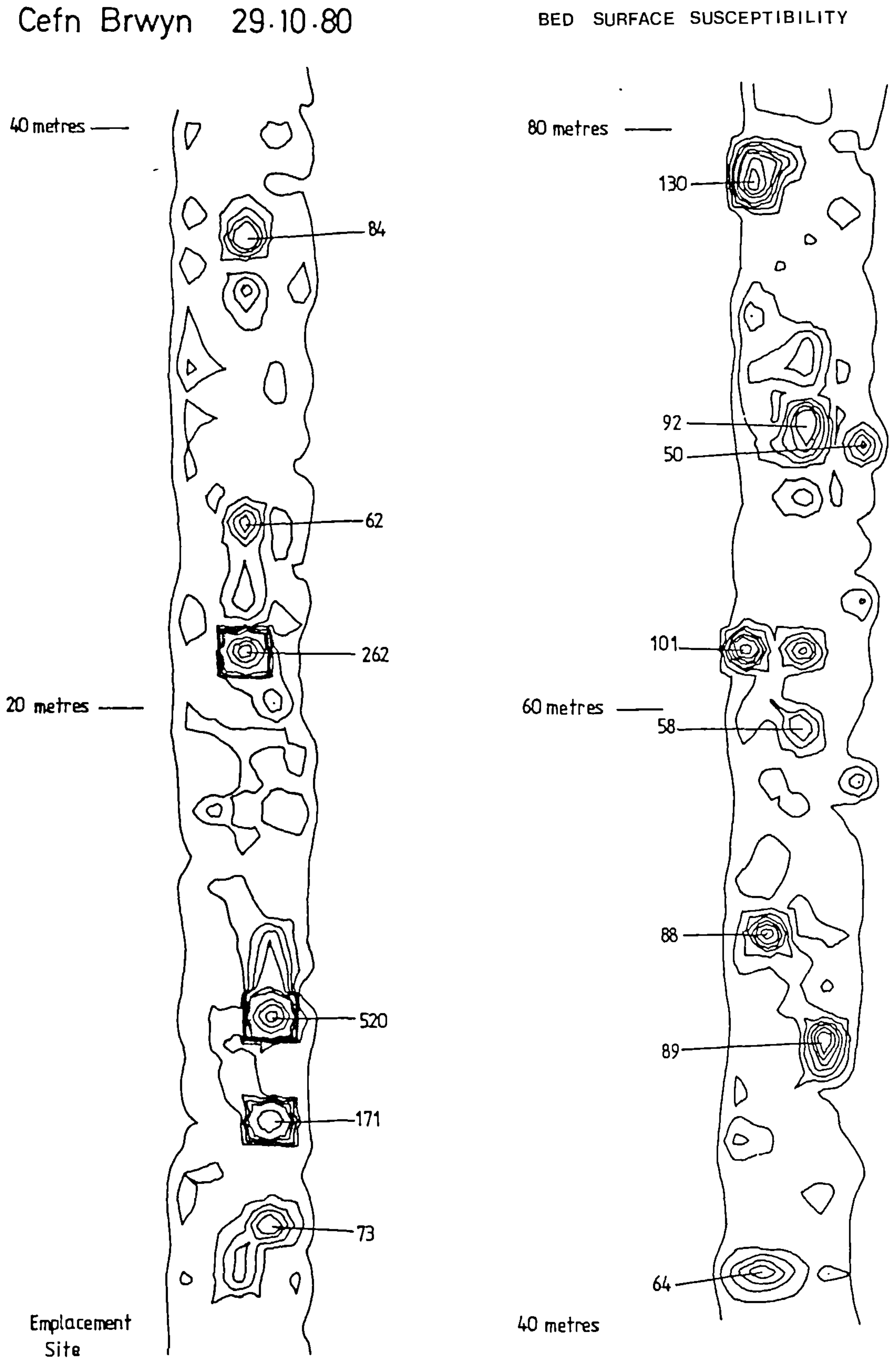


FIGURE 6.24. CEFN BRWYN : AN EXAMPLE OF THE TYPE OF PLOT WHICH MAY BE PRODUCED FOR THE STUDY REACHES BUT ILLUSTRATING THE MARKED CHANGES CAUSED BY THE NATURE OF HETEROGENEOUS BED-MATERIAL MATRICES. THE CONTOUR PLOT IS DERIVED FROM A RAW DATA MATRIX OF READINGS SPACED 0.5m DOWNSTREAM BY 1.0m ACROSS STREAM.

The results discussed above illustrate two key features of transport in gravel bed rivers : The importance of channel bar features for storage of transient bed sediment loads; and, the supply variations of sediment as a result of limitations imposed by armour sediment stabilisation. Discrete sedimentation zones or bars have received relatively little study with regards transport processes (Church and Jones, 1982). However, this data confirms those hypotheses of Meade et al. (1981), and to some extent those of Mosley (1978) and Laronne and Carson (1976), that transport rates of heterogeneous sediments are intimately related to the occurrence of channel bars. In many cases, as shown in Chapter 7 (Morfiodion) and for the Afon Llwyd (above), channel bars represent major sediment storage elements in the river channel which are moved only sporadically by significantly high flows.

For individual bed features and downstream as a whole, these tracer experiments illustrated a tendancy toward downstream fining of sediments not as a result of abrasion or fracturing of coarse sediments, but as a result of size segregation. For the Cefn Brwyn trace, the effect of the upstream compound crump weir may be likened to that of regulation, producing supply-limitation of sediment from upstream especially of finer size ranges. Size segregation at the majority of flows resulted in the accumulation of coarse sediment in the upstream tracing reaches as a pavement or armour layer (Little, Campbell and Meyer, 1972) which was effective in restricting supply of finer sediments from the reservoir below (Klingeman and Emmett, 1982). However, significantly larger discharge events appeared to activate the bed to a much greater depth (see Laronne and Carson, 1976) perhaps as a result of dilation of the grain bed (Emmett and Leopold, 1965) thereby releasing a much finer size distributed tracer sediment population. Similar field observations have been described by Gomez (1983).

By contrast, the apparent stability of the Dolydd shoal site was indicated by the winnowing of fine tracer clasts and the progressive coarsening of the surface "armour" layer preventing any further transport even at flows as great as those which formed the shoal.

These conclusions are in common with those of Milhous and Emmett (1973), Klingemann and Emmett (opp. cit.) and suggest that the relationship of stream power to bedload transport rate is not a simple one. Future developments should bear in mind those occasions where the whole bed is effectively in transport against those when the coarser fractions are most stable and despite stream competence, only limited transport of segregated finer sediments is likely to occur. For the experiments described above, the critical threshold of discharge for this relationship appeared to lie at approximately 17 to 18 cumecs for Cefn Brwyn and 90+ cumecs at the Dolydd site. In both cases, these flow ranges were also associated with the initial phase of transport when all of the tracer material appeared to have been moved. These flows were well above critical discharge calculations based upon the Schoklistch approach, of 4 and 6 cumecs respectively.

In addition to these temporal variations in sediment transport, the efficacy of quasi-stable bed storage features means that considerable spatial variations in transport rate may occur in gravel bed rivers. These are important considerations to bear in mind when sampling bed material load.

CHAPTER 7

TRACER APPLICATIONS AND OBSERVATIONS OF PROCESS AND PATTERN IN PIEDMONT CHANNEL STORAGE AREAS, MORFODION, RIVER SEVERN

7.1. INTRODUCTION

This chapter is concerned with the observation of sediment movement from channel storage areas in the piedmont reaches (Newson, 1981) of the River Severn.

The sediment transport system of the piedmont channel is characterised by a higher ratio of suspended- to bed-load of which the latter is derived partly from upland flood supply, but mainly by the reworking of the piedmont floodplain and channel stored load. This coarse load moves slowly and sporadically through the system in response to upland flooding (Ferguson and Werrity, 1983), and in so doing may become incorporated into sedimentary accumulations, termed bars. The apparent stability of these features means that material held in storage within them may remain immobile during the majority of high flows. The consequence of this persistence of form is possible increased bank erosion during floods contributing to suspended sediment sources.

Information regarding form and activity in these reaches is still largely speculative (Lewin 1978, 1981; Church and Jones, 1982) and our present understanding is derived from comparitively few studies from largely higher energy environments than those experienced in the U.K. (see Ashmore, 1982; c.f. Table 7.1). The very nature of channel change in these reaches poses problems for the observer; changes in bed morphology occur at extreme discharges when the water is turbid, therefore observation is difficult and measurement almost

RESEARCHERS		AREA	DISCHARGE CHARACTERISTICS (cumecs)	LOCAL RIVER SLOPE ($m.m^{-1}$)
Bluck	1971	R.Endrick, Scotland	Q_b - 8	-
Church & Jones	1982	South River, Baffin Island	Q_b - 100	
		Bella Coola, British Columbia	Q_b - 1000	
		Peace River, " "	Q_b - 10000	see below
		Similkameen River, " "	Q_b - 850	
		Lillooet River, " "	Q_b - 850	
Church	1983	Bella Coola, British Columbia	Q_{max} - 1050	0.0025
Fahnestock	1963	White River, USA	Q_{max} - 28	0.008 - 0.18
Ferguson & Werrity	1983	River Feshie, Scotland	Range 8-200	0.009
Grimshaw & Lewin	1980	River Rhiedol, Wales	MAF - 100.4	0.001875
		River Ystwyth, Wales	MAF - 86.2	0.002608
Lewin	1978	River Rhiedol, Wales	MAF - 70.8	0.004
Hein and Walker	1977	Kicking Horse River, B.C.	Range 2.8 - 70 (peak in 1912 - 124)	-
Krigstrom	1962	Hornafjardarfljot, Iceland	Q_{max} - 400 - 500	0.0064 - 0.0105
		Austurfljot, "	Q_{max} - 242	
Leopold and Wolman	1957			mean braided 0.0073
		eg		mean undivided 0.0022
		Horse Creek, Wyoming	Q_{mean} - 1.8	
			Q_b - 11.0	meandering 0.011
		Cotton Wood Creek, Wyoming	Q_b - 23.0	braided 0.004
		New Fork River, Wyoming	Q_b - 37.0	0.0029
Martini	1977	Irvine Creek, Ontario	Q_{max} - 710	0.0047
Smith	1970	South Platte River, Colorado	Range 16-2128	0.0011 - 0.0021
Smith	1974	Kicking Horse River, B.C.	93	braided 0.0034 - 0.0054
				undivided 0.0072

*Q_b - Bankfull discharge Q_{mean} - Mean Discharge Q_{max} - Maximum observed discharge MAF - Mean annual flood discharge

TABLE 7.1. Some case studies of channel bar formation - sediment transport processes in gravel-bed rivers.

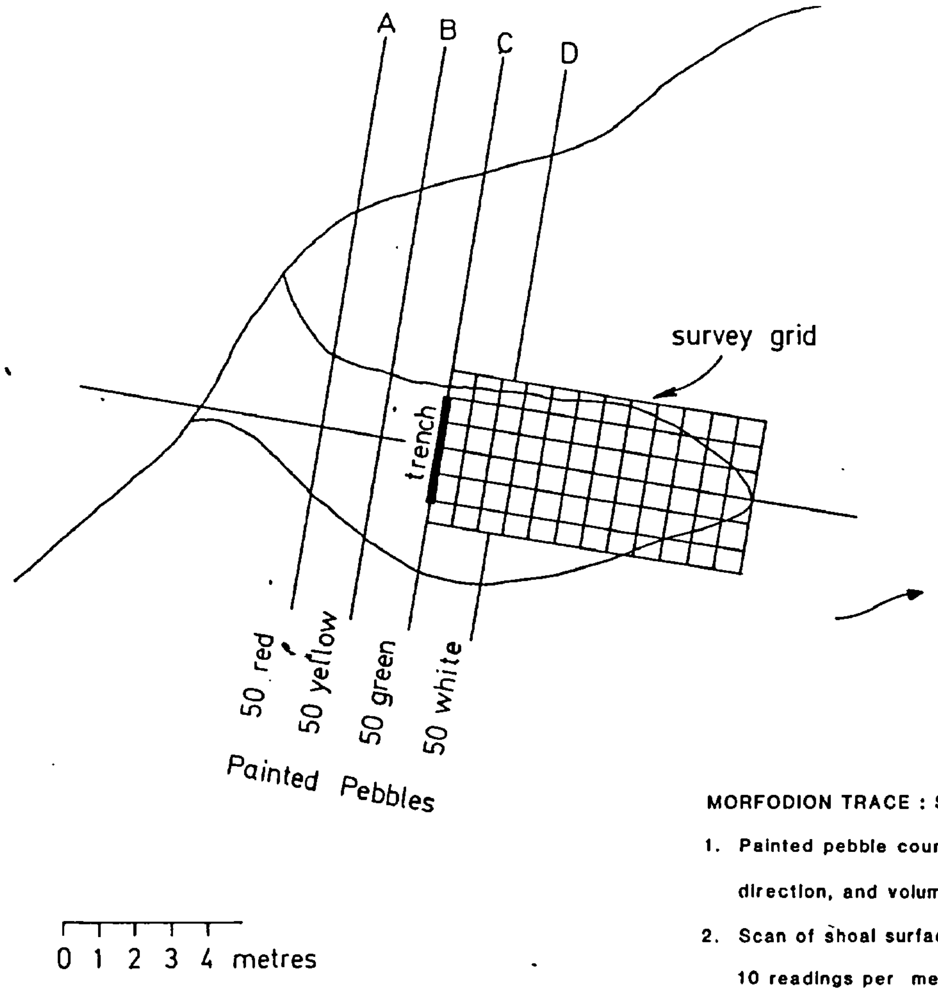
impossible (Smith, 1974; Hein and Walker, 1977; Rust, 1978).

In the gravel bed reaches of the U.K., the natural bed-load flux rates are not so large (Lewin, 1981) and events, whilst still of high magnitude, peak over longer time periods and may induce more uniform bedform development. These reaches, therefore, would appear to offer better scope to observe gravel-bed fluvial processes operating over a much longer time period.

7.2. SITE DESCRIPTION

The Morfodion trace site lies on the River Severn approximately 3Km downstream from Llanidloes, mid-Wales (c.f. Figure 4.1.; National Grid Reference : SN 974857). The river at this point is approximately 30 metres wide with a 500 metre floodplain; it has an actively meandering channel confined within steep to vertical banks of fine floodplain sediment 1 to 2 metres in depth, underlain by fine to coarse gravels which also form the bed of the channel. The contemporary fluvial geomorphology of the area is discussed above (Chapter 4.7.).

The magnetic tracing technique was used to observe the sediment system of a diagonal bar lying at the entrance to a meander bend (Plate 7.1.). The bar is orientated obliquely to the channel with its upstream side anchored to the left concave bank (Figure 7.1.). As with many bar features, the Morfodion shoal is composed of sediments generally finer than those found in the adjacent thalweg, although the surface is composed of a lag concentration of much coarser sediment approximately one grain diameter in thickness, which may be called an armour layer (Figure 7.2.). This grades towards the proximal end of the shoal into material comparable in size with that of the thalweg. At the distal end, the shoal topography steepens sharply



MORFODION TRACE : Survey Procedure.

1. Painted pebble count as an indication of initial movement direction, and volume.
2. Scan of shoal surface X downstream from trench. 10 readings per metre square.
3. At each grid square, pebbleometer classification of accessible trace clasts.
4. Topographic survey across presurveyed transects.
5. Monitoring of downstream shoal sites for longer distance transport.

FIGURE 7.1. MORFODION (RIVER SEVERN) : SKETCH PLAN OF SHOAL SITE INDICATING EMPLACEMENT AREA AND SUEVEY ZONES.

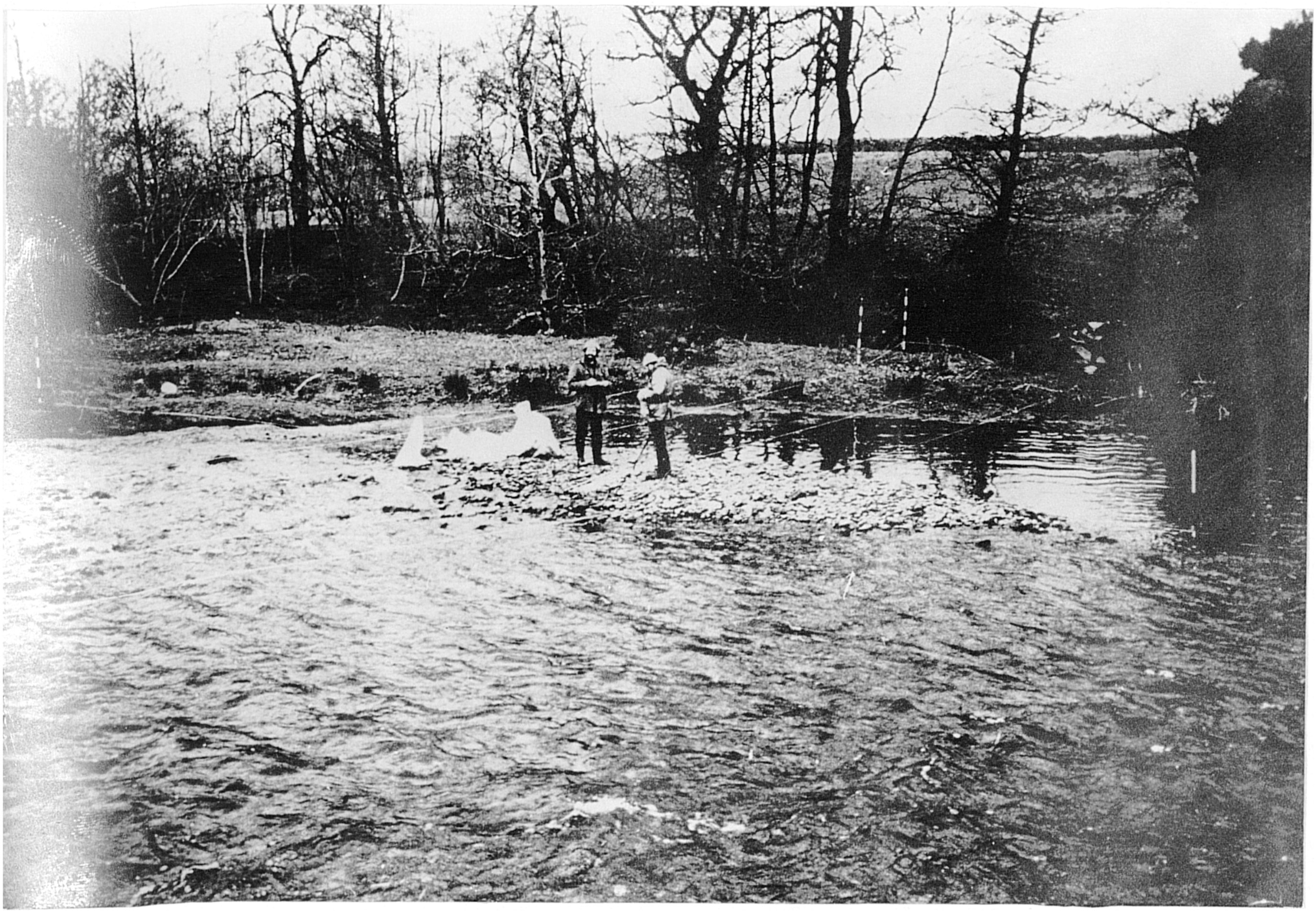


PLATE 7.1. MORFODION (RIVER SEVERN) : STUDY AREA.

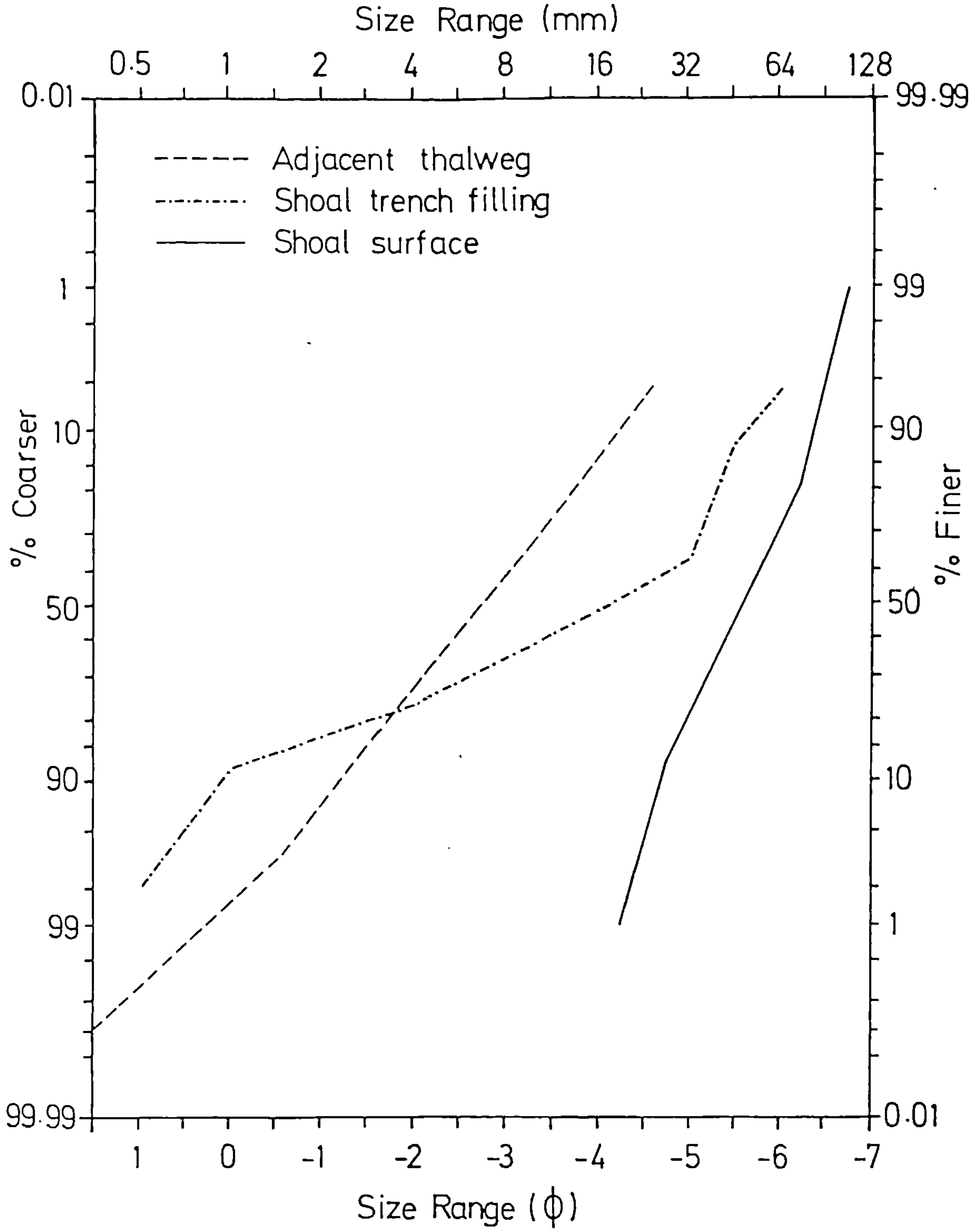


FIGURE 7.2. MORFODION : BED-MATERIAL PARTICLE SIZE DISTRIBUTION CURVES ILLUSTRATING THE CHARACTERISTICS OF THE SHOAL SURFACE, SUB-SURFACE AND ADJACENT THALWEG.

into a deep pool which separates it from the left, concave bank.

7.3. STUDY AIMS

Three main aims were identified:

i) to implement and test the efficacy of the tracing technique in piedmont gravel reaches;

ii) to assess the role of piedmont sediment storage features in sediment supply and transport processes in gravel-bed rivers; and,

iii) to observe the processes and patterns of sediment movement in these reaches.

7.4. EXPERIMENTAL PROCEDURE

Prior to tracing, a large volume of bed material (some 150Kg) was removed from a shoal upstream of the Morfodion site and enhanced in the laboratory in the manner described in Chapter 3. The magnetically enhanced bed-material was emplaced at the shoal site on the 22nd January, 1980.

The tracer sediment was emplaced such that both surface and cross-sectional changes in the shoal could be monitored. This was carried out by excavating a trench across the shoal, oblique to the main flow line and normal to the shoal long axis, with dimensions of 0.3 x 0.3 x 4.0 metres (Figure 7.1; Plate 7.2.). As material was excavated, size proportions were established such that the characteristics of both surface and subsurface material could be replicated with the emplaced tracer. A finer layer of tracer sediment was firmly reconstructed within the trench, followed by a layer of coarser clasts to replicate the armour layer across the shoal surface. The size distribution of the tagged material introduced for the trace is given in Table 7.2..

SIZE RANGE (mm)	WEIGHT (Kg)	PERCENT %
90 +	-	-
45.0 - 89.9	21.7	19
22.4 - 44.9	29.6	26
11.2 - 22.3	27.4	24
5.6 - 11.1	19.4	17
2.8 - 5.5	11.4	10
1.4 - 2.7	3.4	3
1.4	1.2	1
TOTAL	114.1	100

TABLE 7.2. MORFODION (RIVER SEVERN) : PARTICLE SIZE CHARACTERISTICS OF TRACER MATERIAL EMPLACED.

PLATE 7.2. MORFODION : TRENCH EXCAVATION AND TRACER EMPLACEMENT.

In addition to the magnetic tracer sediment emplaced, 200 painted pebbles were placed on the shoal surface, as shown in Figure 7.1., to provide the first indication of sediment movement (to be notified by the farmers owning the land nearby).

Following emplacement, in order to estimate subsequent tracer movement from the trench, surface susceptibility surveys were taken on transects across the shoal (Figure 7.1.) to establish background levels. Subsequent to movement of the tracer, a five tier survey procedure was followed (Figure 7.1.) based on surface susceptibility changes over a grid of one x one metre grid squares together with repeated topographic surveys. Where magnetic tracer clasts were identified on or near the shoal surface, they were analysed for size using a pebbleometer*. However, at all stages of the experiment, the amount of shoal surface disturbance was kept to a minimum so as not to encourage artificial sediment movement. In addition, downstream shoals were monitored for surface susceptibility changes as likely sites for the destination of tracer material.

7.5. RESULTS

Monitoring of tracer movement began with the onset of increased flows in late August/September 1980, following a period of stabilisation under low flows during the post emplacement period, April - August 1980 (Figure 7.3.). No movement in response to increased flows immediately after emplacement (February/March 1980) was observed; all painted tracer pebbles remained in position until the period prior to the survey of 9.9.80. Downstream increases in surface susceptibility, implying tracer movement, were quickly apparent by comparison with the initial surveys (Figure 7.4.). Changes in the surface susceptibility

* Institite of Hydrology.

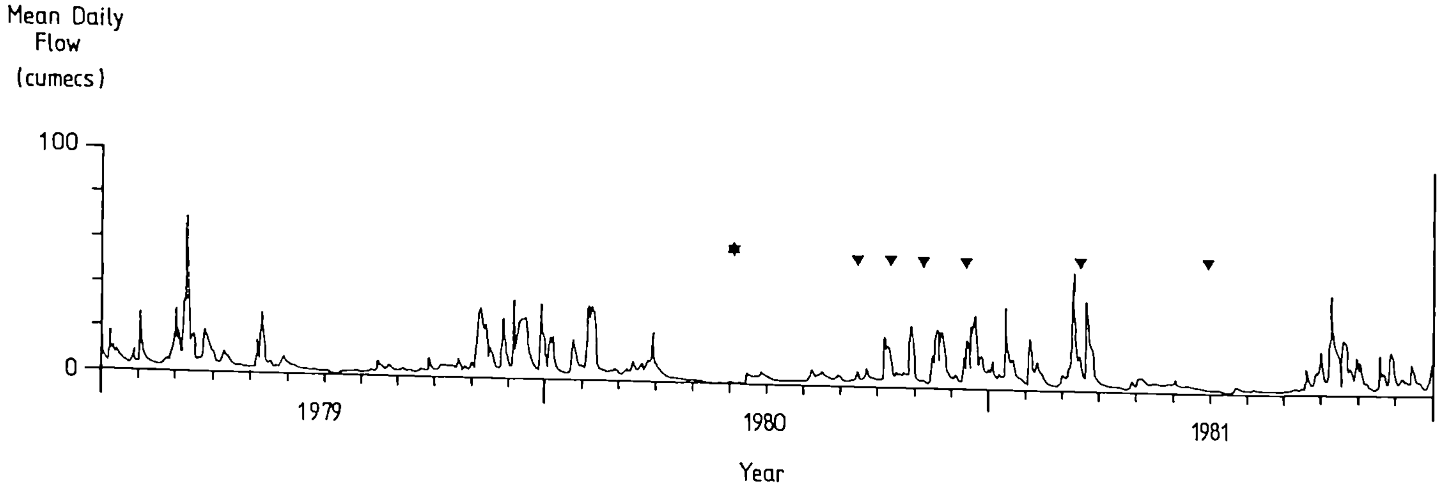


FIGURE 7.3. MORFODION : STUDY PERIOD HYDROGRAPH AS RECORDED AT DOLWEN (Severn Trent Water Authority Gauging Station).

Nb. asterisc denotes emplacement date, triangles denote survey dates.

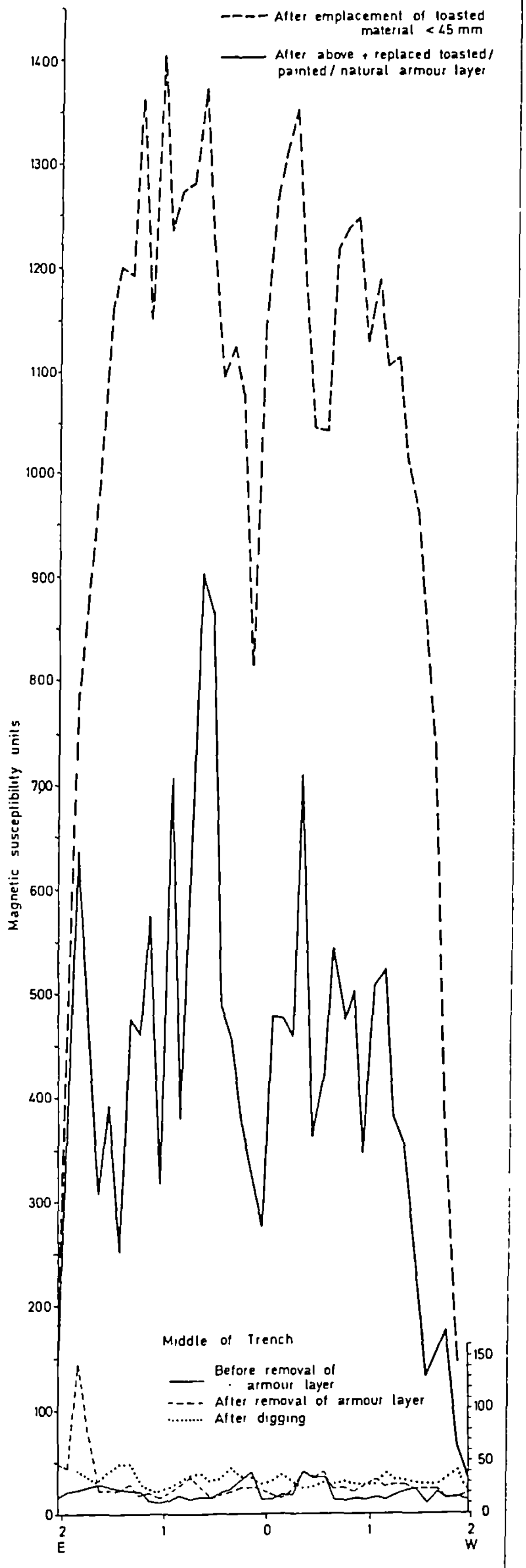
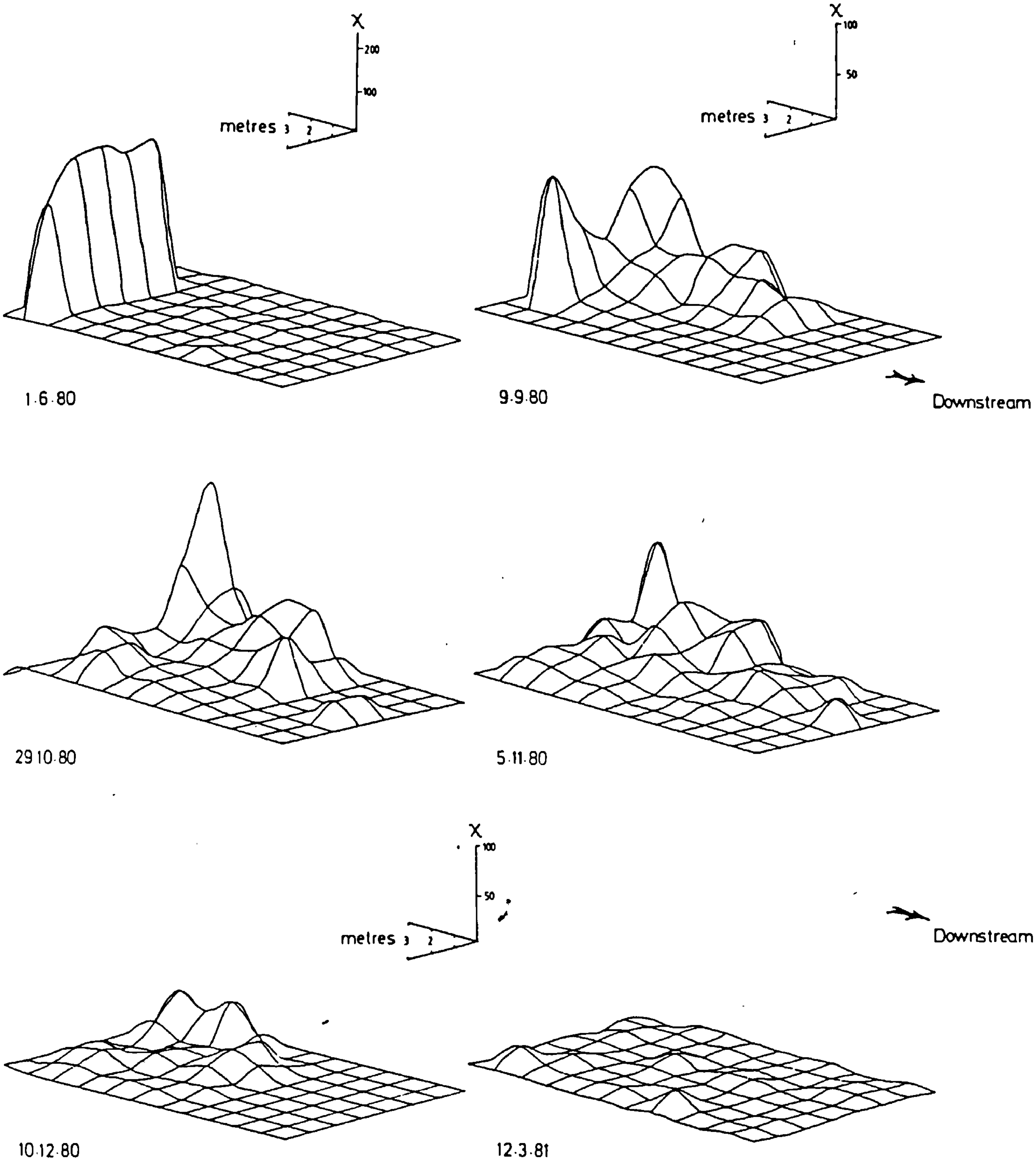


FIGURE 7.4. MORFODION : BED-SURFACE SUSCEPTIBILITY VARIATION ILLUSTRATING THE CHANGES BEFORE AND AFTER EMPLACEMENT OF TRACER MATERIAL

of the shoal and of the tracer particle size found in the immediate surface or subsurface of the shoal (ie. in the lag layer and one to two grain diameters below, to approximately 10cm.) are shown in Figures 7.5. and 7.6.. These are three dimensional isoplots derived from metre square grid data and computer plotted using SOLID (Simpleplot Package, Bradford Univ. Computer Centre).

The grid values of surface susceptibility used in Figure 7.5. have been derived by averaging the search loop readings for each metre square of the survey grid (ten readings per square metre at each survey). A major constraint in determining tracer size data in this fashion was in not disturbing the fabric of the shoal. Where tracer clasts were identified (by their enhanced susceptibility values or pink colouration from treatment), those readily accessible were removed, sized by pebbleometer and carefully replaced in their original position. However, during successive surveys the number of tracer clasts appearing towards the surface declined thus introducing a bias into any sedimentological analysis of the data. As a result, only the statistically reliable D_{84} characteristic (the characteristic for which 84% of the size range is finer) and mean tracer size have been plotted (Figures 7.6. and 7.7.).

The first observation of tracer movement was associated with increased, but still relatively low discharges (previous peak flow : 12 cumecs). From the plots of shoal surface susceptibility, tracer concentrations were seen to remain high over the seeded trench with some loss from the shoal convexity and subsequent increases diagonally downstream at, and towards, the shoal foreset slope. This was associated with a clear pattern of downstream and across bar fining sequences (Figure 7.6.), with most of the tracer sediment located within the interstices of the lag surface layer and accumulating at the head



MORFODION : POST FLOOD SOLID ISOPLOT OF MEAN SURFACE SUSCEPTIBILITY. (X).

FIGURE 7.5. MORFODION : ISOPLOTS OF MEAN SURFACE SUSCEPTIBILITY PER METRE SQUARE (see Figure 7.1. for grid location).

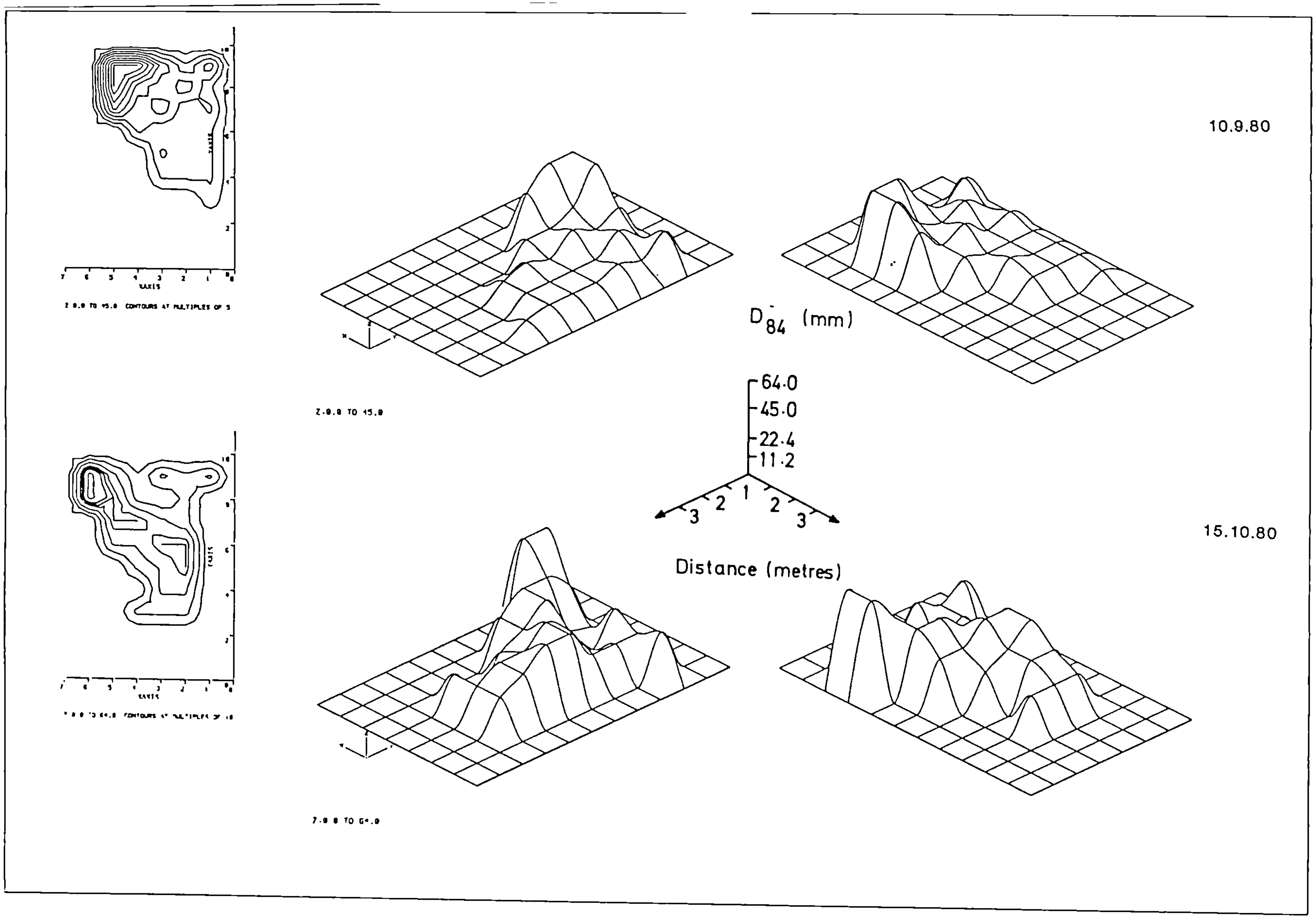


FIGURE 7.6. MORFODION : ISOPLOTS OF D_{84} OF RECOVERED TRACER MATERIAL SIZE FOR EACH SURVEY DATE.

of the foreset slope.

Subsequent tracer movement occurred in response to the first major flood events of the winter season, 1980 (previous peak flows : 73 cumecs (6.10.80) and 34 cumecs (27.10.80)). Topographic surveys (Figure 7.8.) carried out following the recession of flood water levels indicated that the shoal had been excavated across the former convex surface and extended at the foreset. Surface susceptibility surveys (Figure 7.5.) showed noticeably decreased readings over the upstream, true right bank side of the seeded trench area but increased at the left bank towards the foreset slope. Recovered tracer clast size across the shoal surface were noticeably increased, approaching the characteristics of the original emplaced tracer material, which was taken to imply a disturbance across the whole shoal surface and transport of all tracer sizes across the convex slope. At the upstream, right bank area of the seeded trench, the recovered tracer size characteristics remained coarse retaining similar armour sediment size characteristics to the original emplaced tracer material, suggesting that this area was largely undisturbed during the previous storm flows. However, the much reduced surface susceptibility values indicated some burial as a result of upstream supply of sediment. By contrast, tracer clasts recovered over the rest of the seeded trench area were noticeably reduced in size, suggesting that the original emplaced armour layer had been eroded and a new armour coat formed over the much finer sized remaining subsurface tracer material. Downstream fining in size of the tracer material was not so readily apparent in this instance (Figure 7.6.) although fining sequences were observed at the time of survey in association with the vertical development of the foreset. Both fining upward and fining downstream on the foreset were observed in the field, particularly on the avalanche

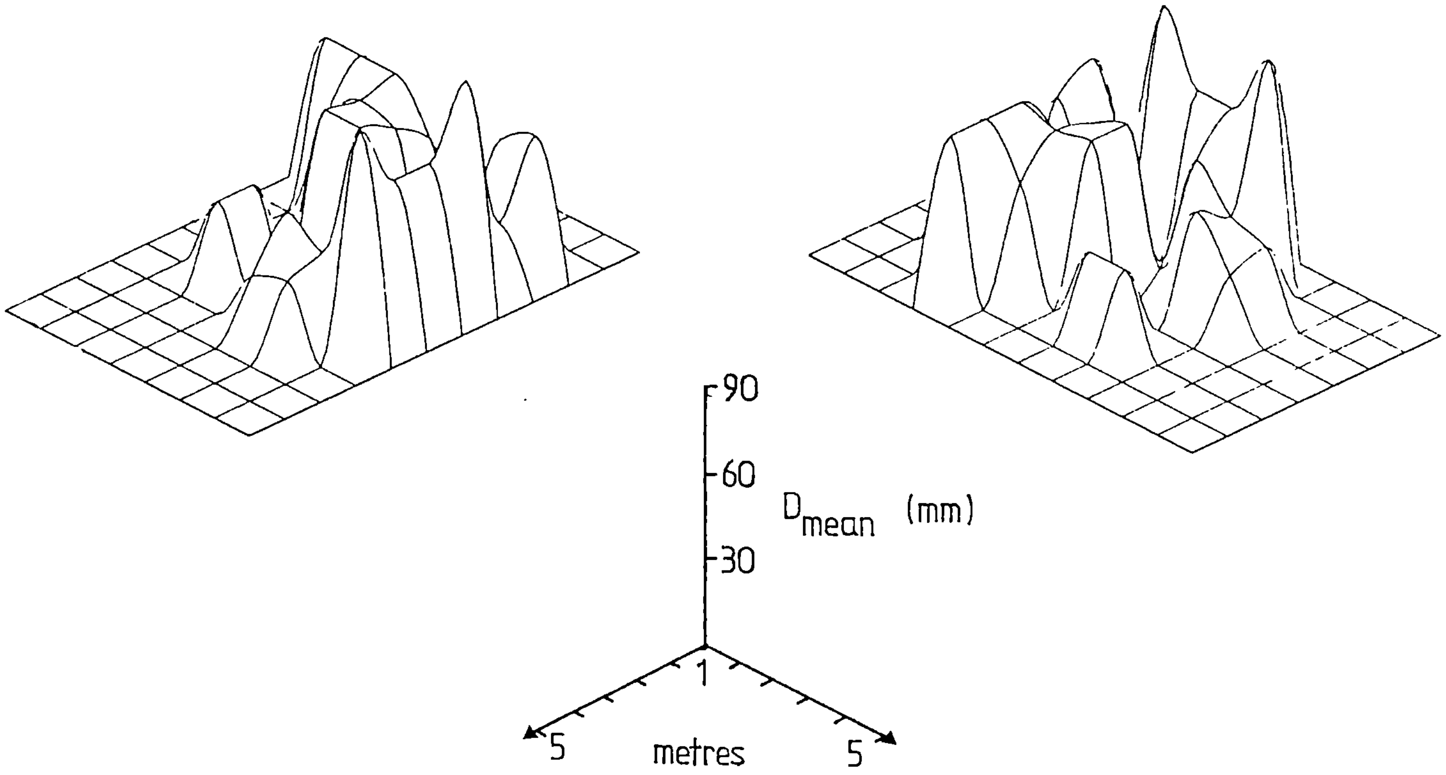


FIGURE 7.7. MORFODION : ISOPLOTS OF MEAN SIZE OF RECOVERED TRACER MATERIAL 10.12.80 (see text).

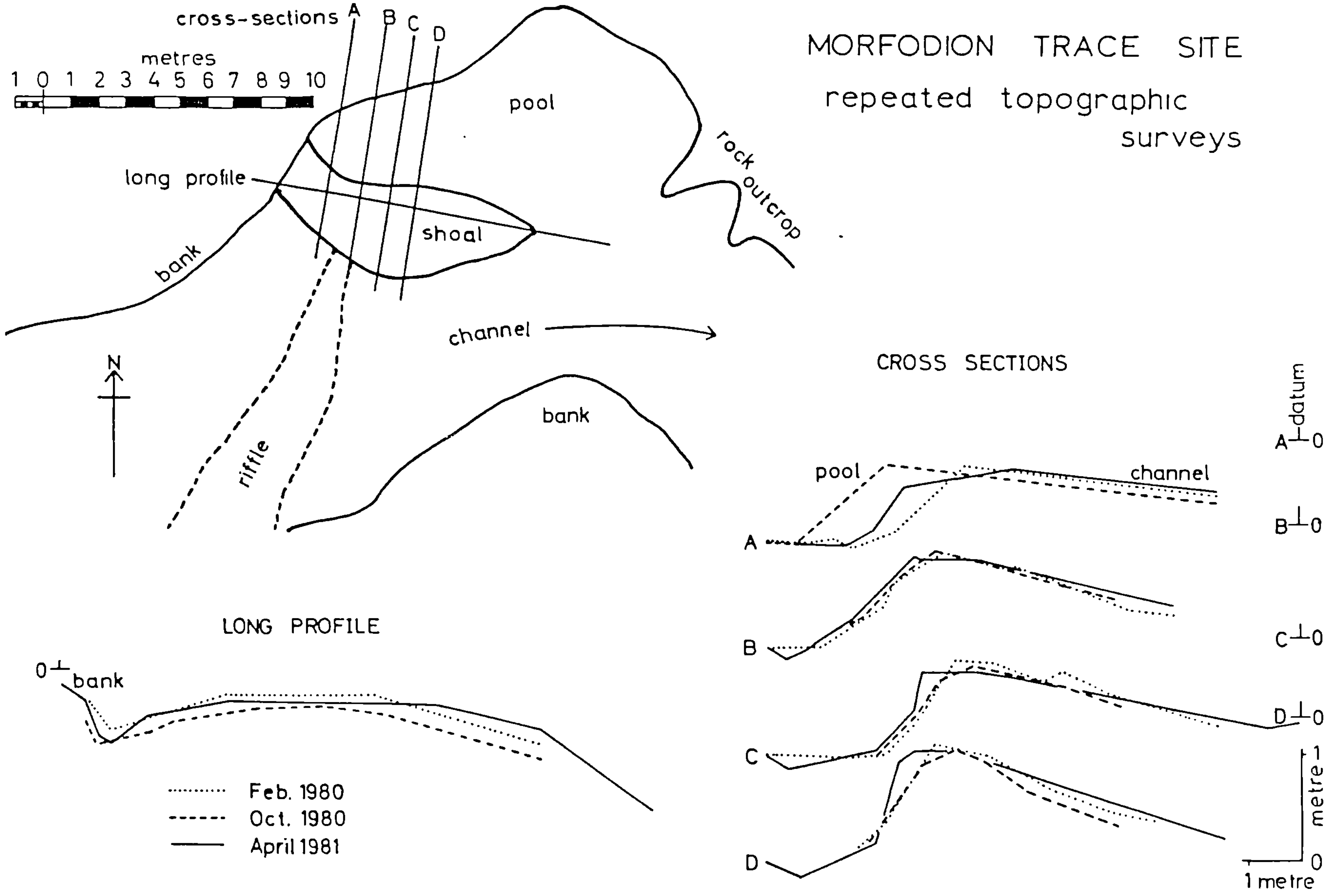


FIGURE 7.8. MORFODION : SHOAL SKETCH PLAN AND REPEATED TOPOGRAPHIC SURVEYS FOR THE STUDY PERIOD.

face of the foreset where larger tracer clasts had collapsed to, and were in the process of being buried at, the foreset base. These patterns are not so obvious in Figure 7.6. largely because of the averaging processes involved in presenting the data (providing data per metre square).

Surveys following this reaffirm a trend of transport to, and accumulation at the shoal foreset of tracer material; the zone of maximum concentration (as indicated by surface susceptibility data) remaining at the upstream, true left bank position close to the original emplacement site (c.f. Figure 7.5. surveys : 5.11.80 to 12.3.81). A gradual reduction of surface susceptibility was observed over the survey grid until by March 1981, the surface susceptibility data had returned to near background levels. Throughout this time, surface susceptibility surveys of downstream shoals likely to receive any transported tracer sediment were carried out in a "Wolman 100" approach (c.f. Chapter 6). For the most part these remained at background levels (see section 7.6.3.) suggesting that the bulk of the emplaced tracer material was buried at the foreset position recorded initially in October 1980. This was supported by topographic surveys (Figure 7.8.) which showed that an overall aggradation of the shoal had occurred in the period to March 1981, with a consequent extension of both the shoal foreset and the distal end.

Given the scale and pattern of tracer movement, as implied by the data described above, an attempt to locate areas of high tracer concentrations within the shoal was made by augering. The extraction of sediment cores was attempted at intervals of 0.3 metres to a depth of 1.0 - 1.2 metres across the original shoal transects using a 0.2 metre diameter powered Stihl Auger (Figure 7.9.). In addition to

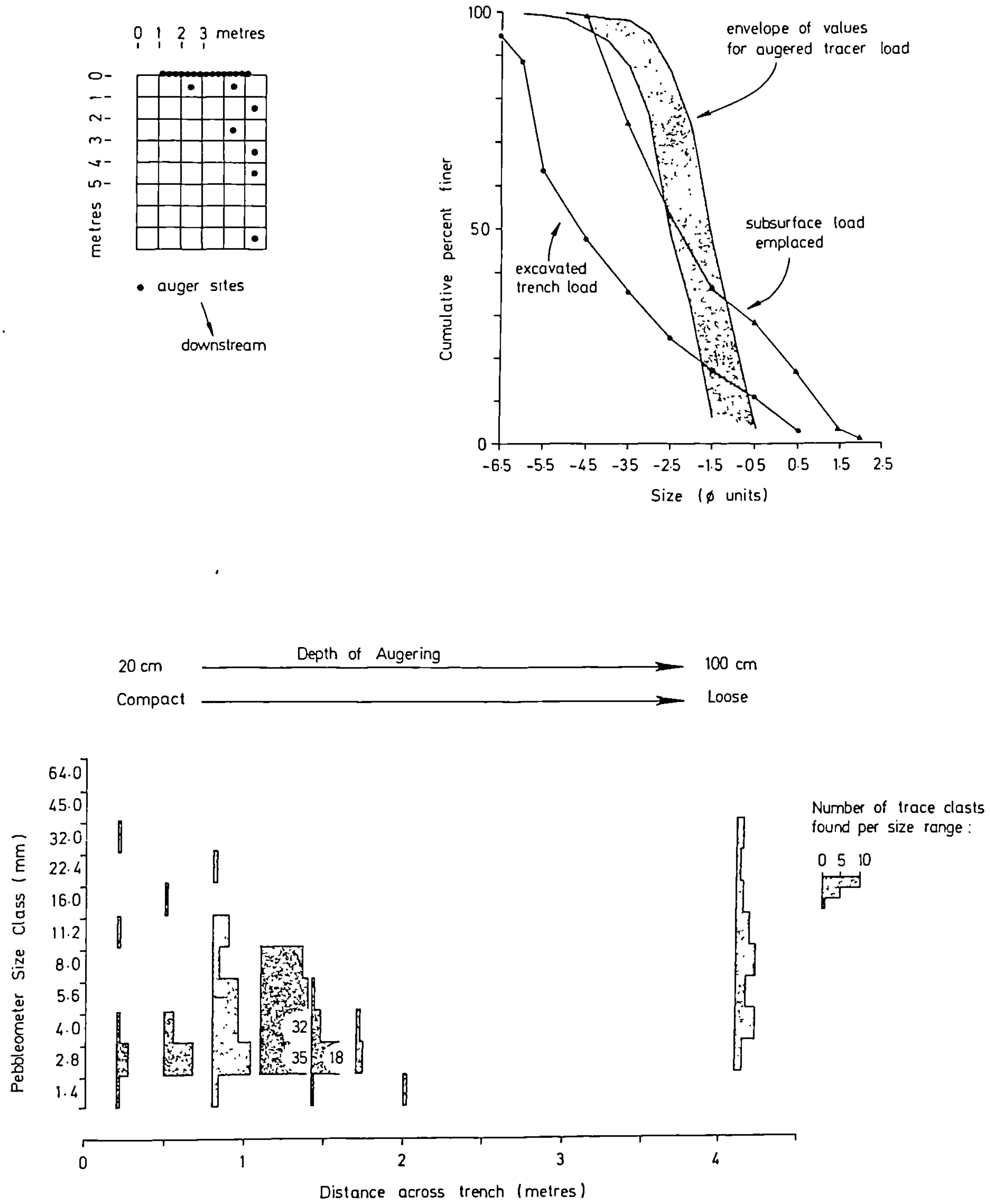


FIGURE 7.9. MORFODION : VARIATIONS IN TRACER MATERIAL RECOVERED BY AUGER SAMPLING OF THE SHOAL SEDIMENTS.

the tracer data derived from these sediment cores, the depth of augering and its apparent ease or difficulty, allowed some qualitative comment as to the induration and fabric of the shoal sediment matrix. These are related to the volume and size data of tracer sediment extracted from the cores in Figure 7.9..

Tracer material was recovered from cores taken within the upstream two metres of the emplacement trench and from only one core at 4.1 metres down the trench length (ie. at the right bank and left bank positions respectively). The largest volume of tracer sediment was recovered from the first five cores extracted at positions between 0 to 1.75 metres along the trench transect. In terms of size, the recovered tracer material exhibited similar characteristics to those of the original emplaced material confirming those observations made above, that for much of the study period, little transport of tracer sediment had been effective here and the originally emplaced matrix had remained relatively undisturbed. No tracer material was found in cores extracted from the convexity of the shoal, but rather at the original shoal foreset position (indicated in Figure 7.8.) where tracer material was found mixed throughout the complete length of the 1 metre core.

A series of additional cores were extracted at positions downstream of the emplacement trench area to test for burial of tracer or any other positions within the shoal. These showed a consistently high proportion of tracer material in those cores extracted from the upstream, left bank area of the shoal. The volume of tracer material increased with depth, most occurring below 0.5 metres in depth and at approximately 0.5 metres inshoal from the current foreset position. This concentration of tracer material consisted predominantly of finer sized clasts (D_{84} = 10mm). Some previous observations made above would suggest that coarser tracer material would be found at some depth below that reached by these cores. However, the local water conditions and the liability of disturbing the shoal further, prevented a more detailed investigation.

7.6. DISCUSSION

7.6.1 PATTERNS OF TRACER MOVEMENT

Observations of downstream fining and fining upward sequences in individual gravel bar sediments have been reported by Smith (1974) Hein and Walker (1977), Martini (1977), and, more recently, by Church and Jones (1982). The trends observed in many bar types are often rudimentary and frequently associated with the development of an 'avalanche front' or foreset. Similar trends were observed at Morfodion although they were influenced in varying degrees by discharge magnitude. The most apparent cross-bar and downstream sorting observed during the study period was associated with a relatively low storm discharge event, early on in the winter season of 1980/81. The position of tracer material on examination of the bar surface suggested that much of the movement had occurred as a result of filter flow or winnowing redistributing some of the finer emplaced material from the upper convexity of the shoal, downstream and within the interstices of larger armour clasts.

Sediment transport was closely associated with the development of the shoal at its foreset and consequently, most of the tracer size variation occurred here. The pattern of tracer distribution prior to burial illustrated a fining upward sequence at the foreset, but given the size ranges involved this was most likely associated with slumping of coarser clasts (Martini, 1977; Ashmore, 1982). At Morfodion, this would seem to be determined by the height of the foreset and its slope into the adjacent pool. Repeated topographic surveys (Figure 7.8.) indicated that foreset slopes of 80° and approximately 1 metre in depth were common for much of the experimental period.

Following the initial transport of tracer by winnowing, the across bar sequence of subsequent transporting flows indicated that the whole of the tracer size range was in motion across the bar. Deposition at the foreset illustrated a sequence of coarse sediments locally to the trench, and almost immediately at the foreset base, with subsurface sized material transported diagonally across the shoal to the foreset with limited downstream sorting. From coring and topographic observations, it would seem that subsequent supply of sediment from upstream is deposited at, and causes collapse of the foreset producing a matrix of mixed subsurface sized material over coarse armour clasts ie. turning the original matrix over on itself to be subsequently buried.

Similar processes to those described above were observed in a sand bedded meander by Dietrich et. al.(1979); fining sequences (upward and downstream) at the avalanche face were associated with the patterns of primary and secondary flow circulation related to particle fall velocity (Stokes Law).

7.6.2 PATTERNS OF BAR DEVELOPMENT AT MORFODION

The pattern of tracer movement, and therefore bar development, would appear to be strongly determined by sediment stability aswell as hydraulic characteristics. The tendancy for bars to appear for some time 'static' has been observed in many cases, particularly so with an attached diagonal bar, where its oblique orientation to flow "imparts a high measure of form resistance.." (Church and Jones 1982). At Morfodion, observations would suggest that sediment stability decreases in the directions of bar development, such that those areas most stable are its points of attachment to the concave bank, and particularly to the upstream riffle. Attachment to the concave bank was clearly evident here. The upstream stability of the bar was inferred from two sources. The most apparent evidence for this was the remnant wedge of tracer in the upstream end of the original trench position. This combined with the qualitative assessment of stability from the augering observations illustrated the static nature atleast of this part of the bar.

Downstream, sediment stability decreased markedly, and was reflected both by the ease of penetration of the Stihl auger and also the pattern of sediment movement (redistribution to the foreset) here. Similarly, slumping at the foreset could be easily initiated whether by human or livestock disturbance.

The pattern of bar development inferred from the movement of tracer as described above suggests a sequence of incremental stages of a prograding foreset. With continued supply of sediment, this would produce a continued extension of the bar downstream. However, calculations from repeated topographic surveys showed that whilst 17 m^3 of material had moved at the site during the period to June 1981, the net movement was only 3.3 m^3 (gain) and would suggest only a limited supply of sediment. Indeed, from the topographic surveys (Figure 7.8.) it is possible to interpret a distinction between those flows that are sediment supplying to the bar, and those which are eroding and restructuring it. The initial period of high flows (October, 1980) were involved mainly in excavating material from the shoal surface and providing redeposition locally downstream. A later sequence of events in March, 1981 however, were observed as supplying material to the bar, aggrading its upstream ramp and prograding the foreset, thereby burying any tracer material previously deposited here. Whilst the range of flows in both cases was similar

A.

DATE OF SURVEY	9.9.80	29.10.80	5.11.80	10.12.80		12.3.81
Date of previous flood	7.8.80	6.10.80	27.10.80	10.12.80	2.2.81	11.3.81
Daily Discharge (cumecs) -						
Minimum	5.96	4.25	15.02	3.41	3.02	36.22
Mean	7.92	22.43	24.86	11.36	22.80	53.39
Maximum	12.16	73.05	34.19	45.71	93.16	75.04

B.

	OCTOBER 1980				MARCH 1981		
Date	Mean	Max	Min	Date	Mean	Max	Min
1	4.12	4.71	3.61	5	5.90	6.23	5.44
2	3.82	3.82	3.82	6	5.84	7.08	5.44
3	3.63	3.82	3.61	7	10.26	15.46	8.94
4	4.01	4.74	3.82	8	10.70	14.17	8.61
5	4.01	4.71	3.82	9	18.94	26.12	8.61
6	22.43	73.05	4.25	10	33.69	46.49	23.32
7	16.93	18.69	15.46	11	53.39	75.04	35.22
8	18.88	20.17	18.21	12	26.31	34.19	20.17
9	19.19	20.68	17.73	13	22.77	25.55	19.17
10	19.45	21.71	17.73	14	17.65	20.68	15.46

TABLE 7.3. MORFODION : FLOOD DISCHARGE CHARACTERISTICS; A) FOR ALL SURVEYS; and, B) MAIN STORM PERIODS OCTOBER 1980 and MARCH 1981.

(Figure 7.3.; Table 7.3.; maximum flows of approximately 75 cumecs), flows were much higher and persisted for a much longer time period during March, 1981 than for those observed for October, 1980. A simplified model of the development of the Morfodion shoal is presented in Figure 7.10..

7.6.3 CRITIQUE OF THE TRACER METHOD AT THIS SCALE

At the same time as emplacing the magnetic tracer, painted pebbles were positioned along the survey transects shown in Figure 7.1.. These were used to complement the new tracer methodology in the first instance by allowing local observers to assess the initial period of movement, and secondly, as an instant check as to the extent and position of the transported tracer.

Subsequent to the first trace, these were monitored with the same grid pattern described above, or where appropriate, their distance downstream from the survey sites. Recovery rates for each colour range at the onset of the winter season are shown in Table 7.4.. The lack of recovery from the foregoing discussion, would suggest that many of these painted pebbles had been buried (and indeed, some were relocated during augering operations); some had travelled upstream (see Thorne and Lewin, 1977). The magnetic tracer has obvious attractions.

However, the magnetic tracer was not without problems of its own in this particular case. When surface susceptibility readings across the shoal began to decrease, surveys were extended to likely downstream sites. One of these, the nearest and most likely receptor site was a point bar slightly downstream on the same meander bend as the study shoal. It consisted predominantly of finer material

TABLE 7.4. MORFODION : RECOVERY RATES OF PAINTED PEBBLES FOLLOWING THE FIRST FLOOD EVENT

DATE	COLOUR	NUMBER EMPLACED	NUMBER RECOVERED	%
15.10.80	White	50	8	16
	Green	50	2	4
	Red	50	-	-
	Yellow	50	7	14

TABLE 7.5. MORFODION : SUMMARY OF THE MAGNETIC CHARACTERISTICS OF NON-TRACER MATERIAL FOUND DOWNSTREAM OF THE STUDY SITE.

	SIRM	S	χ
TRACER MATERIAL	178.0 ± 8.9	0.89 ± 0.045	11.6 ± 0.58
NON-TRACER MATERIAL	111.2 ± 2.9	0.81 ± 1.88	16.1 ± 17.3

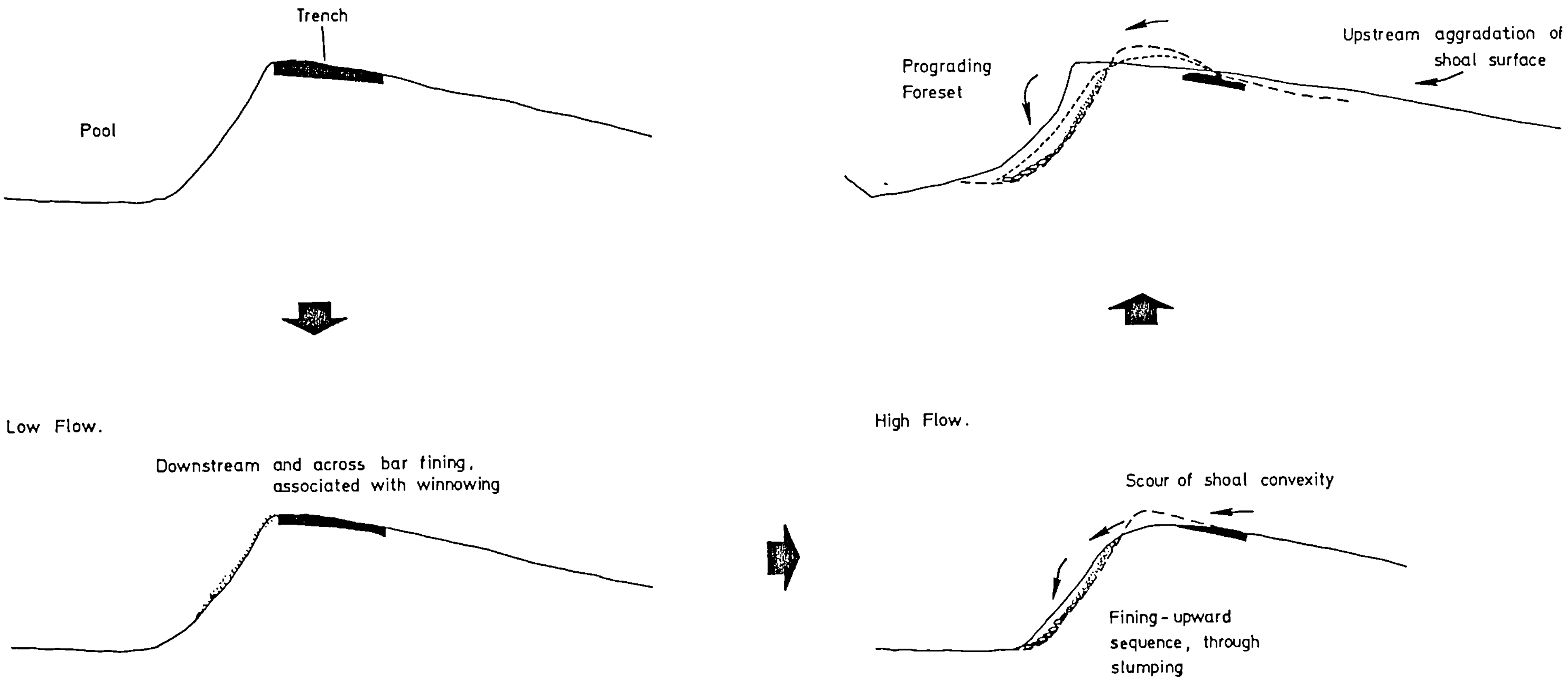


FIGURE 7.10. MORFODION : HYPOTHETICAL MODEL OF SHOAL DEVELOPMENT BASED UPON OBSERVATIONS REPORTED IN TEXT.

(< 32mm) than the study site, and was scattered with material resembling in colour the tracer material used. Surface susceptibility readings were much higher as compared to the study shoal, averaging 39 rather than the background values of 16 - 25, recorded on the trace shoal. With detailed inspection of the material, it could be further recognised as building rubble, shaped very convincingly by downstream transport. Samples were taken for analysis in the laboratory to see how closely their magnetic characteristics resembled those of the tracer. These are summarised in Table 7.5.. The range of values for 'S' and susceptibility is quite large, though their average values are quite comparable to those of the tracer. The discriminating characteristics here are the much lower values for SIRM. These remain fairly consistent.

The field surveys, however, are based on magnetic susceptibility alone, so the range of values recorded here constitutes an immediate problem for the technique. Whilst on an emergent shoal surface this material can be distinguished fairly rapidly using texture and mineralogy, under submerged or buried conditions this is not so easily done and surveys would have to be followed up by detailed sampling and analysis. This extends both the field and laboratory time involved with the technique, and whilst samples may be rapidly processed using the field-portable fluxgate magnetometer (Chapter 3.), this detracts from the essentially simple application of the magnetic tracer.

Despite these reservations, the magnetic tracing technique has been used successfully to describe the patterns of gravel bar evolution. When interpreted in conjunction with topographic surveys, patterns of tracer movement could be used to distinguish between the effects of storm flows in moving sediment already stored at the bar as against those supplying additional sediment in aggrading the bar. In many instances, such information has been lost due to a combination of tracer burial and the dynamic nature of many observed reaches (see for example, Butler, 1977).

7.7. SUMMARY CONCLUSIONS

As with observations within many other environments (see introduction), the major phases of bar development at Morfodion were associated with sudden, episodic events. The progression observed above though is at a completely different scale both temporally and spatially, and for the most part the bar represents a much more stable bedform than those previously described. The pattern of development is uniform with flood events lasting the period of days rather than hours (Smith 1974) despite their sporadic nature. This was the case atleast for the range of discharges monitored; up to 75 cumecs).

However, tracing within this reach of the River Severn has been confined to the observations of transport at storage features as opposed to transport to storage features which may involve considerable spatial differences in transport rates within a single reach. Further downstream on the River Severn, the observations of Thorne and Lewin (1979) support this. However, these inferences are based on the results of a 15 - 47% recovery rate of painted pebble tracer. Whilst observing that trace material was transported onto bars and into temporary storage prior to permanent deposition, their results are interpreted as a considerable loss downstream. Results from the Morfodion shoal study suggest that much of this may be accounted for by the patterns of bar development and the processes of tracer burial during floods. This raises questions as to how tracer experiments should be carried out. For example, if tracer material is emplaced across channel through a bar and riffle sequence, then it seems reasonable to expect more rapid movement off the riffle. The spatial distribution downstream then would raise serious doubts as to estimates of sediment transport rates (see Meade, Emmett and Myrick, 1981; Andrews, 1979a).

Given the static nature of the Morfodion deposit, with both limited growth and transport, then the impact of any increased flows will be borne predominantly by the channel banks, erosion and transport will be effective more particularly on the finer sediment size ranges and increased suspended loads (c.f. Chapter 8.).

CHAPTER 8

MAGNETIC TRACING: SUSPENDED SEDIMENT SOURCE ANALYSES FOR THE UPPER SEVERN CATCHMENT

8.1.1 INTRODUCTION

This chapter is concerned with the identification of source contributing areas for suspended sediment load in the area of the Severn catchment outlined in Chapter 4. Suspended load yield has been analysed for two storm periods during the winter 1979, and for a number of storm events from October 1980 to March 1981. Particular attention is paid during the latter period to suspended load passing at the Abermule gauging station. Certainly, the implications of bed-load flux rates suggest that piedmont suspended load may be of significance here (see Chapter 7). An analysis of source areas is made together with an attempt at matching characteristics from the suspended load. The implications of this is discussed with reference to the dominant local hydraulic engineering and land-use types.

8.1.2 UNDERSTANDING AND PREDICTING SUSPENDED SEDIMENT YIELD:

Attempts to analyse sediment source contributions and routing systems have arisen for two main reasons. Firstly, the inability of conventional methods to identify variations in the relative importance of different source contributions to given suspended sediment concentrations; and, secondly, as a result of an increasing awareness of the ways in which sediment load quality may vary in response to management practises over different potential source areas in a catchment.

For sometime geomorphologists have persisted in using rating equations of load-discharge relationships, the variations involved over time and space may be considerable. For British rivers, these relationships are based on limited measurement over short time periods; the only comprehensive monitoring of suspended sediment and solute yields are those discussed by Webb and Walling (1982). These results, point towards the inadequacies of conventional approaches; for the River Creedy, 50% of suspended sediment is transported in 0.8% of time, or approximately three days per year and by only 9% of the discharge. The implications discussed by Webb and Walling, and indicated by others, suggest a variety of factors in addition to transporting efficiency must be appreciated. Threshold, complex response and recovery are terms loosely associated with the phrase sediment availability which, if it is to be modelled accurately, must involve an appreciation of sediment sources. Much of the increased suspended load in British rivers is attributed to recent changes in environment eg. urbanisation (Gregory and Walling , 1973).

Further understanding of these processes also has considerable practical application aswell as being an aid to understanding dendudation rates and landform development. Environmental problems arising through erosion, storage, transport and deposition are well documented. Identification of erosion source area holds considerable potential for tracing soil particle associated pollutants (the term pollutant may be used loosely here to include herbicides and pesticides together with toxic waste and radio-nuclide spillage transport). For example the Illinois Environmental Protection Agency (1978) quote 90% of organic Nitrogen and Phosohorous originating from upland practises is delivered to streams in particle -associated form. This together with high solute loadings of nitrate-nitrogen and phosphorous may also contribute to cultural eutrophication downstream (Troake et. al., 1976; Van Vlymen, 1980), in addition to possible health risks (Burfield, 1977; Gass, 1978).

Similarly, in the event of toxic spillage, an understanding of sediment routing processes would be required to predict longterm effects or to be able to implement controls.

Equally important, are the ecological implications of suspended sediment delivery. Sokolovaet al. (1978) have used source and sediment characteristics to determine longterm agricultural potential in floodplain depositional areas. There may also be detrimental effects; for example, the silting of navigable reaches or of reservoirs, or the implications of both high turbidity and reduced water depths on freshwater ecology (Wilkin and Hegel, 1982).

Two broad directions of approach have been used to identify and quantify sediment source areas. Continuous field surveys of individual eroding sites, such as that by Imeson (1970, 1978) provides one approach. These inevitably depend on detailed and extensive fieldwork. Faster estimates have been aided by aerial photography and large scale maps, supported by limited fieldwork on individual representative source types (Mosely 1980). Alternatively, detailed analysis from soil plot experiments may be used to provide soil loss equations which can be used to predict yield and delivery ratios according to the morphometric characteristics of the drainage basin (Roehl,1962; Glymph,1957; ASCE, 1970). However, despite considerable effort in these areas such equations remain unreliable for many terrains and soil types (Pickup, 1981). As Lewin (1982) points out:

> "Although individual studies have identified both sediment sources and the process domains which exploit them, it remains difficult to specify which are dominant in terms of gross down-channel sediment yields"

Even more important, the pattern should be appreciated as both scale and event dependent.

As a complementary approach to field survey and mapping procedure, some attempt to identify the natural diagnostic characteristics of the yielded load, or regional soil and regolith types, may be used

MINERALOGY AND GEOCHEMISTRY			
Klages & Hsieh	1975	Gallatin River, Montana USA.	XRD of clay and silt sized sediments relating yielded sediment and alluvium deposits.
Wall & Wilding	1976	Maumee River Basin, USA.	XRD of clay sized sediments, gasometric and calorimetric analyses of silica, aluminium and carbonates.
Lewin & Wolfenden	1978	mid-Wales, Great Britain.	Atomic adsorption spectrophotometry of toxic waste derived from nineteenth century spoil heaps and transported in sediment associated form.
Lewin et. al.	1978		
Sokolova et. al.	1978	Soviet Central Asia, USSR.	XRD and clay mineralogy from sediment sources characterising new deposits for long-term agricultural planning.
Wood	1978	River Rother, Sussex, Great Britain.	XRD of clay and silt sized sediments derived from soils, alluvium, bedrock and suspended load.
RADIOISOTOPE TRACING			
Wise	1980	Review of applications.	Routing models of Caesium^{137}, estimation of sedimentation rates.
Loughran et. al.	1981	Maluna Creek Basin, Australia.	Estimates of sediment erosion, storage and transport using Caesium^{137} concentrations.
Fullen	1982	North York Moors, Great Britain.	Fe^{59} tracing studies of soil erosion.
Wilkin & Hebel	1982	Middlefork River, Illinois, USA.	Estimates of erosion, redeposition and delivery of sediment using Caesium^{137} concentrations.
PHYSICAL CHARACTERISTICS			
Grimshaw & Lewin	1980	River Ystwyth, Wales, Great Britain.	Characterisation of sediment filtrates and source types using Munsell soil colour charts and related to sediment and stream discharge variations.
MAGNETIC CHARACTERISATION			
Oldfield et. al.	1979	Jackmoor Brook, Devon, Great Britain.	Use of natural magnetic characteristics to relate sediment sources to filtered suspended loads.
Walling et. al.	1979	Jackmoor Brook, Devon, Great Britain.	Detailed magnetic analyses of source-sediment type yielded through several storm events.

TABLE 8.1. APPROACHES TO SOURCE IDENTIFICATION OF SUSPENDED SEDIMENTS.

to fingerprint likely source areas with and between flood events. Table 8.1. presents a summary of some recent technique used to this end. These fall into four main categories:

i) Mineralogy and Geochemistry (eg. heavy minerals, N+P concentrations)

ii) Radionuclide Tracing : Cs^{137} or nuclear spillage

iii) Sediment Physical Characteristics - colour, size etc.

iv) Mineral magnetic characteristics of sediments and source areas.

8.2. ANALYSIS OF THE MAGNETIC PROPERTIES OF SUSPENDED LOAD

Processes operative within soils such as weathering, gleying, etc., give rise to magnetic properties readily distinguishable between surface and subsurface samples within a monolith and also between soil and regolith types (Chapter 3). The attempt to use these characteristics to identify sediment source from surface, subsurface and channel origin has been discussed for the Jackmoor Brook catchment, Devon U.K. by Walling et al (1979) and Oldfield et al. (1979).

Using measures of χ, SIRM, $(Bo)_{CR}$) and 'S', these studies illustrated the ease of source differentiation for a range of floods for the winter of 1977-78. The relationship between sediment source and stage, and the dominance of surface soil sources at and immediately after peak flows enabled quite simple modelling of suspended load yield and transport determined by stage and sediment availability.

Whilst all the characteristics are interrelated in diagnosis, these studies illustrated the potential of small changes in backfield parameters $(Bo)_{CR}$ in this case) in rapidly determining source changes.

The main aims of this study were:-

i) to identify the magnetic characteristics of the sediment source contributing areas described in Chapter 4;

ii) to identify the magnetic characteristics of the suspended sediment loads in the catchments against those of the source areas outlined in i); and,

iii) to identify the inter- and intra- storm variability in source area contributions to the suspended sediment load of the catchments.

Due to the fact that only two studies of this type have been carried out before, the guidelines for sampling were limited. This analysis was carried out in two parts. Firstly, a pilot study of the magnetic characteristics of source area sediments related to suspended sediments collected from two winter storms in 1979 was undertaken. This provided the basis for a much more detailed sampling programme of both sources and suspended load sediments for the following winter flood season. In both cases, the collection of suspended load was made by IOH operations. For the pilot study, this was carried out on a flood by flood basis with IOH staff collecting samples for the whole catchment with USDH bomb samplers. For the main period of analysis, an eight hour time-based vacuum sampler was installed at the Abermule gauging station, again operated by IOH. These samples were interspersed with samples taken by cableway and USDH samplers. Upstream, samples were collected at the sites shown on Figure 4.1. using a wading rod and USDH bottle sampler. Whilst the main study site was at Abermule, this provided a suitable spatial context with which it was possible to identify a range of erosion source regimes.

8.3. SUSPENDED SEDIMENT SOURCE IDENTIFICATION : A PILOT STUDY

It is apparent that the main sources of sediment inferred from previous studies in this area and elsewhere in Britain are river channel bed and bank areas (c.f. Chapter 4). The pilot study was used to test this hypothesis, and to test the application of the technique to conventional sampling methods for the Upper Severn Basin.

Source samples were collected in the following manner:

i) as distinguished by soil-vegetation and morphometry changes within the main confines of the river valleys;

ii) readily apparent erosion sources - bank scars and cliff cut faces; and,

iii) channel bed material and bank sediments.

This material was then air dried and packed into clean or distilled water washed 10cc sample holders. These, together with moistened filter samples of suspended load, were analysed for χ, SIRM, $(Bo)_{CR}$ and 'S' variations using the instruments and techniques described in Chapter 3.

8.3.1 PILOT STUDY : RESULTS

A total of 45 filtrate samples were collected for isolated storm events November - December 1979. Analysis of these samples was restricted as a result of the low weight of sediment retained per sample (ranging between 0.0075g to 0.300g and most between 0.05 - 0.15g). The low sample weights precluded any measures of χ, and for weights below approximately 0.025g there was a clear reciprocal relationship between sample weight and SIRM (Figure 8.1.). Up to this weight, there is a sharp exponential decline in the relationship of SIRM - Filtrate weight suggesting that these lower weights may be contributing to artificially high SIRM components.

Considering the discrepancies in SIRM and the lack of χ measures for this limited data set, further analyses were restricted to the normalised hysteresis components $(Bo)_{CR}$ and 'S' for both sediments and

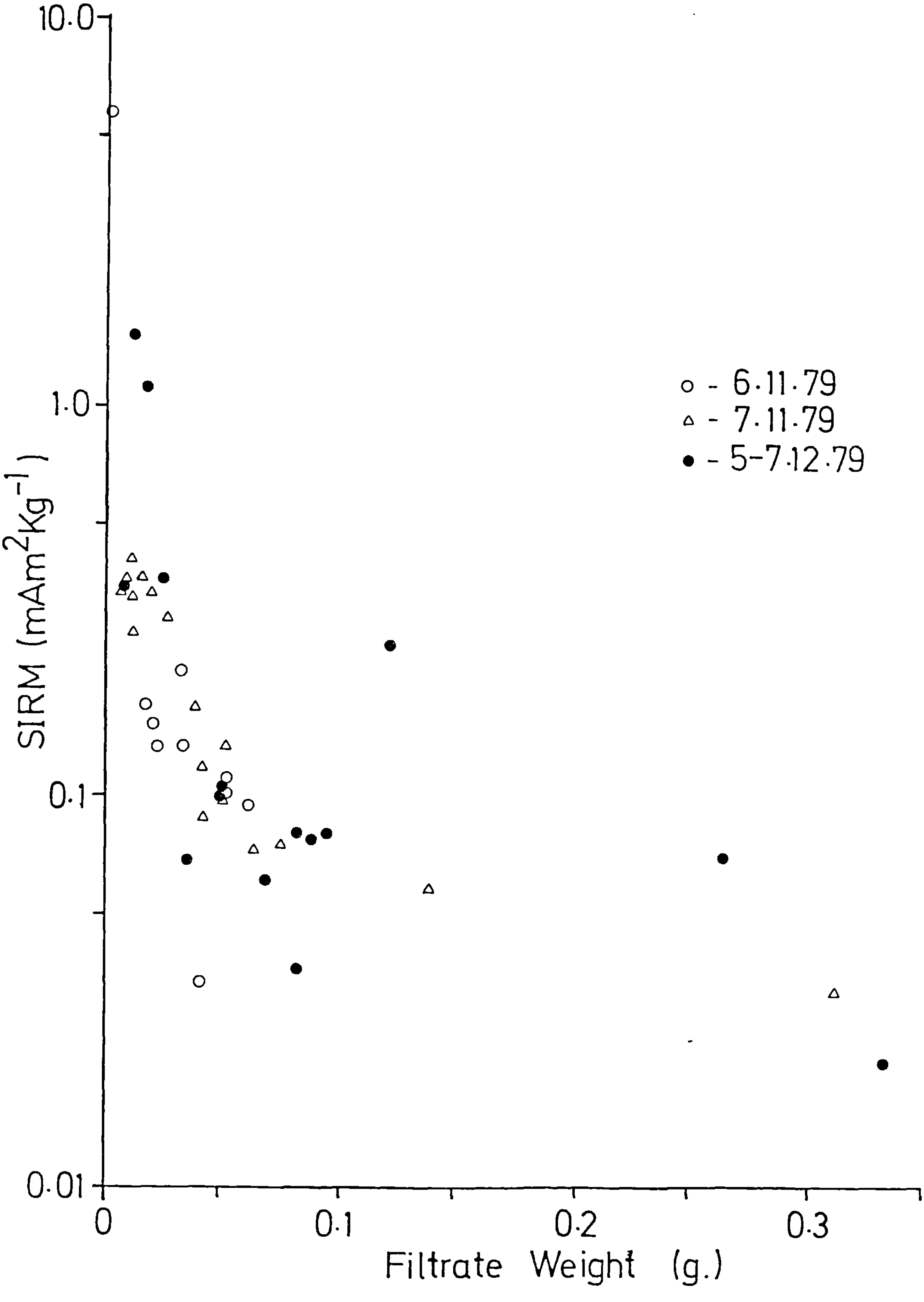


FIGURE 8.1. RELATIONSHIP OF WEIGHT OF SUSPENDED SEDIMENT RETAINED AGAINST SIRM FOR SAMPLES DURING THE PERIOD NOVEMBER - DECEMBER, 1979.

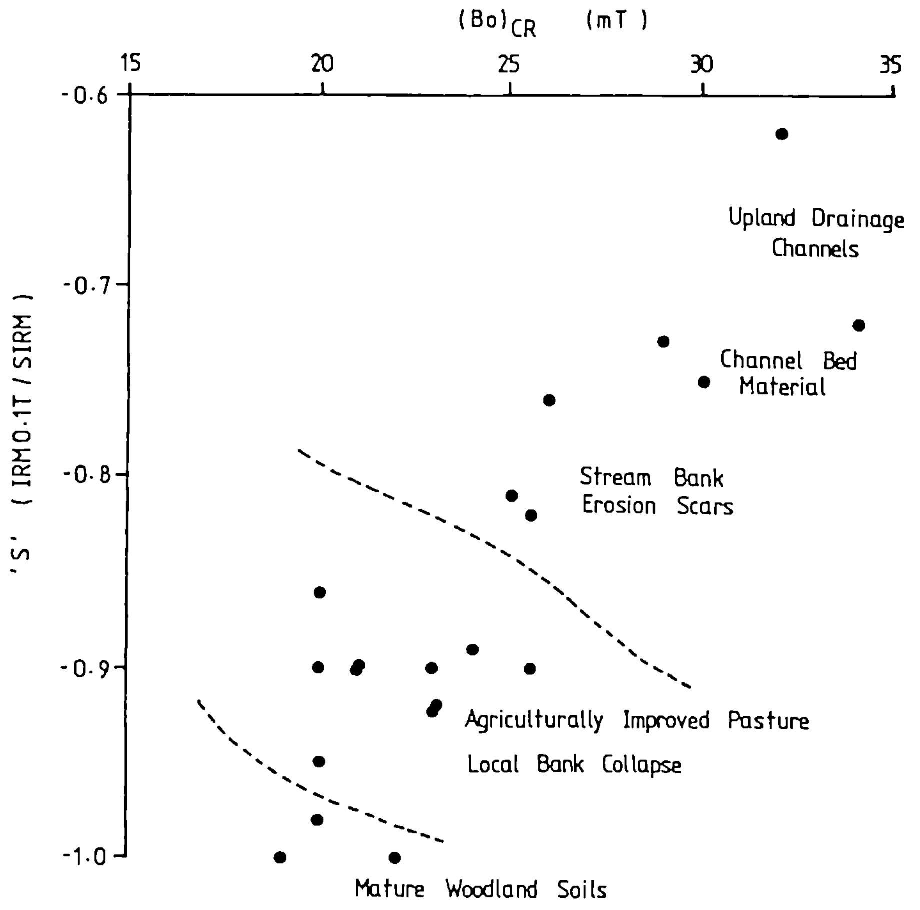


FIGURE 8.2. PILOT STUDY : RELATIONSHIP OF $(Bo)_{CR}$ TO 'S' FOR SAMPLES OF LIKELY SOURCE MATERIAL.

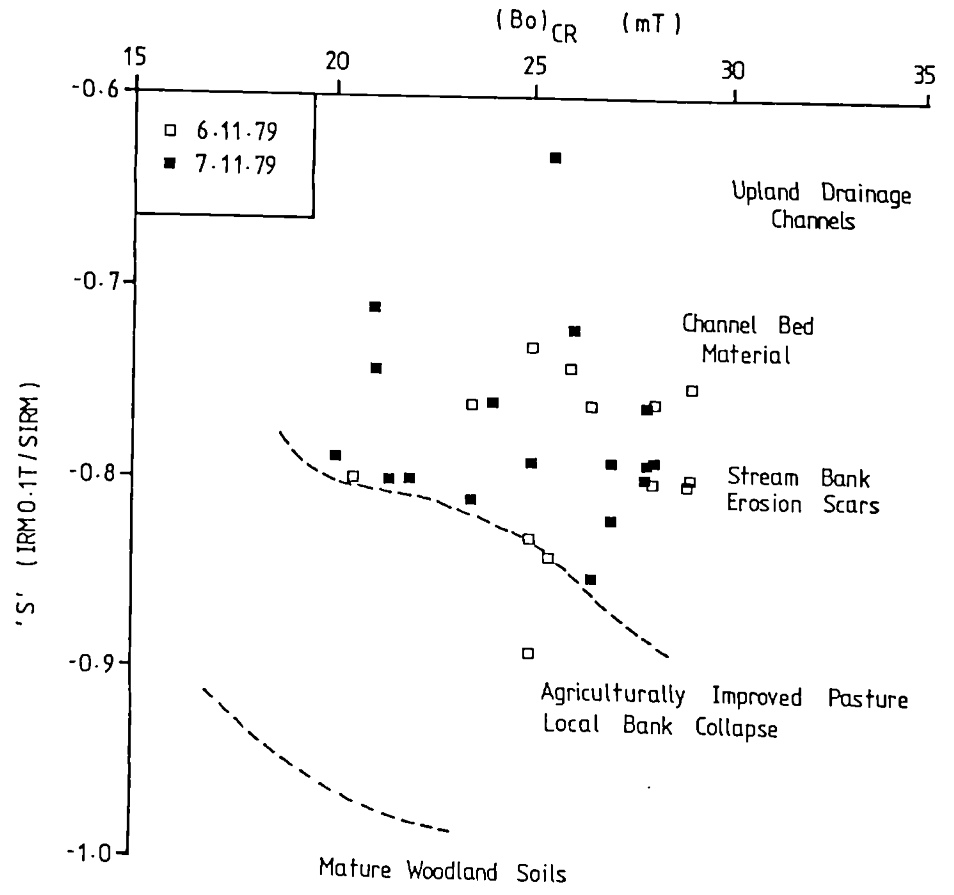


FIGURE 8.3. PILOT STUDY : RELATIONSHIP OF $(Bo)_{CR}$ TO 'S' FOR SUSPENDED SEDIMENT SAMPLES RELATED TO SOURCE CHARACTERISTICS (Figure 8.2.)

sources. A detailed compilation of source characteristics is made in section 8.4. As indicated by Walling et al. (1979), it may be that these small change in $(Bo)_{CR}$ etc., could be instrumental in highlighting specific local changes in source area. From an initial survey of sources from upland channel associations to those at Abermule, it was apparent that this might be so. Indeed, as opposed to the variations in SIRM data, matching $(Bo)_{CR}$ and 'S' for the main source samples illustrated anextremely orderly sequence from which it was possible to match the consistant measures of the suspended load, in a crude ordination fashion. These data are shown in Figure 8.2. and 8.3..

On this basis, a preliminary identification of likely sources was made highlighting for the study period, the importance of channel and bank sediments in the piedmont Severn. Given the essentially pastoral land use in the region, together with the extensive floodplain, these inferences would appear reasonable. The pilot study illustrated the potential for a relatively simple source classification procedure based on $(Bo)_{CR}$ and 'S' alone, providing a preliminary source identification for the upper Severn Basin. This framework was used as the basis for subsequent data analyses, though with more intensive sampling it was hoped that the parameter SIRM would be more discriminatory.

8.4. EROSION SOURCE CONTRIBUTING AREAS

Further to the analysis of sediment sources for the Pilot study, a more detailed program of sampling was carried out. This aimed to identify the magnetic characteristics of two main components in the sediment transfer system:

i) Channel derived sediment; and,

ii) Bank and surface derived sediment.

Samples of surface soil, bankside and bed material were collected with respect to each of the main suspended sediment sampling stations. These were then oven dried at 60°C, packed into clean or distilled water washed 10cc sample holders and measured for χ, SIRM, $(Bo)_{CR}$, and 'S'.

Measurements of IRM characteristics for this study and those that follow, were carried out using the 'molepin' magneto meter and pulse magnetizer equipment described in Chapter 3. The maximum field used with this equipment was 0.3 Tesla ie. $IRM_{0.3}$ or IRM_{3000}. Consequently, the parameters $(Bo)_{CR}$ and 'S' are used in a rather special sense here. Whilst they remain comparable with data derived from the pilot study (because samples saturate near to or below 0.3T), the measures do not confirm to previous published procedures (eg. Thompson et. al., 1980).

8.4.1 RESULTS:

8.4.2 BANK AND SURFACE DERIVED SEDIMENTS, NON-CHANNEL SOURCES:

Table 8.2. presents the range of magnetic characteristics observed for the main sources identified within the catchment. The results presented are a compilation of samples taken within similar source units. As with the Jackmoor Brook studies, sources with the highest χ and SIRM, with minimum Bo_{CR}, are found under mature woodland cover, in this case from a small plantation on the slopes of the piedmont Severn (Berthloyd Coppices- G.Ref SN974 853). For the most part, however, the predominant land use is pastoral with over 60% of the land under permanent grass cover. As a result, mature woodland soils are unlikely to have any significant impact on the system.

The main changes in source characteristics are associated with

SOURCE TYPE :	χ	IRM_{3000}	IRM_{3000}/χ	$(Bo)_{CR}$	S
	$(10^{-6}m^3kg^{-1})$	(mAm^2kg^{-1})	(KAm^{-1})	(mT)	
UPLAND SOURCE AREAS					
Valley Complex Podsolic Soils	0.1	0.248	2.5	36.5	-0.70
Drainage ditch - peat banksides					
Drainage ditch -	0.065	0.041	0.63	-	-
loose bed-material & bankside fines	0.185	0.358	2.3	18.0	-1.0
Channel erosion scars - bluff face, Afon LLwyd	0.12	1.1 - 2.42	9 - 20	13.5 - 20.5	-0.87 - -0.90
TRIBUTARY AREAS TO MAIN SEVERN CHANNEL					
Ploughed top-soil	0.241	0.772	3.22	25.5	-0.82
Mature Woodland soils 0-5cm	4.94	10.06	2.03	22	-1.0
5-15cm	3.72	6.34	1.7	19	-1.0
Bankside material	0.28 - 0.59	0.87 - 3.6	3.4 - 6.3	25 - 29	-0.73 to -0.9
PIEDMONT SEVERN SOURCE AREAS					
Pasture	0.45 - 0.92	1.8 - 5.4	2.4 - 6.3	20 - 24	-0.89 to -0.92
Bankside material - Surface	1.38	0.7 - 10.0	0.5 - 7.3	20	-0.9
mid-profile	1.19	0.23 - 6.0	0.2 - 5.0	24	-0.77
base-profile	0.60	0.07 - 0.2	0.1 - 0.2	45	-0.5

TABLE 8.2. MAGNETIC CHARACTERISTICS OF THE MAJOR LIKELY SOURCE CONTRIBUTING AREAS (NON-CHANNEL) OF SUSPENDED MATERIAL.

soil development in/on deposits proceeding downstream or downslope; from minimal values on the peaty Plynlimon upland surface to much more enhanced values on the main Severn piedmont floodplain between Llanidloes and Abermule. The consistent geology underlying the catchment simplifies the picture emerging here, and in any future studies of mixed lithology catchments the results may be much more difficult to interpret.

Erosion scar characteristics within the main Severn floodplain differ markedly from those of the Jackmoor Brook studies which resembled more the characteristics of local bedrock. Instead, the build-up of fine silty alluvium on an upper cohesive bank and its development into a brown alluvial soil (Clwyd Series) has produced much higher IRM_{3000}'s and reduced $(Bo)_{CR}$ and 'S' values. This is particularly so for the uppermost horizons which show a distinct topsoil development and consequently enhanced magnetic features. These are reduced down profile through a weathered B horizon, to extreme values associated with gleying at the interface of the upper cohesive bank with the underlying coarse sediment matrix. The base of profile typically shows much reduced χ and IRM_{3000} values, as compared to the surface features, and increased $(Bo)_{CR}$ and 'S' values typical of gleyed features (c.f. Chapter 3). The frequency of high river levels necessitates that this area be predominantly grassland.

The soil associations of the nearby hillsides (Denbigh, Brown Earth Series) are frequently used to produce barley and fodder crops in addition to pastureland. Consequently, these may be ploughed and drilled in spring and late summer exposing sediments of much lower magnetic characteristics (lower χ and IRM_{3000}, increased $(Bo)_{CR}$ and 'S') reflecting the turning of the soil profile.

Increasing in altitude, soils of the Hiraethog associations may be exposed through forestry operations or bank erosion. Where these

SIZE RANGE (mm)	χ $(10^{-6}m^3kg^{-1})$	IRM_{3000} (mAm^2kg^{-1})	IRM_{3000}/χ (KAm^{-1})	$(Bo)_{CR}$ (mT)	S
A. UPPER SEVERN CATCHMENT - (Source : LTan drainage ditch)					
5.6 - 11.1	0.12	0.132	1.09	29.5	-0.51
2.8 - 5.5	0.12	0.173	1.44	10.5	-0.90
1.4 - 2.7	0.12	0.250	2.08	15.5	-0.92
0.710 - 1.3	0.13	0.340	2.46	15.6	-0.90
0.355 - 0.70	0.15	0.495	3.30	15.5	-0.90
0.180 - 0.350	0.11	0.420	3.82	17.5	-0.83
0.090 - 0.179	0.11	0.451	4.10	19.5	-0.81
0.063 - 0.089	0.09	0.441	4.90	22.5	-0.77
0.063	0.09	0.436	4.84	23.5	-0.72
B. PIEDMONT SEVERN - (Source : River Severn at Caersws)					
5.6 - 11.1	0.16	0.095	0.596	27.5	-0.77
2.8 - 5.5	0.14	0.075	0.537	31.5	-0.68
1.4 - 2.7	0.13	0.098	0.752	45.5	-0.44
0.710 - 1.3	0.13	0.120	0.924	36.5	-0.53
0.355 - 0.70	0.15	0.249	1.66	29.5	-0.73
0.180 - 0.350	0.17	0.471	2.77	23.5	-0.79
0.090 - 0.179	0.19	1.162	6.12	26.5	-0.76
0.063 - 0.089	0.23	1.199	5.21	26.5	-0.78
0.063	0.26	1.398	5.18	26.5	-0.78

TABLE 8.3. MAGNETIC CHARACTERISTICS OF THE MAJOR LIKELY SOURCE CONTRIBUTING AREAS (CHANNEL) OF SUSPENDED MATERIAL.

sediments are exposed under the peaty surface, they exhibit characteristics typical of limited podzolization; low IRM_{3000} and χ and increased $(Bo)_{CR}$ and 'S' in association with the development of an iron pan or iron staining. The main upland channels, for example the Afon Llwyd, may expose scar and cliff faces in outwash gravel deposits. Detailed sampling of the exposure (Plate 8.1.) exhibited characteristics similar to the local bed material but with slightly increased IRM_{3000} values.

8.4.3 BED MATERIAL - CHANNEL SOURCES

Table 8.3. presents a summary of the magnetic characteristics of channel source materials, comparing upland with piedmont locations. These show the variations in χ, IRM_{3000}. $(Bo)_{CR}$ and 'S' related to particle size of bed material arising from the drainage ditch LTan and from the main Severn channel at Caersws. Whilst much of the finer material carried in suspension in the LTan channel is the product of progressive attrition, it is the contention here that this may not be the case for the piedmont Severn and that the majority of bedload fines are derived from bank erosion. An examination of particle size variation in piedmont Severn bed-material (c.f. Chapter 7. Morfodion site data) shows that less than 2% is of a range available for suspension; 98% of available channel material is coarser than 180μm. This may be a result of the isolation of many upland sediment sources from the lower channel as a result of reservoir regulation and impoundment (Grimshaw and Lewin, 1980; Lewin, 1978). Two main features may be drawn from the data in Table 8.3.. The data from both locations show a relationship of increasing IRM_{3000} with decreasing particle size (Figure 8.4.) This relationship may be attributed to a higher concentration of stable single domain to a superparamagnetic magnetite within the finest particle

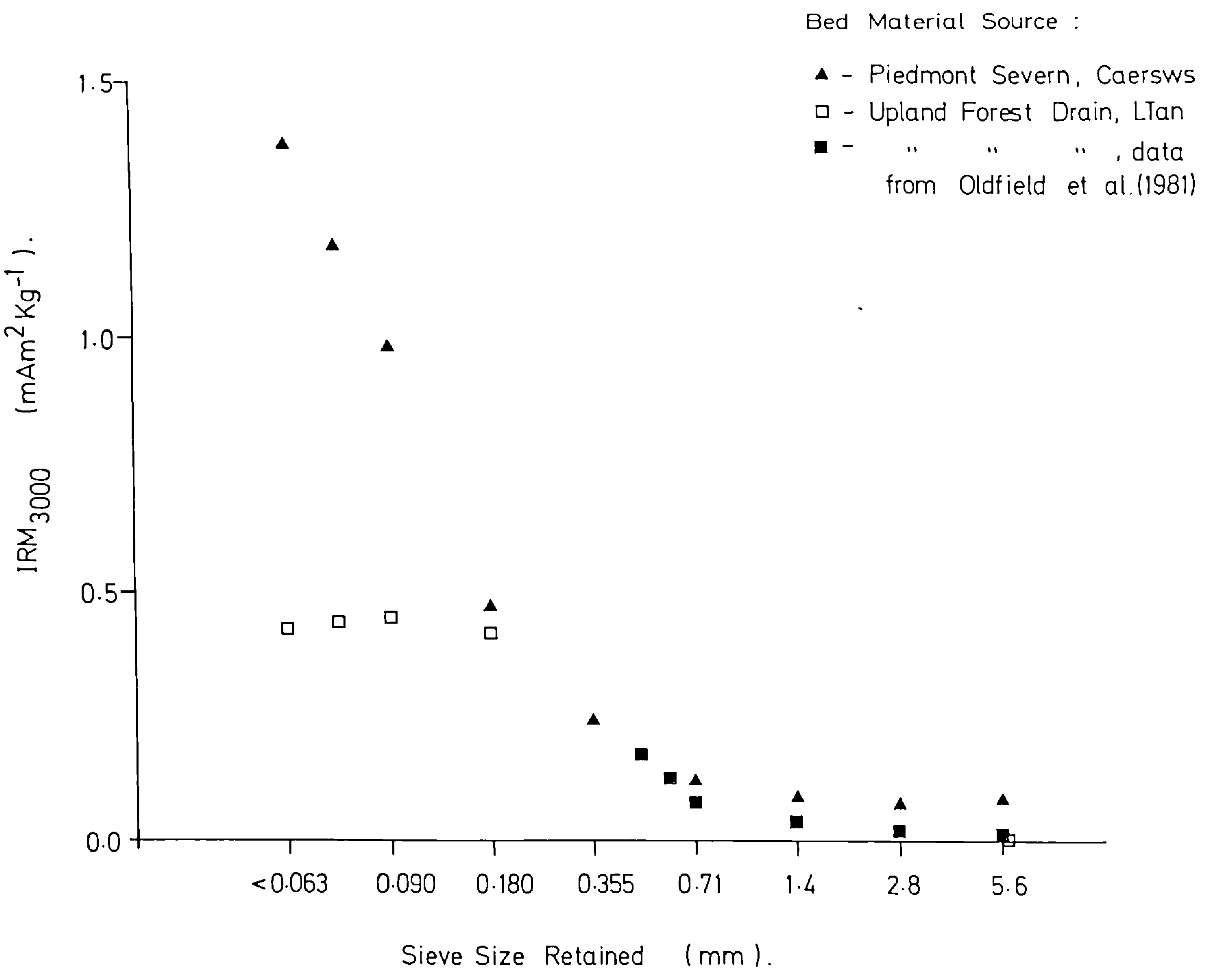


FIGURE 8.4. VARIATION IN IRM_{3000} TO PARTICLE SIZE OF BED-MATERIAL SOURCES.

sizes. Such mineralogy is also associated with much 'softer' IRM curves with the finer particle sizes having much lower coercivities ($(Bo)_{CR}$).

Figure 8.4. shows that for the piedmont channel material, the inverse relationship of IRM_{3000} to particle size is even more pronounced; IRM_{3000} values for the Caersws channel material rising to 1.4 mAm^2Kg^{-1} as opposed to $0.436mAm^2Kg^{-1}$ for the LTan. The Caersws samples also show much more consistent results for $(Bo)_{CR}$ and 'S'. Given that the coarser material is of a similar lithology to that of the upland channels, and there are no distinct bedrock variations, these data may be taken to imply a different source origin for the finer material. The similarity of magnetic characteristics between the finer bed material at Caersws and the local mid-bank sediments (see Tables 8.2. and 8.3.) suggested that these may have a dominant bank source origin.

These data are derived from two extremes of the channel system under study. Under 'normal' circumstances, there would probably be a continuous downstream change in the characteristics of the transportable sediment and a distinction of upland - piedmont sources in this manner would be invalid. On the other hand, there may be more apparent reasons for a definite change such as is observed. Firstly, the upland sediment source areas may be isolated from supplying the piedmont channel as a result of the impoundment of water at Clywedog. Releases from the reservoir may be competent to remove the available fines from the piedmont bed-material and also initiate bank erosion. On the other hand, the natural availability of sediment fines, in recently formed (or forming) floodplain areas, for suspended transport is low in many Welsh rivers draining the Palaeozoic shales (Lewin, 1978; c.f. Chapter 4). As a result, the availability of fines in the floodplain bank deposits may be tapped. In this situation, it is possible to envisage

PLATE 8.1. AFON LLWYD : DOLYDD REACH EROSION SCAR.

a much broader characterisation such as that presented in Table 8.3..

The characteristics of the transported sediment have been analysed to test these hypotheses.

8.5. RESULTS

8.5.1 SUSPENDED SEDIMENT CHARACTERISTICS RELATED TO SOURCE SAMPLES

For the main study period, 131 samples of suspended sediment were analysed : 80 from the Abermule guaging station, and 51 from various sites upstream (c.f. Figure 4.1.).

As observed in the Pilot study, an exponential decline in IRM_{3000} with increasing filtrate weight again occured (Figures 8.5. a, b and c). For the Abermule data, the critical weight would appear to lie between 0.02g and 0.03g. Above weights of 0.03g the IRM_{3000} values remain reasonably consistent between 1-3mAm²Kg^{-1}, accounting for 57% of the data. Further studies suggest that these variations in SIRM or IRM_{3000} for low sediment weights are a result of impurities in the glass fibre filter papers (Smith pers. comm.) aswell as in the sample holders used. Detailed XRF and magnetic analyses of blank filter papers have shown minute ferrimagnetic impurities within the fabric of the filter which when measured with low sediment weights contribute to the IRM retained, thereby artificially increasing the SIRM/IRM_{3000} component. Similar analyses of 'S' and $(Bo)_{CR}$ with sediment weight show no such variation, and may therefore be used as a much more reliable discriminatory measure.

The following analysis is presented in two ways. Firstly, on the basis of the variation between $(Bo)_{CR}$ and 'S' for the complete data set

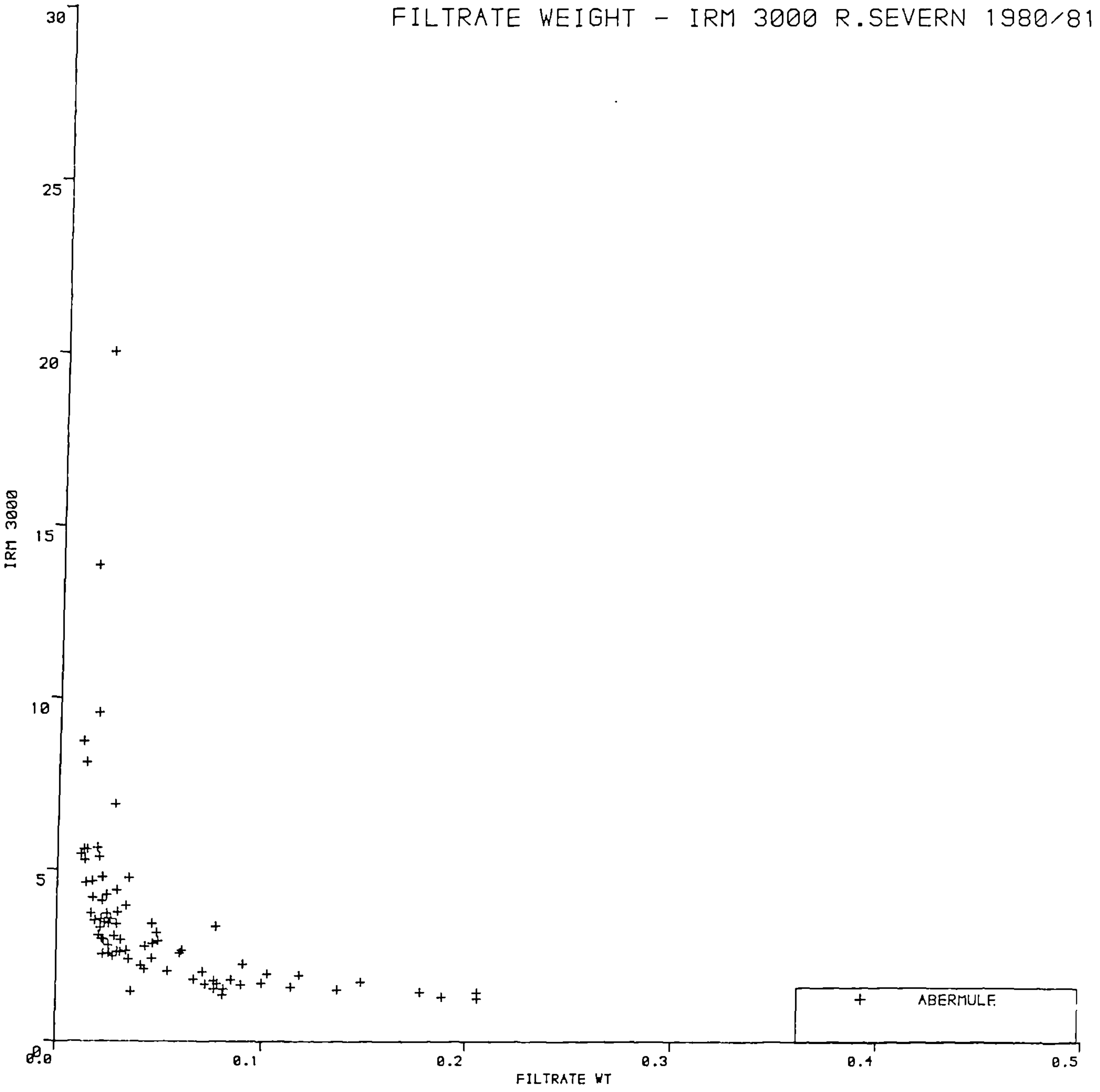


FIGURE 8.5. A). VARIATION OF IRM_{3000} TO WEIGHT OF SUSPENDED SEDIMENT RETAINED FOR THE MAIN STUDY PERIOD - PIEDMONT SITES, ABERMULE.

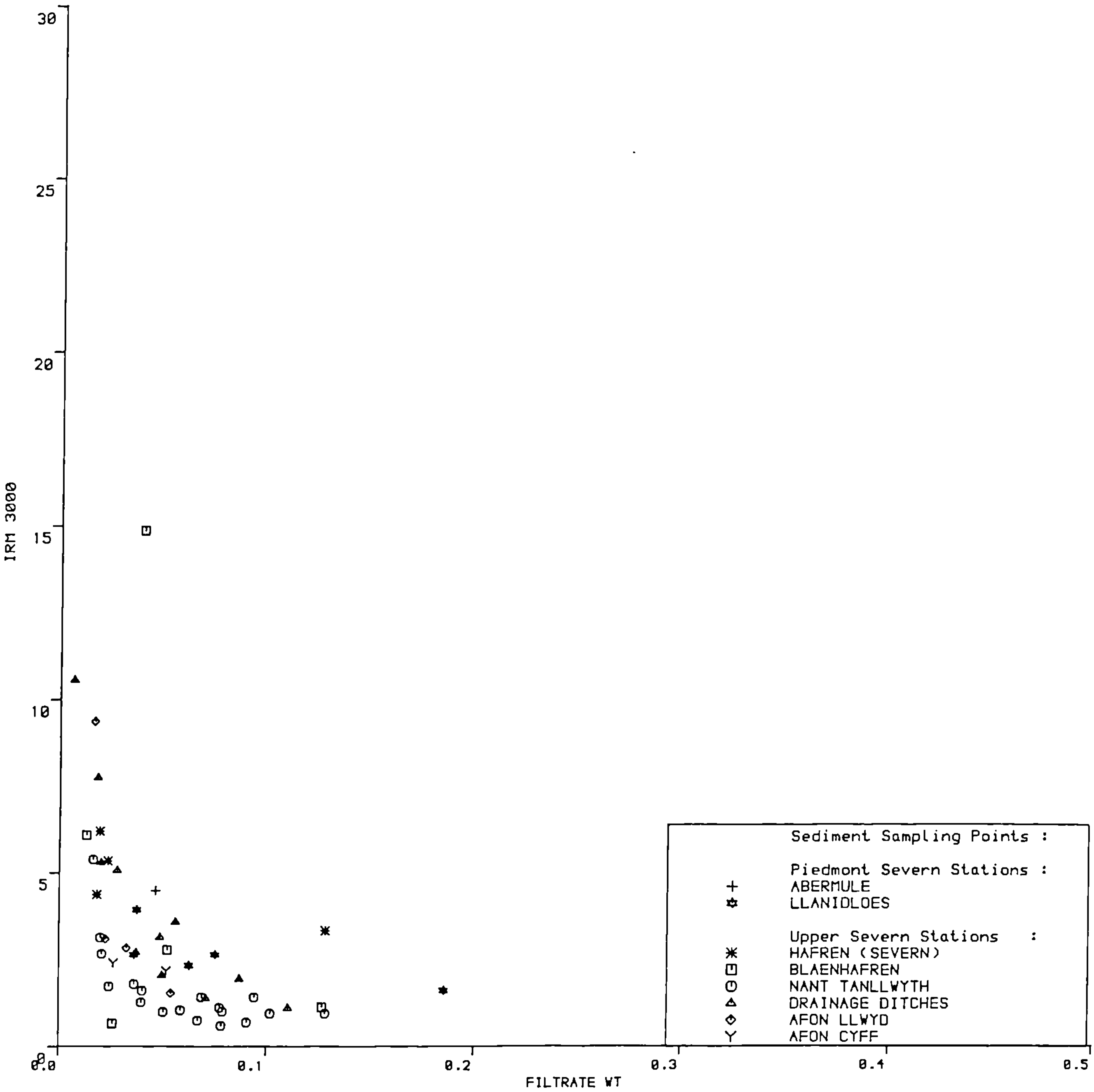


FIGURE 8.5.B. VARIATION OF IRM_{3000} TO WEIGHT OF SUSPENDED SEDIMENT RETAINED - UPLAND SEVERN SITES.

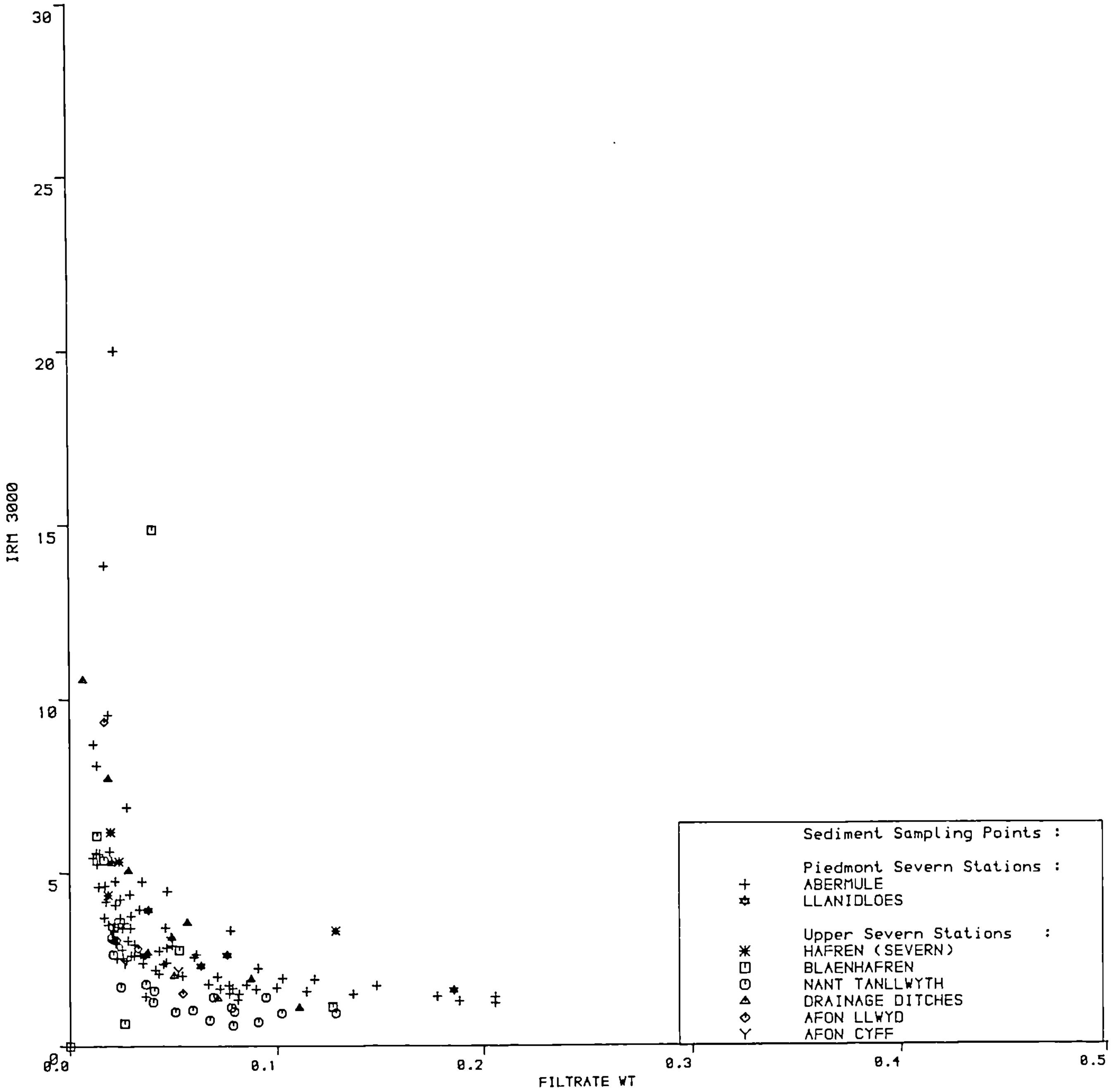


FIGURE 8.5.C. VARIATION OF IRM_{3000} TO WEIGHT OF SUSPENDED SEDIMENT RETAINED - ALL SAMPLING POINTS.

to provide regional source component changes. Secondly, as an individual study of inter and intra-flood variation in source contributing areas for floods monitored during the winter 1980-81.

8.5.2 SUSPENDED SEDIMENT-SOURCE, SPATIAL VARIATIONS

Following the pattern exhibited by the Pilot Study data, both source and sediment data were plotted to show variations in the relationship of $(BO)_{CR}$ to'S'. These are shown in Figs 8.6. and 8.7.. A number of features were immediately apparent from the suspended sediment data.

Firstly, the data showed consistent clustering within the ranges :

$$S = -0.55 \text{ to } -0.90$$
$$(Bo)_{CR} = 20.0\text{mT to } 45.0\text{mT}$$

and secondly;

within the main cluster there were discrete units of apparent source origin. The Abermule samples show clustering within the centre of this group; many of the samples are plotted over one another because their results were so similar (and therefore may only be represented by one single '+'). The remaining samples appeared to cluster around this central core of data. Using the plot of source characteristic data (Figure 8.6.) it is possible to relate these directly and make some inferences as to source origin. Data from the Upper Severn sampling stations showed two distinct groups appearing at each extreme of the Abermule data. At the magnetically "harder" extreme, samples from Blaenhafren and Nant Tanllwyth show much higher $(Bo)_{CR}$ and 'S' values and represent samples taken from the main tributaries draining the Hafren Forestry plantation and from the Blaenhafren felling site respectively. In both cases, the results are consistent with those of the soils of the Hireathog association and reflects the transport of

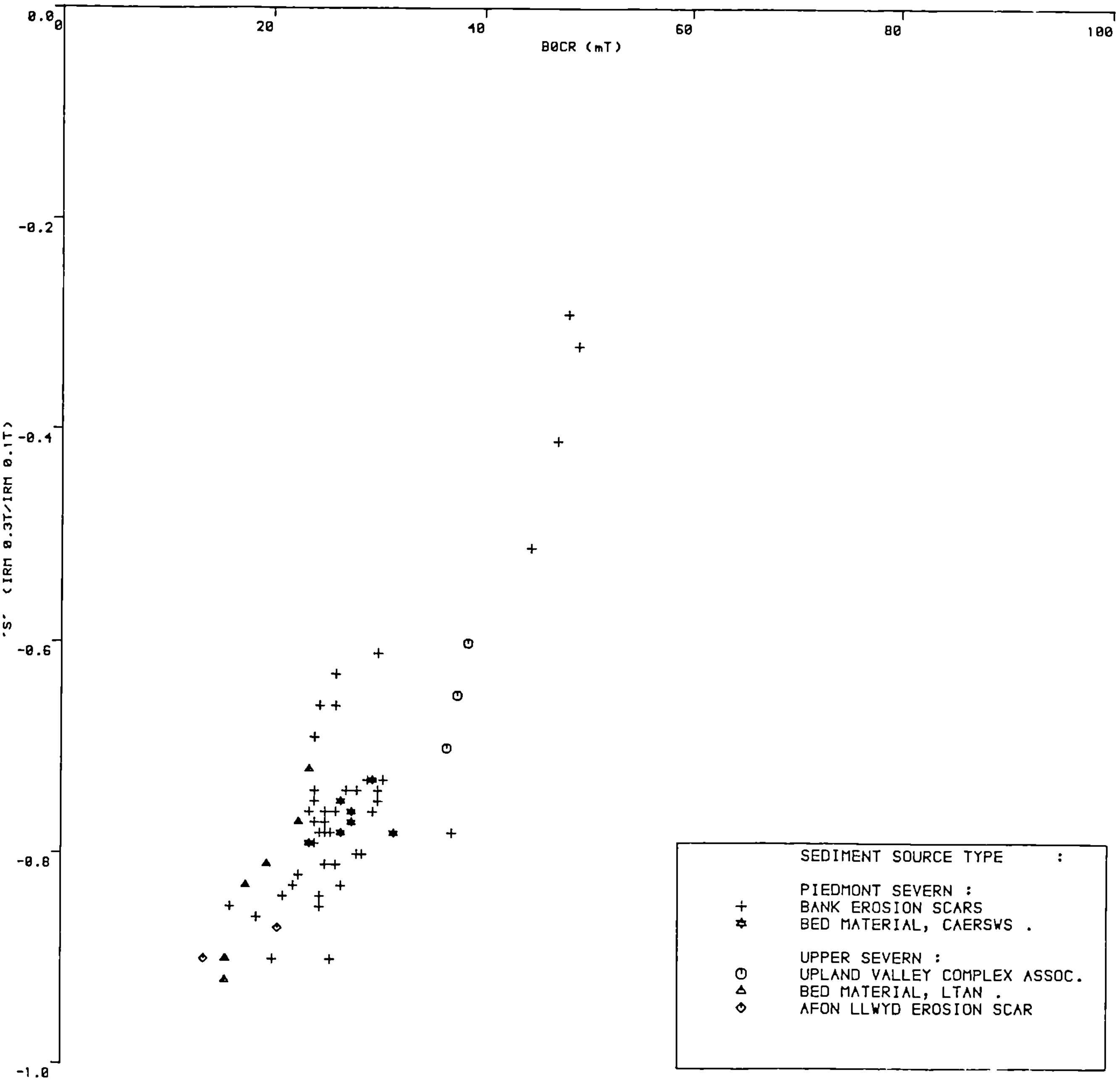


FIGURE 8.6. RELATIONSHIP OF $(Bo)_{CR}$ TO 'S' OF SAMPLED SOURCE MATERIALS.

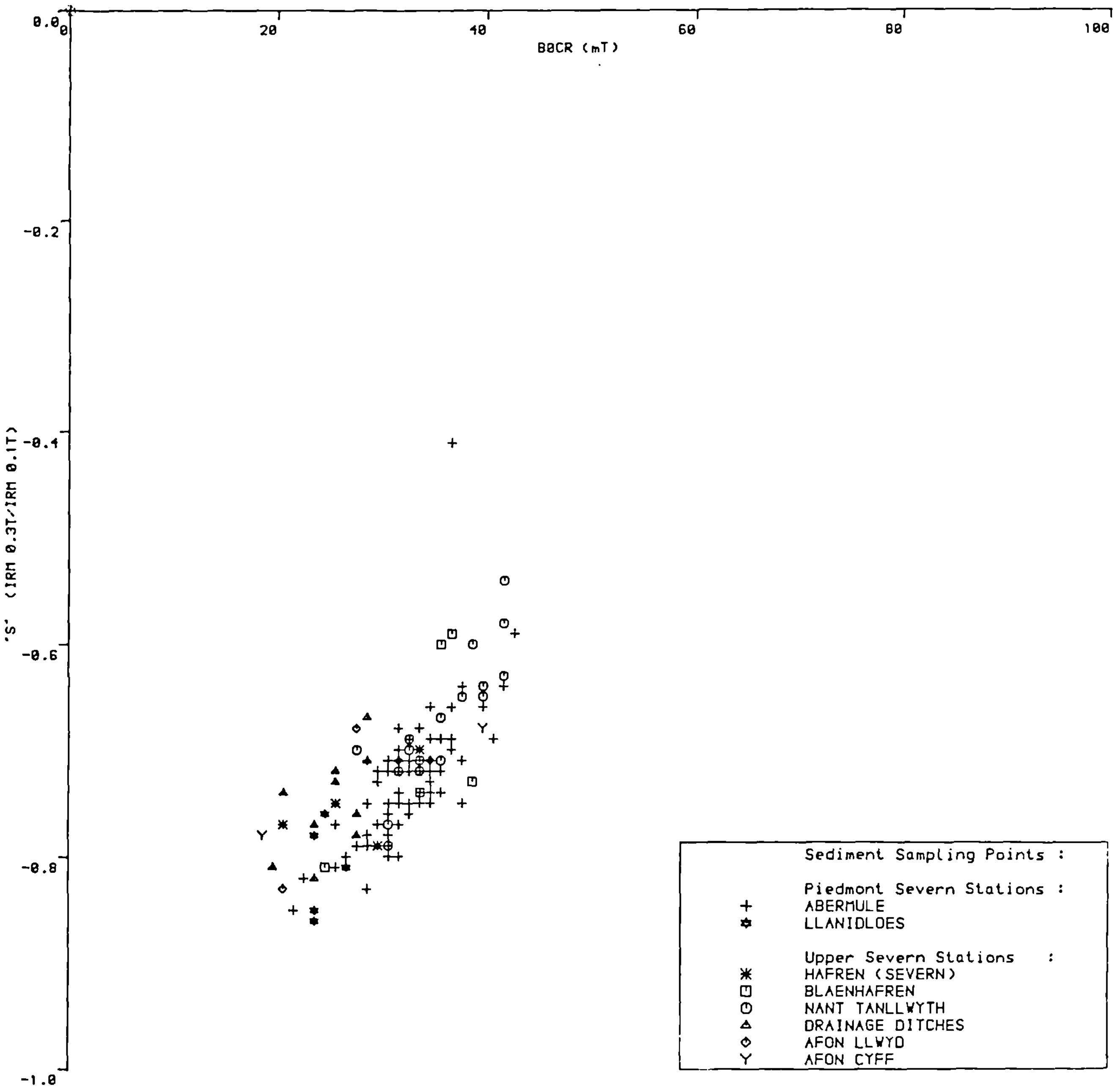


FIGURE 8.7. RELATIONSHIP OF $(Bo)_{CR}$ TO 'S' FOR SUSPENDED SEDIMENT SAMPLES.

iron stained and weakly podsolized sediments. As the Nant Tanllwyth is the main tributary draining this area, it may well be that much of the sediment is being supplied in consequence to felling operations and ground disturbance. No measures of increased yield are available. These inferences are enforced by the lower IRM_{3000} results shown in Figure 8.5..

Samples from Hafren (R.Severn) drainage ditches, the Afon Llwyd and Afon Cyff show similar characteristics to those of the bedload material itself in these channels. With the Afon Llwyd, the distinction between erosion scars and bed material cannot be made on these characteristics alone. The much higher IRM_{3000} values of eroding cliff sediments and of the suspended load (1.1-2.42 mAm^2Kg^{-1} and 1.0-2.5 mAm^2Kg^{-1} respectively) suggests that thesecliff sections are the significant contributors of this load.

The variation in IRM_{3000} results again is an unreliable indicator of source origin for these upland sediments. Even the higher sediment weights such as those sampled in the drainage ditches produce erroneous results. This latter effect may reflect the difficulties of sampling in low water discharge conditions and the liability of including coarser shale fragments in the filtrate which will have similar effects to those discussed previously.

For the Abermule data, 75% fall within the ranges:-

-0.70 to -0.80 for 'S'; and
29.0 to 35.0 mT for $(Bo)_{CR}$

This, in addition to the range of reliable IRM_{3000} results infer a dominant channel bank source origin. The extremes in values, related to topsoil surface wash and bankfull discharge conditions, together with the distinction between bed and bank sediments is discussed with reference to specific flood events in the next section.

The relationships of source-sediment load described above are based upon the characteristics of bulk samples collected from individual source areas. Peart and Walling (1982) observed temporal variations in particle size of suspended load which may be related to seasonal changes in source areas rather than fluctuations in discharge or stage (Ongley et al., 1981). The range of particle sizes analysed in source and suspended loads by Peart and Walling was 0-63μm; for a range of events in some South Devon catchments, they observed size variations commonly finer than 2μm.

We have already seen that for the range of channel materials collected (size ranges 0.063 to 11.1mm), there is a progressive 'softening' (lower $(Bo)_{CR}$ and 'S' values) of the magnetic characteristics parallelled by an increasing IRM_{3000} with decreasing particle size. This may be a contributing factor in the exponential relationship between IRM_{3000} and filration weight. Direct comparisons between the River Severn and those catchments studied by Peart and Walling cannot be made; the size relationships between source and suspended load are dependent upon local material characteristics. Furthermore, sampling of suspended load in broad, shallow upland channels is a difficult task and samples frequently contain coarser bed material which may be transported throughout the full depth ranges of flow. However, for the main Severn channel at Abermule (Grid Ref: SO 157945), if we assume a broad size range of sediment available for suspension between 0 to 90μm, it is possible to infer some magnetic characteristics from the data presented above. Data for material (suspended or avaialable bed sediments) for individual size ranges finer than 63μm is not available primarily due to difficulties in separating beyond this limit; any dispersants used eg. Calgon, may have a detrimental effect on the mineralogy and magnetic characteristics. Secondly, within the time available, the only suspended load samples were those collected on filter papers.

Using the data presented in Table 8.3., the consistent coercivities and 'S' ratios for particles finer than 90μm would suggest that any suspended material derived from these channel sources should be in the range:

26.5mT - (Bo)CR

and, -0.76 to -0.78 - 'S' ratio.

Given the trend in IRM_{3000} values (Figure 8.4.) it is possible to extrapolate values in excess of $1mAm^2Kg^{-1}$ upwards to $3mAm^2Kg^{-1}$ for the finest size ranges. For bed material sources derived directly from bedrock (ie. data from the LTan drainage ditch), sediment in the ranges finer than 90μm have the following characteristics

19-23mT - (Bo)CR

and, -0.8 to -0.72 - 'S' ratio

IRM_{3000} values for these sediments fall in the range 0.43 - 0.45 mAm^2Kg-1.

8.5.3 SEDIMENT-SOURCE INTERRELATIONSHIPS DURING STORM EVENTS

Suspended sediment samples were collected for two storm periods during February 2-14th and March 10-27th, 1981. The results are presented in Figures 8.8. and 8.9., illustrating for each storm period, discharge -suspended sediment variations related to changes in $(Bo)_{CR}$, 'S', IRM_{3000} and actual weight of sediment retained on each filter paper.

The storm hydrographs monitored here span a period of days as opposed to hours (see previous studies at Jackmoor Brook by Walling et al. (1979)). Nevertheless, the volumes of sediment in transport (suspended sediment concentration - lower graph in each case) are much lower than those arising from the Jackmoor Brook study. Furthermore, the suspended sediment response to discharge is not a simple relationship and, for

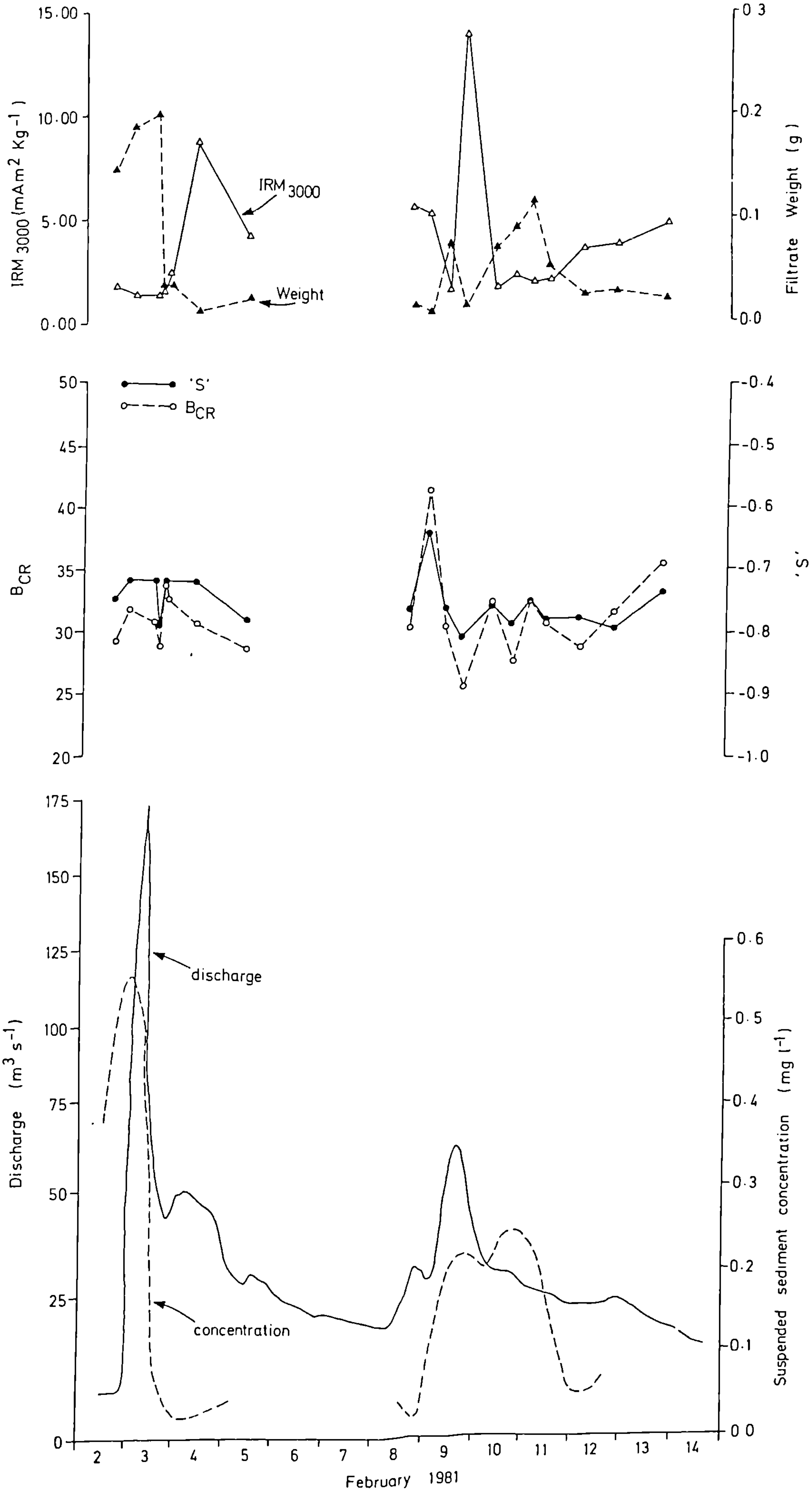


FIGURE 8.8. STORM VARIATIONS IN IRM_{3000}, $(Bo)_{CR}$ AND 'S' OF SUSPENDED SEDIMENT SAMPLED AT ABERMULE, RIVER SEVERN, FEBRUARY 1981.

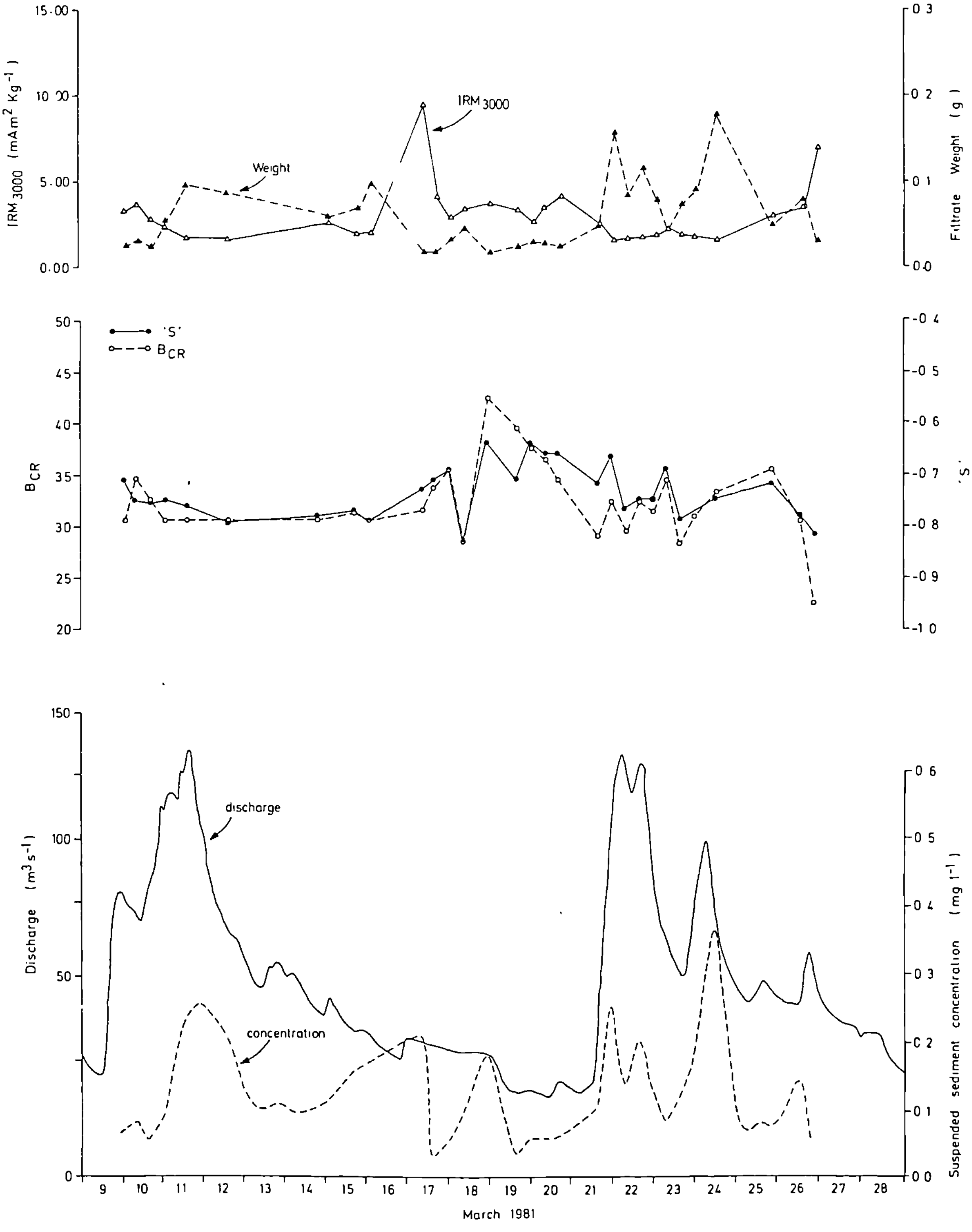


FIGURE 8.9. STORM VARIATIONS IN IRM$_{3000}$, (Bo)$_{CR}$ AND 'S' OF SUSPENDED SEDIMENT SAMPLED AT ABERMULE, RIVER SEVERN, MARCH 1981.

the March 1981 storm period in particular, shows responses apparently independant of stage - discharge variations alone.

Whilst from the foregoing discussions, the dominance of bank and channel source material for sediment supply is illustrated, the range of values suggest that this may be an overgeneralisation. This analysis has been carried out to attempt to identify detailed changes in source-sediment type during individual storm events.

The most apparent feature of the data for both storm periods, is the consistency of the $(Bo)_{CR}$ and 'S' values for much of the observation period in each case. For the March 1981 period, 72% of the data fall within the ranges : 'S' =-0.75 to-0.8

and $(Bo)_{CR}$ = 28.5 to 34.5 mT

This consistency was also noted in the previous discussion. The range of data suggested a predominant mid-bank and bed material source of suspended sediment load.

For both storm periods, the variations observed in IRM_{3000} indicate the complexities discussed above; there is an inverse relationship to filtrate weight and this may reflect either contaminants in the filter paper or definite changes in source material or sediment particle size. The range of $(Bo)_{CR}$ and 'S' values suggest that these peaks in IRM_{3000} are probably artificially high (at these filtrate weights) for their apparent sources. For example:-

	Filtrate weight (g)	IRM_{3000} (mAm^2Kg^{-1})
Feb. 1981-Peak 2	0.0167	13.85
March 1981	0.0184	9.6

During the storm events monitored for the period February 1981, the peaks in IRM_{3000} are associated, in the first instance, with flow and sediment concentration recession; and, in the second instance, with the subsequent rise in suspended sediment concentration. From previous discussions, these increases in IRM_{3000} may be associated with finer particle size ranges being transported in suspension. The source of these particles is difficult to pinpoint using this parameter alone, but, such high values of IRM_{3000} suggest that these sediments may be predominantly bank derived and associated with the enhanced magnetic properties of near surface soil horizons. The peak discharges for both of the February storm events are associated with decreases in both $(Bo)_{CR}$ and 'S' and are consistent with the characteristics of upper bank and surface derived sediments. The slightly higher values of $(Bo)_{CR}$ and 'S' prior to and following the discharge peaks would suggest that, if the source is predominantly bank derived, then the supply of sediments is stage related. At reduced stages, flow is acting upon lower - mid bank sediments which are characterised by increased $(Bo)_{CR}$ and 'S' values (c.f. Table 8.2.).

The IRM_{3000} data from the March 1981 storm hydrograph are much more difficult to interpret and seem to reflect the complexities in sediment response as illustrated by the lower graph in Figure 8.9.

A feature common to both storm periods was the rise of $(Bo)_{CR}$ and 'S' values at the foot of the hydrograph recession limb, and immediately prior to the next rise in stage. Low filtrate weights again pose a problem in the analyses here and preclude any definite comment with regards the parameter IRM_{3000}. The data for both periods (below) was consistently high:

	(Bo)CR (mT)	'S'
February	41.5	-0.64
March	42.5	-0.59

In previous studies, this type of behaviour was associated with source changes from bank to bed derived sediments. However, from the foregoing discussions in this chapter, this is unlikely to be the case here; the range of data is much higher than expected for any bed or bed-rock derived material. Furthermore, this is unlikely to be caused as a result of any change in sediment particle size (these data correspond to material greater than 2.8mm in size) at such reduced flows. Rather, from the range of data for individual sources (Table 8.2.), these results are consistent with two main source areas: lower bank sediments (at the base of profile near to the coarse sediment boundary), or, upstream soil associations such as those discussed for the area of Afon Tanllwyth. The latter area may be discounted as any response of this kind would also be expected at earlier discharge - suspended sediment rises. Perhaps more importantly, the range of values of IRM_{3000} for this source material is much lower than that of the suspended load collected at Abermule. This source change is consistent with a reduction in water stage (from 125+ cumecs to 25 cumecs). The mechanisms of entrainment and supply of sediment may be more complex. Aswell as the operation of reduced stage, the dominance of lower bank sediments as a source suggests two other processes may be of importance. The variations observed in discharge - suspended sediment concentration relationships is consistent with the effects of bank failure as described by Thorne and Lewin (1979) and Thorne and Tovey (1981). The collapse of large ped blocks, combined with the effects of entrainment and undercutting at reduced storm flows, provides the release of basal debris from the bank sediments and the gradual breakdown and incorporation of the ped material. Similar processes were inferred in the study reported by Walling et. al. (1979).

In addition to the mechanism described above, the change in sediment source observed in both Figures 8.8. and 8.9. may also be a response to increased water levels within the floodplain sediments. Floodplain inundation (Hughes, 1980) in this manner provides the release of stored water at reduced flows, augmenting the baseflow component and providing for the entrainment of basal sediment fines. In a much smaller catchment, Wood (1978) has associated such a response to increased throughflow generated from adjacent hillslopes.

However, in terms of total suspended load (as opposed to concentrations), it is the high flow sediment contributing sources which are of most significance. The techniques used above demonstrate the uniform nature of these high flow sources which were quickly and easily matched with those magnetic characteristics identified with the piedmont Severn valley floodplain soils and river bank sediments.

8.6. SUMMARY AND CONCLUSIONS

Suspended sediment samples have been collected by conventional USDH bottle and automatic vacuum samplers at various sites on the River Severn upstream from the Abermule gauging station. The low concentrations of suspended sediment in transport produced unforseen problems in the analysis of magnetic properties. The low weights and naturally paramagnetic properties of much of the local material meant that the simplest measurable parameter, susceptibility, could not be applied here. In addition, at filtrate weights below 0.03g, impurities contained in the filter papers and sample holders begin to dominate over the natural magnetic properties of the filtered sediment. This adversely affected measures of SIRM and/or IRM_{3000}. Any future studies should consider this weakness and, where possible, should consider collecting much larger volumes of streamwater to provide increased sample weights. Alternatively, filter papers are available which are guaranteed free of impurities. However, with such a guarantee, there is also an increase in cost which would probably preclude their use in continuous monitoring experiments.

Despite these limitations, IRM_{3000} and SIRM measures for larger filtrate weights, together with the parameters $(Bo)_{CR}$ and 'S' provided a rapid tool for the broad identification of source types within a limited study of suspended sediment loads.

For upland channels, it was possible to identify three dominant sources of suspended sediment. In deeply eroded drainage channels, the main source identified suggested that sediment was derived predominantly from the Lower Palaeozoic bedrock and from the progressive attrition of bed material to release fines dominated by SSD and SP mineral assemblages (c.f. Chapter 3). Where these channels enter the main stream system, such as the Nant Tanllwyth tributary system, suspended sediment characteristics were associated with the local podsolic and peaty podsolic soils which comprise much of the bank sediments. The main channels are dominated by coarse sediment and bedrock outcrops. The magnetic characteristics of the Tanllwyth suspended load are typified by a hard mineral assemblage with high $(Bo)_{CR}$ and 'S' values. Sediment supply to these channels may also be associated with the development of soil pipe systems which are common in the Hireathog soil complexes (Chapter 4). In larger upland channels, such as the Afon Cyff and Afon LLwyd, sediment sources may have a significant contribution from the erosion of large bluff faces in fluvioglacial outwash deposits (c.f. Plate 8.1. and also Lewin et al. (1974)). The sediment system within these reaches is a complex interplay of bed- and bank - derived sediments: the latter may be readily distinguished by their higher IRM_{3000} values (attributed to the product of insitu weathering processes) as compared to the characteristics of the local bed material. More detailed analyses of suspended sediment concentrations within these reaches would be required to establish the dominance of any one of these sources and the processes supplying material

In the piedmont zone of the River Severn (Abermule) the data collected suggested that sediment sources were largely channel and bank derived This is consistent with the nature of the channel, which is actively meandering through a wide floodplain dominated by pastoral farming activities. The distinction between bed and bank sediments was difficult. Indeed, it is likely that they may be one and the same if the effects of the Clywedog Resevoir (restricting sediment supply to the piedmont zone but augmenting natural flows) combined with the natural lack of sediment fines in contemporary floodplain channel deposits of mid-Wales are applicable here. Further data is required to test this hypotheses. The storm variation in magnetic characteristics of suspended sediment in the piedmont zone were associated with changes in water stage and related to the height of bank being eroded; much more enhanced magnetic characteristics associated with surface soil horizons were observed at or immediately after peak flows. The suspended sediment characteristics of reduced winter flows, between storm events, were typified by much harder magnetic mineral assemblages similar to those of the lower cohesive bank sediments of the piedmont floodplain, and consistent with the magnetic characteristics of gleyed sediments as a result of frequent inundation by high storm flows.

CHAPTER 9

Summary, General Conclusions and Further Research

9.1. Summary

Two essentially separate, but related studies have been carried out:- an investigation of the use of magnetically enhanced tracer sediment in the study of stream bed-load movement; and, an attempt to establish the sources of suspended loads in the mid-Wales reaches of the rivers Severn and Wye using magnetic methods. In addition, these studies combine to describe the nature of sediment transport systems in the gravel-bed reaches of these rivers. In the context of field application to the rivers Severn and Wye, some attempt towards 'whole river' application (and eventual quantification) has been made so as to forge some link between the large volume of data regarding the highly erosive upland reaches (Newson, 1980 a.b.; Lewin et. al., 1974; Painter et. al., 1974) and the effects of engineering and contemporary channel change within the piedmont reaches (Thorne and Lewin, 1979; Hey, 1984). In the past, many geomorphological studies have been too localised and frequently carried out over too short a time period (Pickup, 1981) to provide a solid basis from which we may adequately describe a river regime (Newson, 1981). Such a preoccupation with individual eroding sites in a contemporary context and associations with particular depositional sites, eg. reservoirs and lakes, over greater time scales has produced disparate rates of sediment loss or delivery from a river reach; the former producing much higher rates of loss for the basin as a whole, and the latter, much lower because of the averaging of variable accumulation rates over the time of deposition and the problems of quantifying delivery ratios (q.v. Walling, 1983). Very few data, or indeed techniques, are available to link such aspects of the whole fluvial system, and to monitor the dynamics and interrelationships of both suspended and bed-load sediments over space and time.

Bed-load tracing has been carried out at three stages within the mid-Wales river systems of the Wye and Severn:

1) in the potentially, highly erosive upland drainage ditches (LTan and LTan6'A ditches of the Tanllwyth system, Chapter 5);
2) within the major upland channel tributaries of both river systems (Afon Cyff, River Wye and, Afon Llwyd, River Severn, Chapter 6); and,
3) within the piedmont reaches of the river Severn (Morfodion, Chapter 7)

In addition, suspended sediment source tracing has been carried out on a much wider network of sample sites.

Experimental studies of bedload tracing exhibited two main characteristics. Firstly, the main phase of tracer sediment movement in each case study occurred in response to the first major storm flow event of the winter 1980 (Table 9.1.). Secondly, following transport from the tracer emplacement zones, the importance of storage of sediment for considerable time periods was illustrated. This pattern was consistent throughout all stages of the observed system. At the smallest scale, in forest drainage ditches, the major phase of tracer recovery by trapping (LTAN6'A) occurred in response to the storms of October, 1980. However, in the LTan ditch, whilst the initial transport rate in response to this high flow period was high, bed topography was such that localised storage of sediment was afforded, and provided consistent storage of tracer sediment and lack of subsequent movement even at equally high flows, throughout the following winter flow period.

Within larger upland channels (Afon Cyff and Afon Llwyd) the scale

SITE	OBSERVED MAXIMUM STORM DISCHARGE. (cumecs)	APPROXIMATE CHANNEL WIDTH. (m)	APPROXIMATE CHANNEL SLOPE. $(m.m^{-1})$	CHANNEL TYPE
LTAN	0.018	0.5	0.0095	Forest drainage ditch.
LTAN6'A	0.037	0.5	0.06	"
CEFN BRWYN	18.000	4.0	0.01333	Main upland tributary (Wye)
AFON LLWYD	90.000	8.5	0.01107	Main upland tributary (Severn)
MORFODION	70.000	30.0	0.00125	Piedmont channel (Severn)

NATURE OF TRACER EXPERIMENT:

LTAN - Reconstructed shoal storage area, stabilised under low flows.

LTAN6'A - " " " " " " " "

CEFN BRWYN - Seeded grid area of coarse channel riffle; transport to storage features (C 160metres).

AFON LLWYD - " " " between bridge piers; transport to major shoal reach immediately downstream.

MORFODION - Emplaced trench; vertical and lateral reconstruction including armour layer within a diagonal bar storage feature in the main piedmont Severn channel.

TABLE 9.1. PEAK DISCHARGE AND CHANNEL CHARACTERISTICS AT EMPLACEMENT SITE FOR THE MAJOR SEDIMENT TRANSPORTING EVENT, 6th OCTOBER, 1980.

of bedform development was much greater covering several 10's of metres in channel length as opposed to 0.25-1.0 metre as observed in the drainage ditches. Nevertheless, a broadly similar pattern of tracer movement and deposition occurred. In the case of the Afon Llwyd trace, tracer sediment was moved directly into storage by the October 1980 floods, to be retained within the main bridge shoal for the whole of the study period (21 months). By contrast, the Cefn Brwyn (River Wye) trace highlighted the patterns of sediment movement over a much greater distance to the main storage features initially through a riffle reach of some 160 metres. The movement of tracer through the main channel riffle occurred predominantly in response to flow events of very high discharge magnitude (Table 9.1.). The overall range of flows experienced at the Cefn Brwyn site for the period was 0.1 to 18.6 cumecs. Flows above an identified threshold of 17 - 18 cumecs (peak storm flow) were associated with a much greater recovery of tracer post-flood as compared to lower storm discharges, with the recovered tracer size distribution approaching the characteristics of the total range of sizes available in the bed and suggesting that these higher discharge events were able to activate the bed to a much greater degree and depth. At lower storm flows, the recovered tracer size distributions were much coarser and were thought to reflect both progressive armouring of the bed and gradual burial of tracer material. These aspects of burial and re-exposure of tracer sediments are commonly reported (see for example, Butler (1977) and Klingeman and Emmett (1982)). Downstream variations of tracer recovery in the Cefn Brwyn reach illustrated the importance of medium-large scale bed features (channel bars) to the deposition and storage of bed material.

At Morfodion, within the piedmont reaches of the River Severn storage of sediment in channel bars would appear to be even more long term than that described above. The magnetic tracing technique was used effectively to observe the development of a discrete channel bar feature which was modified at high flows (>70 cumecs) in direct response to the flow but also to the additional supply of sediment from upstream. Observed sediment movement at the site during the beginning of the flood season was essentially due to restructuring that sediment comprising the bar, whilst at similarly high flows later in the winter season (March 1981; peak flow: 75 cumecs) aggradation of the bar in response to sediment supply was observed. Tracer sediment at the bar was only moved very short distances in comparison to the other case studies, engaged in very local progradation of the bar. If these observations are representative of the sediment flux in the channel as a whole (at this point), the sediment supply to the channel is likely to be derived from erosion of the composite bank sediments thus providing a range of fine-coarse gravel sediments which at high flows are likely to be transported fairly rapidly over an armoured bed as a result of sediment overpassing (Everts, 1973).

Tracer recovery rates for each case study reflect the rate and depth of sediment mixing against transport distances, the amount of tracer sediment emplaced and the efficacy of the present generation of instrumentation. The broad range of particle sizes which may be traced using this enhanced magnetic technique, the overall recovery and 'potential' recovery of tracer with improved instrumentation (see later) compare very favourably against other established methods (c.f. Table 9.2.) By far the best tracer recovery rates, up to 71%, were recorded for the forest drainage ditch experiments (Chapter 5). Despite their highly erosive nature, the depth of sediment in storage was small, of the order of 0-10 cms, and sediment mixing did not pose any problems for the highly magnetic sediment used. However, tracer sediment recovered by

RESEARCHER		RECOVERY RATE (%) OF TRACER MOVED.	METHOD	SIZE RANGE (mm.)
ARKELL (this volume)		5.0 - 63.0	Superparamagnetic Enhancement.	1.4 - 90.0 +
BUTLER	1977	35.0	Metal Tag strips	34.0 - 116.0
HASSAN	1983	30.9 - 33.8	Metal Tags & Paint	45.0 - 180.0
HASSAN et. al.	1984	80.0 - 93.0	Ceramic Magnets	45.0 - 180.0
KELLER	1970	48.5	Paint	Gravel - Pebble
LARONNE & CARSON	1976	5.0	Paint	4.0 - 256.0
LEOPOLD et. al.	1966	0.0 - 38.0	Paint	75.0 - 125.0
"		15.0 - 88.0	"	75.0 - 150.0
NIR	1964	4.4	Iron Nails	52.0 - 240.0
SCHICK & SHARON	1974	2.5 - 10.5	Paint	32.0 - 512.0
		64.0	Paint	Pebble
		52.0	Paint	Pebble & Cobble
TAKAYAMA	1965	10.0 - 39.6	Paint	20.0 - 150.0

TABLE 9.2. A COMPARISON OF TRACER RECOVERY RATES WITH SIMILAR EXPERIMENTS (q.v. HASSAN et. al., 1984).

trapping had to be sorted by hand and whilst this was a laborious process it was not unnecessarily lengthy and proved very effective (field recovery rates of 93+%). The high total recovery rates of tracer within ditch systems reflects two things: firstly, the controlled nature of the channels, narrow reaches with fairly simple geometric cross-sections; and secondly, the amount of tracer material emplaced, which on scale represented a completely replaced shoal within the channel bed.

By comparison, the recovery of tracer from experimental studies in the much larger, main channel reaches were considerably lower, of the order of 6-12% per survey. The dominance of sediment storage within medium - large scale bedforms and beneath a coarse armour sediment during each of these experiments necessarily means that tracer sediment will be effectively mixed and buried to much greater depths than those experienced in the drainage ditch channels. For example, at Morfodion, River Severn there was evidence to suggest that the gradual development of the gravel bar would eventually bury the emplaced tracer at depths in excess of 0.5 metre and probably more; the coarsest recovered tracer material was buried below 1 metre auger depth. The instrumentation developed at the time of these traces was not effective enough to locate tracer material at such depths; the effective depth was approximately 0.1 - 0.2 metres depending upon tracer clast size and depth, size and induration of surface materials. Any further research using the magnetic tracing technique must bear this in mind; suggestions for the improvement of instrumentation and technique are made in the following section.

The much greater scale of sediment movement observed for the main channel reach experiments provide some suggestions on hindsight for immediate improvement of the methodology used above. The depth of sediment mixing either within channels or bedforms is important both for the efficacy of instrumentation (above) but also the initial amount of tracer emplaced. At this stage, it is the authors opinion that perhaps much greater detail may have been gained from the experiments if larger volumes of tracer material had been emplaced; some compromise needs to be achieved between those volumes initially used to those volumes of bed-sediment stored within channel bars. An ideal situation would be the emplacement of a complete bedform although the necessary stability requirements may not be achieved. In this context, it may be possible to achieve insitu magnetic enhancement of a channel shoal by using a flame thrower - type device. However, even Moseley's (1978) data for such considerable volumes of sediment yielded recovery rates as low as 5%.

9.2. Conclusions: Sediment Transport Dynamics

Several points are apparent from the foregoing discussion. Firstly, the tracer shoals emplaced in each case may have been unstable at the onset of winter flows. This is a factor inherent in all tracer studies and may create unnaturally high initial sediment transport rates until the tracer actually becomes incorporated into the bed material. However, it is the authors opinion that this is unlikely to be the case for the experiments described above, given the length of time at low flows for each emplacement site to stabilise. Furthermore, if this were the case, the patterns of tracer movement for the early flood events of the study period would have been mirrored for all size ranges of bed material. In many cases this was not so and the coarser fractions of the tracer material added remained static for some time.

However, despite this fact, in some cases the results of tracer movement must reflect, to some extent, the original position of emplacement and its inherent susceptibility to scour to give rise to sediment movement. This is particularly so for the Afon Llwyd and, to some extent, Cefn Brwyn traces; the former reflecting scour processes operating around bridge piers and the latter as a result of the lack of sediment supply from upstream. This compares with the lack of tracer movement once incorporated into the major channel bedforms even at equally high discharges. Whilst Laronne and Carson (1976) point to variations in the bed-load transport rate as a result of small scale bed features, topographic and hydraulic control and the influences of bed structure, such a marked spatial and temporal differentiation of bed-load movement as a result of channel storage has only really been alluded to by Moseley (1978) and Meade et. al. (1981). The effects, as shown herein, clearly question the assumptions made in the application of most sediment transport formulae (Chapter 2) and to the accuracy of data derived from fixed-positioned bed-load sampling devices.

The scale and pattern of tracer movement reflects a similar seasonal pattern to that described by Leopold and Emmett (1976) who suggested that sediment transport rates for the same range of high flows were greater at the beginning of a winter flow season than for any subsequent period. For the traces described above, such a pattern was associated with some of the highest observed discharges for the study period and subsequent to the initial period of quite considerable movement, the incorporation into the streambed as channel bars.

Some continued tracer movement was observed in the Cefn Brwyn study reach where tracer material was still being moved through a long riffle reach. However, the pattern of tracer recovery post-flood in subsequent events to the October, 1980 floods indicated supply variations of tracer sediment as a result of bed surface armour stabilization. Whilst all channel surveys illustrated a trend of downstream size-fining of recovered tracer particles, the data for successive surveys suggested that floods which exceeded a critical threshold of 17 - 18 cumecs were able to activate the bed to a much greater degree and release a greater range of tracer particle sizes for transport. Similar effects were described in the tracer experiments of Laronne and Carson (opp. cit.) and for bed-load material collected during storms, Klingeman and Emmett (1982) showed that the size distributions of bed-load and bed-material were similar at high flows. Whilst the mechanisms of bed material structural break-up are as yet little understood, Emmett and Leopold (1965) attribute the effect of high discharges to a greater dilation of the grain bed, presumably a result of the greater Lift Force generated by such high magnitude events (q.v. Lane and Carlson, 1954).

Subsequent observations suggested that once incorporated into the channel bed, the tracer sediment was stable at those flows less than the flows at which the bed formed. In this instance, this was the first storm flow event of the winter flood season. The size distribution of recovered tracer material for subsequent lower flow storm events reflected the armour surface composition. Further to this, data from the Afon Llwyd trace suggested that once incorporated into bedforms, the recovered tracer sediment size distributions of successive events indicated a progressive coarsening of the armoured surface. This appeared to be the result of vertical rather than downstream winnowing (Chapter 2) given the lack of consistent tracer recovery at sites further downstream.

Similar processes to those described above have been observed by Day (1981) in describing the progressive armouring of flume experimental channels. Day suggests that the coarse surface bed-material re-orientates to provide maximum resistance to flow. For this condition, Church and Gilbert (1975) suggests that the bed will 'underloose' material for a given discharge rather than 'overloose', the condition for which most bed-load formulae are derived (Pickup, 1981). This is presumably made more effective for the sites studied herein as a result of the natural shape of the shale bed material; the dominantly flat pebbles stacking and interlapping to form a tight, imbricate bed structure (q.v. Lane and Carlson, 1954). Further to this, recent results presented by Reid et. al. (1985) suggest that 'particle interlock' may be enhanced over long periods of lower flows as a result of the gravel bedding down and the additional strength which may be added to the bed-material framework as a result of finer particles settling down through the interstices of the surface gravels. The effect is that substantially increased discharges are required to initiate bed-material motion following such periods (Reid and Frostick, 1985). Further fieldwork is required in order to define the differences between the 'particle interlock' processes put forward by Reid et. al. and those of Day's progressive armouring. However, the patterns described by Reid et. al. are consistent with those observed for the Cefn Brwyn reach where tracer material was transported through a coarse gravel riffle; the patterns of tracer recovery exhibited marked differences in size distributions at surveys following periods of lower flows (eg. December, 1980; July, 1981; September, 1982) as compared to those following much higher storm flows (October, 1980; March, 1981). These data would suggest that much more work is required to identify the temporal and spatial changes in bed-material structure in order to be able to accurately predict bed-load transport.

Where tracer material had become incorporated into bedforms, such as observed for the Afon Llwyd and Morfodion reaches, bed-material transport became little more than a local flux with channel bars (Morfodion) advancing sporadically in response to high storm flows. The stability and quasi-permanence of such features has been discussed by Church and Jones (1982) who suggest that the bedform (or sequence of forms) will develop so as to impart the maximum form resistance and stability in adjustment to the hydrological regime and the rate of sediment supply. Observations for the Afon Llwyd shoal confirm Church and Jones' suggestion that the bar stability will increase in response to progressive armouring. As such, the longterm stability of a channel bar will be related to the character of the sediment matrix, the nature of the hydrologic regime, but particularly the frequency with which the largest materials present may be moved collectively and substantially rearranged (Church and Jones, 1982). This may mean that critical flows for bed-material movement in such instances must be in excess of those flows which formed the bar.

Similar observations for the region of Caersws on the River Severn have been reported by Thorne and Lewin (1979) and Milhous and Thorne (1982). Flows within the region of 80 cumecs were associated with general movement of nearly all the bed-material size ranges, whilst for flows between 17-35 cumecs sediment transport occurred as a result of sediment overpassing (Everts, 1973) an intact armour surface. Data collected for the Morfodion reach would suggest that sediment incorporated at such times will have been derived from local bank failure. There is evidence to suggest that gross channel change may occur at extreme events (Lewin, 1981; Church, 1983) although the frequency with which such events would occur for this part of the Severn, atleast, are very rare (Hey, 1984) of the order of once in ten years.

The data described above, and that by Gomez (1983) would seem to confirm those suggestions made initially by Milhous and Klingeman (1973) and since by Emmett (1976) and Klingeman and Emmett (1982), that two relationships are required for a more accurate mathematical description of gravel sediment transport as opposed to those approaches documented in Chapter 2. At high discharges, when the stream is competent to move almost the entire range of bed-material, the data conforms to the conditions required for most of the existing formulae. A second relationship needs to be developed for lower discharges, when the effects of bed armouring are most pronounced. Whilst stream competence may be high at such times, bed-material transport is limited by the availability of material through armour layer restriction. This provides some answer to the problems posed by armouring, once adequate techniques have been developed to quantify the processes (Gessler, 1973; Carling and Reader, 1982). However, further difficulties must be recognised with respect to the role played by medium-large scale bedorms which, as identified above, may store sediment for considerable periods of time under the contemporary fluvial regime.

Armour layer stabilisation, in addition to stable bedforms within channels must play a significant role in generating increased bank erosion during storm flows, particularly at those flows below the critical discharge for bed-load movement identified above where the excess capacity of the flow must be borne by the river banks. This is reflected in the source characteristics of suspended sediment samples collected for the catchment study area (Chapter 8). Within the piedmont reaches especially, the increased bed stability and larger areas of fine flood-plain sediments provides for considerably higher suspended loads as opposed to the uplands. Bearing in mind that data were only collected for a limited period, the analyses suggested that suspended sediment sources were stage related during storm events during the early periods of a winter flood season. However, for later events the characterisation of sources is made more difficult as a result of bank failure and the incorporation of a range of fine sediments from the complete bank-soil profile. Similar processes have been described by Thorne & Lewin (1979) and Thorne (1982). The suspended sediment - source characteristics of the catchments sediments reflects the local soil-bank sediment assemblages of the catchment, the dominant source throughout the study area reflecting channel-bank sediment contributions. However, the distinction between 'channel' or 'bank' origin at any one point in time was difficult with such a limited data set. Bearing in mind the processes outlined above it is feasible that suspended sediments, although initially derived from bank erosion, may be stored as a reservoir beneath the armour surface and released by the processes described by Milhous and Klingeman (1973), Reid et. al. (opp. cit.) and others, during increased flows. As a result, whilst a dominant bank source origin may be inferred from the analyses, the sediments themselves may not have been instantaneously derived from bankside erosion but may be a secondary load essentially channel derived. The nature of fine sediment supply processes, to (c.f. Nanson,1974) and from (Frostick et. al., 1985) the channel, requires much more detailed investigation. In particular, source identification in the manner described herein (see also section 9.4.) needs to be carried out in combination with channel planform surveys and detailed bank erosion measurements. In the uplands, such characteristics reflect those sources initially recognised by Lewin et. al. (1974). In the piedmont reaches, suspended sediment loads may be enhanced as a result of flow regulation which may increase channel armouring and reduce the already limited supply of channel fine sediments for suspension (c.f. Chapter 4.). Abnormal rates of erosion and deposition associated with regulation (Hey, in press) together with already present quasi-stable bedforms may excaserbate flow divergence and bank erosion. Further analysis of historic planform changes needs to be carried out to confirm this hypothesis.

9.3. Application of the Enhanced Magnetic Tracing Technique To Other Areas: Heating Trials with a Variety of Bed Material

The efficacy of any tracing technique must also be measured in its applicability to a variety of areas in addition to its applicability to a variety of sediment sizes and to the immediate problem at hand. With this in mind, enhancement trials with the method detailed in Chapter 3 were undertaken with bed materials other than those of the Plynlimon catchment area.

Samples of bed material were collected from a variety of areas, mainly where sediment transport research was already being undertaken. Five sites were used for the study:

i) Eggleshope Beck : an upland tributary of the River Tees which is monitored as part of a research catchment by the Freshwater Biological Association. The local geology consists of a mixture of peat, till and solifluction deposits over Carboniferous gritstones and limestones;

ii) Narrator Brook : an upland tributary of the River Plym on Dartmoor which is monitored as part of a catchment studies programme by Plymouth Polytechnic. The local geology, granite;

iii) Milliford Bridge : a tributary of the Lymington River in the New Forest, Hampshire and monitored by Southampton University. The river flows through superficial sediments over Oligocene and Eocene sands. Much of the bed material is made up of coarse flints;

iv) Formby sands : an area of coastal sands within Liverpool Bay monitored by Liverpool and Manchester Universities as part of an experiment into coastal sedimentation problems. The sands are derived from Triassic Kauper Marl and sandstone deposits; and,

v) Bodfari sand : a quarried sand from glaciafluvial deposits in Cheshire. This site was used to look into the possibility of using commerically available quarry sands for tracing purposes.

Heating trials were carried out on small crucible samples (approximately 20g. in weight) of seived material within the range -1 to 1 phi. As with the procedure documented in Chapter 3, the main aim of the experiments were to achieve as great an enhancement in magnetic susceptibility as possible. Heating trials were carried out at a variety

SITE	SIZE RANGE	χ_b	χ_e	χ_e/χ_b	T
	Ø	($10^{-6} m^3 Kg^{-1}$)			(^oC)
PLYNLIMON	-2.0	0.07	68.0	971	900
EGGLESHOPE BECK	-1.0	0.05	44.0	880	600
NARRATOR BROOK	0.0,-1.0	0.058	5.28	91	900
MILLIFORD BRIDGE	0.0,-1.0	0.047	47.8	1017	600
FORMBY SANDS	1.0	0.005	2.75	550	600
BODFARI SANDS	1.0	0.027	3.90	144	600

χ_b - Background magnetic susceptibility.
χ_e - Enhanced magnetic susceptibility following heating.
T - Optimum heating temperature.

TABLE 9.3. THE ENHANCEMENT OF MAGNETIC SUSCEPTIBILITY OF BED SEDIMENT MATERIAL FROM OTHER AREAS.

FIGURE 9.1. ENHANCEMENT OF MAGNETIC SUSCEPTIBILITY IN BED-MATERIAL FROM OTHER LOCATIONS :

A) NARRATOR BROOK, DARTMOOR;

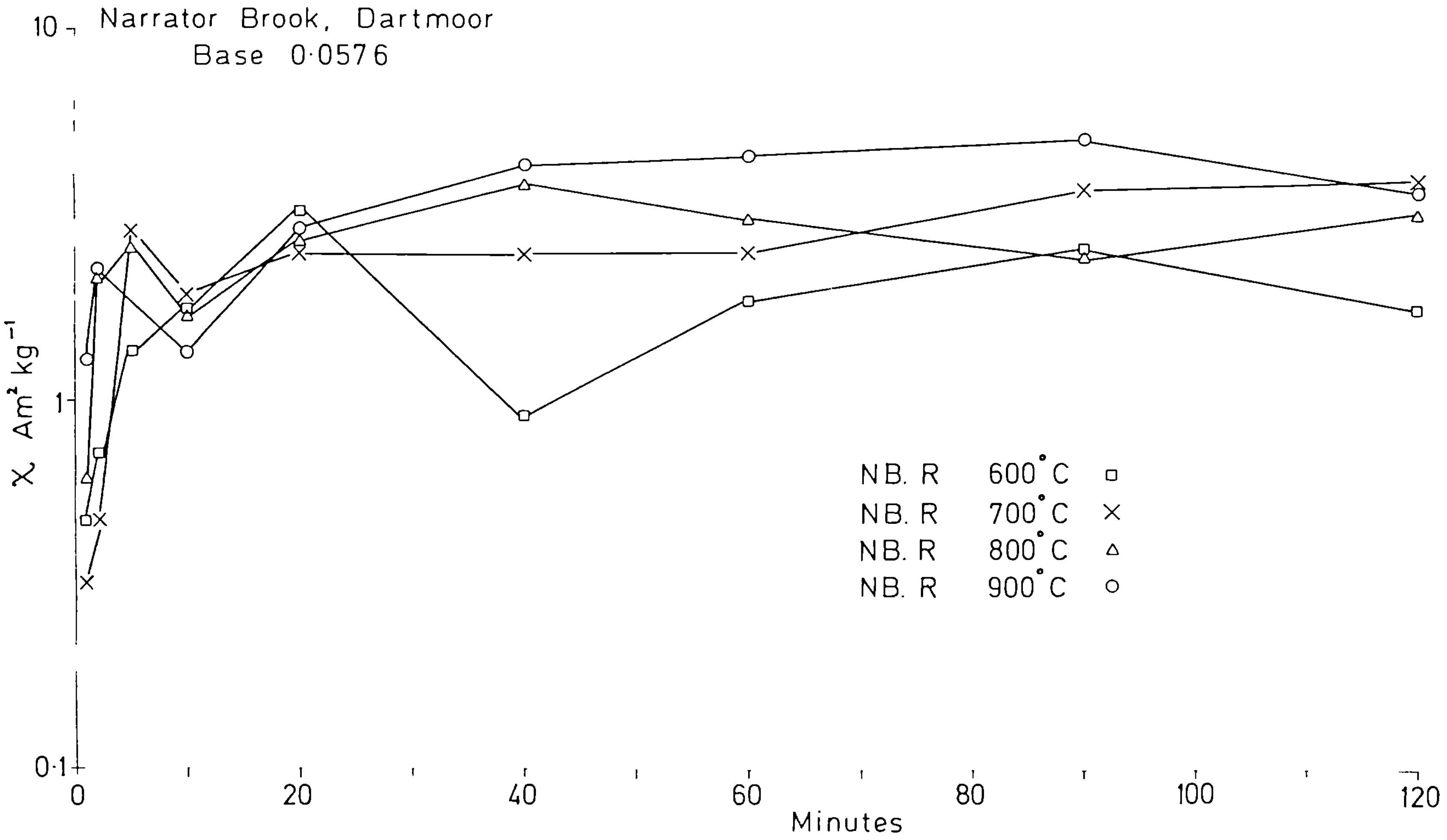
B) EGGLESHOPE BECK, TEESDALE;

C) LIVERPOOL BAY;

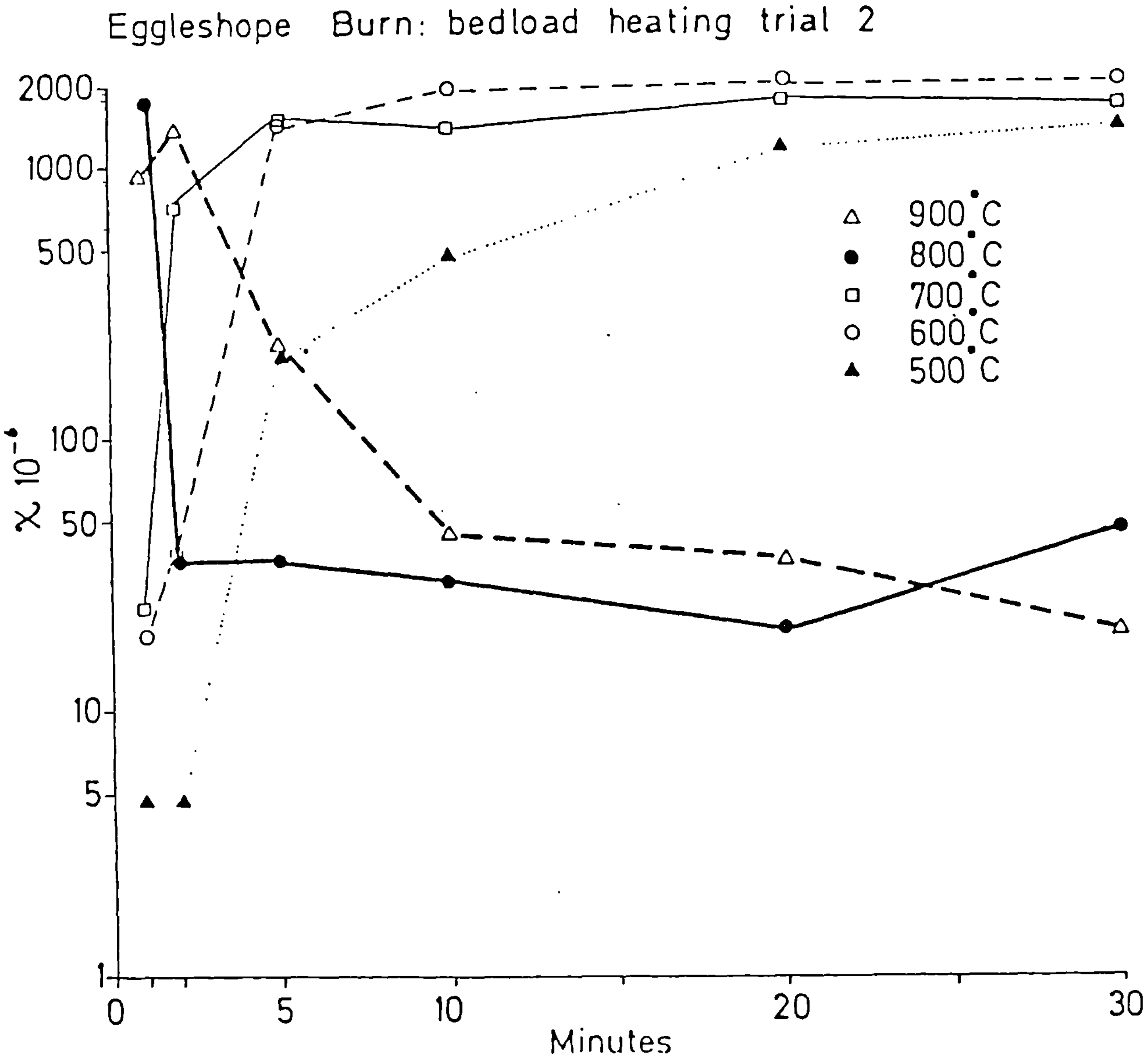
D) BODFARI SAND, CHESHIRE; AND,

E) MILLIFORD BRIDGE, NEW FOREST, HAMPSHIRE.

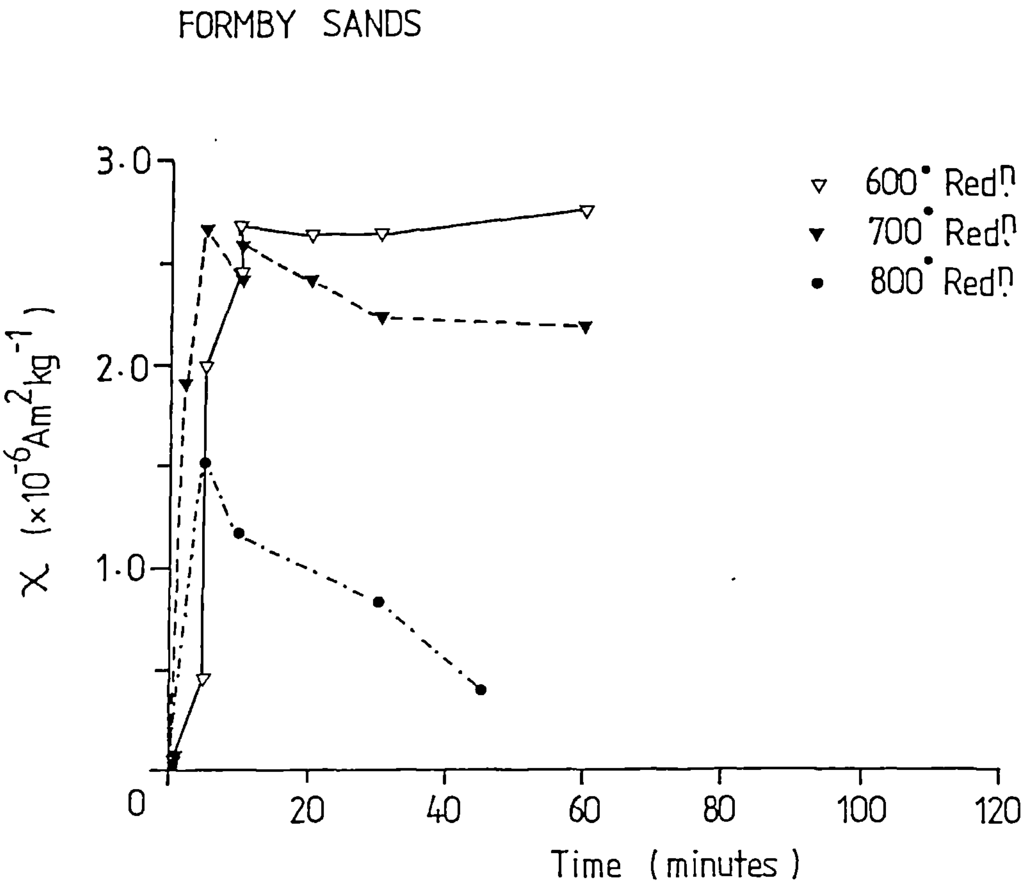
✱ Red^n denotes reducing atmosphere


Narrator Brook, Dartmoor
Base 0·0576
10
1
0·1
χ Am² kg⁻¹
0
20
40
60
80
100
120
Minutes
NB. R 600°C □
NB. R 700°C ×
NB. R 800°C △
NB. R 900°C ○

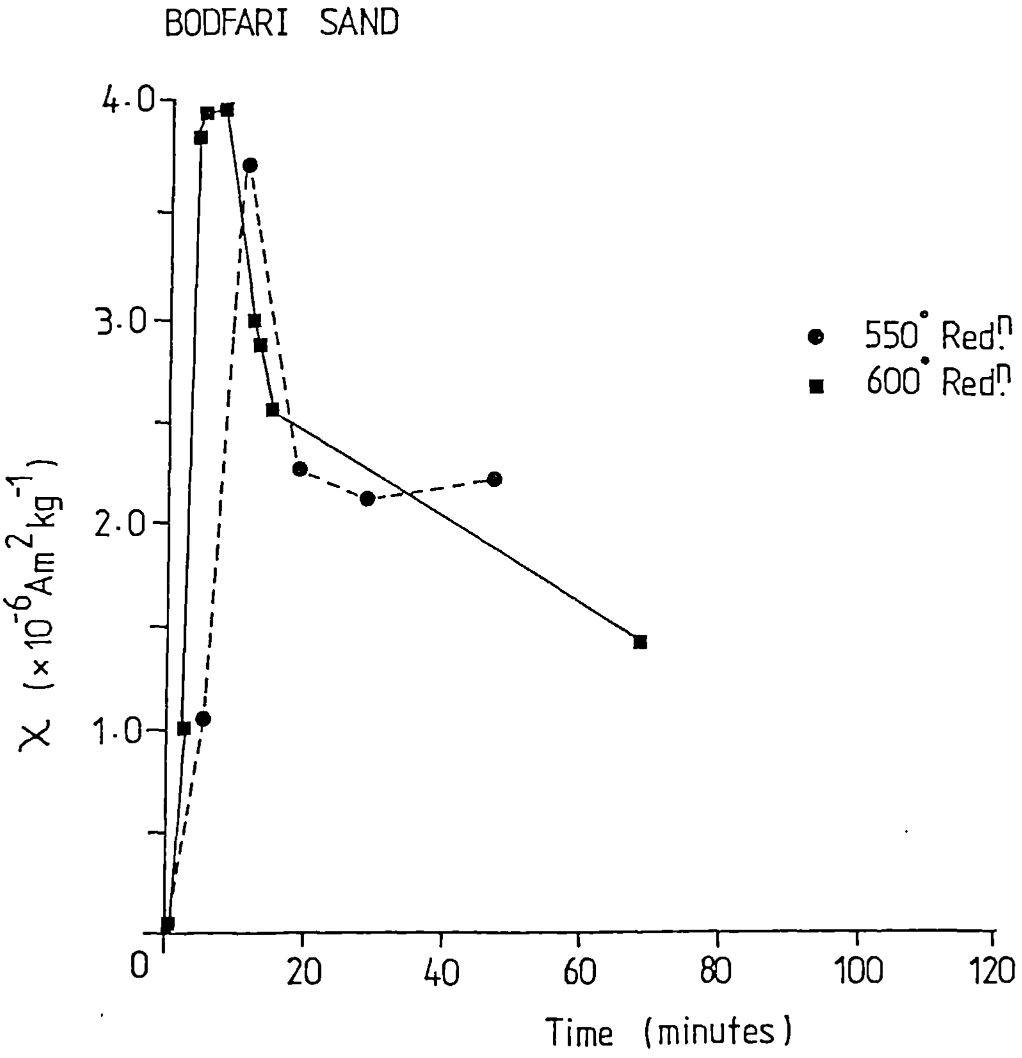
A.


Eggleshope Burn: bedload heating trial 2
2000
1000
500
100
50
10
5
1
χ 10⁻⁶
0
5
10
20
30
Minutes
△ 900°C
● 800°C
□ 700°C
○ 600°C
▲ 500°C

B.

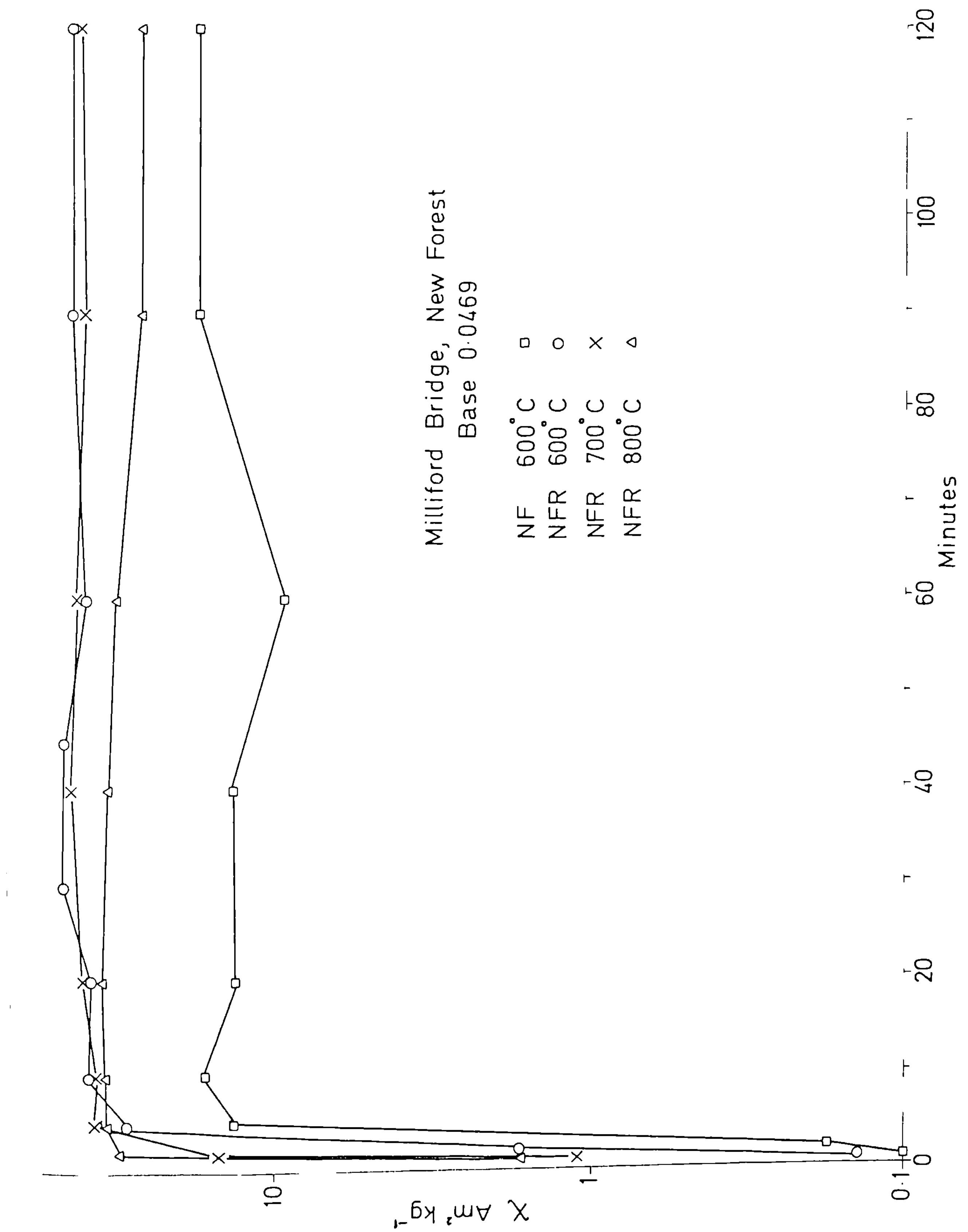

FORMBY SANDS
3·0
2·0
1·0
X (×10^-6 Am^2 kg^-1)
0
20
40
60
80
100
120
Time (minutes)
▽ 600° Redⁿ.
▼ 700° Redⁿ.
● 800° Redⁿ.

C.


BODFARI SAND
4·0
3·0
2·0
1·0
X (×10^-6 Am^2 kg^-1)
0
20
40
60
80
100
120
Time (minutes)
● 550° Redⁿ.
■ 600° Redⁿ.

D.

E.


Milliford Bridge, New Forest
Base 0·0469
NF 600°C □
NFR 600°C ○
NFR 700°C ×
NFR 800°C △
Minutes
0
20
40
60
80
100
120
χ Am² kg⁻¹
0·1
1
10

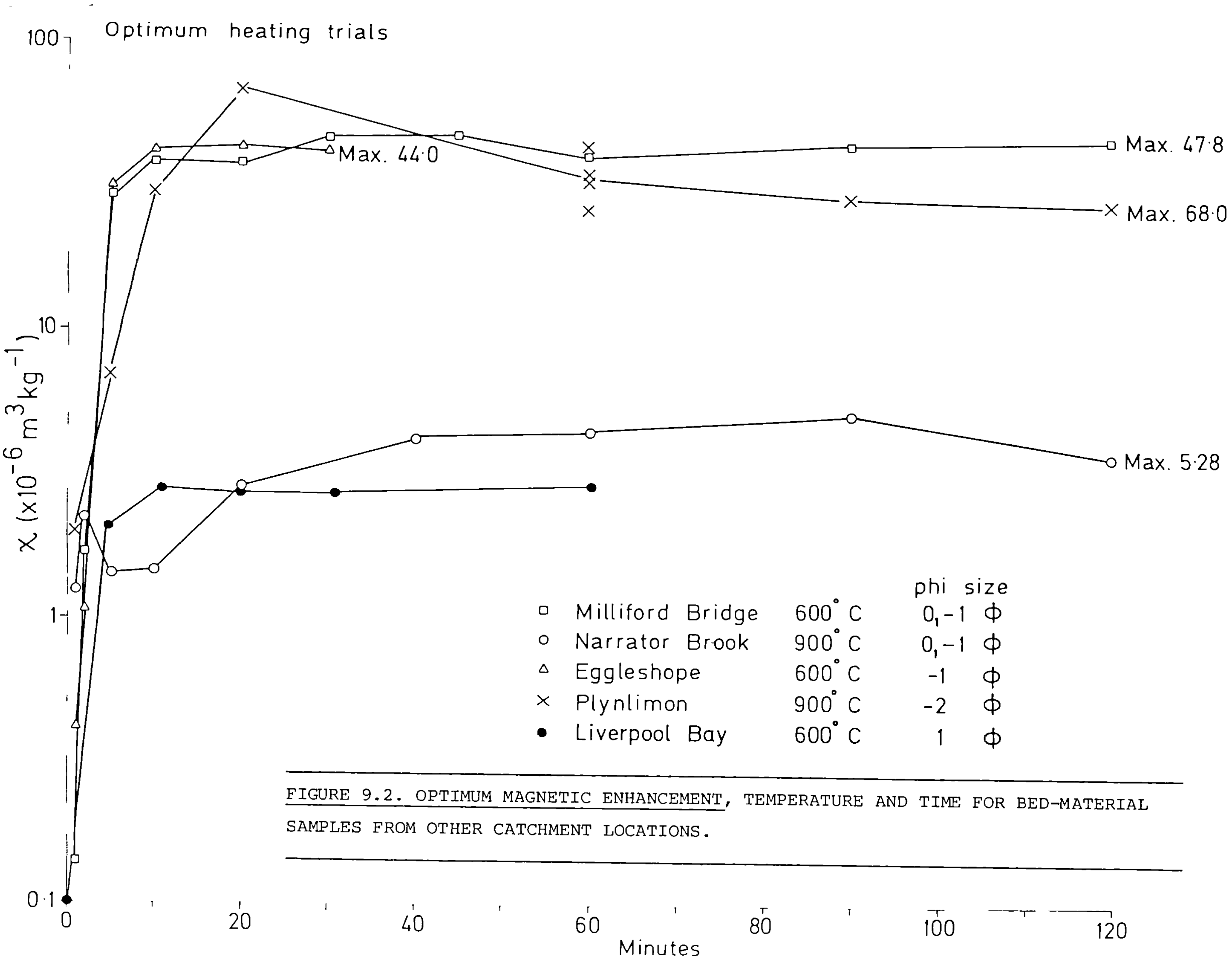


FIGURE 9.2. OPTIMUM MAGNETIC ENHANCEMENT, TEMPERATURE AND TIME FOR BED-MATERIAL SAMPLES FROM OTHER CATCHMENT LOCATIONS.

of temperatures up to 950°C with the material alone and with the material mixed with a 10% by weight addition of anthracite and flour to provide a reducing atmosphere whilst heating. The results are presented in Figures 9.1. and 9.2.. Table 9.3. shows the maximum enhancement achieved for each sample related to similar experiments on the Plynlimon bed material.

In practise, the actual enhancement achieved for the tracer material used in the field trials documented above (Chapters 5 to 7) was between 100 to 150 times the background susceptibility, although the laboratory experiments yielded a much higher value. Using 100-150 as an arbitrary cut off point, the data in Table 9.2. would suggest that the technique is applicable to many lithologies. However, when the material is largely silica as sand grains (Bodfari material) or quartz (Narrator Brook material), the success of the enhancement procedure depends upon the inclusion of some iron mineralogy on the irregular surface of the silica grain.

9.4. Suggestions for Further Research

9.4.1 Bedload Tracing

1) Detection: One of the main constraints of the technique highlighted above was the performance and sensitivity of instrumentation when used to locate magnetic tracer material at depth in a sediment mass. Similar problems were also identified by Ergenzinger and Conrady (1983), and Reid et.al (1984). Variations in surface susceptibility reflect both the sensitivity of the search coil and the nature and geometry of the magnetic tracer being located; small magnetic tracer clasts located close to the rim of the search coil may produce as high a signal as a larger pebble towards the centre of the coil or some way from it. This is largely a technical problem and may be overcome by using an instrument design similar to that

of commercially available underground pipe detectors, the design of which is based upon much larger, paired inductance coils. An example of the use of such instrumentation may be seen in Hassan et. al. (1984) but unlike the instrumentation described herein, there is no numeric output, merely an audible located sound device.

2) Insitu Monitoring : A more dynamic approach to monitoring tracer movement within a flood may be made by installing detection devices within the streambed as reported by Ergenzinger and Conrady (opp. cit.) and by Reid et. al. (opp. cit.). The instrumentation will need any such developments made in line with suggestion 1) above.

Such an approach would be fairly simple to adopt with installations such as the Oak Creek and East Fork River bed-load sampling devices (Klingeman and Emmett, 1982) or indeed smaller bedload traps, to provide data of particle size related to instantaneous hydraulic conditions, in particular the condition for incipient motion or, the variations of stream power to bed-load transport rate as outlined above (section 9.2.) and by Klingeman and Emmett (opp. cit.). From the conclusions made above, it is apparent that such a approach would still need to be integrated with spatial surveys of tracer concentration and topographic surveys to identify eroding and depositional sequences. Such a project is currently being undertaken by the Institute of Hydrology, Plynlimon (Newson, pers. comm.).

9.4.2 Suspended Sediment Source Identification

1) Suspended Sediment Sample collection : two aspects are worthy of attention here. Firstly, one major constraint in the analyses made in Chapter 8 occurred as a result of the low weights of retained filtrate sediment which precluded any measurements of susceptibility and related

parameters (e.g. quadrature Susceptibility) or accurate determination of SIRM or IRM_{3000} for a range of filtrates. A minimum filtrate weight of 0.03g is recommended for future researches for comparable SIRM data ie. larger volumes of stream water may need to be collected. This may need to be increased to encompass susceptibility measurements, although the requirements are clearly catchment source dependent.

For low sediment filtrate weights, impurities in conventionally used glass-fibre filter papers may adversely affect any magnetic measurements. Several other filter bases are available which are guaranteed free of impurities and should be used where local suspended sediment concentrations are low.

2) Suspended Sediments, Source Areas and Particle Magnetic Grain Size:
A major development for any future work of this nature must be in establishing variations in magnetic grain size (c.f. Chapter 3) within and between substrates and for particle size ranges within each substrate. Temporal variations in particle size have been reported by Peart and Walling (1982) and may be of significance to the analyses made in Chapter 8. In addition, suspended sediments derived from sources which have undergone or are undergoing secondary enhancement processes may have considerable variations in magnetic grain size for the same particle size. Both avenues need to be investigated before an accurate determination of source type and variation can be made.

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