The Noon Hill flash floods; July 17th 1983. 
Hydrological and geomorphological aspects 
of a major formative event in an upland 
landscape

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ABSTRACT
The meteorological conditions, pertaining to thunderstorm activity over the north Pennines, which resulted in a major flash-flood are described briefly. The magnitudes of the peak flood discharges generated in three small headwater catchments are estimated using a combination of the slope-area method, culvert geometry, a modified Manning’s equation and boulder transport criteria. The return intervals for the rainfall intensity and the channel discharges are calculated.

The main geomorphological features associated with the flood flows are described. Values of peak shear stress associated with valley sections are estimated and values of the energy expenditure per unit catchment area are proposed. These latter values pertain to thresholds for the generation of apparently ‘unique’ landscape features which may be diagnostic of similar formative events in the Pennine uplands.

The calculated return period for the rainfall intensity and an estimation of the probable recovery period are used to discuss the sensitivity of the Pennine upland landscape to modification by rare high-magnitude events.

Conclusions are drawn as to the role of catastrophic non-uniformitarian precepts in models of upland landscape formation.

KEY WORDS: Floods, Meteorology, Hydrology, Geomorphology, Thresholds, Landscape sensitivity, Recovery period, Catastrophic models, United Kingdom

INTRODUCTION
The basic characteristics of precipitation events which lead to flooding in Britain have been reviewed by Newson (1975) who tabulates a series of references to major floods of which the best known are probably the Lynmouth and various Moray area floods. In this context the West Country and northeastern Scotland seem particularly prone to record rainfalls and associated flooding. Nevertheless, there is a long history of flooding in other upland areas including the northern Pennines. However, for the latter area, reports are few, brief and concentrate on the economic and social effects in the populated coastal or piedmont regions; there being little detailed flood data for sparsely inhabited headwater catchments. In these latter areas economic losses generally are limited but the steep slopes and narrow valleys concentrate floods so that limited damage occurs frequently and the geomorphological response may be spectacular.

Although persistent rain and snowmelt may induce flooding, there is a long tradition of exceptional upland Pennine floods associated with severe thunderstorm activity (Gilligan, 1909; Brooks and Glasspoole, 1928; Duckworth and Seed, 1969; Harvey, in press); possibly intensified by orographic effects (Miller, 1951). However, the geomorphological effects of thunderstorms in the Pennines are poorly documented.

This paper records unusual thunderstorm activity and heavy rainfall of a 'very rare' intensity (Bilham’s 1935 classification). The storm occurred on the 17 July 1983 on the watershed between upper Teesdale and Weardale. Consideration of the geomorphological response is restricted to the channel network and valley bottoms although associated with the
event were major peat landslides which have been described elsewhere (Carling, in press).

Catchment characteristics
Three small catchments were affected by the storm; Langdon Beck, Ireshope Burn and West Grain (Fig 1, Table I). The geology is Carboniferous limestones, shales and sandstones overlain by diamicton and solifluxion clays. The cobble- and boulder-bedded streams are shallow, only a few metres wide and in the upper reaches are closely confined by the valley walls. Further downstream limited floodplain development occurs where valley walls diverge; typically where tributary streams join the mainstem. In the lower courses, channel gradient is reduced and lateral deposition is more extensive as the floodplain broadens to 50–300 m in width. The floodplain is constructed of 1–2 m of coarse gravels overlain by up to 1 m of fine silts and locally these deposits are being reworked by stream meandering.

Wilkinson (1971) gives the mean annual rainfall for the Burnhope catchment (Fig. 1) as 1506 mm and estimated the rainfall in the West Grain catchment as 1618 mm per year. There are no data available for the other catchments but it is unlikely that major differences exist.

The vegetation is rough grazing and heather moorland developed over blanket peat.

METEOROLOGY
In 1983, May was unusually wet (the monthly total at Ireshope Plains being 1414 mm), but was followed by a generally dry summer punctuated by localized intense rainstorms. There was no flooding on the R. Tees or Wear but the Northumbrian Water Authority (NWA) recorded seven upland flooding events in tributary streams between the 7th of May and the 25th of July. Several of these floods were associated with intense thunderstorms. For example, at Hartburn on the 22nd of June, 38 mm of rain fell within 1 1/4 hours, (NERC Flood Studies Report 1975—Return Period ~ 60 years) and on the 12th July 80 mm fell over a 2-hour period at West Woodburn (Return Period ~ 1400 years). The most exceptional event occurred on the 17th of July.

On this latter occasion skies were hazy all day. Light westerly winds shifted to the north-west as the day progressed, and the air temperature rose from a minimum of about 14°C at 03.00 GMT to 24°C by 15.00 GMT. The air pressure fell slowly and steadily all day from 1016-0 mb at midnight to 1010-1 mb at 15.00 GMT. Associated with the westerly wind and pressure drop were altostratus and cumulonimbus cloud developing west of Cumbria from 12.00 GMT and thunderstorms developed on the western flanks of the Pennines. The generally calm conditions (regional wind speed <5 knots) were particularly conducive to rapid formation of deep convective cloud. Strong updraughts and down-draughts, normally associated with this type of weather, were noted at Great Dunn Fell where although the daily wind speed was only 8 knots a maximum gust of 41 knots was recorded. Rainfall was generally light, punctuated by localized heavy falls as described below. There were two centres of intense convectional rainfall, one cell centred over the headwaters of the West Allen and the other over Noon Hill, on the watershed between Weardale and Teesdale (Fig. 1). This paper is concerned with the latter area alone. Heavy rainfall (and some hail) was extremely localized. For example whilst a rain gauge at Greenhills recorded only 12-7 mm in 24 hours, 104-8 mm fell at Ireshope Plains in the period 14.30

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**TABLE I. Basin characteristics**

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Drainage area</th>
<th>Lemniscate ratio</th>
<th>Est. Mean annual flood</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>Langdon Beck</td>
<td>5.93 km²</td>
<td>2.76 km km⁻²</td>
<td>0.27 m³ s⁻¹</td>
<td>35.70</td>
</tr>
<tr>
<td>Ireshope Burn</td>
<td>7.14 km²</td>
<td>3.95 km km⁻²</td>
<td>0.83 m³ s⁻¹</td>
<td>37.28</td>
</tr>
<tr>
<td>West Grain</td>
<td>1.86 km²</td>
<td>3.92 km km⁻²</td>
<td>0.35 m³ s⁻¹</td>
<td>89.97</td>
</tr>
</tbody>
</table>

1 Catchment area above bridge at NGR NY 851330
2 Catchment area above NGR NY 866387
3 Catchment area above West Grain Bridge NGR NY 875372
to 17.00 GMT; the observer stating that the greater part fell within a core period 14.45 to 16.00 GMT. Rainfall was first recorded autographically at Burnhope Reservoir at 14.24 GMT. Unfortunately the recorder failed but a check gauge recorded 87 mm over some 3 hours. This latter station was close to the edge of the intense rainfall as recorded runoff to the reservoir from the Burnhope catchment was small.

Such rainfall totals are relatively unusual (Jackson, 1974). The return period for 104.8 mm over 2 hours is 2500 years and for 87 mm is 400 years (NERC, 1975) whilst Bilham’s (1935) method yields return periods of between 500 and 200 years. Rodda (1970) noted that although a fall of 100 mm over 24 hours has a UK average return period of 100 years, return periods may be significantly shorter in upland areas. Despite these and other uncertainties in
determining precise return periods, it is interesting to note that the 104.8 mm total falls close to Rodda's (1970) curve for maximum falls of rain for given durations in the UK.

HYDROLOGY

Only Langdon Beck is gauged effectively by the NWA at Langdon Beck village (catchment area ~13 km²). In Ireshope Burn and West Grain, man-made structures divert a proportion of the discharge through pipelines to Burnhope Reservoir. Although records of pipeflow are kept, these are not easily related to stream discharge (Wilkinson, 1971). Wilkinson estimated discharges for a two-year period on the West Grain, the smallest of the three streams. He recorded a dry weather discharge of 3–6 l s⁻¹ with storm peak discharges ranging between 0.62 and 1.9 m³ s⁻¹ for storm durations of 51–110 hours. The median summer discharge was 17 l s⁻¹. The streams are all flashy, and Ireshope Burn probably has a similar response and frequency of magnitude of flows as Langdon Beck (Fig. 2). Estimated mean annual floods are given in Table I; these are based on the six parameter catchment equation in the Flood Studies Report (NERC, 1975).

During the storm stream levels rose from base flow to peak in less than 15 minutes and high flows persisted for only two to three hours. Rapid rises persisted downstream although peak stage in the main rivers diminished rapidly downstream so that return periods for the flood in the main rivers were small.

Despite the occurrence of large peat landslides in each catchment, little peat was delivered to West Grain and Ireshope Burn. In contrast, a peat slide of some 30,000 tonnes wet weight was funnelled into the headwaters of the Langdon Beck. The peat, with an initial bulk density of 1.05 t m⁻³, broke up rapidly forming a fine slurry. Initial concentrations of peat there, are estimated as 200 g l⁻¹ or 13 per cent by volume. In terms of the peat concentration, the flow would be classified as a water flood (Costa, 1984) and the peat debris would be expected to have had a minimal effect on hydraulic calculations in the low gradient reaches downstream of Langdon Common Bridge (Location B, Fig. 1).

Owing to the gauging structure being flooded out and the tapping point partially obstructed, the gauged peak discharge in Langdon Beck (Location A, Fig. 1) of 35 m³ s⁻¹ is believed to be a slight overestimate. The recorded hydrograph displayed a practically instantaneous rise to peak discharge. The rapid rise and single peak conform to eye-witness accounts. The flood wave returned to low flows within five hours of peaking; the recession curve for the first three hours being well described by the simple exponential function

\[ Q_t = Q_p e^{-nt} \]  

where \( Q_t \) is discharge at time (t) and \( Q_p \) is the peak discharge.

The gauged peak discharge was checked using the slope-area method (Benson, 1968). For this method the selection of a value of Manning's 'n' is critical (Jarrett, 1984). On the same day a flood was generated in the neighbouring Harwood Beck which is morphologically very similar to Langdon Beck. The peak discharge was recorded accurately at the Harwood gauging station, but a slope-area survey was also conducted. A Manning's 'n' value of 0.05 was selected to match the slope-area estimate to the gauged discharge. Using this value of 'n', the peak discharge in Langdon Beck was estimated independently by two teams (FBA and NWA) as 29 m³ s⁻¹.

Witnesses of the floodwave variously described a 'wall of water carrying peat blocks' overtopping Langdon Common Bridge. Trashline surveys demonstrated that the flow backed up to a depth of +2.12 m and overtopped the bridge structure owing to the constriction imposed by the culvert on the flood wave. Occupants of Valence Lodge were brought from the house by the sound of the approaching flood wave and described the flow as a 'wall of mud with the appearance and consistency of chocolate sauce'.

The section beneath Langdon Common bridge is a box-section rough concrete culvert. Field survey indicated Type V or VI flow (Benson, 1968) through the culvert and discharge calculations using culvert geometry indicated a discharge of 24–27 m³ s⁻¹. Allowing for unmeasured flow over the road of a few cumecs this estimate agrees well with that made at the gauging station, and indicates little downstream attenuation of the flood crest occurred between the culvert and Langdon Beck village.

The return period for a flood of c. 29–35 m³ s⁻¹ from the 13 km² Langdon Beck basin is not great; about 8–20 years (Fig. 2). Comparable magnitude events have occurred in the past, for example in January 1976. The difference is noted, however, in the amount of geomorphological work accomplished. In January 1976 the flood resulted
Effects of flash floods: Noon Hill, 1983

Only some 3.5 km$^2$. A c. 29 m$^3$ s$^{-1}$ discharge of water and peat is therefore equivalent to 8.3 m$^3$ km$^{-2}$ s$^{-1}$.

To confirm the large discharge from a small area, field survey was conducted in the upper reaches of the catchment. The cross-sectional area of the flood was measured every 200 m and the gradient of the left bank trashline surveyed over a 2 km reach upstream of Langdon Common bridge (Table II, Fig. 3). In addition an extra section was surveyed (section 7) at a bedrock control point. Although the standard slope-area method had worked well at the gauging station (channel slope $\sim 0.57^\circ$), it could not be applied accurately to the steeper channel reaches (Fig. 3). Jarrett’s (1984) regime equations based on a modified slope-area method gave good results at the low gradient sections 1 and 3 and at the bedrock

![FIGURE 2. Partial duration series for recorded flood peaks in Langdon Beck between 1969 and 1983. The analysis is based on a normal distribution utilizing Blom’s plotting position. The close square represents the slope-area estimate for the present discharge whilst the open square represents the gauged peak discharge from snowmelt and persistent rain and the whole catchment contributed to the discharge. The geomorphological work accomplished was visually insignificant. In contrast, the level of the trashlines and the amount of channel scour indicated that practically the entire discharge, of the present event, was generated in the Langdon Head area above the confluence with the tributary stream; West Beck. The catchment area of the Langdon Head basin is](image1)

![FIGURE 3. Section locations in Langdon Beck corresponding to Table II. Each section is drawn with the true right bank on the left of the figure. Horizontal lines indicate the distance over which slopes in Table II were calculated. Shaded areas in sections represent approximate bankfull conditions where these could be ascertained](image2)

<table>
<thead>
<tr>
<th>Section</th>
<th>Distance (m)</th>
<th>Width (m)</th>
<th>Maximum depth (m)</th>
<th>Mean depth (m)</th>
<th>Hydraulic radius (m)</th>
<th>Area (m$^2$)</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>2000</td>
<td>50.6</td>
<td>0.95</td>
<td>0.55</td>
<td>0.54</td>
<td>27.83</td>
<td>0.0194</td>
</tr>
<tr>
<td>2</td>
<td>1800</td>
<td>52.3</td>
<td>1.41</td>
<td>0.45</td>
<td>0.44</td>
<td>23.34</td>
<td>0.0215</td>
</tr>
<tr>
<td>3</td>
<td>1600</td>
<td>37.4</td>
<td>0.89</td>
<td>0.64</td>
<td>0.62</td>
<td>23.82</td>
<td>0.0272</td>
</tr>
<tr>
<td>4</td>
<td>1400</td>
<td>43.9</td>
<td>1.75</td>
<td>1.45</td>
<td>1.36</td>
<td>63.89</td>
<td>0.0305</td>
</tr>
<tr>
<td>5</td>
<td>1200</td>
<td>42.0</td>
<td>2.13</td>
<td>0.81</td>
<td>0.78</td>
<td>33.96</td>
<td>0.0168</td>
</tr>
<tr>
<td>6</td>
<td>1000</td>
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<td>0.92</td>
<td>32.52</td>
<td>0.0376</td>
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<tr>
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<td>2.05</td>
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<tr>
<td>8</td>
<td>800</td>
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<tr>
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</tr>
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<td>400</td>
<td>31.3</td>
<td>2.13</td>
<td>1.08</td>
<td>1.01</td>
<td>33.86</td>
<td>0.0459</td>
</tr>
<tr>
<td>11</td>
<td>200</td>
<td>26.4</td>
<td>3.70*</td>
<td>1.71*</td>
<td>1.52</td>
<td>45.24</td>
<td>—</td>
</tr>
</tbody>
</table>

* Depths reflect a degree of bedrock erosion
control point, section 7. Using the data in Table 2, Manning's 'n' was estimated as 0.087 and $Q_p = 34 \text{ m}^3 \text{s}^{-1}$. Excessively high and variable discharge estimates were obtained at other sections which may relate to the unsteady nature of the heavily sediment-laden flood wave in these upper reaches.

The normal slope-area method was also found to be inapplicable in the case of Ireshope Burn and West Grain. To estimate discharges in the latter catchments boulder competence calculations were used in conjunction with an empirically modified Manning's equation which was initially tested against the known flood discharge in Langdon Beck.

### Boulder competence calculations

In the downstream reaches of each catchment, the three major axes of the five largest boulders transported by the flood were measured (Table III). For Langdon Beck, data were used with published procedures (Table IV) to calculate the competent velocity. The competent velocity was taken as equivalent, in shallow flood flow, to the sectionally-averaged velocity (Baker, 1973) and the discharge

### Table III. Largest boulder data (m)

<table>
<thead>
<tr>
<th></th>
<th>Long (m)</th>
<th>Intermediate (m)</th>
<th>Short (m)</th>
<th>(LIS)$^4$ 3</th>
<th>Cory Shape Factor</th>
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</thead>
<tbody>
<tr>
<td><strong>Langdon Beck</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.40</td>
<td>1.00</td>
<td>0.43</td>
<td>0.84</td>
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<tr>
<td>1.37</td>
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<td>0.68</td>
<td>0.95</td>
<td>0.61</td>
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<td>1.10</td>
<td>0.80</td>
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<td>0.60</td>
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<tr>
<td>0.87</td>
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<tr>
<td>$\bar{x} = 1.24$</td>
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<td>0.55</td>
<td>0.82</td>
<td>0.54</td>
<td></td>
</tr>
<tr>
<td><strong>Ireshope Burn</strong></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.15</td>
<td>0.80</td>
<td>0.65</td>
<td>0.84</td>
<td>0.68</td>
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<tr>
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<tr>
<td>0.80</td>
<td>0.60</td>
<td>0.60</td>
<td>0.66</td>
<td>0.87</td>
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<tr>
<td>1.80</td>
<td>0.55</td>
<td>0.30</td>
<td>0.67</td>
<td>0.30</td>
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<tr>
<td>$\bar{x} = 1.09$</td>
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<td>0.48</td>
<td>0.68</td>
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<tr>
<td><strong>West Grain</strong></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>1.65</td>
<td>1.25</td>
<td>0.37</td>
<td>0.91</td>
<td>0.26</td>
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<td>1.18</td>
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<tr>
<td>$\bar{x} = 1.39$</td>
<td>0.99</td>
<td>0.62</td>
<td>0.92</td>
<td>0.55</td>
<td></td>
</tr>
</tbody>
</table>

### Table IV. Comparative discharge estimates for Langdon Beck using boulder data

<table>
<thead>
<tr>
<th>Method</th>
<th>Boulder data</th>
<th>Reference</th>
<th>Velocity ($m \text{s}^{-1}$)</th>
<th>Discharge ($m^3 \text{s}^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Initial overturning</td>
<td>Semi-axes</td>
<td>(Helley, 1969)</td>
<td>3.87</td>
<td>90</td>
</tr>
<tr>
<td>(theory)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Initial sliding</td>
<td>Semi-axes</td>
<td>(Mears, 1979)</td>
<td>3.38</td>
<td>79</td>
</tr>
<tr>
<td>(theory)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Initial transport</td>
<td>$d_s$ intermediate axes</td>
<td>(data from Fig 4 Carling, 1983)</td>
<td>1.20</td>
<td>28</td>
</tr>
<tr>
<td>(empirical)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Initial motion/transport</td>
<td>$d_s$ intermediate axes</td>
<td>(Costa, 1983, equation 8)</td>
<td>4.35</td>
<td>101</td>
</tr>
<tr>
<td>(empirical)</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>

**Note:** Manning's equation was initially tested against the known flood discharge in Langdon Beck.
FIGURE 4. (a) Channel reach in West Grain from which all deposits have been scoured-out; (b) Boulder jam in the West Grain; (c) Chute cut through overbank silts and gravels in Langdon Beck; (d) Boulder berm formed in flow-separation zone at a valley-wall expansion in the West Grain (flow right to left); (e) Boulder jam responsible for stepped-bed morphology in Langdon Beck.
estimated using the continuity equation \( Q = AU \).
Each estimate was evaluated by comparison with the gauged discharge (35 m\(^3\) s\(^{-1}\)) and the slope-area discharge (29 m\(^3\) s\(^{-1}\)). All the methods yielded excessive discharges, the most accurate being too great by a factor of two to three (Table IV).

An alternative and, in this case, accurate method is that derived from field data collected in a similar Teesdale stream which yields an estimate 3 per cent less than the slope-area method. The procedure is based on calculating the critical shear stress for the mean boulder data from an equation derived from data given by Carling (1983, Fig 4).

\[
\tau_0 = 52.37 \, d_{5}^{1/3}
\]  
(2)

where \( \tau_0 = N \, m^{-2} \) and \( d_{5} \) is measured in metres.

The Duboy's relationship for shear stress on the channel bed is

\[
\tau_0 = \gamma RS
\]  
(3)

Manning's equation is

\[
\bar{U} = \frac{R^{2/3}}{S^{1/2}} \, n^{-1}
\]  
(4)

and the continuity equation is

\[
Q = AU
\]  
(5)

By simultaneously solving equations (2)–(4) the variables \( \tau_0 \) and \( R \) were eliminated to give \( U \) in terms of \( d_{5}, S, \gamma \) and \( n \) only.

\[
\bar{U} = 14 \, d_{5}^{-2/9} \, \gamma^{-2/3} \, S^{-1/6} \, n^{-1}
\]  
(6)

Assuming the peat load had little effect on the stream hydraulics in the downstream reaches, \( \gamma \) was set equal to 9807 N m\(^{-3}\) and considering equation (5), the discharge was calculated from

\[
Q = 3.06 \times 10^{-2} \, A \, d_{5}^{-2/9} \, S^{-1/6} \, n^{-1}
\]  
(7)

where \( A \) is the surveyed cross-sectional area, \( d_{5} \) is the mean boulder diameter, \( S \) is the water surface slope over the reach and 'n' was taken as 0.05. The cross-sectional area is measured in square metres and the boulder diameter in metres.

The revised method was applied in Ireshope Burn and West Grain with apparent success (Table V). The values, i.e. 30 and 16 m\(^3\) s\(^{-1}\), are believed to closely represent the discharge in the ungauged catchments being more likely to be slight underestimates rather than over-estimates.

All the published methods applied to the boulder data appeared to over-estimate the competent velocity and hence the discharge in Langdon Beck. The fact that large boulders in shallow and steep streams apparently move at less than theoretical critical values has been documented previously in the UK (e.g. Fearnsides and Wilcockson, 1928; Miller, 1951; Carling, 1983).

### Time of concentration

The time of concentration of the floodwave in Langdon Beck can be approximated. An eyewitness reported the peat-laden flood wave reached Langdon Common Bridge at 15.00 GMT ± 10 mins. As the first rainfall was recorded at 14.28 GMT a maximum of about 42 minutes were available from the first rainfall to the flood wave reaching the bridge.

The average celerity through downstream reaches was calculated as c. 1.2 m s\(^{-1}\) (Table IV) and the calculated travel time is in accord with the arrival time at Langdon Beck village (3.3 km downstream from Langdon Common Bridge) where the flood peaked shortly before 16.00 GMT.

The velocity coupled with the typical depths in downstream reaches (Table II) indicate that reach-averaged Froude Numbers (\( \bar{U}^{2}/gD \)) are subcritical. This result, although possibly initially surprising, is not incompatible with a recent compilation of Froude Number data for flow regime in steep channels (Jarrett, 1984).

### Geomorphic Effects

Newson (1980) divides the geomorphic response to major floods into slope effects and channel effects. The slope effects within the catchments consisted primarily of major peat landslides.

Within the upper reaches of Langdon and West Grain catchments the existing channel deposits were completely evacuated (Fig 4a). Severe erosion of the shale bedrock occurred and boulders were plucked from the jointed limestone. The till slopes adjacent to the channels were undermined, cracking

<table>
<thead>
<tr>
<th>Table V. Estimated and recorded discharges</th>
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<tbody>
<tr>
<td>Langdon Beck</td>
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<tr>
<td>Adventurism and recorded discharges</td>
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<tr>
<td>------------------------------------------</td>
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<tr>
<td>Modular rating curve</td>
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<tr>
<td>35</td>
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<td>34</td>
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<td>22</td>
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<tr>
<td>Slope/Area method</td>
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<tr>
<td>29–34</td>
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<tr>
<td>Culvert geometry</td>
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<td>24–27</td>
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<tr>
<td>Boulder transport</td>
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<td>28</td>
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<tr>
<td>30</td>
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<td>16</td>
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Effects of flash floods: Noon Hill, 1983

Developed in the turf and gravel fans developed where gullies debouched on the hillside or in the valley bottom.

Downstream, where the valleys widened or were obstructed by diversion dams, gravels were deposited infilling the channels and narrow valley floor to a depth of 1 m. This effect was most well developed in West Grain where erosion and deposition reaches alternated. A conservative estimate of 10000 tonnes of gravel deposited in the West Grain valley was made from field survey data.

At some locations the gravel infilling the channel had been elevated above the bank level and, at expansions in the valley width, formed boulder berms (Costa, 1984; Jarrett and Costa, 1985) (Fig 4d). These are characteristic of flood flows overloaded with coarse bedload (e.g. Scott and Gravlee, 1968). In the West Grain, large boulder-jams developed similar to those described by Krumbein (1942) (Fig 4b), and detailed sedimentological investigation indicates that these may be low-viscosity debris-flow lobes.

In the lower reaches of the catchments, chutes were cut across channel bends (Fig 4c) and often the main channel was infilled with gravel and the post-flood discharge diverted down the chutes. Chute bars, large point bars and gravel splays one pebble thick were widely developed on the inside of channel bends. Below tributary streams, which represented a local increase in the supply of debris, gravel deposition occurred in the channel to levels above the banks, although overbank deposition was often absent in straight reaches. The downstream margins of these deposits were often marked by small boulder jams (Fig 4e). Intense turbulence and possibly a hydraulic jump below the jams and the sudden reduction in bedload owing to storage above the jam resulted in extensive channel scour immediately downstream resulting in a morphological sequence reminiscent of the erosion-deposition sequence noted in the narrow headwaters.

In one case a longitudinal gravel bar developed downstream of an obstruction on the flood plain. The obstruction was apparently an old longitudinal bar of similar size and composition, so that the present deposit formed a congruous extension to the previous deposit.

Peat blocks were stranded across the valley bottom. Many of these, up to 2 m in diameter, had been rolled along the streambed and had been worn into characteristic spindle-shapes. Peat slurry was deposited in sheltered localities but this was rare and the absence of fine peat, sand or silt deposits immediately after the flood was notable.

DISCUSSION

Hydrology and boulder transport

Although it was possible to estimate the discharge in downstream reaches the question remains as to the magnitude of the peak discharge in the upper channel reaches. The presence of a large quantity of peat debris and locally high bed-load transport rates preclude exact analysis. Except for the boulder-jams in the West Grain, there was no evidence that the flows in the upper-catchment channels were mechanically cohesive debris-flows (sensu Lewis, 1984). The evidence indicates a 'debris torrent' (Miles and Kellerhals, 1981) would be appropriate terminology with sediment concentrations in the 'high' and locally the 'extreme' categories as defined by Beverage and Culbertson (1964).

Because of the difficulties in modelling the hydrological response of such floods, semi-empirical reconstruction methods to estimate peak discharges are important (Costa, 1983; Williams, 1983). In this respect equation 6 was found to be applicable to the present investigation. Although the close agreement with the gauged discharge may be fortuitous and equation 2 is empirically-derived, it has been shown to have a theoretical basis.

Further investigation is required into the processes of boulder entrainment in steep catchments. Macro-turbulence is often cited as a mechanism following the investigations of Matthes (1947) and Baker (1973) reviewed the concept with respect to threshold transport velocities which appear to be less than theoretically derived values. However, recent work by Jackson (1976) indicates that macroturbulent structures may only develop in relatively deep flows, throwing some doubt on its role in shallow streams. The importance of local flow accelerations, variable degrees of particle protrusion and the undermining of particles may play a more important role in initial motion of particles in debris-torrents than has been previously realized.

General Geomorphology

The observed channel changes, for example, the infilling of meander bends by bedload deposition and the straightening of the stream-course by chute incision is similar to changes described on the Hoaraoak Water in the 1952 Exmoor flood (Anderson and Calver, 1980) and in the Howgill...
Fells in 1982 (Harvey, in press). Recovery to date has also been similar to the Exmoor flood, there having been a partial reoccupation of the previous channel with local exhumation of its previous cross-sectional geometry. Most meander chutes have been abandoned although a few are now the main streamline. The overloading of the channel with coarse bed material has resulted in frequent ‘overbank’ discharges by moderate events, which have eroded cobble splay and have deposited silts in previously open-work gravels. Where sediments were completely evacuated from the channel bed major slope slumping occurred during the winter months resulting in high suspended sediment loads. This delayed response was owing to unstable slope mobilization by frost action and winter rainfall. The importance of the wedge of stream-bed sediments in controlling slope-movement in narrow defiles does not seem to have been described previously, although Anderson and Calver (1977) noted extensive under-cutting of valley walls.

The alternating of distinctive eroded and depositional zones along headwater stream-courses consequent to major flash floods has been described in the UK only by Gilligan (1909). However, these features, and the small boulder-jams formed further downstream, are similar to a general group of bed-forms referred to as ‘stepped-depositional’ features (Bowman, 1977). Whittaker and Jaeggi (1982) review the origin of these bed-forms in upland-streams and conclude, from flume-studies, that they develop during catastrophic flow conditions probably beneath hydraulic jumps at peak discharge. However, the mechanism would not seem appropriate for the formation of the large jams (Fig 4b) which morphologically resemble debris-flow boulder-lobes.

The stepped-bed effect is probably transitory because stored-material is being mobilized and entrenched as obstructions are by-passed by the streams. Although it seems unlikely that the larger depositional features will be eroded, the channels appear to be rapidly readjusting to a plan-form similar to that pertaining before the flood. The implication is that although some sort of geomorphological threshold was exceeded, in that the boulder-jams are likely to be persistent, the channels either are adjusted to conditions of ‘maximum’ discharge (e.g. Chorley and Morgan, 1962) or, more likely, are insensitive (Brunsden and Thomes, 1979) to the magnitude of large events. Hence the magnitude of the event (i.e. near maximum rainfall for the given duration) was manifest in the quantity of material transported and deposited in the upper metamorphosis downstream.

**Magnitude-frequency and the concept of effectiveness**

The spatially variable response of the basins is typical of what Schumm (1973) described as ‘complex response’. The response is dependent upon the forcing functions, but also upon the availability of sediment, the opportunity for storage and the elapsed time. The latter factor applies both to the duration of the event and also in the concept of ‘recovery’ since the impulse ceased. Following a major event the resulting depositional and erosion forms will readjust to some degree; our appreciation of which is partially dependent upon what post-event time-slice we choose to study.

In very small temperate basins, basin-wide synchroneity in the hydrological response is unusual although, occasionally as here, a high magnitude event can occur within a small part of a small basin. In the present example, a large amount of work was accomplished in moving sediment into the valley bottoms but work achieved in terms of inorganic sediment yield from the catchments was probably small. The return period of the flood peak at the basin outlet was not great (8–20 years for Langdon Beck) and is therefore of little value unless other factors are taken into account and may be quantified.

Wolman and Gerson (1978) and latterly Newson (1980) have suggested that catchment response be related to a concept of ‘effectiveness’. An event may be ‘effective’ in terms of the mass of material moved or it may be ‘effective’ in shaping the landscape (Wolman and Gerson, 1978) but it is rarely clear how much change constitutes a significant impact. The assessment of ‘recovery periods’ (Brunsden and Thomes, 1979) is a useful quantitative means of assessing the geomorphological impact, but the problem is recognizing recovery within the human time-framework. Many rare events may have a visually dramatic impact but the actual effect on shape or amplitude of the landscape can be negligible. The argument of necessity returns to the need to quantify storage and transport, for specific parts of the landscape after formative events, in order to obtain comparable measures of effectiveness in individual units. Fortunately some quantitative measures of the magnitude of the threshold event can be derived.
Landscape forming thresholds for diagnostic landforms

Within the present context it would appear that a major threshold was exceeded for both slopes and valley-floor processes. Large peat landslides and valley-wall slumping testify to the slope effects. However, slopes distant from gullies and flushes, and the interfluves were unaffected. Other 'unique' features clearly associated with the flood and not 'normal' processes occurring each year are the alluvial fans, boulder jams, bars, chutes and channel fills. Of these only the latter two are likely to be modified to any degree by fluvial processes of the order of the mean annual flood. The presence of a 'fossil' bar in one valley also indicates the occurrence of a previous high magnitude event and the preservation potential of these features.

Because the energy input was localized and abrupt there was a rapid damped lineal response down-valley in the headwater regions as sediment was stored en route. The channels only demonstrated entrenching into the bedrock in the upper reaches. Downstream large sediment inputs from gullies minimized erosion, despite the large shear stresses, as the channels were overloaded with bedload. Sensitivity to erosion therefore decreased progressively along the valley axis although locally there were exceptions association with valley convergences and boulder-jams for example. Figure 5 demonstrates the approximate first-order exponential decay of peak shear stress with distance downstream in Langdon Beck. The shear stress is in excess of that required to move boulders which accounts for the calibre and quantity of sediment transported.

In attempting to define (i) a geomorphic threshold value for catchment response and (ii) a recovery period for the event, it is fortunate that, three valleys were affected to varying degrees and that bedload data are available for the West Grain (Wilkinson, 1971).

It was evident, from the amount of sediment deposited in the valleys and the number and size of the flood features, that the catchments were affected unequally and this could not be related to differences in sediment supply. Diagnostic flood deposits of a size comparable with, and containing boulders as coarse as those shown in Figs 4b and e are rare in other small streams but are found for example in the Harthope and Mosedale valleys (Milne, 1982; Rose and Boardman, 1983). In this respect a major but not unique threshold appears to have been exceeded in the West Grain. In the upper reaches of Langdon Beck similar but smaller and fewer features are evident whilst Ireshope Burn had no major features. It is possible therefore to rank the catchments subjectively in terms of geomorphic work done and also objectively in terms of potential energy (E) expressed as kJoules per unit catchment area;

$$E = \gamma QS/A$$  (8)

Using data from Table I, a catchment area of 3.5 km$^2$ for Langdon Beck above the West Beck confluence, and discharge values of 30, 28 and 16 m$^3$ s$^{-1}$ for Ireshope Beck, Langdon Beck and West Grain respectively, the following tabulated data are obtained (Table VI).

The calculations would indicate that c. 2.0–3.0 kJ km$^{-2}$ was an adequate threshold value in the study area to promote large-scale landform changes in the valley bottom.

Wilkinson (1971) gave the bedload erosion rate in West Grain measured over two years as 21 m$^3$ km$^{-2}$ yr$^{-1}$. Assuming 1.8 tonnes m$^{-3}$ for freshly deposited coarse bed material (Newson and Leeks, 1985) the c. 10 000 tonnes in the West Grain valley at present could be removed in as little as 142 years at a steady rate of removal although the large
TABLE VI. Potential energy and associated effects

<table>
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<tr>
<th>Location</th>
<th>Potential Energy (kJ km(^{-2}))</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ireshope Burn</td>
<td>1.5</td>
<td>local chute development, bank erosion, boulder transport, gravel splays</td>
</tr>
<tr>
<td>Langdon Beck</td>
<td>2.8</td>
<td>as above but also: channel over-loading, widespread chute dev., small boulder-jams and berms, bedrock erosion</td>
</tr>
<tr>
<td>West Grain</td>
<td>7.6</td>
<td>as above but also: major berms and boulder jams</td>
</tr>
</tbody>
</table>

Boulder jams would persist. However, the time factor is likely to be substantially less as the removal rate will be initially higher than that recorded by Wilkinson and will then decline exponentially. Brunsden and Thomes (1979) introduced a transient-form ratio or sensitivity index, expressed as

\[
TF_r = \frac{\text{mean relaxation time}}{\text{mean recurrence interval}} \tag{9}
\]

Utilizing the recurrence interval of Bilham's method for 104.8 mm of rainfall i.e. 500 years, we obtain \( TF_r = 142/500 = 0.28 \), indicative of a generally insensitive landscape despite the magnitude of the event and the current visual impact. However, the limitation to the sensitivity concept and measure is that the event made large quantities of sediments available for transport by lesser events and in this respect may be viewed as a significant modifier to the landscape (Schumm, 1973).

CONCLUSIONS

Although this paper is based on only one detailed case study it is important to consider that the major impulses for landscape formation in many small steep basins in the northern Pennines may be related to formative events which are considered rare and catastrophic within a human time-framework (Return periods c. \( 10^2 - 10^3 \) years). These events, as typified by the Noon Hill floods are formative in that they not only enhance certain erosional and depositional features which are characteristic of the uplands i.e. gullies, landslide scars and valley-depositional features but also, in the sense indicated by Schumm (1973), they move substantial quantities of sediments from one store to another down the energy slope making this material available for more frequent processes to transport and modify existing flood-forms. Otherwise many upland Pennine landscapes may be usefully viewed as insensitive, in that 'visual' recovery is fairly rapid perhaps within one-tenth of the return period of the event sequence. A portion of the time between events may therefore be taken up with complex readjustments so that typical process studies will not necessarily represent steady-state conditions in the catchment but some stage on an exponential recovery curve between events. In this respect the variability in erosion rates noted between similar lithologies and landscapes (Newson and Leeks, 1985) may be partially explained. It is useful therefore to view upland landforms only as existing within a dynamic equilibrium over a long time span consisting of a series of formative events and recovery periods (i.e. \( > 10^3 \) years). In the intermediate time-span (\( 10^2 \) years) fluvial processes of sediment transport cannot necessarily be viewed as time-independent.

Despite this assertion, it is probable that the uniformitarian concepts of Wolman and Miller (1960) may be applied pragmatically over very short time-scales (\( 10^1 \) years?). For example, the maintenance of upland channel plan-form is relatively insensitive to change in that negative feedback is rapid. Major threshold changes in the valley-slope system however must be viewed within a conceptual framework encompassing catastrophic non-uniformitarian precepts.

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