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Review

A review of the evolving perceptual model of hillslope flowpaths at the Maimai catchments, New Zealand

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Abstract

The Maimai catchment has been the site of ongoing hillslope research since the late 1970s. These studies have facilitated the development of a detailed perceptual model of hillslope hydrology at Maimai. This perceptual model has grown in complexity beyond analytical description; nonetheless it provides a very useful case study of hillslope hydrological processes and encapsulates much of what field hydrologists have come to recognize as the dominant hillslope runoff processes in steep, humid catchments. No single research approach has resolved the complexities of streamflow generation in this highly responsive catchment. Yet, each data set reviewed in this paper adds to the cumulative understanding of catchment behavior by providing alternative (and sometimes conflicting) interpretations of hillslope subsurface flow. Initial dye tracer studies of macropore flow provided insight into hillslope flow processes, but suffered from the limitations of a single-method approach. Subsequent water isotopic tracing studies showed clearly that stored soil water and groundwater comprised the majority of channel stormflow; notwithstanding, isotope-oriented approaches did not enable the development of a mechanistic understanding of hillslope processes. An integration of tensiometer recording and tracer techniques was required for later reconciliation of different process interpretations concerning the role of macropores and old/new water ratios. Although single throughflow pits continued to be the indicator of subsurface flow timing and magnitude for several published studies at Maimai, subsequent whole hillslope trench studies showed that flow varied widely across a slope section—making the single pit observations of the previous studies suspect. Most recent observations demonstrate that small depressions in the bedrock surface may exert a significant control on water mobility and mixing. In particular, the bedrock topography appears to determine spatially the pathway of rapid saturated subsurface water flow and tracer breakthrough at the hillslope scale. The Maimai catchments in New Zealand provide a historical perspective on the issues faced by hillslope/small catchment hydrologists since the mid 1970s and highlights the advantages of multiple repeat experiments for testing hypotheses and improving our mechanistic understanding of subsurface flow. © 2002 Published by Elsevier Science B.V.

Keywords: Hillslope hydrology; Isotope tracing; Macropores; Subsurface stormflow; Runoff generation

1. Introduction

Hillslope hydrological investigations have been conducted in a variety of geographical, hydrogeological, and climatic settings and have been reviewed recently by Bonell (1998). While extensive in location and scope, most catchment-based hillslope investigations have not continued much beyond a typical 2–3 year funding cycle. This frequently limits the opportunity for formal testing of hypotheses proposed by previous researchers at a site and limits the

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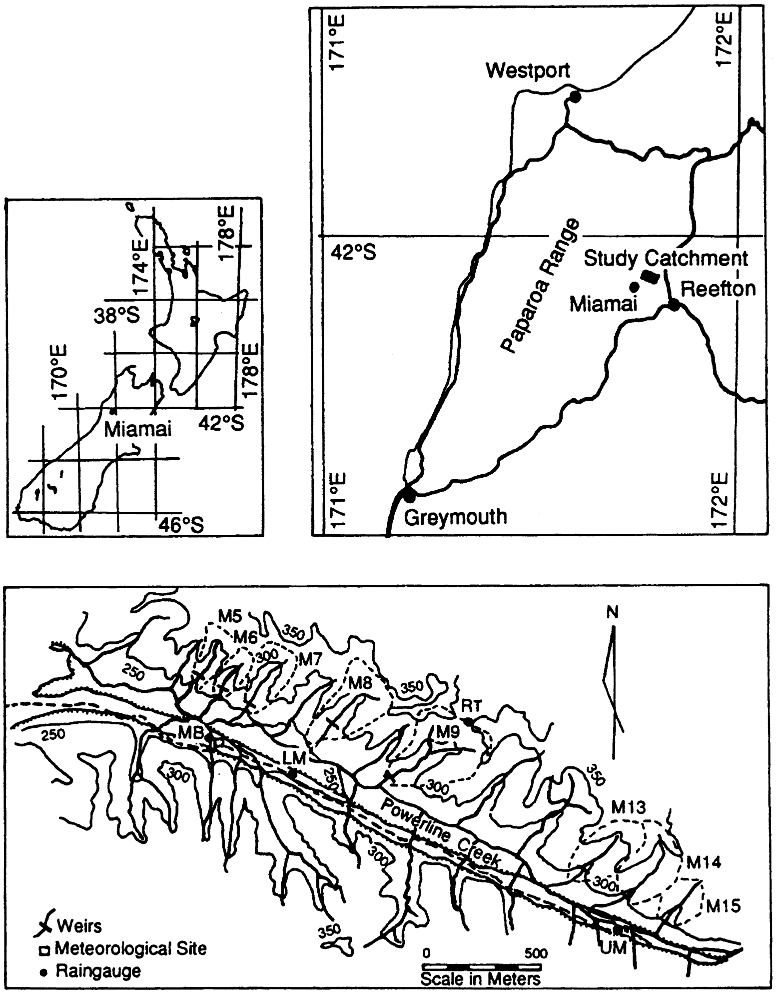

Fig. 1. Maimai study catchment location, South Island, New Zealand (from Rowe et al., 1994).

cumulative understanding of a single hillslope. Therefore, hillslope hydrological observations and 'perceptual models' (Beven, 1993) derived from them are rarely 'challenged' by other research scientists carrying on increasingly intensive work at the same hillslope or catchment. One notable exception is the Maimai research catchment in New Zealand. Maimai has been the site of ongoing hillslope investigations by several research groups since the late 1970s. These studies have facilitated the evolution and development of a detailed perceptual model of hillslope hydrology at Maimai. This perceptual model has now grown in complexity defying analytical description; nonetheless it provides a very useful case study of hillslope hydrological processes and encapsulates much of what field hydrologists have come to recognize as the dominant hillslope runoff processes in steep, humid catchments.

The goal of this paper is to synthesize the development of an evolving perceptual model of subsurface flow at the Maimai catchment. We hope that this review will: (1) provide a comprehensive overview of studies on subsurface flow within a well-characterized humid, temperate, forested catchment; and (2) chronicle how the evolution of the Maimai perceptual model has been affected by the methods used, the magnitude and frequency of events studied, and the scale of inquiry of specific studies. We hope to show

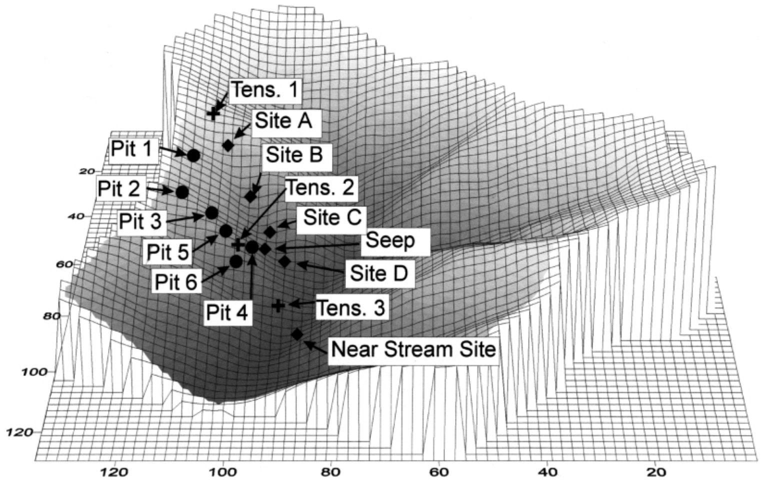

Fig. 2. M8 catchment study site locations. Sites A–D and Pits 1–6 are original study locations from Mosley (1979). Pearce et al. (1986) and Sklash et al. (1986) reactivated seven of Mosley's measurement sites and added suction lysimeters and maximum-rise piezometers at the near stream site (SL and Well B) and near the M8 stream weir (SL and Well A). McDonnell (1990a,b) extended the instrumentation with tensiometer networks installed at the near stream site, and upslope of Pit 5 and Pit A.

the value of working in a cumulative fashion at a single research site. This approach has yielded, in the case of the Maimai catchment, a rich understanding of hydrological processes.

2. Maimai: the quintessential steep humid catchment

2.1. History of the Maimai catchment

In 1974, the New Zealand Forest Research Institute initiated a multi-catchment study at Maimai in the Tawhai State Forest, near Reefton, North Westland, on the South Island of New Zealand (Rowe et al., 1994; Rowe and Pearce, 1994). This was done in response to concerns regarding the possible impacts of forest management activities on the quantity and quality of water supplies and aquatic habitats. Data were gathered on streamflow characteristics, stream sediment yield, and slope stability of mixed beech forest catchments. Although the conversion of native forests to exotics and the harvesting of mixed forest stands did not prove economically viable (Rowe et al., 1994), the long-term studies of the Maimai catchments continued. The data collected provide a useful comprehensive long-term study of hydrological processes in forested catchments.

2.2. Physical characteristics of the Maimai catchment

The Maimai study area consists of eight small catchments (1.63–8.26 ha) located to the east of the Paparoa Mountain Range situated on south-facing slopes draining into Powerline Creek (informal name) (Fig. 1). The catchments lie parallel to each other and share similar topographic characteristics. Slopes are short (<300 m) and steep (average 34°) with local relief of 100–150 m. Stream channels are deeply incised and lower portions of the slope profiles are strongly convex. Areas that could contribute to storm response by saturation excess overland flow are small and limited to 4–7% of catchment areas (Mosley, 1979; Pearce et al., 1986). The 3.8 ha M8 catchment has been studied most intensively and reflects the physical characteristics of the other watersheds (Fig. 2).

Frontal systems from the Tasman sea pass regularly across the Paparoa Range (from westerly and northerly directions), creating a climatic regime with frequent and occasionally prolonged periods of rainfall. Mean annual precipitation is approximately 2600 mm, producing an estimated 1550 mm of runoff.

The summer months are the driest; average monthly rainfall from December to February is 165 mm and is between 190 and 270 mm per month for the rest of the year. Typically, there are 156 rain days per year with few temperature extremes and only about 2 snow days per year (Rowe et al., 1994).

In addition to being wet, the catchments are highly responsive to storm rainfall. Quickflow (QF as defined by Hewlett and Hibbert, 1967) comprises 65% of the mean annual runoff and 39% of annual total rainfall (P) (Pearce et al., 1986). The quickflow response ratio (R index = QF/P) is roughly double that of the most responsive basins documented in the eastern United States (Hewlett and Hibbert, 1967). Pearce et al. (1986) note that the R index (averaged for runoff events from rainfalls of greater than 25 mm) is 46%, compared with 3–35% for 11 basins distributed between Georgia and New Hampshire (Hewlett et al., 1977).

2.3. Vegetation and soils of the Maimai catchment

The vegetation is mixed evergreen beech forest (*Nothofagus* spp.), podocarps, and broadleafed hardwoods. It is multi-storied, with a canopy 20–36 m high, a dense fern and shrub understory, and a fern and moss ground cover. Rowe (1979) reported annual interception losses of 26% for the undisturbed mixed evergreen forest. Mean evaporation rates for the M8 catchment are 0.46 and 0.28 mm/h for summer and winter, respectively (Pearce and Rowe, 1981).

The study area is underlain by a firmly compacted, moderately weathered, early Pleistocene conglomerate (Old Man Gravels). The conglomerate is comprised of clasts of sandstone, granite, and schist in a clay-sand matrix and is poorly permeable with seepage losses to deep groundwater estimated at 100 mm/yr (O'Loughlin et al., 1978; Pearce and Rowe, 1979). Overlying soils are classified as Blackball hill soils. The typical soil horizon is characterized by a thick, well developed organic horizon (average 17 cm), thin, slightly stony, dark grayish brown A horizon, and a moderately thick, very friable mineral layer of podsolized, stony, yellow-brown earth subsoil (average 60 cm). Silt loam textures predominate. Profiles examined by Webster (1977) showed that the organic humus layer had a mean total porosity and macroporosity of 86 and 39% by volume, respectively, with an infiltration rate of 6100 mm/h. The mineral soils are very permeable and promote rapid translocation of materials in suspension or solution (Rowe et al., 1994). The total porosity averages 45% by volume, with average bulk densities of 1.5 g/cm^3 and saturated hydraulic conductivities on the order of 10–300 mm/h. The wet and humid environment, in conjunction with topographic and soil characteristics, result in the soils normally remaining within 10% of saturation by volume for most of the hydrologic year (Mosley, 1979). As a result, the soils are strongly weathered and leached, with low natural fertility.

The thinness of the soils promotes the lateral development of root networks and channels. Soil profiles at vertical pit faces in the Maimai M8 catchment reveal extensive macropores and preferential flow pathways which form along cracks and holes in the soil and along channels created by live and dead root (Mosley 1979; 1982). Lateral root channel networks are evident in the numerous tree throws that exist throughout the catchments. Preferential flow also occurs along soil horizon planes and the soil-bedrock (Old Man Gravels) interface (Mosley, 1979, 1982; McDonnell, 1990b, 1997; Woods and Rowe, 1996; Brammer, 1996; McDonnell et al., 1998).

3. Early studies of macropore flow

3.1. Hydrometric observations at the catchment and hillslope scale

Mosley (1979) conducted the first comprehensive hydrologic process study of the Maimai catchments. This study included a series of hydrometric and dye tracing experiments in the M8 catchment to monitor streamflow and subsurface flow through the soil mantle at a variety of topographic positions. Streamflow was measured at three sites along the stream channel. Seven 2–3 m wide pits were dug to the underlying bedrock and troughs were installed at the base of the pits to intercept and measure subsurface flow (Fig. 2).

Hydrometric observations were made during 12 storm events of varying characteristics. At the three stream channel sites (Sites B–D) and one upslope hollow site (Site A), stream and subsurface flow

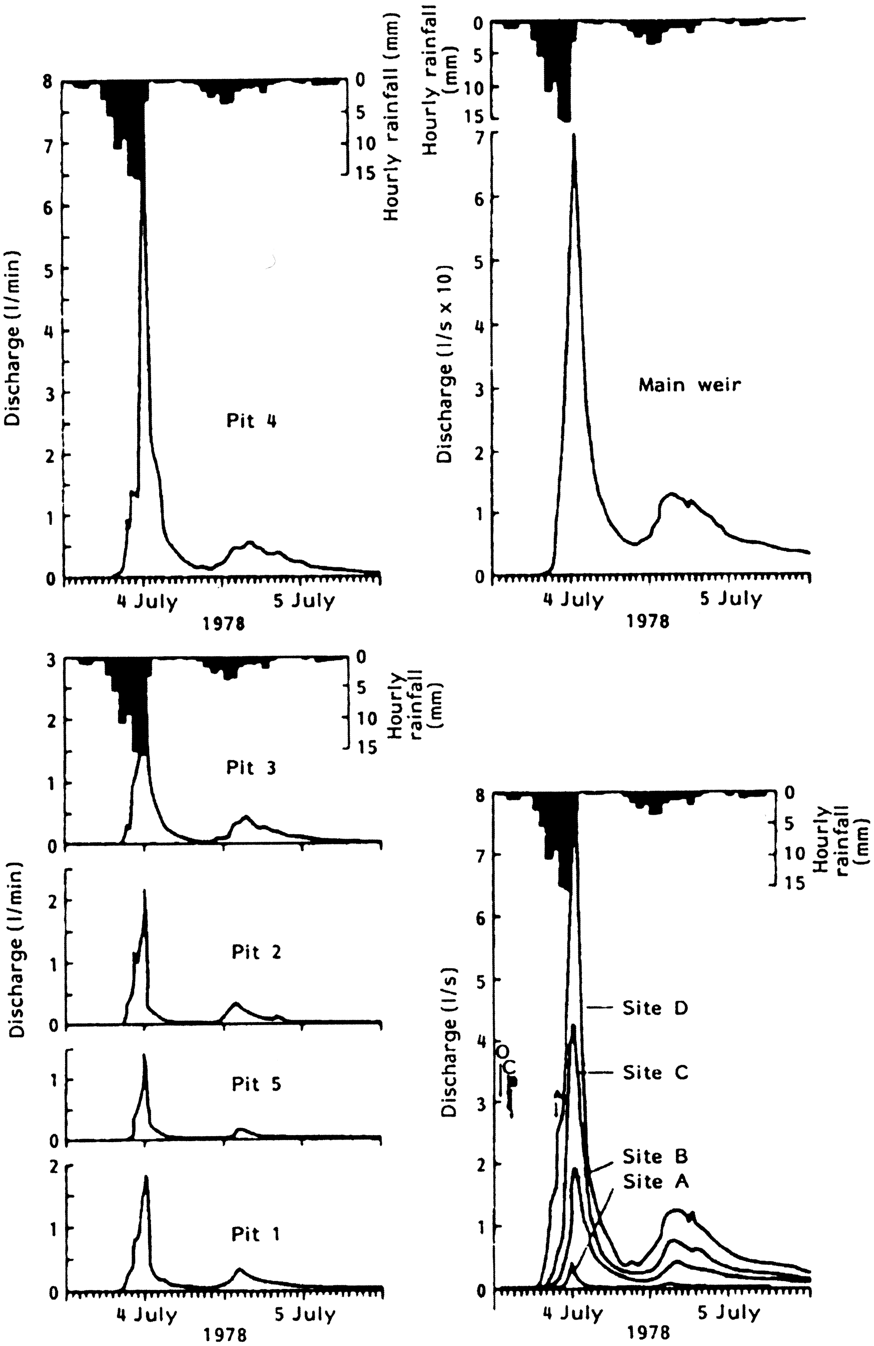

Fig. 3. Comparison of hydrograph response from the throughflow pits (1–6), zero-order stream sites (A–D) and at the M8 main weir as shown in Fig. 2, 4–5 July 1978 (from Mosley, 1979).

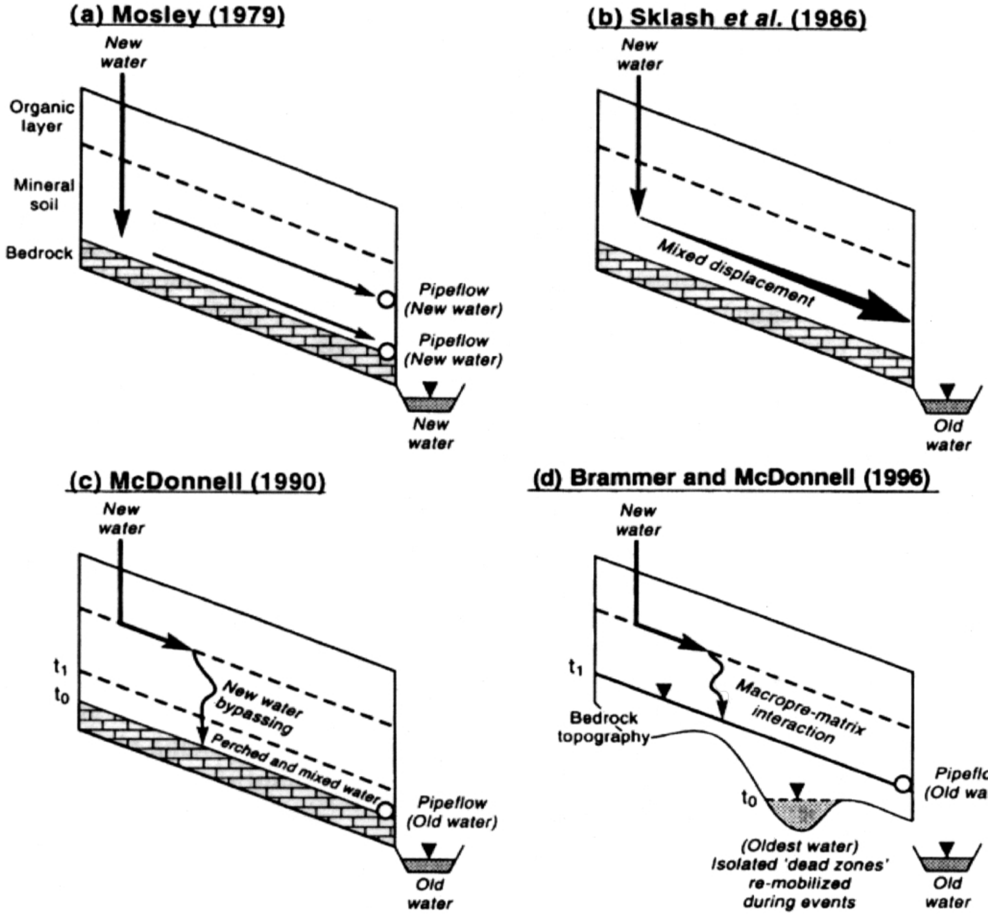

Fig. 4. Maimai conceptual models: (a) Mosley (1982); (b) Mosley (1982); (c) Sklash et al. (1986); (d) McDonnell (1990a); (e) Brammer et al. (1996).

hydrographs were closely aligned in time and flow volume increased in a downslope direction (Fig. 3). There was a close coincidence in the time of the peaks and an increase in discharge and total flow volumes in a downslope direction. This indicated that water was moving considerable distances through the soil during the rising limb of the hydrograph (Mosley, 1979). Mosley (1979) noted that if subsurface flow at each site was derived from the immediate surrounding contributing area only, peak discharge and total flow volume would be independent of distance from the catchment divide.

A rectilinear slope profile stretching from the valley bottom to the ridge top was sampled to monitor soil profile moisture content. Nine sites along the profile were sampled at 20 cm increments to a maximum depth of 60 cm five times during a precipitation event. The data suggested that there was

> … a rapid movement of water vertically in the soil profile and in the downslope direction, such that a saturated wedge, whose shape was modified by the shape of the bedrock surface, built up in the soil mantle. The wedge almost intersected the ground surface at the foot of the slope and tapered off in the upslope direction, becoming restricted to only a thin layer above the bedrock surface at the top of the slope (Mosley, 1979 page 799).

Prior to Mosley's (1979) investigation, stormflow generation was attributed primarily to near stream source areas. Subsurface stormflow was relegated to a secondary role in most forest hydrological studies worldwide. Also, at the time, most research on subsurface flow was conditioned by the theoretical work of Freeze (1972) who stated that subsurface stormflow contributions to peak discharge could only occur under specific and very limited conditions. Mosley sought to investigate the relative importance of subsurface stormflow as a streamflow generating mechanism in the steep humid Maimai catchments with their convex slopes and incised stream channels.

3.2. Hydrometric observations at the pit face scale

In addition to flow measurements, Mosley (1979) made a number of important visual observations of hillslope flow. He noted that the pit faces displayed points of concentrated seepage during storm events, usually at the base of the B horizon, at which high rates of outflow were observed. At one site, Mosley (1979) observed that water gushed out of two pipes discovered at the base of the B horizon, at a rate approaching 20 l/s. The significance of this preferential/macropore flow mechanism was further evaluated via eight dye tracer trials conducted at various pit faces (Fig. 2). The first four trials were conducted at the end of a low intensity storm event. Dye (sodium dichromate, rhodamine B, and lissamine green) was applied to the soil surface 1–4 m upslope of the pits using a cylinder inserted into the humus layer. The first traces of dye emerged through root holes in the pit faces 1.5–23 min following application. Dye later appeared along the base of the B horizon. Dye velocities were calculated as 0.17–0.81 cm/s. Lateral spread of the dye of up to 1 m was also noted. The fourth trial was conducted at a newly-excavated pit with additions of water (79 1 total) applied 4 m upslope prior to dye application. Under these conditions, dye appeared 30 s after application, indicating a travel velocity of 1.1 cm/s.

Further dye tracer experiments were conducted at the end of a higher intensity storm event with rhodamine B dye simply sprinkled onto the soil surface 1 m upslope from the pits. Dye first appeared at the base of the humus layer and within seeps in the B horizon 60–82 s after application. Travel velocities ranged from 1.2 to 2.1 cm/s. In general, Mosley (1979) found that maximum dye travel velocities were up to 300 times greater than the measured saturated hydraulic conductivity (252 mm/h) for the mineral soil. This indicated that there was some downslope movement of water occurring over distances of up to several tens of meters during the rising limb of the storm hydrograph. Mosley reasoned that this could not be accomplished by saturated flow through the matrix alone. Where there was low root density and few channels in the A and B horizons, water appeared downslope at the base of the A horizon. At sites with greater density of roots and channels in the mineral soil, water reappeared from seeps and at the soil-bedrock interface; losses to the matrix were minimal (Fig. 4).

Mosley (1979) concluded that subsurface flow through 'macropores' and seepage zones in the soil was the predominant mechanism of stormflow generation in storms with quickflow greater than 1 mm and that this mechanism was capable of contributing to storm period streamflow. Mosley (1979) further stated that even in small storms, subsurface flow from all parts of the watershed appeared to contribute to stormflow. Mosley attributed delayed flow throughout the hydrograph recession to unsaturated zone drainage from the soil mantle and possibly saturated flow from the soil bedrock interface.

Mosley conceded that water/tracer additions

> … were clearly artificial, but except in the immediate vicinity of the cylinder it is considered that movement of the dye was similar to that of water applied by an intense rainfall event (1979, page 800).

Mosley (1979, page 806) concluded that "this flow was of 'new' water, and no evidence for translatory flow was observed." The definition of 'new water' in this case, was that water added by, and during, the precipitation event causing the subsurface flow hydrograph rise (Mosley, 1998, personal communication). This choice of words caused considerable subsequent debate in the literature.

3.3. Subsurface flow velocity estimation

Mosley (1982) continued the subsurface flow trials by sprinkling sites under differing harvesting conditions: undisturbed, logged and logged/burned/planted.

Water was applied as a line source 1 m upslope from pits 2 m long and 0.5–1 m wide. Once again, the pits were excavated to bedrock with intercepting troughs installed at the base. A maximum of 30 l of water was applied, but subsurface flow was often observed before more than 15–20 l had been applied. Rates of application were within the range of subsurface flow discharges observed by Mosley (1979) under natural rainfall.

A comparison of the values for mean velocity, maximum velocity and ratio of output volume to input volume was conducted (Mosley, 1982). High mean subsurface velocities were measured under each site type: 0.45 cm/s for undisturbed, 0.69 cm/s for slash-covered, and 0.64 cm/s for burned sites. The hydrographs of all sites were very similar in shape and, when plotted on a common time base, similar in peak times. The rapid flow velocities and visual observations showed that there were preferential flow pathways along cracks and holes in the soil and both live and dead roots. Profile wetness increased downslope and vertically; the thickness of the saturated zone decreased in an upslope direction.

The observed high flow velocities reported by Mosley (1982) along preferred pathways in the soil profile supported the earlier conclusion of Mosley (1979) that rapid subsurface flow can contribute to storm-period runoff at Maimai:

> … at a mean flow velocity of 0.3 cm/s and with a mean slope length of 30 m, water could travel from divide to stream channel in 2.8 h (Mosley, 1982, page 87).

According to Mosley (1982), this required that the soil had to be saturated before water could enter the macropores—it could then move rapidly over long distances through unsaturated soil without being totally absorbed into the matrix. Mosley (1982) attributed the occurrence of saturation during a storm event to antecedent moisture conditions, rainfall intensity, total rainfall, and location on the hillslope. Furthermore,

> … the experiments indicate that a large proportion of the precipitation onto, or inflow at, the upslope end of a soil block may be absorbed into the matrix and that the proportion appearing as rapid outflow at its bottom end may be very small. This latter proportion should increase during a rain event, with the duration of a rain event, and in the downslope direction (Mosley, 1982, page 90).

3.4. A proposed perceptual model of hydrologic response

From these preliminary studies, a schematic/perceptual model of catchment hydrologic response (subsurface flow paths) was proposed (Fig. 4(A)). It was recognized that all of the illustrated flow pathways were probably active at each site, but the relative importance varied significantly:

> Variability in flow velocity and the proportion of the input appearing as rapid outflow is a function of antecedent moisture conditions and of the relative importance of the various pathways at a given site, which in turn is a function of soil characteristics, macropore network and parent material at base of soil (Mosley, 1982, page 65).

The model considered macropore flow to be a 'short-circuiting' process by which water could move through the soil at rates up to 300 times greater than the measured mineral soil saturated hydraulic conductivity (Mosley, 1979). As a result, Mosley (1982, page 65) stated that water might reach the channel system from a portion of the catchment that varies with antecedent moisture conditions and storm duration and intensity, but which may be the entire catchment in moderate to large rainfall events.

With hindsight, we know that Mosley's perceptual model was limited by the lack of fluorescence intensity measurements (future investigations would show considerable dilution of the applied dye, indicating much more mixing). Mosley's model relied on the assumption of a continuous, well-connected macropore flow system present within the soil (Fig. 4(A)). Furthermore, the tracer application intensity of simulated 'events' was that of very long recurrence interval storms; hence application rates may have enhanced macropore flow beyond that witnessed during actual storms.

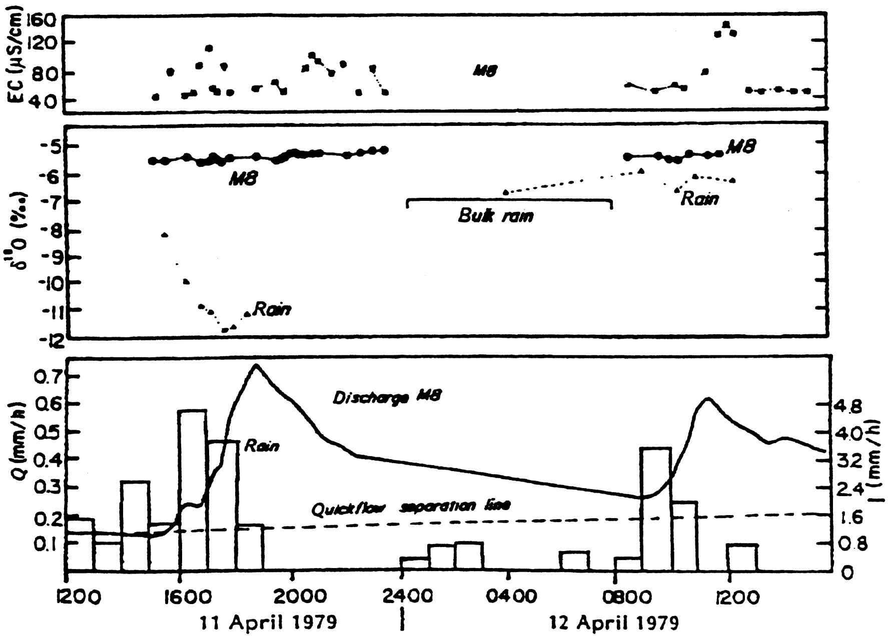

Fig. 5. Storm response in catchment M8, April 11–12, 1979 (from Pearce et al., 1986).

4. Isotope tracing and old water displacement

4.1. Application of the isotope hydrograph separation technique

Pearce et al. (1986) collected weekly samples of M8 rainfall, soil water and streamflow for electrical conductivity (EC), chloride (Cl^-), deuterium (δD), and oxygen-18 ($\delta^{18}O$) composition from 1977 to 1980. Streamflow was sampled from the M8 catchment, the undisturbed control catchment (M6), and Powerline Creek. Rainfall was sampled at two sites within the study area. Seven measurement sites (Pits 1–3, 5, Sites A and D, and the seep) used by Mosley (1979) in M8 were reactivated and instrumented with suction lysimeters and piezometers. Two additional sites within M8 (near stream site and at the catchment outflow) were similarly instrumented and sampled (Fig. 2).

Pearce et al. (1986) found that the $\delta^{18}O$ values for the weekly rain samples ranged from 3 to 12‰ and displayed some seasonality, where isotopically heavier values occurred during the summer and lighter values were observed in the winter months. Both the stream and groundwater samples followed rainfall trends but with smaller seasonal variation. The M8 $\delta^{18}O$ stream samples followed groundwater values indicating similar water. The decreased temporal variations in stream and groundwater suggested to Pearce et al. (1986) that: (1) most of the mixing of old and new waters occurred on the hillslope; and (2) subsurface water discharge to the stream was an isotopically uniform mixture of stored water. Samples collected at higher-than normal flow rates showed no deviation from the seasonal isotopic trends, despite large associated fluctuations in rainfall δD or $\delta^{18}O$. This indicated that only small contributions of new water occurred at high flow rates. Pearce et al. (1986) interpreted the conclusions of Mosley (1979, 1982) to mean that new precipitation accounted for storm discharge and formed most of the streamwater. As a result, Pearce et al. (1986) stated that their interpretation directly refuted Mosley's (1979) determination that rapid transmission of new water through macropores formed the majority of stream runoff:

> Our results indicate that in at least this hydrologic environment, which is apparently highly suited to throughflow of storm rainfall,

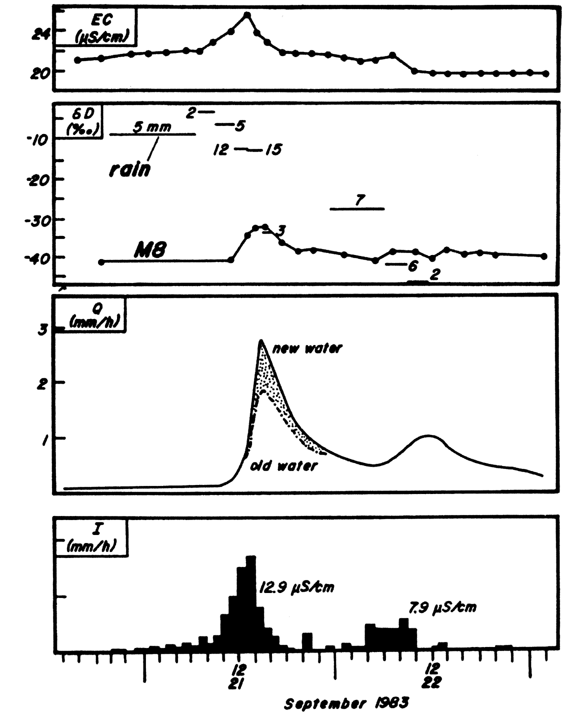

Fig. 6. Stream M8 δD and EC response, September 21, 1983 (from Sklash et al., 1986).

> processes which deliver 'old' water are largely responsible for hydrograph generation. Currently favored runoff mechanisms, which involve rapid flow of 'new' water over the ground surface or through the soil matrix, soil macropores, or other rapid transit pathways, cannot explain the streamflow response (Pearce et al., 1986, page 1263).

4.2. Sampling at the hillslope and catchment scale

Pearce et al. (1986) sampled two small storms immediately after logging of the M8 catchment in April 1979 (Fig. 5). The $\delta^{18}O$ values of the storm runoff fluctuated only slightly from baseflow $\delta^{18}O$; new water inputs were only 3% of storm runoff. The electrical conductivity and chloride data confirmed the low contribution of rain (new water) to stormflow. Stream EC rose to 80–100 μS/cm from an initial value of 47 μS/cm during the first event. EC rose to over 140 μS/cm during the second event the following day, lagging the hydrograph peak. The changes in EC showed increases in total solutes rather than dilution. Cl^- concentrations in streamflow remained constant throughout the events. The increases in EC and the consistency of Cl^- concentrations suggested that the storm runoff response was predominantly water that had a substantial period of contact with the soil. The contribution of 3% new water could be accounted for by direct precipitation onto the stream channel. Therefore, the combination of isotope compositions and solute concentrations provided strong evidence that, at least in small events, and under moderately wet antecedent conditions, rapid throughflow of infiltrated rainwater was not an important storm runoff mechanism.

Sklash et al. (1986) extended the Pearce et al. (1986) hydrograph separations into other stream sections of the M8 catchment, including six of Mosley's throughflow pits, for several storm events in September 1983 (Fig. 6). Events sampled had return periods of between 4 and 12 weeks (i.e. small, high frequency events). Isotope hydrograph separations of the M6 and M8 streams indicated that old water dominated runoff from all of the events. New water was approximately 15–25% of the stormflow. New water contributions to quickflow in M8 could be accounted for by flow from less than 10% of the catchment area; not much larger than the area capable of generating saturation overland flow as determined by Mosley (1979) and the permanently saturated area of 5% estimated by Pearce and McKerchar (1979). The EC and Cl^- data again supported isotopic results by indicating no significant dilution of the stream by new water. Flushing of water with high EC and Cl^- values into the stream was pronounced in M8. This was characterized by marked increases in EC and Cl^- on the rising limb of the hydrograph even though concentrations in the rain were considerably lower than those in the stored water (Fig. 6). The effect was much decreased in the second of two closely spaced storm events, indicating a flushing response. Stored water appeared to dominate outflow as shown by variations in δD, EC, and Cl^-concentrations in subsequent storms, reflecting the differing contact times between stored water and the soil matrix.

Table 1
Catchment M8 old and new water contributions to throughflow September 21, 1983 (from Sklash et al., 1986)

Site	δD_0% old water δD	δD_N% new water δD	δD_P% δD of throughflow at peak discharge	Time of sample (h)	% Old water	% New water
Pit 1	−43.1[a]	−12.2[c]	−33.7	14:10–17:52	70	30
Pit 2	−43.1[a]	−12.2[c]	−31.4	14:10–17:52	62	38
Pit 3	−43.1[a]	−12.2[c]	−29.2	14:10–17:52	55	45
Site A	−42.3	−12.2[c]	−32.9	14:02	69	31
Pit 5	−45.5	−12.2[c]	−43.5	14:06	94	6
Seep	−44.0	−12.2[c]	−41.3	14:01	92	8
Site D	−39.1[b]	−12.2[c]	−32.4	14:18	75	25
M9	−41.3	−12.2[c]	−32.5	14:58	70	30

[a] Based on SL5S value.
[c] Weighted average rain.
[b] Low flow value on September 23.

4.3. Sampling at the pit face scale

The sampled hillslope water displayed large spatial variability in δD composition (Sklash et al., 1986). The pit locations (used originally by Mosley, 1979) represented a cross section of topographic positions and hillslope hydrologic conditions. The values of δD during both low-flow periods and in response to storm events showed that pit throughflow was dominated by old water (Table 1). Deep suction lysimeter δD values (Table 1, location shown in Fig. 2) were much lighter isotopically than shallow lysimeter and throughflow δD samples taken at the same location. This suggested that much of the throughflow arrived laterally from thinner soil profiles. At the same sites, the EC and Cl^- values showed noticeable vertical differences in solute flushing (Sklash et al., 1986). Some locations (pit 5 and the seep) had old water percentages in excess of 90%, even though Mosley (1979) had examined the same sites and observed rapid dye tracer breakthrough and gurgling macropores:

> The δD values of throughflow sampled at selected sites in the instrumented subcatchment of catchment M8 confirm that the throughflow was dominated by old water both between and during storm runoff events. These findings refute conclusions drawn from traditional hydrometric and dye tracer tests of storm runoff generation in the Maimai catchments and raise considerable doubt about results of similar tests in other research watersheds (Sklash et al. 1986, page 1282).

4.4. A proposed perceptual model of hydrologic and isotopic response

Sklash et al. (1986) measured large water table rises in the mid-slope and near stream maximum-rise piezometers during storms (Fig. 6, locations shown in Fig. 2). The piezometers located near the valley bottom had the highest response, close to total profile saturation. Visual observations confirmed that overland flow occurred only in valley bottom areas. Sklash et al. (1986) hypothesized that two mechanisms could possibly account for the large water table rises: (1) conversion of capillary fringe into phreatic water (as had been observed elsewhere by Sklash and Farvolden, 1979; Gillham, 1984; Abdul and Gillham, 1984); or (2) rapid lateral inflow of displaced old water into areas of deep soil from areas of shallower soil (Fig. 4(B)). Both mechanisms appeared to be triggered by new water infiltration, but old water from the saturated zone still dominated storm runoff (Pearce et al., 1986; Sklash et al., 1986). The response of the maximum-rise piezometers in the M8 catchment was consistent with the concept of groundwater ridging. Sklash et al. (1986) postulated that saturated wedges on the lower slopes and groundwater ridges in the valley bottoms developed quickly as infiltrating rain converted the tension-saturated zone into phreatic

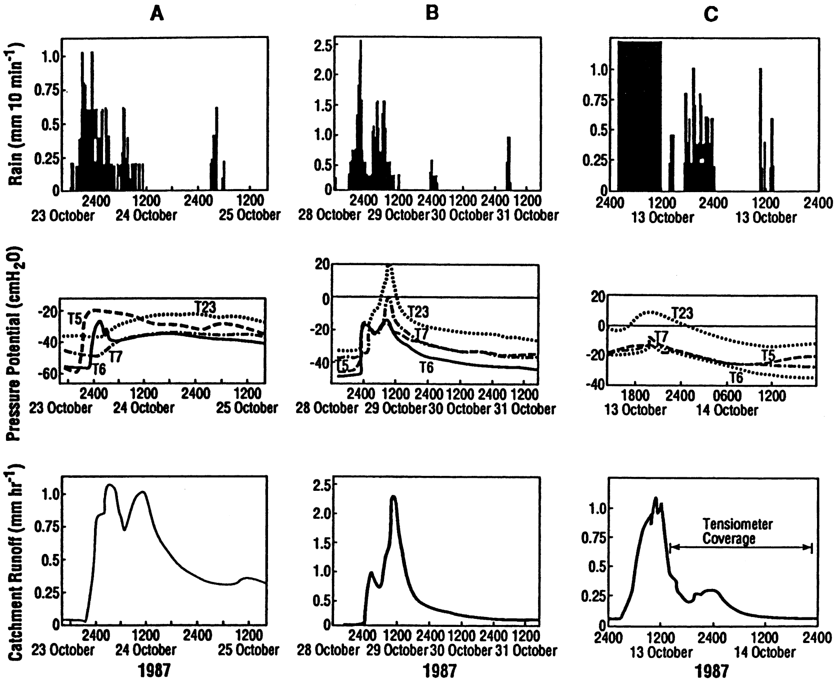

Fig. 7. Pressure potential response for tensiometers located in instrumented hollow, showing relationship between matric potential (ψ) and rainfall-catchment runoff condition. Three storms are shown, having rainfall totals of (A) 25 mm, (B) 58 mm and (C) 103 mm. Soil depth 1–1.5 m, slope angle 35–40°. Tensiometers T5, T6, T7, and T23 inserted at 170, 410, 820 and 1080 mm below soil surface, respectively (from McDonnell, 1990b).

water. This perceptual model negated the need to invoke rapid transmission of new water down slope via macropores in order to explain the stream flow response, since stored water was the main component discharged into the stream channel during events (Fig. 4(B)).

With hindsight, we know that the Pearce et al. (1986) and Sklash et al. (1986) perceptual model was limited by the lack of any soil physics data to confirm that M8 soils indeed had a tension saturated zone. Only small events (return periods of <6 months) were studied and no direct evidence of groundwater ridging was presented apart from point observations from maximum-rise wells. Pearce et al. (1986) and Sklash et al. (1986) discounted pipe flow measurements of Mosley entirely—likely because significant amounts of macropore flow did not necessarily occur during the small-magnitude events that they monitored. For larger events, the macropore flow phenomenon was observed and measured by several other later investigators (McDonnell, 1990a,b; Woods and Rowe, 1996; McDonnell et al., 1998).

5. A Combined hillslope, soil physics, isotopic, and chemical approach

McDonnell (1990a) and McDonnell et al. (1991a,b) combined isotope and chemical tracing with detailed tensiometric recording in near-stream, mid-hollow, and upslope hollow positions in the M8 catchment (Fig. 2) in an effort to explain the discrepancies between the perceptual models offered by Mosley (1982), Pearce et al. (1986) and Sklash et al. (1986). Pearce et al. (1986) and Sklash et al. (1986) showed that storm events were dominated by pre-event old water; however, simple order of magnitude calculations with Darcy's equation and estimates of the volume of water stored in the near-stream zone prior to precipitation events showed that the near-stream zone alone was insufficient to explain the peak discharge of pre-event old water during many storms

(McDonnell, 1991a). First, not enough old water was in a near-stream discharge position to account for all the old water in the event hydrograph, despite rapid near-stream response to precipitation (McDonnell, 1991a). Second, no capillary fringe was observed—the soils drained and wetted up along a linear moisture release curve based on characteristic curves developed for small intact cores extracted from each catchment position. However, in the hydrologically active mid-slope hollows (Pit 5 as shown in Fig. 2), McDonnell (1990b) found that matric-potential response was highly variable for different storm magnitudes, intensities, and pre-storm matric-potential conditions. Tensiometric measurements revealed an erratic infiltration-potential relationship (McDonnell, 1989). During a low magnitude (25 mm) rainfall event on October 23–24, 1987 (Fig. 7(A)), tensiometric data showed that a semi-constant wetting front propagated vertically through the profile with strong matric-potential response lags with depth. Although some bypass flow seemed to occur in the upper soil horizon (<50 cm), as evidenced by the response of tensiometer T5 (see Fig. 7 caption for explanation), rainfall depth and soil-moisture content were low enough so that the lower soil depths did not receive appreciable moisture from above, until streamflow response had subsided (Fig. 7(A)). Therefore, a slope water table did not develop.

During a larger magnitude (58 mm) rainfall event on October 29, matric-potential in the lower soil horizons (>75 cm) responded almost instantaneously to infiltrating rain (Fig. 7(B)). This response was a function of disequilibrium in soil pressure potentials during wetting, caused by the presence of soil macropores (McDonnell, 1990a). Furthermore, much of the matrix exhibited unrequited storage during this type of wetting, indicative of a two-component flow system of rapid macropore flow and slow matrix flow. In other words, vertical bypassing of water to depth allowed soil water potentials to reach 0 cm H_2O while the wetting front was still moving downward through the upper soil. For the largest event monitored (103 mm of rainfall) on October 13, matric potential remained relatively constant throughout the profile, during the limited period of tensiometer coverage (Fig. 7(C)). McDonnell (1990b) observed that most of the soil profile remained saturated during this episode.

5.1. Groundwater development and longevity

Generally, when rainfall intensities were low, but pre-storm soil water content was high, McDonnell (1989) found that additional rainfall rapidly filled the available soil-moisture storage, and perched water-table conditions quickly developed at the soil-bedrock interface. On the other hand, if short-term rainfall intensities were high, rainfall bypassed the upper soil horizons and moved to the profile base via vertical cracks, so that tensiometers in the lower half of the soil profile responded ahead, or independent of the upper tensiometers. Water table longevity at the soil-bedrock interface was very short and showed a close correspondence with hillslope throughflow rate, as measured by McDonnell (1989, 1990a). Down slope drainage of perched water was extremely efficient and showed no lag with recorded throughflow for the selected storms. McDonnell et al. (1991a) noted that this indicated that lateral saturated flow was rapid and moved through pipes located at the soil-bedrock interface (corroborated visually using dye tracers). The rapidity of tensiometric recession in the lower half of the soil profile in events with perched water-table conditions supported the idea of rapid down slope drainage through pipes. The interconnectedness of pipes in those zones was assumed by McDonnell (1990a) to be high enough to account for the rapidity of water-table decline and pore pressure dissipation.

5.2. Mechanics of preferential flow

McDonnell (1990a) reasoned that in the steeply sloping hollow zones (where much of the M8 stream runoff originated), bypass flow leads to the generation of runoff long before the moisture status of the soil matrix would predict. This was due to the matric-potential disequilibrium within the soil. McDonnell (1990a) noted that it was important to distinguish between the conductivity of the matrix (K^*) and that of the soil with macropores (K_{sat}). If the flux density of the rain (V_o) is greater than K^*, local ponding will eventually occur leading to vertical bypassing, whether or not V_o is greater than K_{sat}. Therefore, McDonnell (1990a) found that it was not unrealistic for 5–10 mm/h rainstorms to create localized ponding on a soil purported by Mosley (1979) to have a K_{sat} of

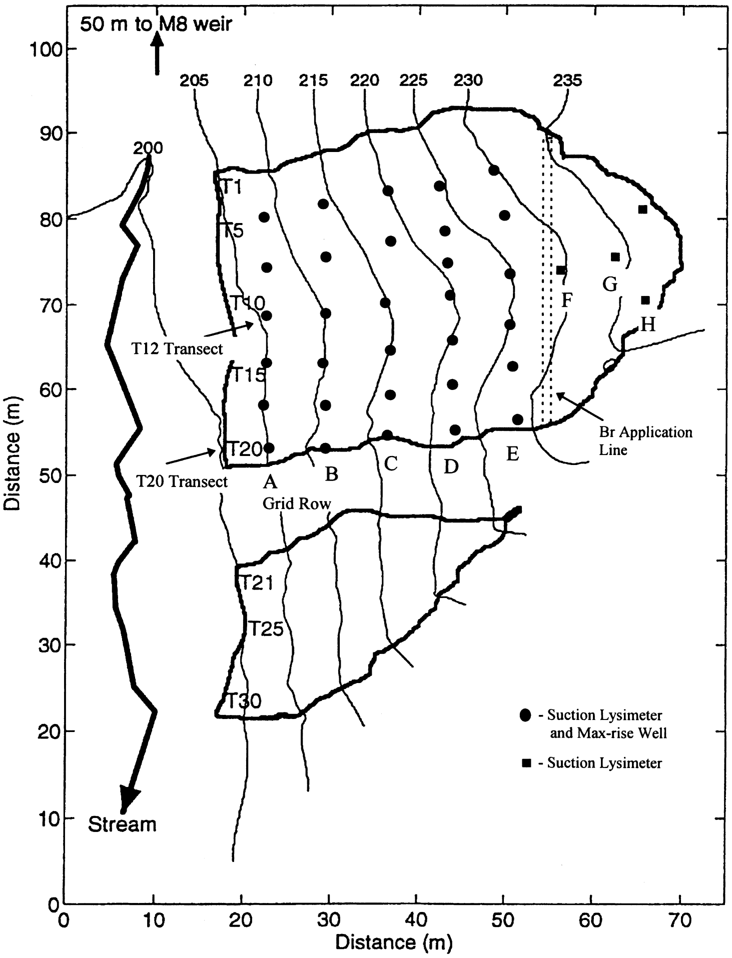

Fig. 8. Trenched study hillslope from Woods and Rowe (1996) showing the topographic contours and original trough locations. Also shown is the distribution of wells, lysimeters and Br application line from the subsequent study of Brammer (1996).

100–200 mm/h. It simply meant that K^*, the appropriate matrix property, was less than 5–10 mm/h. McDonnell (1990a) argued that local bypassing required only that $V_o > K^*$. Again, the key feature being the development of soil water potentials >0 cm H_2O at the soil-bedrock interface even though vertical wetting front propagation was still occurring downwards through the upper soil.

5.3. A perceptual model of macropore flow and old water displacement

McDonnell (1990b) noted that as infiltrating new water moved to depth, free water perched at the soil-bedrock interface, and water 'backed-up' into the matrix, where it mixed with a much larger volume of stored, old matrix soil water (Fig. 4(C)). Matric-potential evidence from the responsive mid-slope hollow (Pit 5, Fig. 2) showed that this water table was dissipated by the moderately well-connected system of pipes at the mineral soil-bedrock interface. The flux relationship between vertical bypass infiltration and lateral pipeflow was equal, because there was a significant time delay between water-table perching and subsequent dissipation of positive pore-pressure in the soil. McDonnell (1990a) reasoned that this delay was the critical process necessary

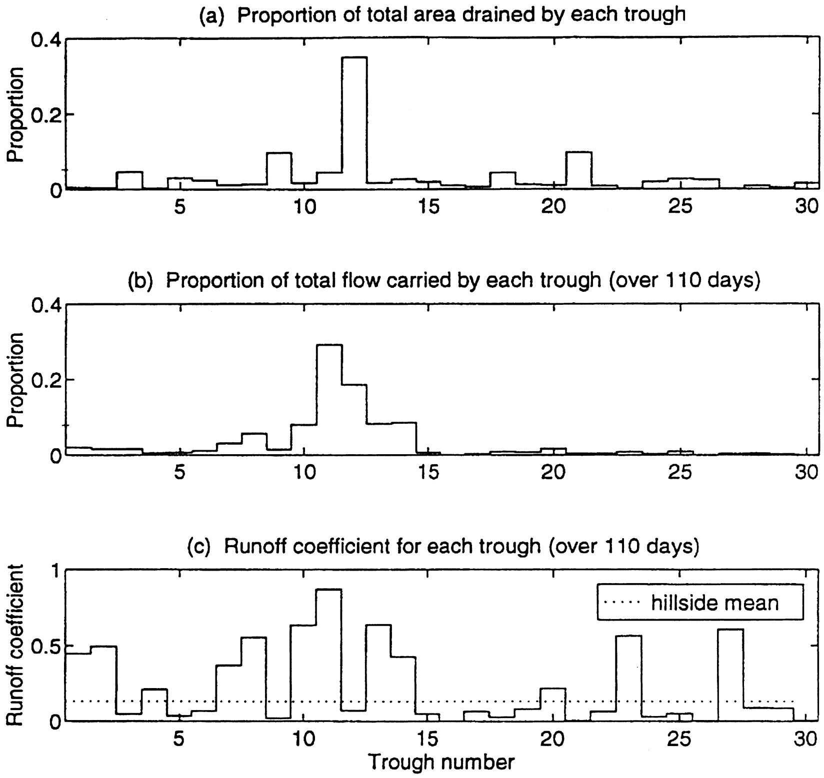

Fig. 9. Spatial distribution of trough characteristics: (a) subcatchment area; (b) total flow over entire study period; (c) runoff coefficients (from Woods and Rowe, 1996).

to shift the new water signatures to that of old water at the hillslope scale.

Isotopic data from Pit 5 throughflow (Sklash et al., 1986; McDonnell et al., 1991b) showed that old water dominated subsurface flow at these mid-slope hollow sites by up to 85%. McDonnell (1990a) reasoned that the pipes distributed this mixture of newly bypassed rainfall and mixed stored water downslope to the first order channel bank. The shift from new to old water was expected to occur on the slope, as indicated in Fig. 4(C). Stewart and McDonnell (1991) showed that between-storm matrix water varied in age from approximately one week at the catchment divide (near Site A; Fig. 2) to over 100 days at the main M8 channel margin, supporting the notion of very short hillslope water residence times. With hindsight and subsequent hillslope excavations by a 1992 Earth Watch research team (McDonnell, unpublished data) we know that soil pipes at the soil-bedrock interface are not continuous beyond about 25 cm, thus affecting the applicability and acceptance of the above-stated pipe flow component of the perceptual model. Furthermore, the disconnection between essentially point tensiometer measurements and pipeflow was a speculative component of this reasoning.

6. Whole hillslope trenching and flow collection

6.1. The trench excavation

Woods and Rowe (1996) established a subsurface collection system along the base of a hillslope hollow on the left bank of the stream draining the M8 catchment (Fig. 2). A vertical face 60 m long and 1.5 m high was cut across the toe of the hillslope. Thirty subsurface flow

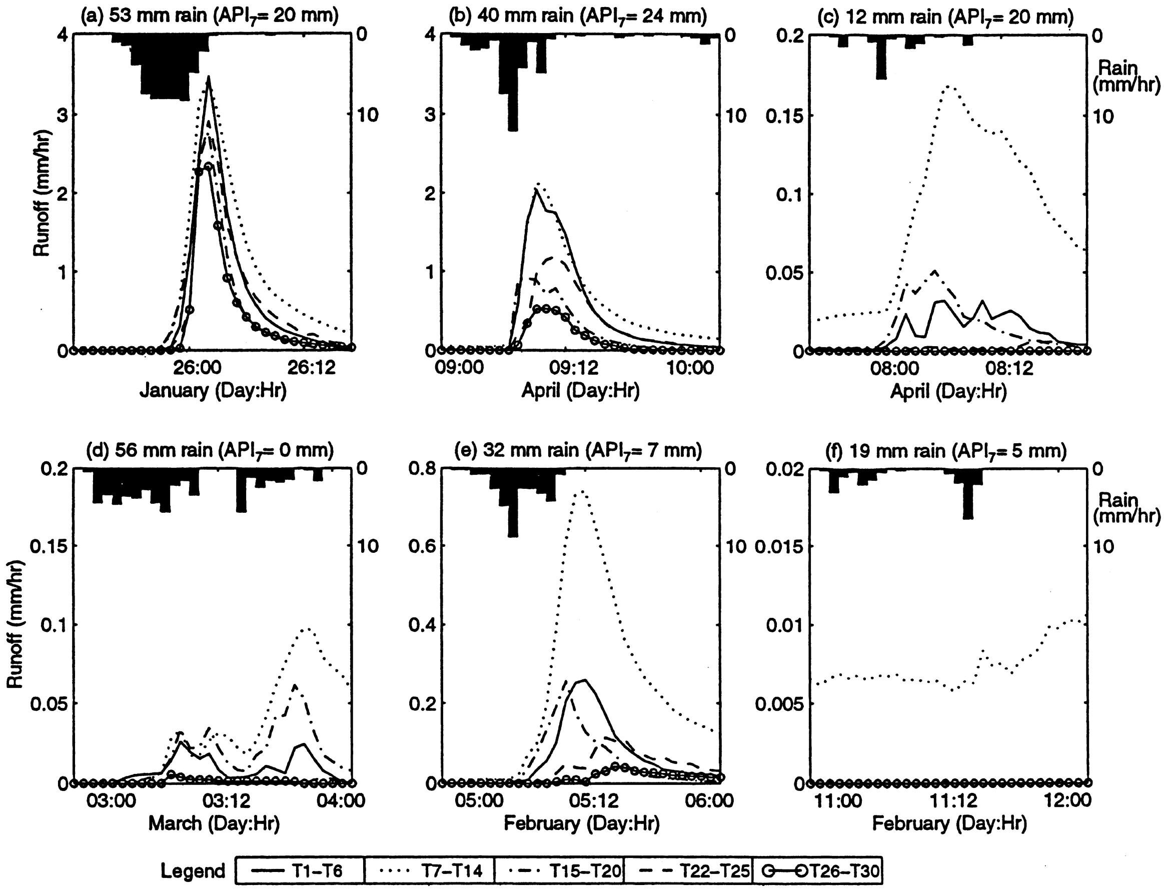

Fig. 10. The effects of storm size, pre-storm wetness, and topographic features on runoff from groups of troughs. Runoff is most uniform in space for large storms on an already wet hillslope. Storm size decreases across the page and pre-storm wetness (API_7) decreases down the page. Troughs T7–T14 drain the most convergent section of the hillslope (from Woods et al., 1997).

collection troughs were installed end-to-end, across the base of the excavated face at the soil-bedrock interface (Fig. 8). The troughs were sealed to the cut face and covered. Flow collected in the troughs was routed to tipping bucket flow meters and recorded from November 1992 through 1993.

6.2. Subsurface flow volumes

Subsurface flow from the hillslope was highly variable in both magnitude and timing. Neighboring collection troughs showed unexpected differences in flow rates. Fig. 9 shows the spatial distribution of the subcatchment areas and proportion of total flow among the troughs. Woods and Rowe (1996) noted that the troughs draining the central hollow displayed higher-than-expected volumes of flow. Consequently, the amount of subsurface flow drained by each trough did not always correspond with the estimated upslope subcatchment (surface topographically derived trough catchment) area (Fig. 9). In particular, Woods and Rowe (1996) estimated that trough 12 (Fig. 9) would drain 35% of the hillslope, yet it did not dominate hillslope flow. Trough 11 drained only 5% of the total hillslope area but yielded the highest flow volumes during events (Woods and Rowe, 1996). This suggested that surface topography alone did not determine the trough sub-catchment divide or the subsurface contributing area. Trough runoff coefficients were also variable across the hillslope (Fig. 9(c). When the troughs were grouped by topographic feature (convergence and divergence), variability in flow across the hillslope cut face was reduced but still observed (Woods and Rowe, 1996).

6.3. Development of a new topographic index of spatially variable subsurface runoff

Woods et al. (1997) developed a new topographic index to describe the time-varying spatial pattern of subsurface runoff at the trench face. They found that the degree of variability changed markedly with antecedent moisture conditions. Fig. 10 shows 30 h runoff hydrographs for six events. Runoff rates were calculated as the volume rate per unit subcatchment area. Five sets of adjacent trough data from Woods and Rowe (1996) were grouped based on common topographic shape characteristics (convergent, divergent, and planar). Troughs T7–T14 drained the most convergent portion of the hillslope and delivered proportionately more runoff than other groups under drier conditions with smaller storms (Fig. 10). Trough runoff response across the trench was most uniform for larger storms and under wetter antecedent conditions. As storm size and antecedent precipitation decreased (shown in Fig. 10 moving across and down, respectively) the difference in runoff response increased until only T7–T14 responded under a 19 mm event with a API_7 of 5 mm. The divergent and planar groups of troughs delivered less runoff both at storm peaks and during storm recessions under the range of conditions shown in Fig. 10.

6.4. Development and application of a new index

Woods et al. (1997) modified the TOPMODEL index of Beven and Kirkby (1979) to allow for the prediction of time varying source areas of subsurface runoff observed in the Maimai trench data. The model

> … uses an approximation of Darcy's law for the case of shallow topographically driven flow, a nonlinear expression for space- and time-driven recharge, and makes a quasi-steady state assumption that subsurface runoff balances recharge (Woods et al., 1997, page 1064).

Rainfall, soil depth, and saturated hydraulic conductivity were assumed to be uniform in time and space. Three assumptions distinguish the Woods et al. (1997) model from TOPMODEL: (1) the soil lies above an impermeable layer; (2) saturated hydraulic conductivity does not vary with depth; and (3) the pattern of recharge can be estimated using a simple nonlinear function of local and catchment-average saturated zone thickness. In assumption (3), recharge was spatially variable rather than a catchment-wide average and relates to the depth to the saturated zone. The most significant departure from standard hillslope modeling in this approach was the application of a simplified treatment of spatially variable recharge. The model was calibrated for troughs T26–T30 with 6 months of 6-h data and applied to adjacent trough groups.

Woods et al. (1997) concluded that the variability of subsurface runoff depends on both topography and wetness. Their index of spatially variable subsurface runoff assumed a parametric relationship for recharge as a function of topographic position and catchment wetness. The model also produced an index for saturated zone thickness. Numerous assumptions were necessary to limit the complexity of the model. Some of the assumptions included: uniform soil depth, variable recharge/transpiration related to depth to water table, persistent water table between storms, and no variation in K_{sat} with depth. Understandably, complex catchment heterogeneity cannot be fully represented in a hydrologic model; however, some of these assumptions might significantly impact hydrologic behavior of hillslope runoff. Mosley (1979, 1982), Pearce et al. (1986), Sklash et al. (1986), and McDonnell (1990a,b) have indicated that the soil depths at Maimai vary significantly and often follow catenary sequences with shallow soils on the ridges and deepest soils in the hollows (range: 0.2–1.8 m, McDonnell, 1990a). Soil depth variability was previously thought responsible for much of the inconsistency in subsurface flow not apparent from surface observation (Mosley, 1979; Pearce et al., 1986; McDonnell, 1990a). Woods et al. (1997) attributed some of the hydrological variability between the divergent/planar hillslope sections and the hollow sections to variable recharge of the water table. A larger proportion of precipitation was assumed to reach a water table that was closer to the soil surface. This assumption of a persistent water table between storms was contrary to evidence presented by Mosley (1979) and McDonnell (1990a). Furthermore, assumptions regarding the nature and spatial distribution of

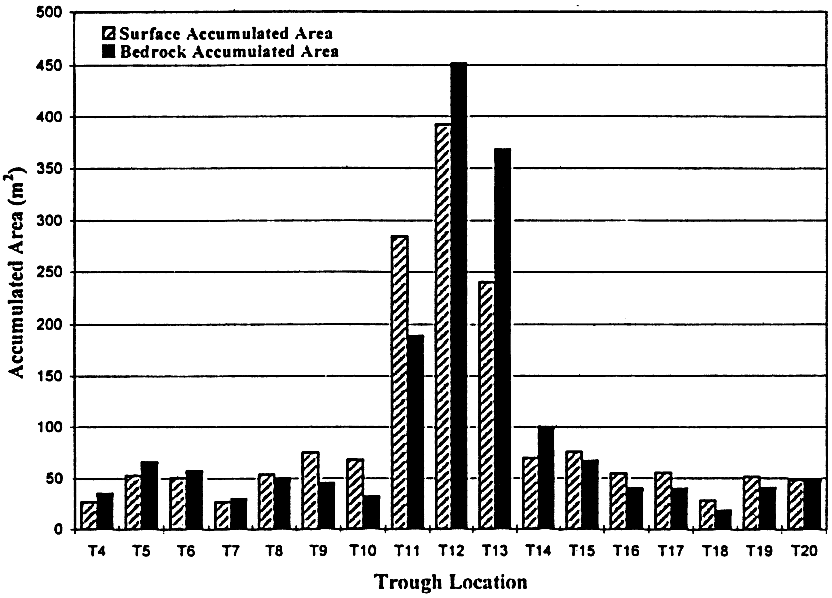

Fig. 11. Accumulated areas calculated for each trough subcatchment based on surface topography and bedrock topography (from Freer et al., 1997).

transpiration and the efficiency of flow rather than moisture concentration in hollows versus shoulder sections remain to be tested.

6.5. A perceptual model of hillslope flow: a single pit does not make a hillslope

The spatial variability across the trough face, even when grouped by topographic features, revealed that flow data from small troughs are difficult to extrapolate to larger scales. The observations of the Woods and Rowe (1996) study suggest that a model of hillslope response must not only adequately describe the spatial variability, but also take into account the effect of scale and the physical controls on the production of runoff. For subsurface flow, Woods and Rowe (1996) reasoned that it was not an 'effective' contributing area that determined the runoff (as is the case with Horton overland flow and saturation overland flow), but rather the size of the vertical saturated soil moisture storage.

Although useful, the Woods and Rowe (1996) study treated the hillslope as a 'black box' and did not collect data on flow paths and mechanisms controlling flow to the troughs. Woods and Rowe (1996) documented the extreme spatial variability in runoff across a seemingly planar hillslope, as indicated, but not tested, by Mosley (1979, 1982). Woods and Rowe (1996) could not explain why this variability occurred, and thus advocated grouping trench sections. Woods and Rowe (1996) clearly demonstrated the dangers of extrapolating data from single throughflow pits to larger scale hillslopes and watersheds.

7. Hillslope scale bromide tracing

7.1. One hydrologist's signal is another's noise

McDonnell (1997) argued that the bedrock topography and not the surface topography might explain the unexpected flow distribution at the Woods and Rowe (1996) trench face. For instance, Woods and Rowe (1996) concluded that there was

> … some positive correlation between flow and area but troughs 12, 9 and 21 did not dominate flow as might be expected (page 68).

Since water perches at the soil-bedrock interface, this surface may have determined the ultimate pathway of flow and the spatial pattern of seepage at the

Table 2
Subsurface contributing area and Br^- recovery percentages for the seven monitored events (modified from McDonnell et al., 1998)

	Rainfall event						
	1	2	3	4	5	6	7
Peak subsurface saturated area (%)	80	30	25	50	60	65	60
Bromide recovery (% of total)	30.1	1.1	1.4	26.1	20.9	8.5	6.7
Total rainfall (mm)	41.5	30.2	24.4	83	62	74	61
Rainfall duration (h)	18.3	28.2	16	74.3	22.2	63	37.8
API_7	10.8	6.4	5.1	4.1	0.1	9.1	0
Peak intensity (mm/h)	6.6	5.0	5.2	9.8	10.0	10.4	10.8
Total rainfall for entire study period (45 days): 376.1 mm							

trench. Therefore the process of interest may have in fact been saturation from below; a wetting up from the bedrock surface into the soil profile. McDonnell (1997) argued that differences in flowpaths not immediately related or predictable by standard topographic surveys or the topographically-based modeling approaches of Woods and Rowe (1996) may have resulted from the discrepancy between the bedrock surface topography and the soil surface topography. McDonnell (1997) pointed to the differences in the surface and bedrock upslope area distributions (Fig. 11) to explain the higher than expected contributions from trough 13 and some of the variability across the trench face. There is a shift in area at the central portion of the trench toward troughs 13 and 14 (Fig. 11). McDonnell (1990a) demonstrated that water tables were extremely short-lived. Thus he suggested, between events, the surface topography might determine unsaturated zone flow direction with gravity and matric potential together controlling the gradient of total potential (McDonnell, 1997). Bedrock topography, on the other hand, was hypothesized to control saturated subsurface flow. In a reply to McDonnell (1997); Woods and Rowe (1997) agreed that bedrock topography is more suitable than surface topography for delineating the area draining to each trough. However, Woods and Rowe stated that when troughs are grouped in sets of four or five, as in their earlier work (Woods and Rowe, 1996), the discrepancy between bedrock and surface delineated upslope-contributing areas is negligible. Furthermore, the sensitivity of the accumulated area calculations to variations in the surveyed and interpolated surface upslope of the troughs may limit the reliable estimates of single trough accumulated area (Woods and Rowe, 1997).

7.2. The Br^- line source injection

During March to May 1995, Brammer et al. (1995) performed a hillslope–scale tracer experiment on the trenched hillslope described by Woods and Rowe (1996). The objective of the study was to determine the hillslope-scale flowpath and mixing processes controlling flow and tracer breakthrough at the trench face under natural rainfall conditions. In particular, the role of surface topography versus bedrock topography in controlling hillslope-scale flow was examined. A 5 m grid of suction lysimeters and co-located maximum-rise wells was installed in a 30 m × 30 m plot 5 m upslope from the trench face (Fig. 8). A line source of 3 kg LiBr was applied 35 m upslope from the trench face. Antecedent moisture conditions were relatively high; 17 mm of rainfall had occurred in the previous two days. The first rainfall event occurred 12 h after the tracer application. Seven storm events, ranging in magnitude from 24.4 to 41.5 mm, were recorded and sampled during the 45-day period.

7.3. Hillslope water-table response and Br^- recovery

McDonnell et al. (1998) reported that the spatial distribution of subsurface saturation and water table development on the slope appeared to be controlled by depressions in the bedrock topographic surface. Peak piezometer response for each event was greatest along the transect 12A–12E (Fig. 8) that extended up the central hollow. The center of the grid (position 12C) was the area of greatest and most persistent water

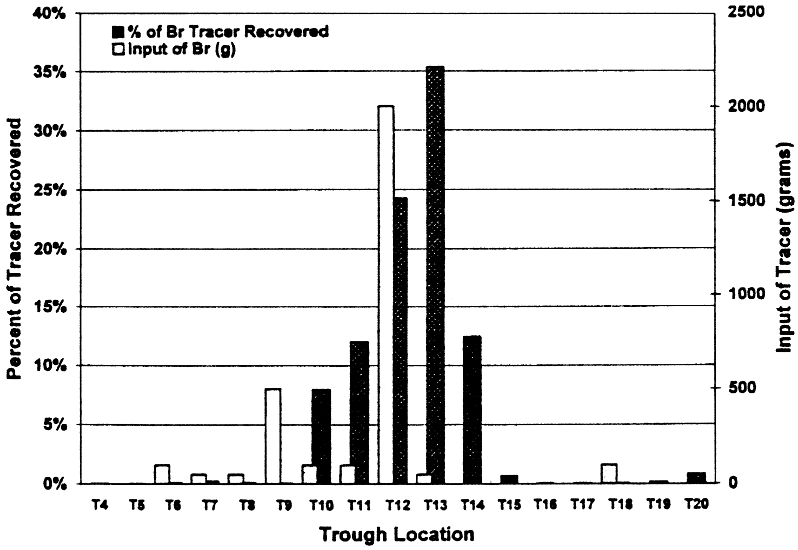

Fig. 12. Histogram of estimated direct tracer input into trough subcatchment areas computed based on surface topography. Tracer recovery is also plotted by trough and shows clearly that recovery was not related to amount applied (McDonnell et al, 1998).

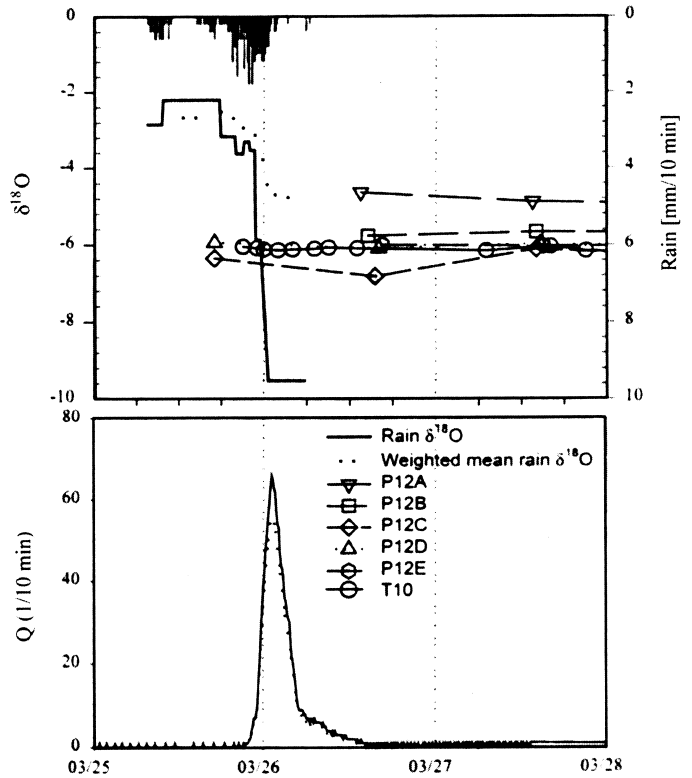

Fig. 13. $\delta^{18}O$ values from rainfall, maximum rise wells, trough 10 and a two-component hydrograph separation for trough flow. The faint dashed line below the hydrograph denotes the split between old and new water (from McDonnell et al., 1998).

table development—saturated conditions persisted typically 3–5 days after each event. Soil depths were more shallow downslope of position 12C toward the trench face. Consequently, Brammer (1996) and McDonnell et al. (1998) interpreted this bedrock depression as a reservoir or local impedance to the downslope movement of saturated subsurface stormflow.

The extent of subsurface contributing area was related directly to rainfall intensity and amount (Table 2). The connected area of subsurface saturation ranged from approximately 1250 m^2 or 82% of the grid area during the first event, to 30% and 25% of the grid area during the 2 smaller subsequent events. The remaining events displayed subsurface saturation extents similar to the first event. Br^- recovery at the troughs varied as an apparent function of rainfall intensity/duration and time since tracer application (McDonnell et al., 1998). The first event accounted for the largest recovery of Br^- (30%) (Table 2). Low rainfall amounts and intensities in events #2 and #3 resulted in lower recovery (<4%). Subsequent events (#4–7) were more representative of local precipitation conditions. Events #4 and #5 resulted in 26 and 22% recovery, respectively, while the last two measured events resulted in decreasing amounts of tracer recovery (~8%) as the tracer moved off-site.

Fig. 12 shows the estimated amount of Br^- tracer that was applied directly to each trough subcatchment (Freer et al., 1997). This was calculated from the line application of Br^- across each trough sub-catchment based on surface topographic maps in Woods and Rowe (1996). The calculated total recovery of Br^- across the face is also shown in Fig. 12 for each trough subcatchment. Recovery for the entire hillslope was extremely high (82%), reflecting the wet, steep nature of the site. The variation of tracer input across the face did not correspond well with tracer recovery. No tracer recovery occurred in troughs T6, T9 and T18, which received approximately 700 g (23%) of the tracer input. Troughs with subcatchments that did not receive large or any input of tracer (T10, T11, T13, and T14) were responsible for large percentages of Br^- recovery (7–35%). McDonnell et al. (1998) attributed this to the spatial pattern of bedrock topography and the problems associated with delineation of subcatchments based on surface topography where subsurface topographic controls and not surface topography might regulate tracer pathways and mass export distribution. There was no significant correlation between percentage of Br^- recovered and subcatchment areas and volumes based on surface topography ($r^2 = 0.44$ and 0.50, respectively) (McDonnell et al., 1998).

Despite the high proportion of tracer recovery, there was little evidence of 'new water' in high flow troughs during the 7 events. A hydrograph separation from a high runoff, high Br^- export trench section (Fig. 13) showed that negligible new water was exported from the trench during rainfall (McDonnell et al., 1998). Since 1998, several papers have questioned the seemingly paradoxical findings of tracer breakthrough with minimal new water. In other words, "How can rapid breakthrough of Br^- and concentrations of up to 20 ppm Br^- in hillslope runoff occur with minimal new water detectable at the trench face?" Reflecting on these data, we addressed this Br/$\delta^{18}O$ conundrum by calculating the minimum detectable fraction of new water in trench runoff. We assume that tracer was distributed uniformly over the hillslope in order to calculate the expected delivery of Br^- in trench runoff. If one assumes that 10% new water contribution to hillslope runoff is the minimally detectable amount of new water that we might realistically be able to achieve with a two-component isotope hydrograph separation (Genereux, 1998) and we assume, conservatively, an evenly distributed line source of bromide (100 m long and 2 cm wide), then the surface area of the 3 kg bromide source applied by Brammer (Table 2) would be 2 m^2. If we then estimate the new water passing through the 2 m^2 application line as the amount of rainfall falling on this 2 m^2 area of the 1200 m^2 trenched hillslope, then:

$$0.05\ \mathrm{m} \times 1200\ \mathrm{m}^2 = 60\ \mathrm{m}^3,$$

(total rain × hillslope area = volume of rain water entering hillslope)

$$60\ \mathrm{m}^3 \times 0.4 = 24\ \mathrm{m}^3,$$

(total rainfall on hillslope × runoff ratio = hillslope runoff)

$$0.1 \times 24\ \mathrm{m}^3 = 2.4\ \mathrm{m}^3,$$

(proportion of rain water in trench runoff × total hillslope runoff = volume of rainwater in trench runoff)

$2.4 \text{ m}^3/60 \text{ m}^3 = 0.04,$

(volume of rainwater in trench runoff/total rain on hillslope = new water runoff ratio)

$2.4 \text{ m}^3/1200 \text{ m}^2 = 0.002 \text{ m} \times 2 \text{ m}^2$

$= 0.004 \text{ m}^3$ trench run off from application line

(new water in trench runoff/hillslope area = runoff depth × application line area = trench runoff from application line)

$2.76 \text{ kg} \times 0.004 / 24 \text{ m}^3$ total trench flow = 4.6 ppm,

(Br^- on application line × new water from application line/total trench runoff = mean expected concentration in trough runoff)

The results of this simple exercise are consistent with the results presented by Brammer (1996) where he found rapid delivery of Br^- in trench runoff with small proportions of new rainwater. The highest recorded trench outflow concentrations of Br^- were 20 ppm, detected in the first 41.5 mm event with higher concentrations of Br^- (50 ppm) detected in sampled wells and lysimeters on the hillslope (Brammer, 1996). The analysis completed above demonstrates that the rapid delivery of Br^- with little new water is possible. The combined isotopic and applied tracer approach highlights the consistency of the observations and independent corroboration of this finding.

7.4. Controls on subsurface saturation

A lack of correlation between soil depth, water table response, and the surface-derived upslope accumulated area indicated that physical factors, other than surface topography alone, controlled movement of water and tracer on the hillslope (McDonnell et al., 1998). The first event described by Brammer (1996) showed tracer breakthrough within 6 h of rainfall onset. This rapid movement of tracer 35 m downslope from the application line to the trench face indicates the existence of preferential flow pathways. Matrix flow based on a simple Darcy calculation using K_{sat} values from Mosley (1979) and McDonnell (1989) indicate that it would take approximately 30 days to cover this straight-line distance. McDonnell (1990a) suggested that extended macropore networks might explain old water dominated rapid runoff response. Maximum flow velocities reported by Mosley (1982) of $V_{max} = 0.42$ cm/s suggest that water could travel the 35 m from the top of the slope to the troughs in roughly 2.5 h. Data from McDonnell et al. (1998) indicated that preferential flow at the soil-bedrock interface (but not necessarily continuous connected pipes) occurred over the entire hillslope. Movement along the soil-bedrock interface allowed water to be translated rapidly downslope and laterally across the hillslope (Brammer, 1996). These rapid flow paths appeared to be dependent upon extent of subsurface saturation. No downslope-saturated wedge could be delineated from the transient groundwater data; rather complete, albeit of variable depth, hillslope water table development occurred (McDonnell et al., 1998).

7.5. A bedrock topographic index for subsurface runoff

Freer et al. (1997) investigated the correlation between the surface topography and the bedrock topography of the trench hillslope through computation of the TOPMODEL topographic index $\ln(a/\tan\beta)$ of (Beven and Kirkby, 1979), where a is the upslope contributing area and $\tan\beta$ was the local slope angle. This index is based on the multidirectional flow algorithm of Quinn et al. (1991) derived with a high resolution topographic survey data (approx. 800 points, Woods, personal communication) and bedrock surface mapping (99 separate points) (Brammer, 1996). Higher index values were associated with increased subsurface saturation propensity and potentially higher rates of flow (Freer et al., 1997). Each index (topographic surface and bedrock surface derived) predicted cumulative trench flow patterns well. Variability between events for individual troughs was greater than the variability between the two indices (Freer et al. 1997).

Correlations of runoff patterns across the trench face to both surface and bedrock based index distributions were worse for small precipitation events. The spatial correlation between the surface and bedrock indices overall was high. The distribution of bedrock-accumulated areas at the trench face appeared to match the distribution of recovered Br^- tracer (McDonnell et al., 1998). There was a shift in accumulated area in the central trench portion toward troughs T13 and T14 when the bedrock-accumulated area was used. McDonnell et al. (1998) suggested that this may explain the higher-than-expected contributions from trough T13 and some of the variability witnessed by Woods and Rowe (1996) across the trench face, given that saturated subsurface flow appeared to relate more closely to the bedrock-accumulated area and the bedrock topographic surface. The contrast between trough flow from T11 and T12 that Woods and Rowe (1996) had difficulty rationalizing appeared to be explained by the bedrock-accumulated area in Fig. 12. The explanation appears to lie in the impermeable bedrock surface itself—not only did it control the spatial pattern of mobile subsurface stormflow, but flow along it appeared to enable the remobilization of Br^- tracer that had been trapped in small depressions (Fig. 4(D)) following the first event after the line source application.

7.6. A bedrock topography perceptual model

The hillslope maximum-rise piezometer response, subsurface flow Br^- concentrations, and flow and breakthrough characteristics at the trench face support the hypothesis of bedrock surface control on subsurface storm flow (McDonnell et al., 1998). The qualitative interaction of the mobile and immobile regions of both soil and lateral subsurface flow systems is shown in Fig. 4(D) Rapid lateral movement within this saturated zone appears to occur along narrow ribbons of highly mobile flow that exploit depressions in the bedrock surface (McDonnell et al, 1996; Burns et al., 2001). The conclusions and perceptual model are corroborated by other watershed studies at Panola mountain, Georgia (McDonnell et al., 1996) and Plastic Lake, Ontario, Canada (Peters et al., 1995). What remains uncertain in this perceptual model is possible communication between bedrock water and transient subsurface flow as identified recently in other hillslopes by Montgomery et al. (1997) and Onda et al. (2001). While these studies have shown upward hydraulic gradients through underlying saprolite or sandstone bedrock, it remains to be seen whether or not this might occur at Maimai, where the cemented conglomerate is extremely tight and not fractured (at the hillslope scale). Pressure wave induced effusion of old water may be an alternative explanation for the rapid old water mobilization at the hillslope scale (Torres et al., 1998); however, the advected transport of Br^- off the hillslope at Maimai argues against a pressure wave induced displacement. This is a fruitful area for further study at Maimai and elsewhere.

8. Conclusions and outlook for the future

This paper synthesizes the progression of ideas in the development and ongoing modification of a perceptual model of hillslope flow at the Maimai catchments. Each data set reviewed reveals not only a cumulative understanding of catchment behavior, but alternative interpretations of the hillslope subsurface flow system. The initial single-technique approaches of Mosley (1979) show the limitations and often misleading interpretations from dye tracers alone. Subsequent isotopic studies by Pearce et al. (1986) and Sklash et al. (1986) showed clearly that stored water comprised the majority of channel stormflow; however, their isotope-oriented approach did not enable them to develop a mechanistic understanding of the processes. Studies that followed (McDonnell, 1990a,b; McDonnell et al., 1991a,b) combined soil physics techniques with isotopic and chemical tracing to reconcile the earlier different process interpretations. Although single throughflow pits continued to be the indicator of subsurface flow timing and magnitude, further study by Woods and Rowe (1996) showed that flow varied widely across a hillslope section—making the single pit observations of the previous studies highly suspect. Furthermore, a new index based on these findings was able to predict the spatial and temporal dynamics of flow from groups of troughs at the trench face (Woods et al., 1997). Observations by McDonnell et al. (1998) revealed significant bedrock surface control on mobile subsurface flow timing and tracer breakthrough. While previous works have treated the soil-bedrock interface as a smooth linear boundary, McDonnell et al. (1998) demonstrated that small depressions and micro-topographic relief in the bedrock surface exert large control on water mobility and mixing. In particular, the bedrock surface appeared to determine the pathway of mobile subsurface water flow and tracer breakthrough during events at the hillslope scale.

The preponderance of data from the Maimai catchments and the hillslope units within it suggest that the following features are common at the site:

- high infiltration rates well in excess of maximum precipitation intensities,
- transient water table development on hillslopes,
- topographic convergence of water flow into hollows,
- vertical bypass flow to depth with large events and more uniform wetting front propagation with small and protracted events,
- rapid lateral throughflow response to precipitation following threshold water table development,
- two domains of lateral throughflow that include rapid (albeit disconnected) pipe flow at the soil bedrock interface and slow, more uniform matrix flow,
- old water dominated throughflow (resident soil water as opposed to new rainfall),
- high degree of throughflow variability across seemingly planar hillslopes, and
- bedrock topographical control on the spatial distribution of mobile subsurface saturated flow.

These points notwithstanding, some of the key aspects of hillslope hydrology at Maimai remain unresolved, including:

- relationships between pressure wave propagation and particle velocity in the context of rapid effusion of old water,
- inter-event connections and between event disconnections among different landscape units within catchments (i.e. hillslopes, hollows, and riparian zones), and
- how process finding from <10 ha headwater catchments scale to larger fractions of the Maimai landscape and whether or not these larger landscape units are simply a linear superposition of the many smaller <10 ha subcatchments that dominate the landscape.

These issues are at the forefront of small catchment hydrology research (McDonnell et al., 2001; McDonnell and Tanaka, 2001; Uhlenbrook et al., 2001) and the development and application of hydrological and hydrochemical models. The work to date at Maimai has focused on hillslope units almost exclusively. The relative roles of hillslopes, riparian zones, and hyporheic zones in streamflow generation and composition are poorly understood at the site. Little research has attempted to integrate hillslope, riparian, and hyporheic approaches with catchment runoff in order to estimate their relative roles in stormflow and baseflow generation, the partitioning of old and new water, impact on stream water residence time, and subsequent control of stream chemistry. Furthermore, little is understood about the integration of flowpaths and flow sources through time.

Recent research has indicated that on the event time scale, hillslope runoff signatures may not be apparent in stream stormflow in many catchments. At the Maimai watershed, the volume of the riparian zone in relation to the volume of the hillslope zone is small in comparison to other catchments such as Panola mountain, Georgia (Burns et al., 2001) and Sleepers river, Vermont (McGlynn et al., 1999). Among the Maimai sub-catchments, the ratio of the volume of the riparian reservoir to the hillslope reservoir increases with increasing basin scale. The Maimai watersheds provide an optimal environment for examination of the riparian reservoir across spatial scales and the deconvolution of the relative roles of hillslopes, riparian zones, and hyporheic zone/stream channels. Related research at other catchments, and catchment intercomparisons with alternative geological settings, morphological histories, and climatic regimes will greatly accelerate understanding of these processes and linkages (Jones and Swanson, 2001). The rich understanding of hillslope hydrology gained at the Maimai catchments clearly demonstrates the advantages of benchmark research catchments and successive intensive investigation utilizing innovative techniques and formulation of hypotheses built upon prior research.

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References

Abdul, A.S., Gillham, R.W., 1984. Laboratory studies of the effects of the capillary fringe on stream flow generation. Water Resources Research 90, 691–698.

Beven, K.J., 1993. Prophecy, reality, and uncertainty in distributed hydrological modeling. Advances in Water Resources, 16, 41–51.

Beven, K.J., Kirkby, M.J., 1979. A physically-based variable contributing area model of basin hydrology. Hydrological Sciences Bulletin 24, 43–69.

Bonell, M., 1998. Selected challenges in runoff generation research in forests from the hillslope to headwater drainage basin scale. Journal of the American Water Resources Association 34 (4), 765–785.

Brammer, D.D., 1996. Hillslope hydrology in a small forested catchment, Maimai, New Zealand. MS Thesis, Department of Forestry, State University of New York, Environmental Science and Forestry, Syracuse, NY.

Brammer, D.D., McDonnell, J.J., Kendall, C., Rowe, L.K., 1995. Controls on the downslope evolution of water, solutes and isotopes in a steep forested hillslope. EOS, Transaction of the American Geophysical Union 76 (46), 268.

Burns, D.A., McDonnell, J.J., Hooper, R.P., Peters, N.E., Freer, J.E., Kendall, C., Beven, K., 2001. Quantifying contributions to storm runoff using EMMA analysis and hydrologic measurements at the Panola Mountain Research Watershed, (Georgia, USA). Hydrological Processes 15 (10), 1903–1924.

Freer, J., McDonnell, J.J., Beven, K.J., Brammer, D.D., Burns, D., Hooper, R.P., Kendall, C., 1997. Topographic controls on subsurface storm flow at the hillslope scale for two hydrologically distinct small catchments. Hydrological Processes 11 (9), 1347–1352.

Freeze, R.A., 1972. Role of subsurface flow in generating surface runoff: upstream source areas. Water Resources Research 8, 609–623.

Genereux, D., 1998. Quantifying uncertainty in tracer-based hydrograph separations. Water Resources Research 34 (4), 915–919.

Gillham, R.W., 1984. The effect of the capillary fringe on water table response. Journal of Hydrology 67, 307–324.

Hewlett, J.D., Hibbert, A.R., 1967. Factors affecting the response of small watersheds to precipitation in humid areas. Proceedings of first International Symposium on Forest Hydrology, pp. 275–253.

Hewlett, J.D., Cunningham, G.B., Troendle, A.C., 1977. Predicting stormflow and peakflow from small basins in humid areas by the R-index method. Water Resources Bulletin 13, 231–253.

Jones, J.A., Swanson, F.J., 2001. Hydrologic inferences from comparisons among small basin experiments. Hydrological Processes, 15 (12), 2363–2366.

McDonnell, J.J., 1989. The age, origin and pathway of subsurface stormflow in a steep, humid headwater catchment. Doctoral thesis, University of Canterbury, Christchurch, NZ.

McDonnell, J.J., 1990a. A rationale for old water discharge through macropores in a steep, humid catchment. Water Resources Research 26, 2821–2832.

McDonnell, J.J., 1990b. The influence of macropores on debris flow initiation. Quarterly Journal Engineering Geology 23, 325–331.

McDonnell, J.J., 1997. Comment on "The changing spatial variability of subsurface flow across a hillside" by Ross Woods and Lindsay Rowe. Journal of Hydrology (NZ) 36 (1), 97–100.

McDonnell, J.J., Owens, I.F., Stewart, M.K., 1991a. A case study of shallow flow paths in a steep zero-order basin: a physical-chemical-isotopic analysis. Water Resources Bulletin 27, 679–685.

McDonnell, J.J., Stewart, M.K., Owens, I.F., 1991b. Effects of catchment-scale subsurface watershed mixing on stream isotopic response. Water Resources Research 26, 3065–3073.

McDonnell, J.J., Freer, J., Hooper, R., Kendall, C., Burns, D., Beven, K., 1996. New method developed for studying flow on hillslopes. EOS 77 (47), 465–472.

McDonnell, J.J., Brammer, D.D., Kendall, C., Hjerdt, N., Rowe, L.K., Stewart, M., Woods, R.A., 1998. Flow pathways on steep forested hillslopes: The tracer, tensiometer, and trough approach. In: Tani (Ed.). Environmental Forest Science. Kluwer Academic Publishing, Boston, pp. 463–474.

McDonnell, J.J., Tanaka, T., 2001. On the future of forest hydrology and biogeochemistry. Hydrological Processes 15 (10), 2053–2055.

McDonnell, J.J., Tanaka, T., Mitchell, M., Ohte, N., 2001. Foreword: hydrology and biogeochemistry of forested catchments. Hydrological Processes 15 (10), 1673–1674.

McGlynn, B.L., McDonnell, J.J., Shanley, J.B., Kendall, C., 1999. Riparian zone flowpath dynamics during snowmelt in a small headwater catchment. Journal of Hydrology 222, 75–92.

Montgomery, D.R., Dietrich, W.E., Torres, R., Anderson, S.P., Heffner, J.T., Loague, K., 1997. Hydrologic response of a steep, unchanneled valley to natural and applied rainfall. Water Resources Research 33 (1), 91–109.

Mosley, M.P., 1979. Streamflow generation in a forested watershed, New Zealand. Water Resources Research 15, 795–806.

Mosley, M.P., 1982. Sursurface flow velocities through selected forest soils South Island, New Zealand. Journal of Hydrology 55, 65–92.

O'Loughlin, C.L., Rowe, L.K., Pearce, A.J., 1978. Sediment yields from small forested catchments, north Westland-Nelson, New Zealand. Journal of Hydrology (NZ) 17, 1–15.

Onda, Y., Komatsu, Y., Tsujimura, M., Fujihara, J., 2001. The role of subsurface runoff through bedrock on stormflow generation. Hydrological Processes 15 (10), 1693–1706.

Pearce, A.J., McKerchar,, A.I., 1979. Upstream generation of storm runoff. In: Murray, D.L., Ackroyd, P. (Eds.). Physical Hydrology—New Zealand Experience. New Zealand Hydrological Society, Wellington, pp. 165–192.

Pearce, A.J., Rowe, L.K., 1979. Forest management effects on interception, evaporation and water yield. Journal of Hydrology (NZ) 18, 73–87.

Pearce, A.J., Rowe, L.K., 1981. Rainfall interception in a multi-storied evergreen mixed forest: Estimates using Gash's analytical model. Journal of Hydrology 48, 341–353.

Pearce, A.J., Stewart, M.K., Sklash, M.G., 1986. Storm runoff generation in humid headwater catchments 1 where does the water come from? Water Resources Research 22, 1263–1272.

Peters, D.L., Buttle, J.M., Taylor, C.H., LaZerte, B.D., 1995. Runoff production in a shallow soil Canadian Shield, basin. Water Resources Research 31 (5), 1291–1304.

Quinn, P., Beven, K., Chevallier, P., Planchon, O., 1991. The prediction of hillslope flowpaths for distributed hydrological modeling using digital terrain models. Hydrological Processes 5, 59–79.

Rowe, L.K., 1979. Rainfall interception by a beech-podocarp-hardwood forest near Reefton, north Westland, New Zealand. Journal of Hydrology (NZ) 18, 63–72.

Rowe, L.K., Pearce, A.J., 1994. Hydrology and related changes after harvesting native forest catchments and establishing *Pinus radiata* plantations Part 2. The native forest water balance and changes in streamflow after harvesting. Hydrological Processes 8, 281–297.

Rowe, L.K., Pearce, A.J., O'Loughlin, C.L., 1994. Hydrology and related changes after harvesting native forest catchments and establishing *Pinus radiata* plantations Part 1. Introduction to the study. Hydrological Processes 8, 263–279.

Sklash, M.G., Farvolden, R.N., 1979. The role of groundwater in storm runoff. Journal of Hydrology 43, 45–65.

Sklash, M.G., Stewart, M.K., Pearce, A.J., 1986. Storm runoff generation in humid headwater catchments 2. A case study of hillslope and low-order stream response. Water Resources Research 22, 1273–1282.

Stewart, M.K., McDonnell, J.J., 1991. Modeling baseflow soil water residence times from deuterium concentrations. Water Resources Research 27, 2681–2694.

Torres, R., Dietrich, W.E., Montgomery, D.R., Anderson, S.P., Loague, K., 1998. Unsaturated zone processes and the hydrologic response of a steep, unchanneled catchment. Water Resources Research 34 (8), 1865–1879.

Uhlenbrook, S., Leibundgut, C., McDonnell, J.J., 2001. Foreword to the Special Issue: runoff generation and implications for river basin modelling. Hydrological Processes in press.

Webster, J., 1977. The hydrologic properties of the forest floor under beech/podocarp/hardwood forest, North Westland. Masters in Agricultural Science thesis, University of Canterbury, NZ.

Woods, R.A., Rowe, L.K., 1996. The changing spatial variability of subsurface flow across a hillside. Journal of Hydrology (NZ) 35 (1), 51–86.

Woods, R.A., Rowe, L.K., 1997. Reply to "Comment on 'The changing spatial variability of subsurface flow across a hillside', by Jeffrey J. McDonnell. Journal of Hydrology (NZ) 36 (1), 101–103.

Woods, R.A., Sivapalan, M., Robinson, J.S., 1997. Modeling the spatial variability of subsurface runoff using a topographic index. Water Resources Research 33 (5), 1061–1073.