2 An Evolving Perceptual Model of Hillslope Flow at the Maimai Catchment

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2.1 INTRODUCTION

Hillslope hydrological investigations have been conducted in a variety of geographical, hydrogeological and climatic settings and have been reviewed recently by Bonell (1993) and Buttle (1994). While extensive in location and scope, most catchment-based hillslope investigations have not continued much beyond a typical two- to three-year funding cycle. This frequently limits the opportunity for formal testing of hypotheses proposed by previous researchers at a site and limits the cumulative understanding of a single hillslope. Therefore, hillslope hydrological observations or *perceptual models* (Beven 1991) are rarely "challenged" by other research scientists carrying on increasingly intensive work on the same hillslope or catchment. One notable exception is the Maimai research catchment in New Zealand. Maimai has been the site of ongoing hillslope research by several research teams since the late 1970s. These studies have facilitated the evolution and development of a very detailed yet qualitative perceptual model of hillslope hydrology at Maimai. This perceptual model has now grown in complexity to defy analytical description; none the less it provides a very useful case study of hillslope hydrological processes and encapsulates all that the field hydrologists have come to recognize as the dominant hillslope runoff processes in steep, humid catchments.

The goal of this chapter is to synthesize the development of ideas relating to an evolving perceptual model of subsurface flow at the Maimai catchment. We hope that this chapter 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 the value of working in a cumulative fashion at a single research site. We feel that this approach has yielded, in the case of the Maimai catchment, a rich understanding of hydrological processes.

2.2 MAIMAI: THE QUINTESSENTIAL STEEP HUMID HILLSLOPE

2.2.1 History of the Maimai Catchment

In 1974, the 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 adverse effects on the quantity and quality of water
supplies and aquatic habitats (large-scale forestry management operations). Data
were gathered on streamflow characteristics, stream sediment yield and slope
stability of mixed beech forest catchments. Though the conversion of native forests
to exotics and the harvesting of mixed forest stands proved to be economically
inviable, the long-term studies of the Maimai catchments continued. The data
collected provides a useful comprehensive and long-term study of hydrological
processes in forested catchments.

2.2.2 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) (Figure 2.1). The catchments lie parallel to
each other and share similar topographic characteristics. Slopes are short (<300 m)
and steep (average 34°C) 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 overland flow are small and limited to
4–7% of catchment areas (Mosley 1979; Pearce *et al.* 1986). The M8 catchment has
been the most intensely studied and is typical of the physical characteristics of the
other Maimai watersheds (Figure 2.2).

Passage of frontal systems from the Tasman Sea across the Paparoa Range (from
westerly and northerly directions) occurs regularly and creates 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; monthly rainfall from December to February averages
165 mm and for the rest of the year between 190 and 270 mm. On average, there are
156 rain days per year with little temperature extreme and only about two 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 (QF/P) is roughly double that of the most responsive basins
documented in eastern United States (Hewlett and Hibbert 1967). Pearce *et al.* (1986)
note that the R index for QF/P averaged for runoff events from rainfalls of greater
than 25 mm is 46% compared with 3–35% for 11 basins distributed between Georgia
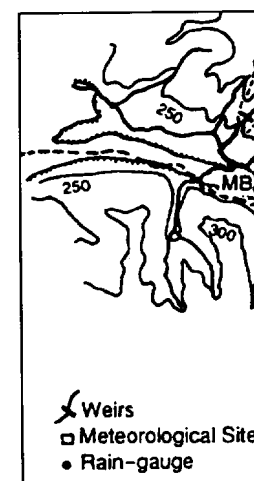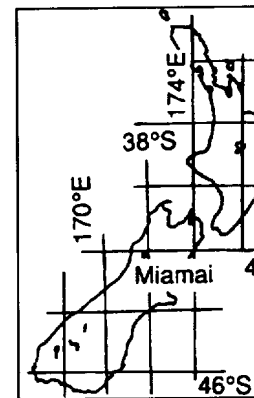and New Hampshire that Hewlett *et al.* (1977) studied.

Figure 2.1 Maimai study
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2.2.3 Vegetation and

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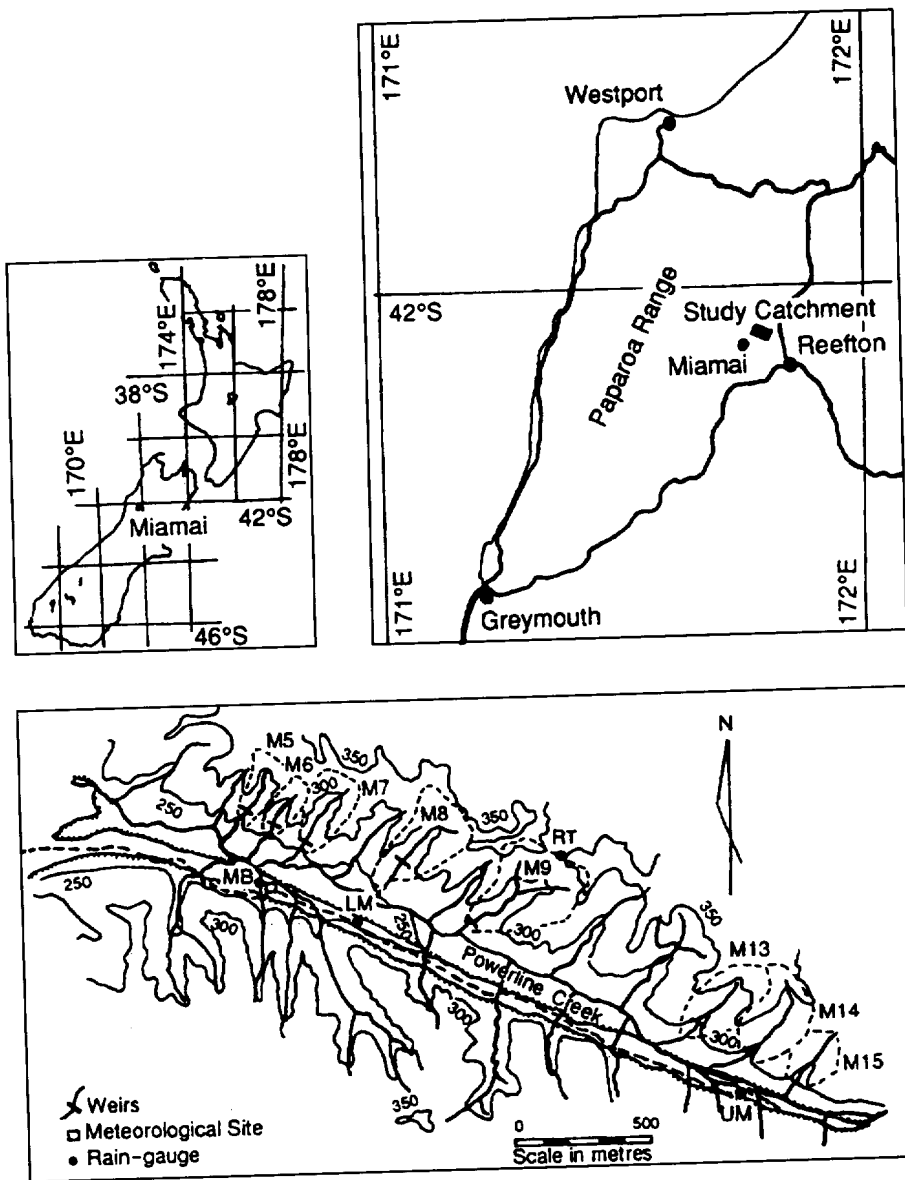
𝄢 Weirs
▫ Meteorological Site
● Rain-gauge

Figure 2.1 Maimai study catchment location, South Island, New Zealand (from Rowe *et al.* 1994)

2.2.3 Vegetation and Soils of the Maimai Catchment

The vegetation is a mixed evergreen beech forest (*Nothofagus* spp.), podocarps, and broadleafed hardwoods. It is multi-stored, with a canopy of 20–36 m high, a dense fern and shrub understory and a fern and moss ground cover. Annual interception
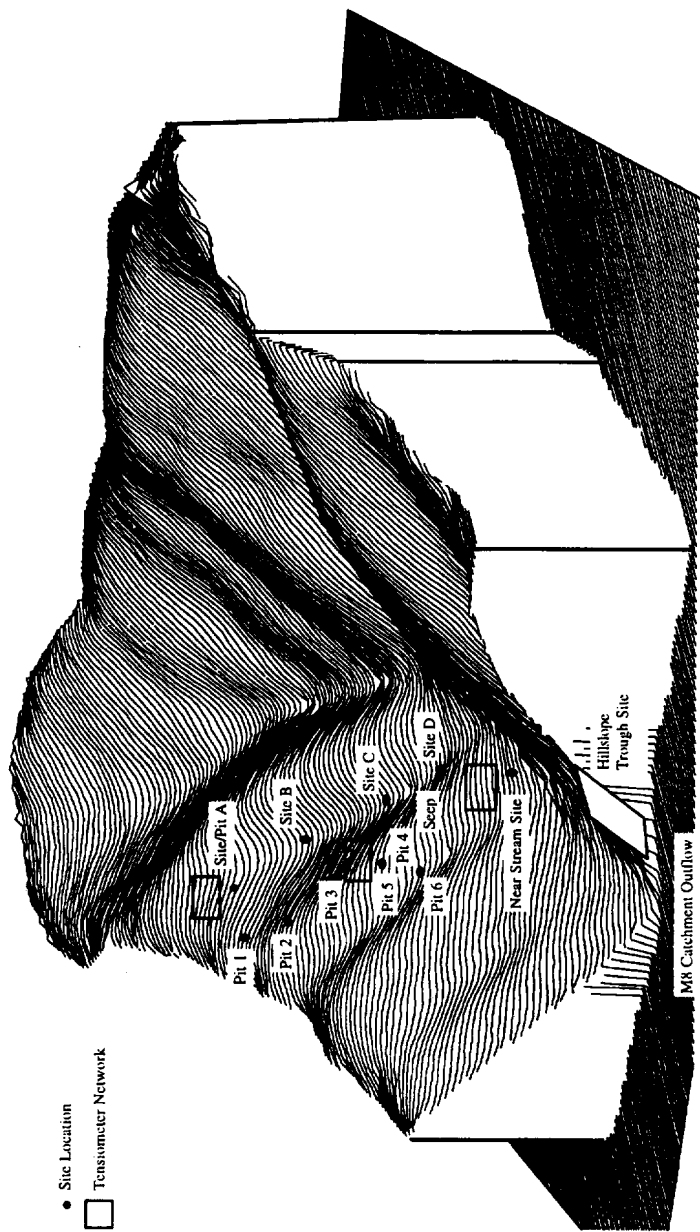
Figure 2.2 M8 catchment study site locations. Sites A–D and Pits 1–6 are original study locations from Mosley (1979). Pearce *et al.* (1986) and Sklash *et al.* (1986) reactivated seven of Mosley's measurement sites and added suction lysimeters and max-rise piezometers at the near-stream site (SL and Well B) and near the M8 stream weir (SL and Well A). McDonnell (1990a) extended the instrumentation with tensiometer networks installed at the near-stream site, and upslope of Pit 5 and Pit A

losses are estimated to be
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losses are estimated to be 26% for undisturbed mixed evergreen forest (Rowe 1979). 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 comprises clasts of sandstone, granite and schist in a clay–sand matrix and is effectively impermeable with seepage losses to deep groundwater estimated at 100 mm year (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 greyish brown A horizons and a moderately thick, very friable mineral layer of podsolized, stony, yellow-brown earth subsoils (average 60 cm). Silt loam textures predominate. Study profiles examined by Webster (1977) showed that the organic humus layer averaged a 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 averaged 70% by volume, with average bulk densities of $0.80 t/m^3$ and saturated hydraulic conductivities of 250 mm/h. The wet and humid climatic environment, in conjunction with topographic and soil characteristics, result in the soils normally remaining within 10% of saturation (Mosley 1979). As a result, the soils are strongly weathered and leached, with low natural fertility.

The thin nature of the soils promotes the lateral development of root networks and channels. Soil profiles reveal extensive macropores and preferential flow pathways at vertical pit faces in the Maimai M8 catchment which form along cracks and holes in the soil and along live and dead root channels (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.

2.3 EARLY STUDIES OF MACROPORE FLOW

2.3.1 Hydrometric Observations at the Catchment and Hillslope Scale

The first comprehensive hydrologic study of the Maimai catchments was conducted by Mosley (1979). This study included a series of hydrometric and dye-tracing experiments in the Maimai 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 long pits were dug into the underlying gravels, and troughs were installed at the base of the pits to intercept and measure subsurface flow (Figure 2.2).

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 hydrographs were closely aligned in time and increased in a downslope direction (Figure 2.3). There was a close coincidence in the time of the peaks and an increase in discharge and total flow volumes downslope. This indicated that water was moving considerable distances
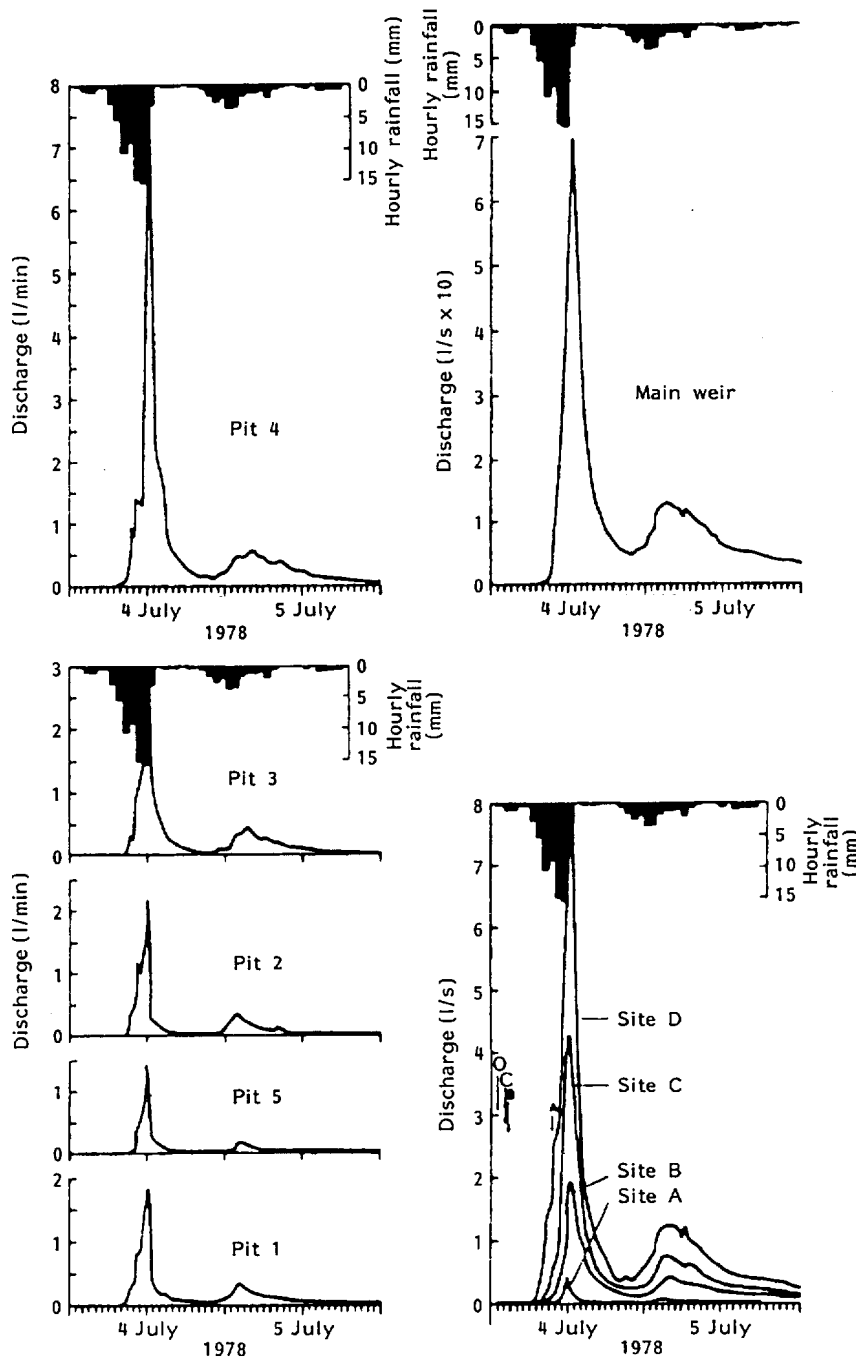
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Figure 2.3 Comparison of hydrograph response from the throughflow pits (1–6) zero-order stream sites (A–D) and at the M8 main weir as shown in Figure 2.2, 4–5 July 1978 (from Mosley 1979)

through the soil during the rising limb of the hydrograph. 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.
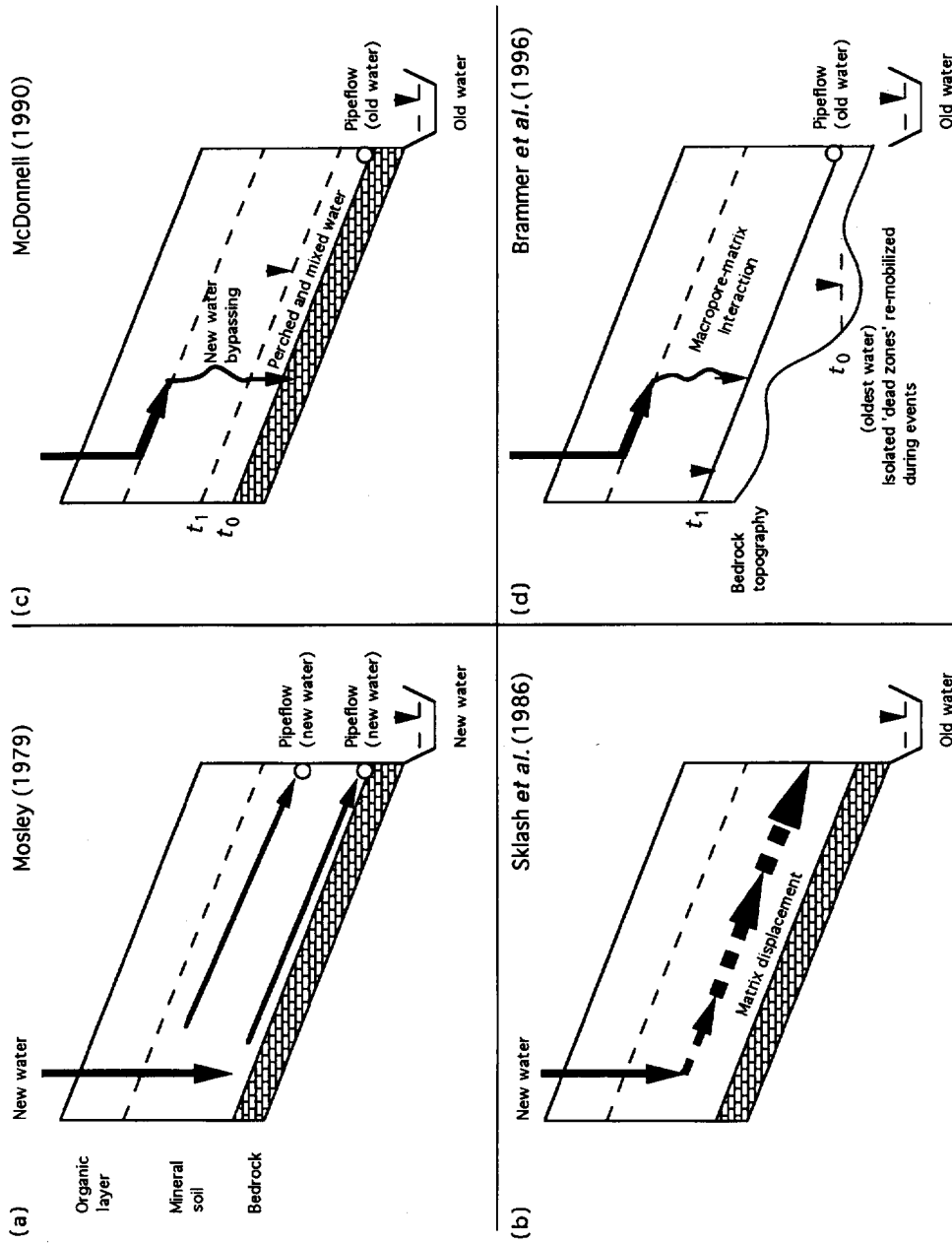
2.3.2 Hydrometric Observations at the Pit Face Scale

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 significant 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 preferential/macropore flow mechanism was evaluated via eight dye-trace trials, conducted at various pit faces (Figure 2.2). 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 at arrival times of 1.5–23 min following application. Dye appeared later 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 l 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 1.2–2.1 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 metres during the rising limb of the storm hydrograph. Mosley reasoned that this could not be accomplished by saturated flow through the mineral soil alone. Where there was a lack of root density and channels in the A and B horizons, water appeared downslope at the base of the A horizon. At sites with a 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. Mosley concluded that this flow was of "new" water and that subsurface flow through macropores and along long discontinuities in the soil profile was capable of contributing to storm period streamflow.

2.3.3 Subsurface Flow Velocity Estimations

Mosley (1982) continued the subsurface flow trial by sprinkling sites under different 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

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again, the pits were excavated to bedrock with intercepting troughs installed at the base. A maximum of 30 l 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).

A comparison of the values for mean velocity, maximum velocity and ratio of output volume to input volume were conducted. 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 for burned sites. The hydrographs of all sites were shown to be very similar in shape and if 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 roots, both live and dead. Profile wetness increased downslope and vertically; saturated wedge thinning was observed in an upslope direction.

2.3.4 A Proposed Perceptual Model of Hydrologic Response

From these preliminary studies, a schematic/perceptual model of catchment hydrologic response (subsurface flow paths) was proposed (Figure 2.4(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, p. 65). The model considered macropore flow to be a totally "short-circuiting" process by which "new" water could rapidly appear in the stream hydrograph.

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-conducted macropore flow system present within the soil. Also, the trace application intensity of simulated "events" was that of very long recurrence interval "storms"; hence application rates may have actually *induced* macropore flow.

2.4 ISOTOPE TRACING AND OLD WATER DISPLACEMENT

2.4.1 Application of the Isotope Hydrograph Separation Technique (Long-Term Catchment Sampling)

Pearce *et al.* (1986) sampled weekly 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,

Figure 2.4 (*opposite*) Maimai conceptual models: (a) Mosley (1979); (b) Sklash *et al.* (1986); (c) McDonnell (1990a); (d) Brammer *et al.* (1996)

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 (Figure 2.2).

Pearce et al. (1986) found that the $\delta^{18}O$ values for the weekly rain samples ranged from -3 to -12 per mil (‰) and displayed some seasonality; 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 variations. The groundwater samples followed M8 $\delta^{18}O$ stream 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 or from other times despite large associated fluctuations in rainfall δD or $\delta^{18}O$, indicating that only small contributions of new water occurred at high flow rates. This observation directly refuted Mosley's determination that rapid transmission of new water formed the majority of stream runoff.

2.4.2 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. 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 streamflow. Stream EC rose to $80–100\,\mu S/cm$ from an initial value of $47\,\mu S/cm$ during the first event. EC rose to over $140\,\mu S/cm$ during the second event the following day, lagging the hydrograph peak. The changes in EC indicated increases in total solutes rather than dilution, which would be expected if water from the storm rainfall had dominated the runoff. Cl^- concentrations in streamflow remained constant throughout the events. The increases in EC and the consistency of Cl^- concentrations indicate that the storm runoff response was predominantly water which had a substantial period of contact with the soil. The contribution of 3% of new water could be accounted for by direct precipitation onto the stream channel. 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 the mechanism that produces storm runoff.

Sklash et al. (1986) extended the Pearce et al. (1986) hydrograph separations into two first- and one second-order stream, and six throughflow pits, for several storm events in September 1983. Events sampled had return periods of between four weeks and three months (i.e. 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

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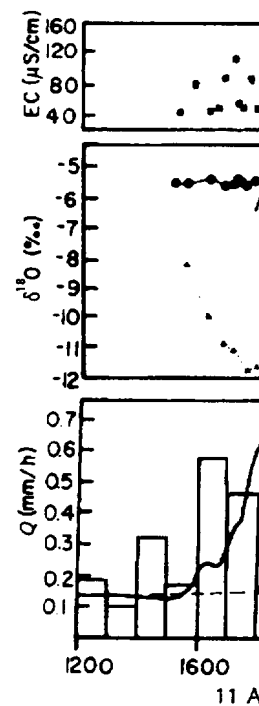2.1; location shown i
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Figure 2.5 Storm resp

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 by Mosley (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 (Figure 2.5). The effect was much decreased in the second of closely spaced storm events. Stored water appeared to dominate outflow as shown by variations in EC and Cl⁻ concentrations in subsequent storms, reflecting the differing contact times between stored water and the soil matrix.

2.4.3 Sampling at the Pit Face Scale

The collected hillslope water displayed large spatial variability in $\delta^{18}O$ composition. The pit locations (used originally by Mosley 1979) represented a cross-section of hillslope positions and topographic 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 2.1). Deep suction lysimeter δD values (Table 2.1; location shown in Figure 2.2) were much lighter isotopically than shallow lysimeter and throughflow δD samples. This suggested that much of the throughflow
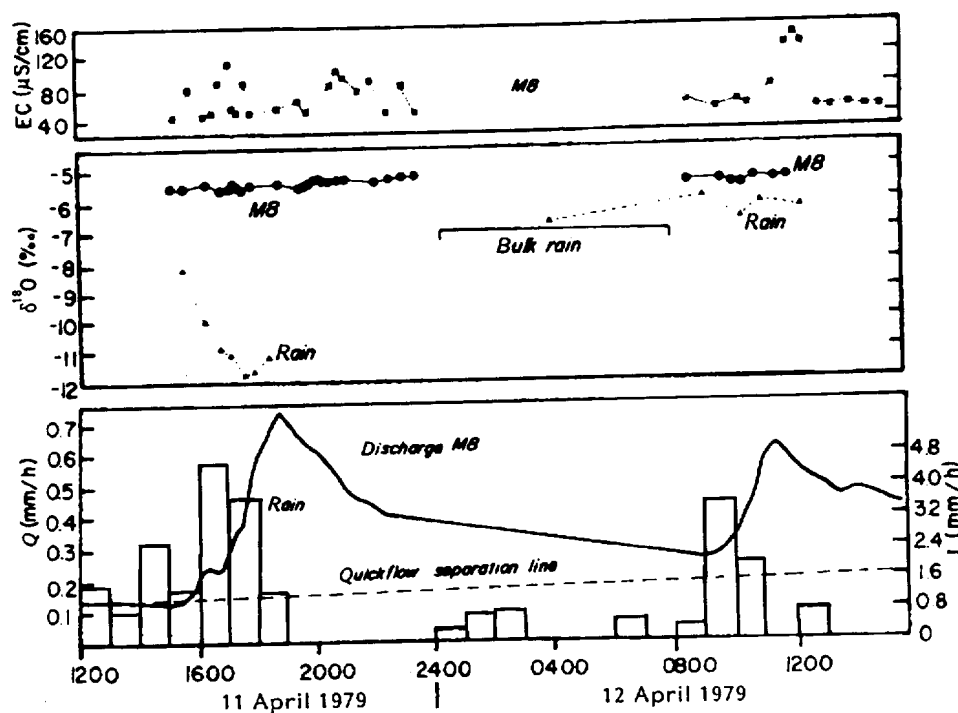

Figure 2.5   Storm response in catchment M8, 11–12 April 1979 (from Pearce *et al.* 1986)

Table 2.1 Catchment M8 old and new water contributions to throughflow 21 September 1983 (from Sklash *et al.* 1986)

Site	δD_O(%)	δD_N(%)	δD_P(%)	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

δD_O, old water δD; δD_N, new water δD; δD_P, δD of throughflow at peak discharge.
[a]Based on SL5S value.
[b]Low flow value on 23 September.
[c]Weighted average rain.

arrived laterally from thinner soil profiles (Figure 2.6). At the same sites, the EC and Cl⁻ values showed noticeable differences in the relative influence of solute flushing at the hillslope pit sites.

2.4.4 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 max-rise piezometers during storms (Figure 2.6; locations shown in Figure 2.2). The piezometers located near the valley bottom had the highest response, close to achieving surface 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. Both mechanisms appear 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. Saturated wedges on the lower slopes and groundwater ridges in the valley bottoms were thought by Sklash *et al.* (1986) to develop quickly as infiltrating rain converted the tension-saturated zone into phreatic water. This perceptual model negated the need to invoke rapid transmission of new water downslope via macropores in order to explain the streamflow response, since stored water was the main component discharged into the stream channel during events (Figure 2.4(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
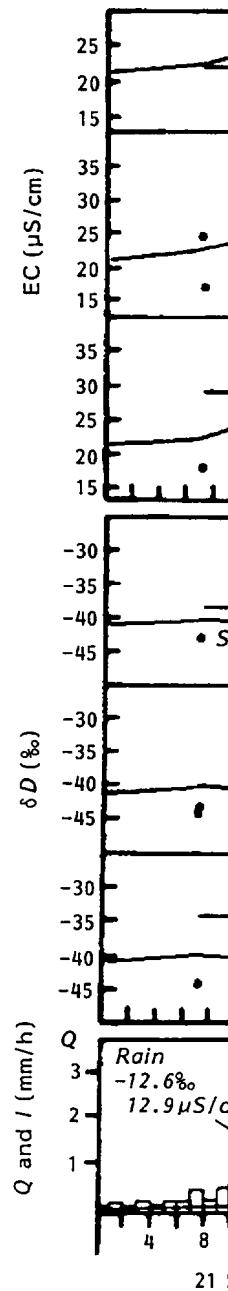
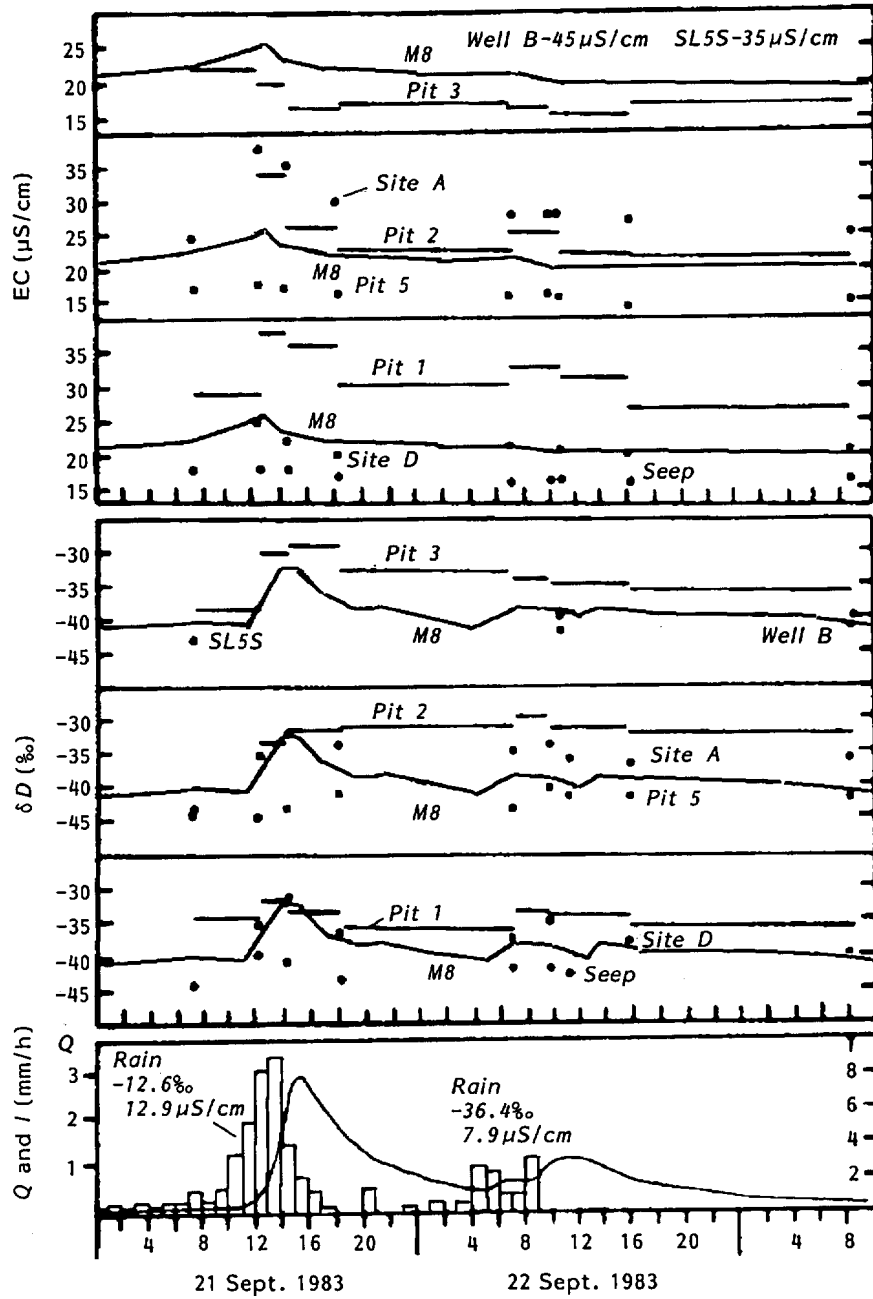**Figure 2.6** Compariso
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Figure 2.6 Comparison of stream (M8) and hillslope δD and EC response, 21 September 1983, storm (from Sklash *et al.* 1986)

apart from point observations from maximum-rise wells. Pearce *et al.* (1986) and Sklash *et al.* (1986) discounted pipeflow measurements of Mosley entirely, possibly because significant amounts of macropore flow did not necessarily occur during the small-magnitude events monitored.

2.5 A COMBINED HILLSLOPE SOIL PHYSICS, ISOTOPIC AND CHEMICAL APPROACH

2.5.1 Soil Potential Response in Hydrologically Active Hollows

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 (Figure 2.2) in an effort to explain the discrepancies between the perceptual models offered by Mosley (1982), Pearce *et al.* (1986) and Sklash *et al.* (1986). In the hydrologically active mid-slope hollows, (e.g. Pit 5 as shown in Figure 2.2), McDonnell (1990b) found that soil 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 even on 23–24 October 1987 (Figure 2.7(a)), tensiometric data showing a semi-constant wetting front propagate through the profile with strong soil 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 Figure 2.7 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 (Figure 2.7(a)). Therefore, a slope water-table did not develop.

During a larger magnitude (58 mm) rainfall event on 29 October, pressure potential in the lower soil horizons (> 75 cm) responded almost instantaneously to infiltrating rain (Figure 2.7(b)). This response was a function of a disequilibrium in soil pressure potentials during wetting, caused by the presence of soil macropores (McDonnell 1991). 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. For the largest event monitored (103 mm of rainfall) on 13 October, soil-pressure potential remained, relatively constant throughout the profile, during the limited period of tensiometer coverage (Figure 2.7(c)). McDonnell (1990b) observed that most of the soil profile remained saturated during this episode.

2.5.2 Groundwater Development and Longevity

Generally, when 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 intensitites 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 was very short and
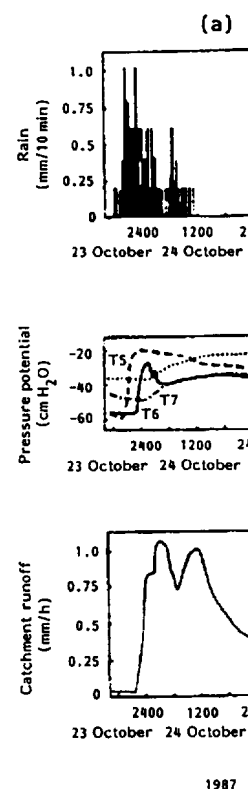
Figure 2.7 Pressure potential showing relationship between Three storms are shown. depth 1–1.5 m, slope are 820 and 1080 mm below

showed a close corre McDonnell (1989, 19 efficient and showed r *et al.* (1991a) noted moved through pipes bedrock interface. Th profile in events with downslope drainage pipes in those zones account for the rapid

2.5.3 Mechanics o

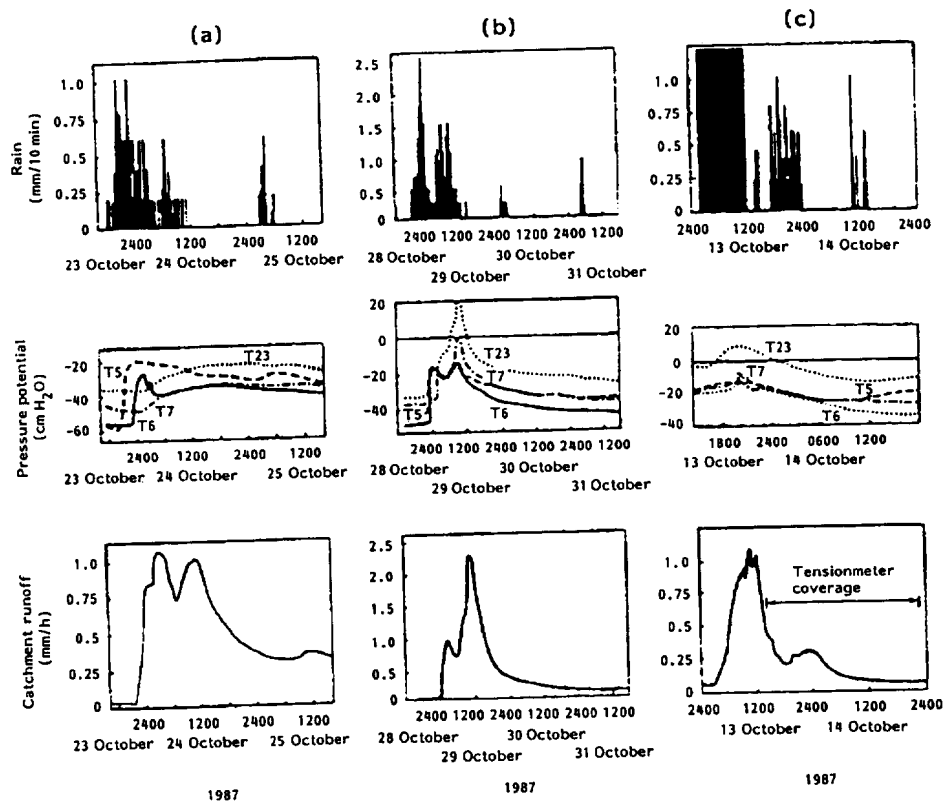McDonnell (1990a) re the M8 stream runo

Figure 2.7 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)

showed a close correspondence with hillslope throughflow rate, as measured by McDonnell (1989, 1990a). Downslope drainage of perched water was extremely efficient and showed no lag with recorded throughflow for select storms. McDonnell et al. (1991a) noted that this indicated that lateral saturated flow was rapid and moved through pipes (corroborated visually using dye tracers) formed at the soil–bedrock interface. 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 downslope drainage through pipes (McDonnell 1989). 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.

2.5.3 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 soil truly "releasing"

water long before wetting along a measured wetting–drying curve would predict. This was due to the pressure 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 would 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 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 $V_o > K^*$.

2.5.4 A Perceptual Model of Macropore Flow and Old Water Displacement

McDonnell (1990b) noted that as invading new water moved to depth, free water perched at the soil–bedrock interface, as water "backed-up" into the matrix, where it mixed with a much larger volume of stored old matrix soil water (Figure 2.4(c)). Pressure potential evidence from the responsive mid-slope hollow (Pit 5; Figure 2.2) showed that this water-table was dissipated by the moderately well-connected system of pipes at the mineral soil–bedrock interface. The relationship between crack infiltration and lateral pipeflow was not linear, because there was a significant time delay between water-table perching and subsequent distribution of positive pore pressure in the soil. McDonnell (1990a) reasoned that this delay was the critical process necessary 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 Figure 2.8. Stewart and McDonnell (1991) showed that between-storm matrix water varied in age from approximately one week at the catchment divide (near Site A; Figure 2.2) to over 100 days at the main M8 channel margin, supporting the notion of very short hillslope water residence time. (A subsequent hillslope excavation project in 1992 by an EarthWatch research team (McDonnell, unpublished data) revealed 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 perceptual model.)

2.6 WHOLE HILLSLOPE TRENCHING AND FLOW COLLECTION

2.6.1 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 (Figure 2.2). A vertical face 60 m long and 1.5 m high was cut across the toe of the hillslope. Thirty subsurface flow collection troughs were installed end-to-end, across the base
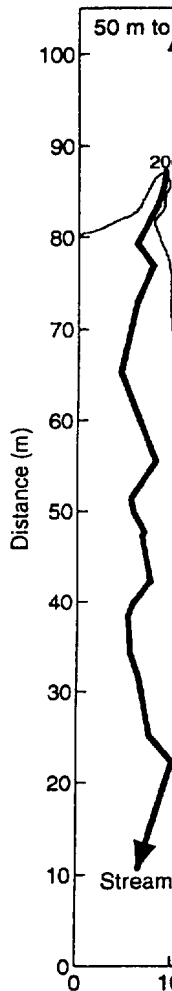

Figure 2.8 M8 hillsl[...]
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of the excavated fac[...]
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2.6.2 Subsurface [...]

Subsurface flow fron[...]
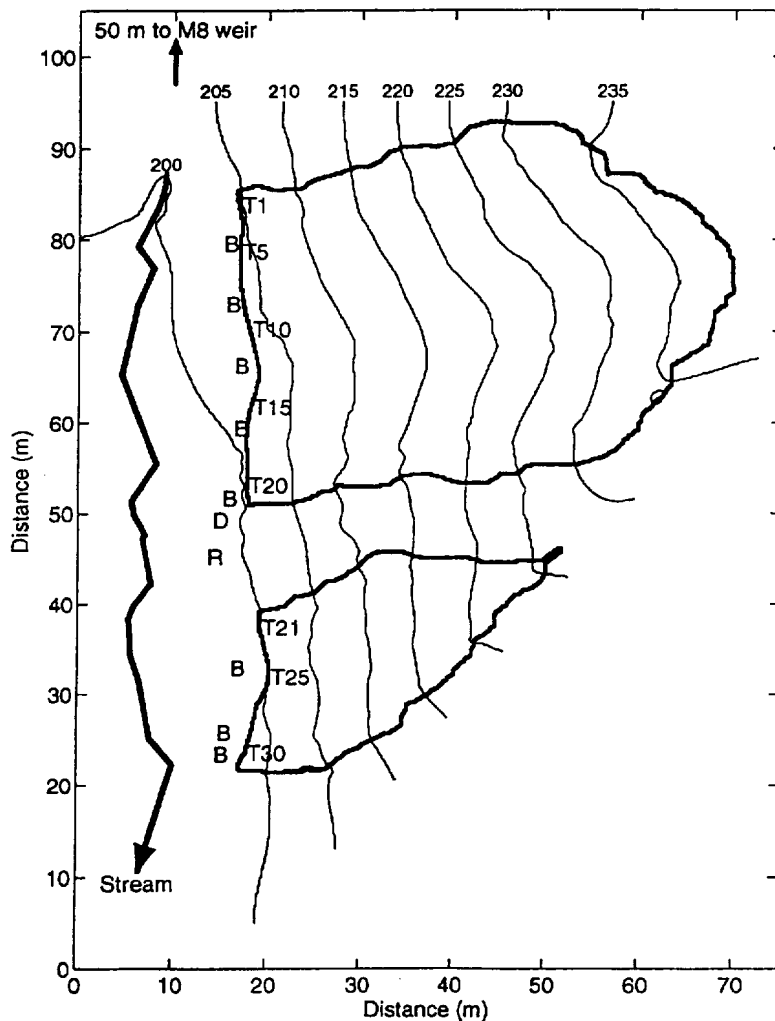timing. Neighbourin[...]

Figure 2.8 M8 hillslope trench study; topography and location of selected troughs and measurement equipment: Tn=trough n, B=tipping bucket stand, D=datalogger, R=raingauge (from Woods and Rowe 1996)

of the excavated face at the soil–bedrock interface (Figure 2.8). The troughs were sealed to the cut face and covered. Flow collected in the troughs was routed to tipping bucket flowmeters and recorded from November 1992 to 1993.

2.6.2 Subsurface Flow Volumes

Subsurface flow from the hillslope was highly variable in both magnitude and timing. Neighbouring collection troughs showed unexpected differences in flow
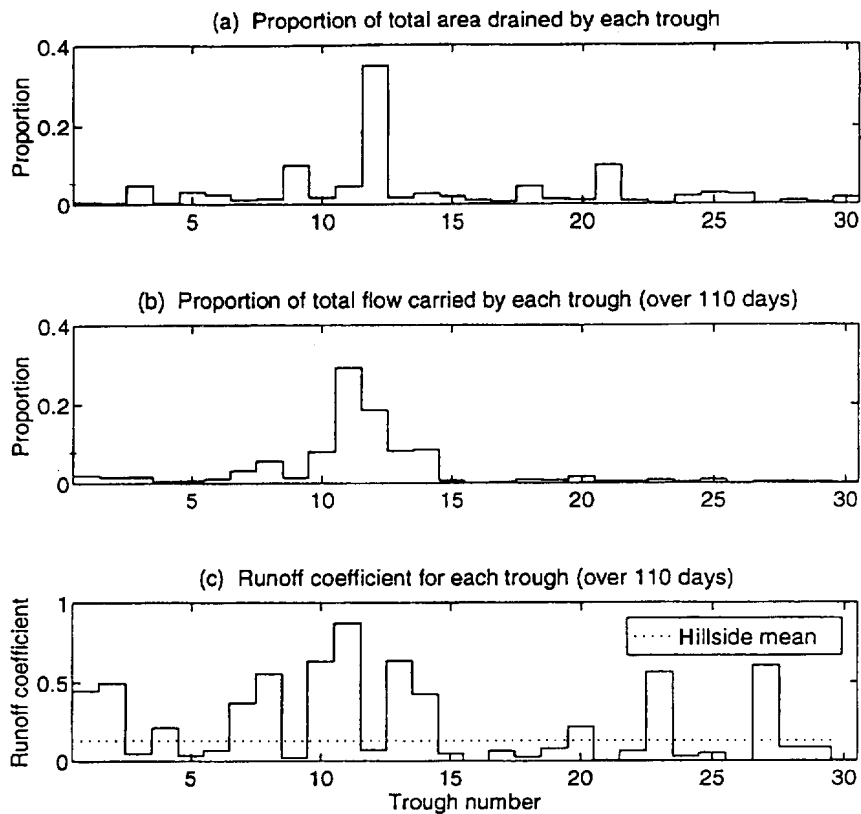
Figure 2.9 Spatial distribution of trough characteristics: (a) subcatchment area; (b) total flow over entire study period; (c) runoff coefficients (from Woods and Rowe 1996)

rates. Figure 2.9 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 (trough catchment) area (Figure 2.9). In particular, Woods and Rowe (1996) estimated that trough 12 (Figure 2.9) would drain 35% of the hillslope, yet 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 subcatchment divide or the subsurface contributing area. Trough runoff coefficients were also variable across the hillslope (Figure 2.9(c)). When the troughs were grouped by topographic feature (convergence and divergence), variability in flow across the hillslope cut face was still observed (Woods and Rowe 1996).

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2.6.3   Isotopic Sampling along Trench Face

Stewart and Rowe (1994) applied a lumped-parameter approach to model the residence time of subsurface water in the trough subcatchments. Rainfall $\delta^{18}O$ was compared to soil water $\delta^{18}O$ collected from suction lysimeters on the slope. Shallow-depth suction lysimeters were the most responsive isotopically to rainfall inputs. Residence times increased while isotopic response to rainfall input decreased with soil depth.

Results of the Stewart and Rowe (1994) model applied to the M8 stream for a sampled storm during January 1994, was discussed in Unnikrishna *et al.* (1995) (Figure 2.10). When this model was applied to the trough system, isotopic results displayed a large variation across the trough face, similar to the flow variations (Stewart, pers. comm., 1994). Low-flow trough (end troughs away from the central hollow) $\delta^{18}O$ values showed that 60% of the flow was event water, but this water had a relatively long residence time. The distribution had a peak at 4–5 hours, a mean of 28 hours; 50% of the water discharged had a residence time of less than 12 hours. Flow commenced about four hours after the onset of rain when sufficient water had infiltrated the soil. The results from the high-flow troughs were very different. The water was initially more depleted isotopically ($\delta^{18}O$ about $-6.2‰$) than the stream, indicating a source from groundwater with a much longer residence time. After 8 hours of rainfall, the $\delta^{18}O$ changed to $-5.0‰$ and remained constant for 16 hours, before gradually shifting back toward the groundwater composition after 36 hours. Stewart and Rowe (1994) assumed that groundwater supplied a near-constant flow to the stream that gradually had soil water added to it after rain has been falling for some hours. Groundwater again dominated the storm hydrograph as the soil-water flow decreased.

2.6.4   A Perceptual Model of Hillslope Flow: A Single Pit does not a Hillslope make

The spatial variability across the trough face, even when grouped by topographic features, revealed that the flow data of small troughs are difficult to extrapolate to larger scales. This study makes apparent the fact that a model of hillslope response must 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 which determined the runoff (as is the case with infiltration excess or saturation excess flow), but rather the size of the saturated soil moisture storage.

Although very useful, the Woods and Rowe (1996) study treated the hillslope as a "black box" and did not collect data on flow paths and mechanisms of flow to the troughs. Stewart and Rowe (1994) restricted their isotope sampling to the stream and specific high-flow and low-flow troughs; they did not investigate the spatial variability across the trough face or soil-water isotopic compositions on the slope.
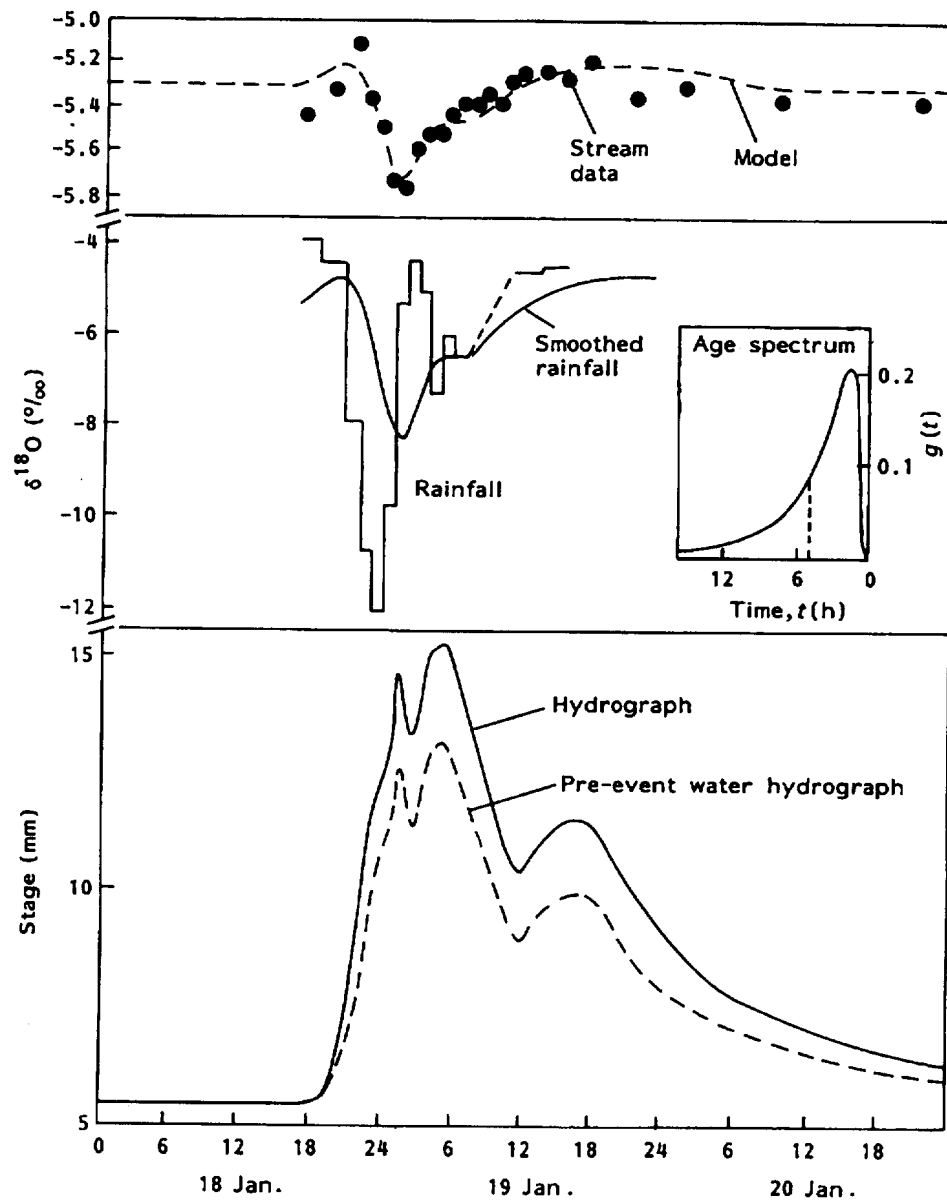
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Figure 2.10 Event based water age spectrum and hydrograph separation for the M8 Maimai catchment (from Unnikrishna *et al.* 1995)

2.7 A HILLSLOPE-SCALE BROMIDE TRACER INJECTION

2.7.1 The Br⁻ Line Source Injection

During March to May 1995, Brammer (1996) and Brammer *et al.* (1996) performed a hillslope-scale Br⁻ tracer experiment on the trenched hillslope described by Woods

Figure 2.11 M8

and Rowe (1996). A 5 m grid of suction lysimeters and co-located maximum-rise wells were installed in a 30×30 m plot 5 m upslope from the trench face (Figure 2.11). A line source of 3 kg LiBr was applied 35 m upslope from the trough face. Antecedent moisture conditions were relatively high; 17 mm of rainfall occurred in the previous two days. The first rainfall event occurred 12 hours after the tracer application.
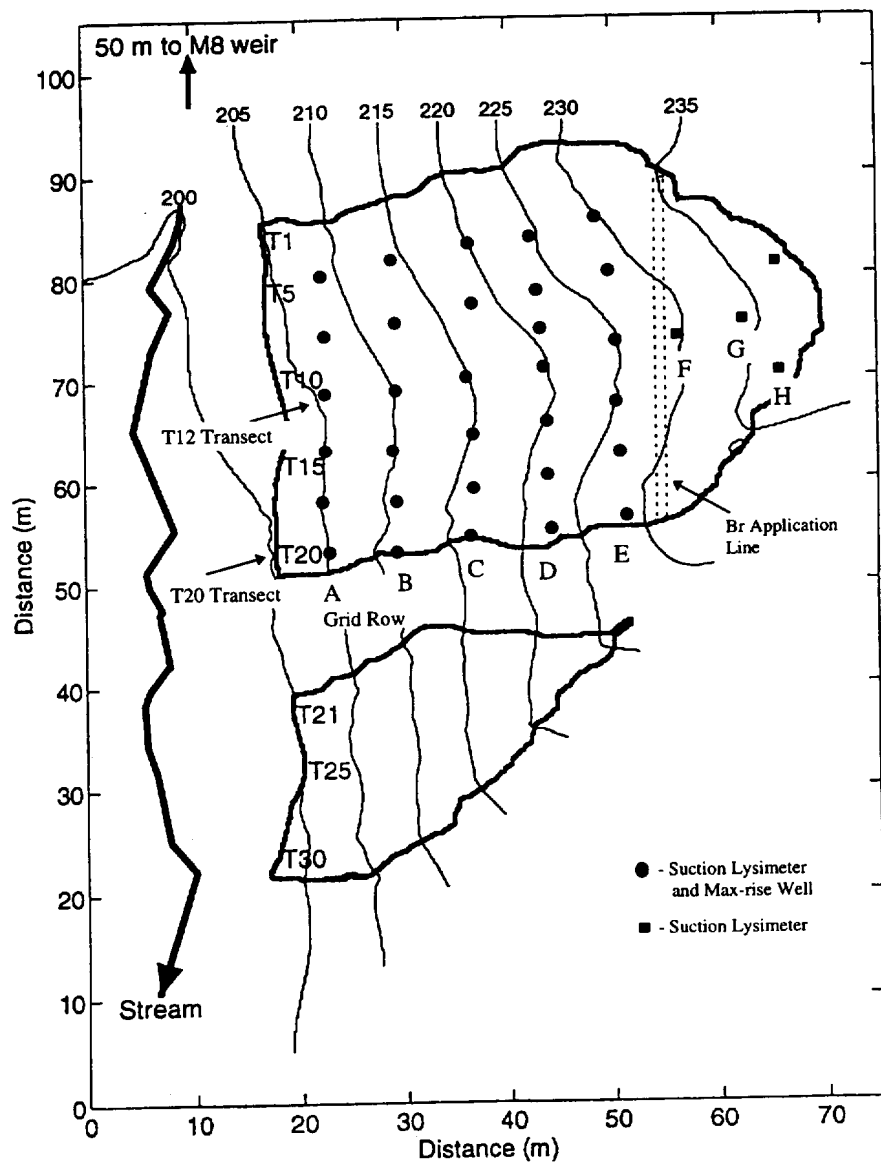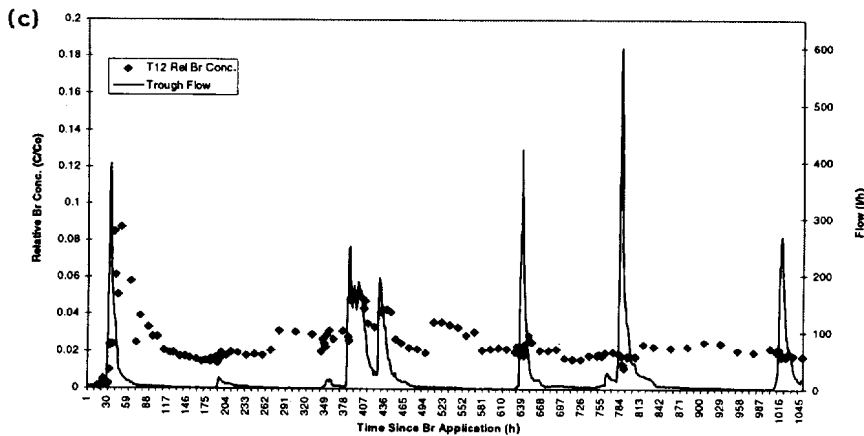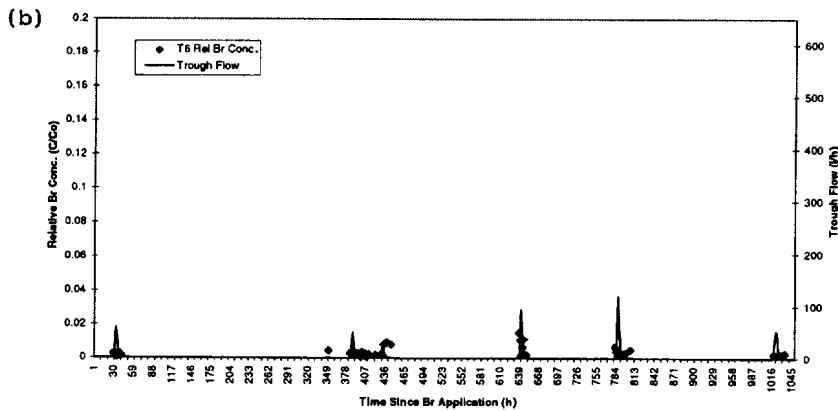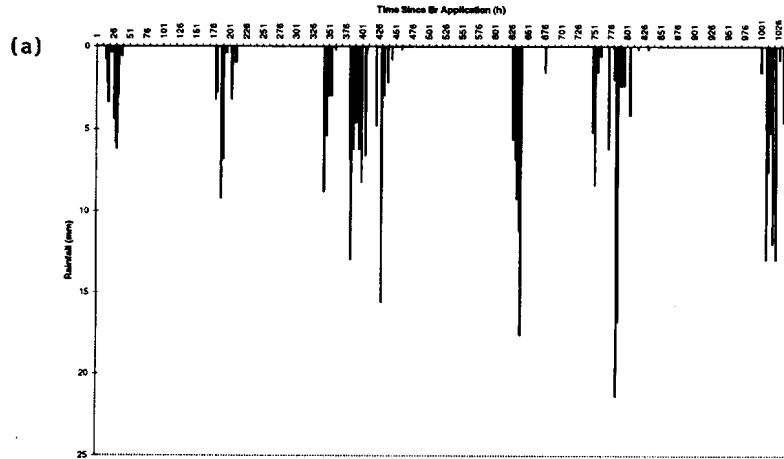

Figure 2.11 M8 hillslope plot (based on figure from Woods and Rowe 1996)

Rainfall for Entire Study Period (24 Mar-7 May 1995)

(a)

(b)

(c)

2.7.2 Surface ver

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2.7.3 Br⁻ Breakth
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Brammer et al. 1996)

2.7.2 Surface versus subsurface controls on Br⁻ export off the plot

Surface topography alone did not fully explain the subsurface flow response (as Woods and Rowe (1996) found) nor Br⁻ concentrations across the trench face. Furthermore, spatial patterns of water-table and Br⁻ concentration on the hillslope grid showed poor correlations with topographic position. Depths to bedrock were measured at a 2.5 m grid-scale over the hillslope. Soil depths within the grid became shallower downslope near the trough face forming an impeding "lip". A midslope bedrock depression appeared to act as a subsurface storage reservoir. The mapped spatial water-table response and groundwater persistence in this zone reflected this drainage pattern. Maximum-rise wells indicated a rapid rise and fall in water-table levels along the edges of the grid. The highest levels were recorded in the hollow with a persistent water-table (of up to four days), detected up the central hollow transect, 15 m upslope from troughs 11 and 12. This may explain the long flow recession times for these troughs, as observed by Woods and Rowe (1996).

Within the grid, the tracer plume reached the first row of the suction lysimeters six days after application, consistent with the average soil hydraulic conductivity on the slope of 30 mm/h. High Br⁻ concentrations were observed at the edges of the grid, apparently due to bedrock topographic control of subsurface flow. In general, Br⁻ concentration did not follow expected spatial patterns based on surface topography. Increased tracer concentrations were not observed in the hollow wells or suction lysimeters, despite their location within the estimated subcatchment areas receiving significant drainage from upslope.

2.7.3 Br⁻ Breakthroughs: Effect of Diffusion into Matrix depending on Antecedent Wetness Conditions

Flow volume variability across the trough face was similar to observations by Woods and Rowe (1996). Similary, Br⁻ tracer arrival and concentrations displayed spatial and temporal variability related to flow. Br⁻ was detected in the outflow from the troughs draining the central hollow (troughs 11–14) 18 hours after the start of rainfall, indicating rapid movement downslope. During the first event, Br⁻ concentrations arriving at the high-flow troughs were the highest measured during the study period, while samples from low-flow troughs (troughs 4–8 and 16–20) did not show detectable Br⁻ concentrations until the third storm event, 14 days after application (Figure 2.12). Subsequent events did show increases in Br⁻ concentration with time, indicating remobilization of Br⁻ tracer from the soil matrix; remobilization appeared sensitive to antecedent soil-moisture conditions. It is important to note, however, that tracer breakthrough at the trough face preceded appearance of Br⁻ in the upslope suction lysimeters by 2 days and maximum-rise wells by 7 days, suggesting rapid downslope preferential flow of 10's of metres.

Figure 2.12 (*opposite*) Comparison of Br⁻ concentration breakthrough curves across the hillslope: (a) study period rainfall; (b) low flow (trough 6); (c) high flow (trough 11) (from Brammer *et al.* 1996)

2.7.4   A flowpath-based perceptual model of 3D hillslope flow across catchment

The maximum-rise well response and subsurface flow Br$^-$ concentrations support the notion of bedrock microtopographic depressions at the soil–Old Man Gravel interface. The less mobile water is mobilized during large events, but effectively isolated from downslope saturated–unsaturated flow between events. The interaction of these mobile and immobile regions of both soil and lateral subsurface flow system is illustrated in Figure 2.4(d)). The maximum-rise wells located in deep soil slope positions were the most responsive to rainfall, but did not contain significant tracer concentrations. The Br$^-$ concentration response of the grid wells and suction lysimeters at the edge of the plot suggests opportunity for inter-microcatchment transfer of water and flow.

2.8   CONCLUSIONS

This chapter synthesizes the progression of ideas in the development and modification of a perceptual model of hillslope flow at the Maimai catchment. Each data set reviewed in the chapter reveals not only a cumulative understanding of catchment behaviour, 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; notwithstanding, 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) demonstrated that an integration and combination of techniques was required for comprehensive hillslope characterization and reconciliation of 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 slope section – making the single pit observations of the previous studies highly suspect. Most recent observations by Brammer et al. (1996) reveal much microtopographical control on subsurface flow timing and tracer breakthrough. Whilst previous works have treated the soil–bedrock interface as a sharp boundary, Brammer et al. (1995) demonstrated that small depressions in the bedrock surface exert a large control on water mobility and mixing; thereby producing localized "dead zones" of longer residence time waters.

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