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Bypass Water Flow through Unsaturated Microaggregated Tropical Soils

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ABSTRACT

Recent evidence suggests that bypass flow occurs in many soils even under unsaturated conditions, but experimental confirmation is lacking. We measured bypass flow in two microaggregated Inceptisols from the humid Atlantic region of Costa Rica under water-application rates below those needed to produce ponding. Water was applied with a constant-head rainfall simulator to undisturbed soil cores taken when soils were near field capacity. Mass of soil core and of water exiting the core (outflow) and time to first appearance of water at the bottom of the core (breakthrough) were recorded. For each application rate, steady-state outflow was reached quickly at a core mass well below that corresponding to soil saturation. (In contrast, classical theory for sandy soils without macropores, based on the Richards' equation, predicts that breakthrough should take more than twice the time actually observed.) Successive step increases in application rate produced successively smaller increases in core mass and decreases in the lag time before outflow increased. Our results suggest that bypass flow will occur in the noncapillary interpedal pore space whenever the application rate exceeds the infiltration rate of individual microaggregates. Because values for matrix conductivity are extremely low (much less than typical rainfall rates), we suggest that bypass flow may be the rule rather than the exception in microaggregated soils with extensive interpedal noncapillary pore space.

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WATER FLOWS downward through soils by two routes: slowly through small pores (micropores), or more rapidly through larger pores. The rapid flow can be termed *bypass flow* in the sense that the water bypasses micropores and thus the majority of the soil's reactive surface (van Genuchten and Wierenga, 1976; Hillel, 1980; Rao et al., 1980; Bouma, 1981; Beven and Germann, 1982; Sollins and Radulovich, 1988). Where it occurs, such bypass flow may be extremely important: fertilizers, pesticides, and other contaminants move downward rapidly with low rates of absorption into the soil, and nutrients already in the soil are little leached (Wild, 1972; Thomas and Phillips, 1979; Bouma, 1981; White, 1985; Sollins and Radulovich, 1988; Solórzano and Radulovich, 1989).

Bypass flow occurs via noncapillary pores, defined here as those pores draining under tensions ranging from hardly measurable up to that corresponding to field capacity. *Non-capillary* thus includes both *macro-* and *meso-pores*, as defined by Wilson and Luxmoore (1988). These large pores account for the difference between water content of a soil at saturation and water content at field capacity (Radulovich et al., 1989). In contrast, capillary pores or micropores remain filled with water after the soil reaches field capacity and drainage from meso- and macropores has practically ceased. Meso- and macropores exist in virtually all soils; even largely impermeable horizons such as fragipans have been shown to contain macropores (Parlange et al., 1989).

It is widely thought that the soil must be close to or fully saturated, with water ponded or nearly ponded on its surface, before bypass flow can occur (Bouma, 1981; Seyfried and Rao, 1987; Parlange et al., 1988; Steenhuis and Muck, 1988; Wilson and Luxmoore, 1988). Consequently, bypass flow has usually been considered to be of practical significance only in soils of low infiltrability (e.g., Seyfried and Rao, 1987), and studies of water conduction via macropores have concentrated on ponded flow (Bouma and Wosten, 1979; Hillel, 1980; Rao et al., 1980; Seyfried and Rao, 1987; Parlange et al., 1988; Steenhuis and Muck, 1988; Wilson and Luxmoore, 1988). There are indications, however, that bypass flow may not require full saturation (i.e., ponding on the soil surface). Wild (1972) reported water flow through cracks in unsaturated soils, and Phillips et al. (1989) showed that water enters simulated macropores in unsaturated soils. In addition, work with zero-tension lysimeters (Russell and Ewel, 1985; Radulovich and Sollins, 1987) and studies of dye penetration (McVoy, 1985; Sollins and Radulovich, 1988) suggest that, at least in microaggregated soils, water flows along preferential paths even without surface ponding. Indirect evidence that bypass flow may occur in unsaturated soils comes from the finding that measured rates of water and solute transport downward through unsaturated soils are sometimes much higher than most hydrological models would predict, even for saturated conditions (Parlange et al., 1988). However, direct experimental confirmation that bypass water flow can occur without visible ponding on the soil surface is still lacking.

For several years we have been studying nutrient and water dynamics on abandoned agricultural land in the lowland humid tropics of Costa Rica (Radulovich and Sollins, 1985, 1987, 1991; Sollins and Radulovich, 1988; Radulovich et al., 1989; Solórzano and Radulovich, 1989). Because of the potential relevance of bypass flow as a critical nutrient-cycling process, we undertook a series of simple experiments to determine whether bypass flow can occur in such soils at water-application rates below those needed to pond water on the surface.

Here we report on the outcome of these experiments and compare results with predictions based on the classical theory of water flow in soils without noncapillary pores. We then suggest a scenario to account for the occurrence of bypass flow in the absence of surface ponding.

MATERIALS AND METHODS

Undisturbed soil cores were obtained during 1988 and 1989 from sites at the La Selva Biological Station near Puerto Viejo de Sarapiquí (Sollins and Radulovich, 1988; Radulovich et al., 1989; Sollins et al., 1992) and at the Centro Agronómico Tropical de Investigación y Enseñanza (CATIE) in Turrialba (Russell and Ewel, 1985; Seyfried and Rao, 1987), both in the humid tropical Atlantic region of Costa Rica. All soils are microaggregated Inceptisols formed in alluvium or colluvium of volcanic origin. The La Selva soils (Helechal consociation) are borderline Oxic Humitropepts/Typic Dystropepts (Sollins et al., 1992). The CATIE soils are Typic Dystropepts (Aguirre, 1971). Infiltrability of both sets of soils is very high, and ponding and overland flow are uncommon (Russel and Ewel, 1985; Radulovich and Sollins, 1987; Seyfried and Rao, 1987; Sollins and Radulovich, 1988). Both sets of soils drain very rapidly; the La Selva soils drain to a virtually constant water content in <1 h (Sollins and Radulovich, 1988). This rapid drainage, commmonly observed in well-aggregated tropical soils (Sánchez, 1976), makes the use of the otherwise controversial concept of field capacity perfectly sound in these soils, at least in the absence of a shallow water table or of granulometric or density discontinuities. Because the La Selva soils are highly microaggregated (Strickland et al., 1988), they behave like sands with regard to infiltration and drainage but like clays with regard to their water-holding capacity (Sollins and Radulovich, 1988).

Two sites were sampled at each location. A forested site at La Selva (forest site) had been abandoned about 1968 after a short period of cropping and grazing and had regrown in mixed secondary forest; a second site (grass site), abandoned about 1982, had been invaded by grasses (*Brachyaria, Ischaemum,* and *Panicum* spp.) plus some ferns (*Hypolepis* and *Nephrolepis* spp.) and shrubs (Sollins and Radulovich, 1988). Sites at CATIE were an annatto (*Bixa orellana* L.) plantation abandoned about 1980, and an area of minimum-input interplanting of cacao (*Theobroma cacao* L.) and several shade species (same site used by Seyfried and Rao, 1987).

All sampling was conducted during the rainy season (May–December) of 1989 in order to assure that the soil was near field capacity. Sites were first cleared of vegetation and litter with machetes, taking care to avoid soil compaction, then wetted to saturation. The soil was covered with a plastic sheet (2 by 2 m) and allowed to drain to field capacity for a period of 48 h so that water would not be absorbed into a dry soil matrix during the experiments (Cassel and Nielsen, 1986; Sollins and Radulovich, 1988). Steel cylinders (107.5-mm i.d.) with waxed walls and sharpened outer lower edges (Sollins and Radulovich, 1988; Radulovich et al., 1989) were used to take from 3 to 20 soil cores of 200-mm length from each sampling site. Cylinders containing intact soil cores were transported to the laboratory in plastic bags placed inside cushioned containers to minimize evaporation and compaction.

In the laboratory, each cylinder was hung from a balance (10-g sensitivity), and a pan was placed underneath to catch outflow. The pan was placed on an electronic balance (10-mg sensitivity) so that outflow mass could be monitored. Water was dripped onto the core with a constant-head rainfall simulator. A nozzle consisting of 29 hypodermic needles provided a uniform pattern of drops across the entire surface. Mass of the core and effluent was recorded through time, as well as time to water breakthrough (time from initial water application to first visual evidence of water exiting the core). The result of this procedure is referred to here as a *water breakthrough curve* (Radulovich et al., 1989) by analogy with the terminology used in solute-transport studies (Nielsen and Biggar, 1961).

Visual observation during the experiments revealed that, in all cases, breakthrough occurred somewhere on the bottom surface of the soil cores and not at the periphery. Preferential flow along the (waxed) inner surface of the steel cylinders was therefore not responsible for the observed breakthrough pattern. Earlier trials with dye (rhodamine B) confirmed that applied solutions flowed through the soil cores via well-defined preferential paths and not along walls (Sollins and Radulovich, 1988; Radulovich et al., 1989).

In an initial series of experiments, three to four cores from each of the four sites were selected at random. Water was applied at first at a constant rate of ≈ 15 mm h^{-1}, chosen to fall within the range of rainfall intensities reported for tropical regions (Jackson, 1989). Mass of the core and effluent were recorded until steady-state flow was established (i.e., outflow rate equaled inflow). Next, the inflow rate was stepped up to a higher constant value, and core and effluent mass were monitored until a new steady state was reached. This procedure was repeated three times or until water ponded on the soil surface. At the end of each

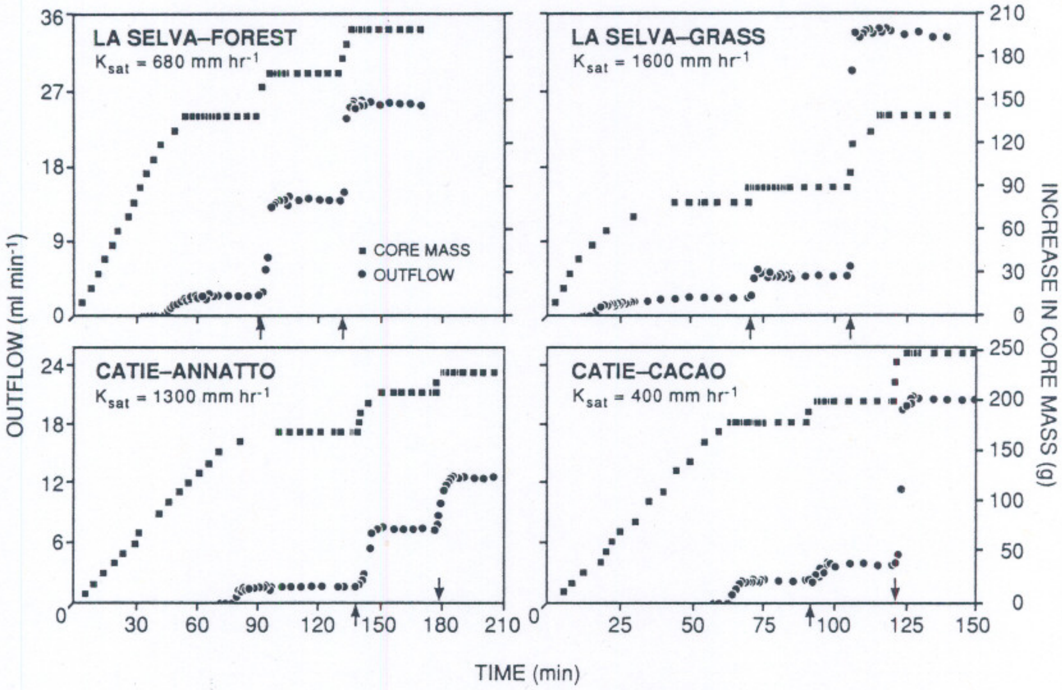

Fig. 1. Rate of water outflow and increase in core mass vs. time after initiation of water application. First outflow point represents breakthrough. Arrows indicate times when application rate was increased. Each graph refers to one core chosen as representative of the site. Initial core weights (at field capacity) were: La Selva — forest = 1816 g; La Selva — grass = 1918 g; CATIE — annatto = 1981 g; CATIE — cacao = 1956 g.

run, water was kept ponded (5-mm head) until steady-state flow was obtained (Radulovich et al., 1989). The saturated hydraulic conductivity (K_{sat}) was then calculated from the ponded flow rate with Darcy's equation, after correcting for the head. The K_{sat} was thus the rate just short of that needed to produce ponding, termed hereafter the *saturated flow rate*. The mass of the fully saturated soil core was calculated as the total core mass minus the mass of the 5-mm head. Several cores from each of the four sites were oven-dried to determine soil bulk density, which averaged 0.73 Mg m^{-3} (SE = 0.01) for the La Selva soils and 0.93 Mg m^{-3} (SE = 0.04) for the CATIE soils.

The computer program LEACHM (Hutson and Wagenet, 1989) was used to compare the outcome of the experiments with predictions based on the macroscopic theory of water flow through soils without noncapillary pores. The water-flow calculation in this program is based on a classical Crank-Nicolson discretization of the one-dimensional Richards' equation. The retentivity function of Hutson and Cass (1987) and the conductivity equation of Campbell (1974) were used to relate matric potential and hydraulic conductivity, respectively, to volumetric water content. Measured values for the soil cores were used for saturated water content and saturated hydraulic conductivity. Values for other constants were chosen to fall within the range considered typical of coarse sands. (We did this because, even though the La Selva and CATIE soils are clay textured, they behave like coarse-sandy soils when at or near saturation. It thus seemed preferable to compare the hydraulic behavior of the La Selva and CATIE soils with that of sandy or coarse-sandy soils.) For initial conditions, we assumed a uniform matric potential of −10 kPa throughout the 200-mm-long column, and that no water drained from the bottom of the column until the bottom layer of soil had saturated. Once this had occurred, the column was assumed to behave as if a water table were present at the bottom (c.f., Hutson and Wagenet, 1989, p. 41).

A second series of experiments, performed only on the La Selva soils, examined the relationship between water-application rate and maximum water velocity inside the cores, as determined by the time until breakthrough and until steady-state flow. Each day, from three to four cores were collected at field capacity from the forest and grass sites as described above, used that same day for analysis of water breakthrough, and then discarded. A single non-ponding rate of water application (17.3, 32.7, 66.7, or 133.3 mm h^{-1}) was used for each core.

RESULTS

Outflow rate and increases in core mass through time for different water-application (inflow) rates are illustrated in Fig. 1 for all sampling sites. Steady-state flow was reached for all cores at application rates well below that required to create ponding. The saturated rate (equivalent to K_{sat}) varied from 20.0 to 240 mL min^{-1} (132–1590 mm h^{-1}) for cores from La Selva soils and from 8.1 to 59.0 mL min^{-1} (54–390 mm h^{-1}) for the CATIE soils. Such extreme variability in infiltration rates is typical of these soils (Radulovich and Sollins, 1985; Sollins and Radulovich, 1988). Once steady-state flow was reached for a given inflow rate, the mass of the core did not increase further (see Fig. 1), which we interpret as meaning that a fixed cross-sectional area of noncapillary pores was conducting water at that inflow rate. When the inflow rate was stepped up, outflow rate and core mass rose rapidly until a new steady state was reached. We interpret the increase in core mass as indicating an increase in the cross-sectional area of pores that were conducting water.

As the inflow rate was stepped up, the mass of the core increased at a steadily decreasing rate. Plotting the rate of increase (slope) against the steady-state flow rate for all cores from the four sites gave a highly

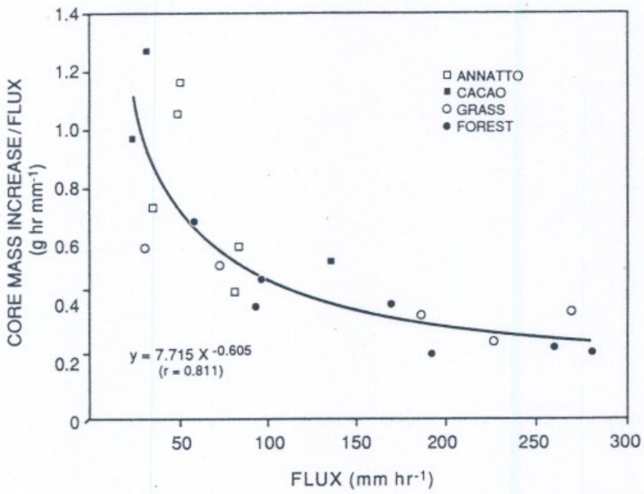

Fig. 2. Increase in core mass per unit flux change vs. flux. Data are from three cores for the annatto site, two for the cacao, three for the grass, and four for the forest, and include all curves shown in Fig. 1.

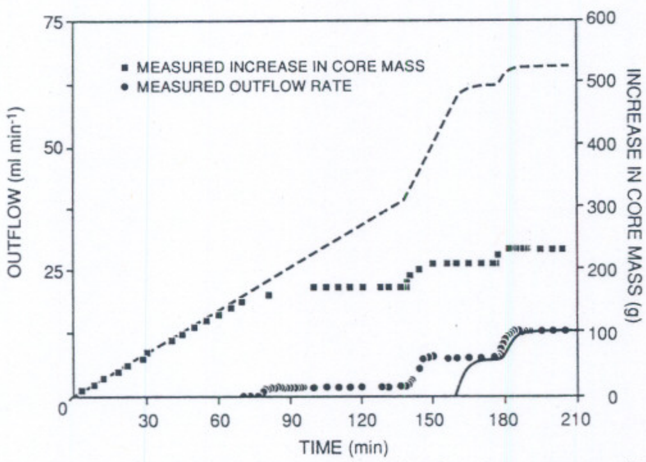

Fig. 3. Comparison of measured results for the annatto soil (same core as used for Fig. 1) with predictions for a coarse sand based on Richards' equation.

significant inverse curvilinear relationship (Fig. 2). That the relationship is inverse means that the lowest (initial) application rate caused the largest increase in core mass; stepping up to successively higher application rates caused successively smaller increases.

Figure 3 compares results for a core from the annatto site (CATIE) with the behavior predicted with the LEACHM model for a sandy soil without macropores. Observed time to breakthrough (75 min) was much less than predicted (160 min). Moreover, LEACHM predicted that the core should contain 475 g of water at the predicted time of breakthrough, whereas the core in fact contained 155 g of water when breakthrough occurred.

In the second set of experiments, we further studied the relation between water-application rate and time to breakthrough. We found that time to breakthrough decreased with increasing inflow rate (Fig. 4). Dividing core length by time to breakthrough then gave a maximum velocity (ignoring tortuosity). These max-

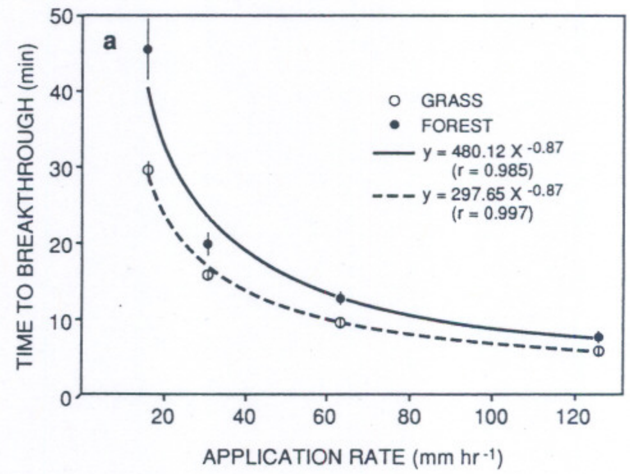

Fig. 4. Relationship between application rate and time to breakthrough. Each point is the mean of three or four cores, ± 1 standard deviation.

imum velocities increased with increasing flow rate and were >10 times the application rates, indicating that water flow was concentrated in a cross-sectional area less than one-tenth that of the total core. This proportion is similar to the proportion of total soil volume accounted for by noncapillary pores in these soils (Radulovich et al., 1989).

DISCUSSION

How can water flow rapidly through large pores in the soil cores without visibly ponding on the soil surface? A picture helps outline a conceptual model, first for microaggregated soils, such as are common in the humid tropics (Fig. 5a), then for soils that do not exhibit microaggregation (Fig. 5b). In the microaggregated soils, noncapillary interpedal pores are abundant and the infiltrability of the microaggregates (micropeds) is very low. Macropores, such as cracks, earthworm burrows, and root channels may also be abundant, especially if the soils are under perennial plant cover or have not been tilled recently. The application rates that we used probably exceeded by a wide margin the hydraulic conductivity of the microaggregates. For example, in some microaggregated Oxisols, matrix conductivity decreased from ≈120 mm h^{-1} near saturation to <0.1 mm h^{-1} near field capacity (≈ −10 kPa in these soils) (Sharma and Uehara, 1968). If the matrix conductivity of our soils is of similar magnitude, an application rate of only 0.1 mm h^{-1} would exceed the capacity of the microaggregates in such soils to absorb and conduct water. Once the hydraulic conductivity of the microaggregates is exceeded, the surfaces of those aggregates will become saturated and water will flow around them and move downward through the soil. (This pore system that surrounds the microaggregates in such soils we term here the *interpedal pore network*.) Once in the interpedal pore network, water may be funneled into larger pores (macropores) or converge into fingers, causing yet more localization of flow and thus further bypassing of the soil matrix. Thus, a soil need not be saturated for water to flow rapidly through the cores, as seen in Fig. 1.

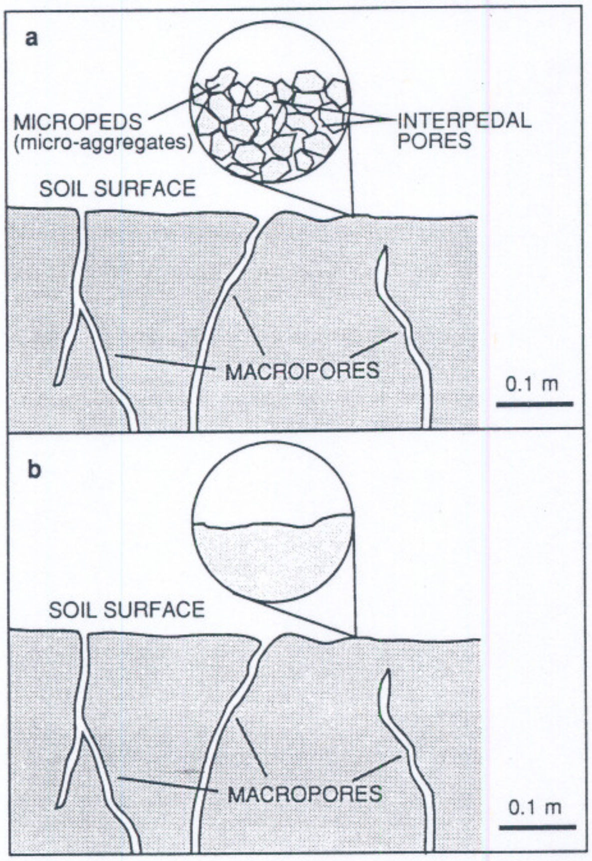

Fig. 5. Schematic representation of the types of pores found in (a) microaggregated soils and (b) other soils. Although both types of soils may contain both micropores and macropores, only the microaggregated soils exhibit large amounts of interpedal porosity.

Many soils lack a strong microaggregate structure and thus lack an extensive interpedal pore system (Fig. 5b). Macropores may be abundant in some such soils, for example, those under forest cover (e.g., Watson and Luxmoore, 1986; Phillips et al., 1989), but the total number (and cross-sectional area) of noncapillary pores is still far less than in microaggregated soils (Fig. 5b vs. 5a). Ponding in nonmicroaggregated soils becomes visible at the surface as soon as K_{sat} of the matrix is exceeded, and runoff occurs as water moves on the soil surface until it encounters, and then flows down, a large pore. Thus, the initiation of bypass flow in nonmicroaggregated soils coincides with the onset of visible ponding, and the extent of bypass flow should be inversely proportional to the incidence of visible ponding (insofar as macroporosity and saturated infiltration rate are related).

It should be noted that pores 0.1 μm or smaller accounted for ≈60% of total porosity in the La Selva soils (Sollins and Radulovich, 1988). For water to move down 200-mm through 0.1-μm pores would take 2700 d, calculated with the method of Radulovich et al. (1989). Thus, the water in such pores must be virtually stagnant, perhaps contributing to the unusually high rates of denitrification reported for some La Selva soils (Robertson and Tiedje, 1990).

The experiments reported here show an inverse curvilinear relationship between application rate and rate of increase in core weight per unit flux (Fig. 2). This relation suggested to us the following scenario for water flow through the soil. At low inflow rates, flow is restricted to the interpedal pore system (smaller noncapillary pores) and to the walls of the larger pores. As the inflow rate is increased, the layer of water flowing down walls of noncapillary pores becomes thicker, causing successively larger pores to fill. The limit to this process is reached when all pores are filled and the saturated flux rate is reached. This scenario extends that of Wilson and Luxmoore (1988), who assumed that mesopore conduction must preceed macropore conduction, whereas ours assumes that water begins to flow along the pore walls as soon as the walls saturate, and may thus begin flowing through noncapillary pores of all sizes nearly simultaneously, although perhaps not at the same rate through all of them.

Our hypothesis that water flows along the walls of unfilled pores is supported by results of our second experiment (Fig. 4), which showed that time to breakthrough decreased with increasing application rate and approached an asymptote at an application rate well below that required to achieve saturation. The traditional concept that successively larger pores become water filled as the application rate is increased would not explain this tendency for time to breakthrough to level off short of saturation. Our model, on the other hand, would suggest that, as flow rate increased, water transport took place farther away from the pore walls and thus incurred less resistance due to friction. The limit to this decreasing function occurred as the water-film thickness reached a threshold beyond which friction remained essentially constant.

CONCLUSIONS

The importance of bypass water flow in altering movement of nutrients and pollutants through the soil has been much discussed. It is usually assumed, however, that bypass flow occurs only when soils are at or near saturation and that bypass flow thus occurs only rarely. Our results show that bypass flow can occur in well-aggregated soils at application rates much lower than the saturated flow rate, indeed at application rates less than typical rainfall rates. Bypass flow may thus be the rule, rather than the exception, for well-aggregated soils (e.g., many tropical and volcanic soils) and forest soils.

In general, such a change in paradigm would require reevaluation of rates of nutrient entry to and loss from the soil matrix (Sollins and Radulovich, 1988; Solórzano and Radulovich, 1989). For forest ecosystems, some basic assumptions about nutrient-cycling pathways would have to be reexamined. For example, nutrients leaching from the forest floor may tend to move more rapidly through the soil profile with less likelihood for root uptake than previously thought (c.f., Sollins, 1989). For agriculture, the change in paradigm would require that strategies for fertilizer application in tropical agricultural soils be redesigned to increase fertilizer incorporation while minimizing leaching to the groundwater.

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