



Investigating raindrop effects on transport of sediment and non-sorbed chemicals from soil to surface runoff

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Abstract

A simple modeling and laboratory investigation was carried out to investigate the raindrop effects on both sediment detachment and chemical transport from soil–water into runoff. Solute movement between soil–water and runoff is usually modeled as either a mixing model or as a diffusion-like process, both of which ignore the important roles of raindrop impact on the transport process. We hypothesized a process-based chemical transport model that incorporated both rain-drop induced mixing and diffusion and tested it using a small-scale experiment in which simulated rainfall fell on soil, pre-saturated with chloride (Cl^-) laden water. We simultaneously observed sediment and Cl^- runoff concentrations trends and the evolution of the ‘shield’ layer composed of relatively heavy particles that resettle after each raindrop-impact. Using recently published and directly measured parameters, the model results generally agreed very well with measured concentrations. The exception was for the early (<5 min) Cl^- transport, which was faster than the model predicted, suggesting that an additional process needs to be added to our model. Even with this deficiency, the model developed here described our experimental results better than popular ‘mixing-layer’ and ‘diffusion’ models. This study provides a new approach to chemical transport modeling by linking the rain-controlled processes with similar soil erosion processes.

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1. Introduction

When raindrops impact the soil, they eject sediments, soil–water, and associated chemicals

from the soil surface. According to the USDA (1989), erosion is the source of 99% of the total suspended solids, 80% of the total phosphorus, and 73% of the total Kjeldahl nitrogen in the waterways of United States and undoubtedly around the world. A thorough understanding of the raindrop effects on sediment detachment and chemical transport from soil to runoff is needed to develop models that can be used

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to predict sediment and chemical transport and, ultimately, provide solutions to non-point source pollution control.

Although many models have been developed that attempt to simulate raindrop effects on sediment detachment, research continues in earnest (e.g. Gao et al., 2003; Parlange et al., 1999; Sander et al., 1996, 1998; Rose et al., 1994; Sharma et al., 1993). Similarly, many models have been developed to predict chemical transfer between soil–water and runoff (e.g. Zhang et al., 1999, 1997; Wallach, 1991; Ahuja, 1990; Steenhuis et al., 2001, 1994; Steenhuis and Walter, 1980). The chemical transfer models can be generally grouped into two categories, ‘mixing-layer’ models and ‘diffusion-like’ models, none of which explicitly account for the influences of raindrop impacts.

Rose (1985), Hairsine and Rose (1991), and Rose et al. (1994) developed a physically based, or process-based, soil erosion model (Rose model) that includes raindrop effects on sediment detachment. The Rose model has been successfully applied to a variety of laboratory studies (e.g. Gao et al., 2003; Heilig et al., 2001; Sander et al., 1996). Obviously, transport of sediment-bound chemicals in the surface soil could be predicted directly by this model. One of the unique characteristics of Rose model is the development of a shield layer, which protects underlying soil during prolonged events (e.g. Rose, 1985; Rose and Hairsine, 1988). The shield layer develops on top of the soil due to the preferential accumulation of rapidly settling, heavy particles after repeated raindrop-ejection of soil into the overland flow.

Heilig et al. (2001) demonstrated the fundamental conceptual basis of the Rose model with a simple experiment that used a man-made soil of visually contrasting white clay and black sand to observe the formation and protective nature of the surface shield. In this paper we used a similar approach to develop a solute transport model, which considers both raindrop detachment and diffusion. To test the model, we modified the simple experiments developed by Heilig et al. (2001) and Gao et al. (2003) by adding a chloride (Cl^-) solution to the clay–sand soil. The objective of this investigation was to improve our understanding of the fundamental mechanisms involved in raindrop induced solute transport.

2. Theory

During rainfall, raindrop detachment/mixing near the soil surface and diffusion of solutes from deeper in the soil are two important processes controlling solute transport from soil to runoff. Considering only these two processes and assuming they can be superimposed, the governing equation for solute transport can be written as:

$$\frac{\partial(DC)}{\partial t} + \frac{\partial(qC)}{\partial x} = e_s + e_d \quad (1)$$

where D is the ponding water or runoff depth at the surface, C is the solute concentration in runoff, e_s is the solute transport rate of raindrop controlled processes, e_d is the solute transport rate of diffusion, and q is the overland flow rate.

The diffusion transport can be approximated by the theory developed by Wallach and his colleagues, which assumes that there is a thin, laminar boundary layer at the soil surface that controls solute transfer between soil and runoff (Wallach et al., 1988; Wallach, 1991; Wallach and Van Genuchten, 1990). The diffusion rate, e_d , can be written as:

$$e_d = k_d C_s \quad (2)$$

where C_s is the solute concentration in the soil and k_d is the diffusive mass transfer coefficient, which can be determined theoretically following Appendix C of Wallach et al. (1988).

During rain-induced soil erosion, each raindrop ejects both soil-particles and soil–water into runoff. The Rose model accurately describes this processes with respect to sediment detachment and transport in runoff (e.g. Gao et al., 2003; Parlange et al., 1999; Sander et al., 1996, 1998). Rose (1985, 1983) proposed the following equation to relate the sediment detachment or entrainment, e , to rainfall intensity, p :

$$e = aP^b \quad (3)$$

where a is the soil detachability (M l^{-3}), P is the rainfall rate, and b is a constant. Relatively recent studies have shown that the constant $b=1$ (Sharma et al., 1995; Jayawardena and Bhuiyan, 1999; Gao et al., 2003). Noting that the soluble chemical is in the soil–water, Eq. (3) can be revised to represent

the ‘solute entrainment’:

$$e_s = \frac{aP\theta}{\rho_s} C_s \quad (4)$$

where θ is soil–water content, ρ_s is the density of the solute, C_s is solute concentration in soil–water.

3. Experimental methods

The experimental set-up is simple and designed to investigate raindrop effects on both soil particle detachment and chemical transport to surface runoff. The physical set-up, shown in Fig. 1, is similar to that used by Heilig et al. (2001) and Gao et al. (2003). A small, cylindrical, Plexiglas container (column, area=45 cm²) was filled with man-made soil, saturated with a solution of Cl⁻, and leveled 3 m below a computer-controlled rainmaker. After carefully establishing an initial ponding depth of 7 mm on the soil, rain was simulated over the cylinder and overflow from the four holes in the side of the chamber maintained a constant ponding depth. Experiments

were run at three different rainfall intensities: 30, 52, and 74 mm h⁻¹. Experiments were replicated in duplicate or triplicate to show the characteristic scatter of the data when experiments are duplicated. Outflow was systematically collected from one of the holes on the container with a 0.1 ml pipette. The first sample was taken just before applying the rain and subsequent sampling occurred every 1–3 min depending on the rainfall rate. Later samples were typically collected at longer intervals as the dynamic behavior of the system slowed. Each experiment continued until the ponded water was completely clear of sediment, 1.5–3 h depending on the rainfall rate. The sediment concentration of each sample was measured using the same spectrophotometer and methods used by Heilig et al. (2001). The Cl⁻ concentrations were measured by Digital Chloride-meter (Buchler Instruments, M442-5000).

The soil, 90% sand (180–212 μm) and 10% clay (hydrous Kaolin supplied by Englehard Corp, NJ), was mixed and saturated with Cl⁻ solution. In order to minimize clay aggregation, Cl⁻ was dissolved from NaCl. The initial concentration was 29,820 ppm;

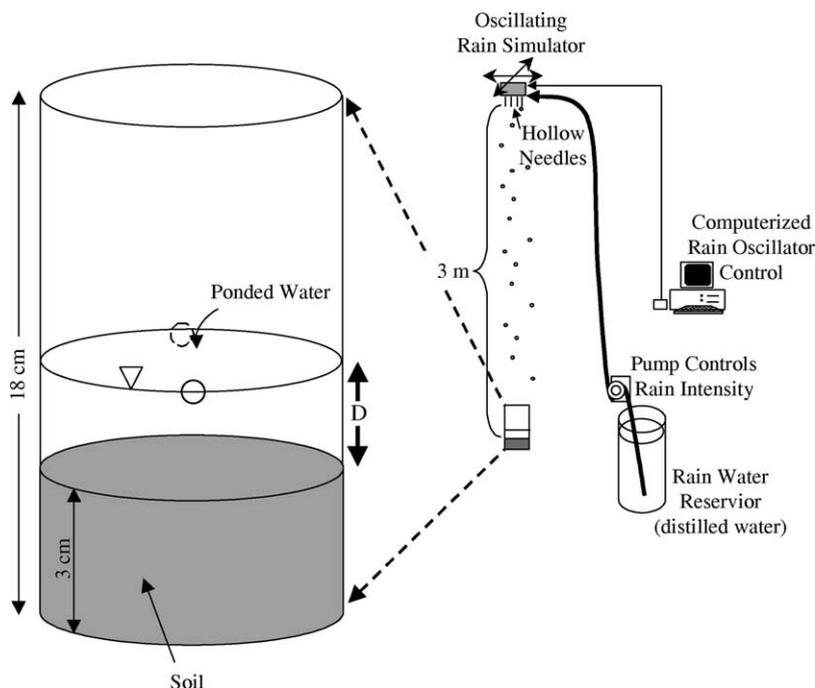


Fig. 1. Schematic of the experimental apparatus and set-up (not to scale).

this large concentration was to ensure that Cl^- would not drop below detection limits before relevant processes were observed and/or measured. The saturated water content was $\theta_s = 0.365$. The soil was put into the container and placed under the rainmaker, where the soil surface was smoothed and leveled. Preponding was achieved by placing a strip of paper on the soil surface to prevent mixing during the preponding process. Water was carefully pipetted onto the paper until 7 mm deep, at which point the paper was carefully removed with minimal disturbance to the soil.

The computer-controlled rainmaker used in the experiments is the same as Gao et al. (2003). The simulated rainfall consistently had uniformity of > 0.9 and the rain intensities were tested twice before and once after each experiment.

4. Application of the theory to the experiments

Under our experimental conditions, Eq. (1) reduces to:

$$D \frac{dC}{dt} = e_s + e_d - PC \quad (5)$$

For notational simplicity we combined the constants in Eqs. (2) and (4) into a combined or effective transfer coefficient that takes the form:

$$k_e = k_d + k_s \quad (6)$$

where $k_s = (aP/\rho_b)\theta$, Eq. (5) can be rewritten as:

$$D \frac{C}{t} = k_e C_s - PC \quad (7)$$

Letting C_s be the chemical concentration in the soil–water near the soil surface and following Wallach et al. (1988), C_s can be expressed as:

$$C_s = C_0 \exp(h^2 D_E t) \operatorname{erfc}(h \sqrt{D_E t}) / \theta \quad (8)$$

where $D_E = D_s / \theta$, $h = k_e / D_s$ and C_0 is the original concentration of solute in the soil. Eq. (7) can be solved by using Eq. (8) and the initial condition $C = 0$ at $t = 0$. The normalized solute concentration, C/C_0 , in the runoff as a function of dimensionless time,

$T = Pt/D$, is:

$$\frac{C}{C_0} = \frac{\zeta}{1+w} [\exp(wT) \operatorname{erfc}(\sqrt{wT}) - \exp(-T)] + 2 \sqrt{\frac{w}{\pi}} E(\sqrt{T}) \quad (9)$$

where $\zeta = k_e J / (P\theta)$ and $w = k_e^2 D / (P\theta^2 D_E)$, are dimensionless groups, P is the rainfall rate, and $E(x) = \exp(-x^2) \int_0^x \exp(y^2) dy$ is Dawson's integral (Abramow and Stegun, 1970). Eq. (9) describes the change of non-absorbed solute concentration in runoff.

To determine k_d for our experiments, the diffusion constant, D_s , was based on the published Cl^- -water diffusion coefficient ($D_w = 1.6 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$) and the Millington and Quik (1961) tortuosity model, as also used by Wallach et al. (1988), Ahuja (1990), and Zhang et al. (1997):

$$D_s = D_w (\theta^{10/3} / \theta_s) \quad (10)$$

where θ is the soil–water content and θ_s is the saturated water content (0.365 for our soil). For our experiments $D_s = 4.2 \times 10^{-6} \text{ cm}^2 \text{ s}^{-1}$. This value was used to determine the transfer coefficient, k_d , determined theoretically following Appendix C of Wallach et al. (1988): $k_d = 4.6 \times 10^{-5} \text{ cm s}^{-1}$ ($P = 74 \text{ mm h}^{-1}$), $k_d = 3.8 \times 10^{-5} \text{ cm s}^{-1}$ ($P = 52 \text{ mm h}^{-1}$), and $k_d = 2.9 \times 10^{-5} \text{ cm s}^{-1}$ ($P = 30 \text{ mm/h}^{-1}$). Heilig et al., (2001) used the same soil that we used in these experiments and determined the soil detachability, $a = 400 \text{ mg cm}^{-3}$. The effective solute transfer coefficients, k_e (Eq. (6)), for our experiments were: $k_s = 2.1 \times 10^{-4} \text{ cm s}^{-1}$ ($P = 74 \text{ mm h}^{-1}$), $k_s = 1.5 \times 10^{-4} \text{ cm s}^{-1}$ ($P = 52 \text{ mm h}^{-1}$), and $k_s = 0.9 \times 10^{-4} \text{ cm s}^{-1}$ ($P = 30 \text{ mm h}^{-1}$).

5. Results and discussion

To corroborate our soil detachability, taken from Heilig et al. (2001), we applied the Rose model to our sediment data. The Rose model, as applied to these experiments, requires a value for the theoretical maximum depositional depth, a.k.a. 'shield' depth. Our shields developed very quickly and were almost complete within 5–10 min and did not grow substantially over the next several hours.

Table 1
Experimental results for clay with three different rainfall intensities

Clay	$P=74$ mm/h			$P=52$ mm/h		$P=30$ mm/h		
	Run 1	Run 2	Run 3	Run 1	Run 2	Run 1	Run 2	Run 3
M^* (mg/cm ²)	800	740	870	610	600	500	530	520
Shield depth (mm)	7.7	7.5	7.6	5.0	5.9	4.0	4.3	4.2
R^2		0.94			0.91		0.95	

Because our experimental soil was level, i.e., not sloped, the observed maximum depositional shield depth should be approximately equal to the theoretical maximum. On a sloped surface, by contrast, the depositional material saltates down slope with each raindrop impact and, thus, the shield is never ‘complete’ and new material is continually eroded. Thus, we were able to measure the maximum shield depth directly by dissecting the soil after each experiment to determine the mass of the full shield, M^* (mg cm⁻²) (Table 1). As observed by Heilig et al. (2001), the M^* values and associated shield depths were somewhat positively correlated to the rain intensities (Table 1). This might be because high rain rates increase the probability that multiple drops will fall very near one another at the same time, creating deeper ‘craters’ than drops acting independently and thus creating deeper shields and larger M^* ’s. Interestingly, the use of a cylindrical container improved the shield uniformity from that achieved by Heilig et al. (2001) with their cubical container. Also note that within our precision of measurement, the shield depth in these experiments was the same as observed when the soil was saturated with salt-free water, i.e., Heilig et al. (2001), indicating that the presence of the salt did not have an obvious impact on raindrop–soil interactions.

Fig. 2 shows the experimental sediment results for the three different rain intensities. Using the average M^* for each rainfall rate and soil detachability, a , derived from Heilig et al. (2001), we applied the Rose model to our data with good overall results (Fig. 2), $R^2 > 0.9$ in general. This reinforces our justification of applying the Heilig et al. (2001) soil detachability to our solute transport model. It is interesting that the results shown in Fig. 2 do not exhibit the slight systematic error described by Heilig et al. (2001). We

think this is because we directly sampled the ponded water rather than collecting outflow as Heilig et al. (2001) did. In all plots, clay concentration initially increased very quickly to a peak concentration and decreased to zero for long times, indicating that the shield was complete and prevented on-going erosion. These results agree with previous studies by Heilig et al. (2001), Parlange et al. (1999), Sander et al. (1996), and others.

Cl^- concentrations in runoff at different times are plotted in Fig. 3 for the three rainfall rates. Our model, Eq. (9), generally agreed well with our experimental Cl^- data (Fig. 3). Unlike data from similar experiments by other researchers (e.g. Zhang et al., 1997; Ahuja and Lehman, 1983), our data show the initial, rising limb of the concentration curve. This difference is expected because our experiments used a constant ponding water depth so that at $t=0$ the solute concentration was, ideally, zero. Most other published experiments start with no initial ponding so, although they are unable to sample as early as we did. We expect our early concentrations to be lower than theirs because our initially ponded water dilutes the early concentrations. Our model appears to be inappropriate for predicting the short, rising limb of the concentration-time curve, which suggests further work is needed to fully understand the relevant processes. We anticipate that explicit inclusion in the model of soil–water pumping by raindrops will capture the early behavior better. On the other hand, the time durations to the peak concentrations are very short compared to the whole experimental duration and they are short compared to the sediment concentration peaks (Fig. 2). This may lend some justification for the ‘instantaneously mixing’ assumption used in other solute transport models (Zhang et al., 1997; Steenhuis et al., 2001, 1994; Steenhuis and Walter, 1980). The average R^2 for the model performance associated

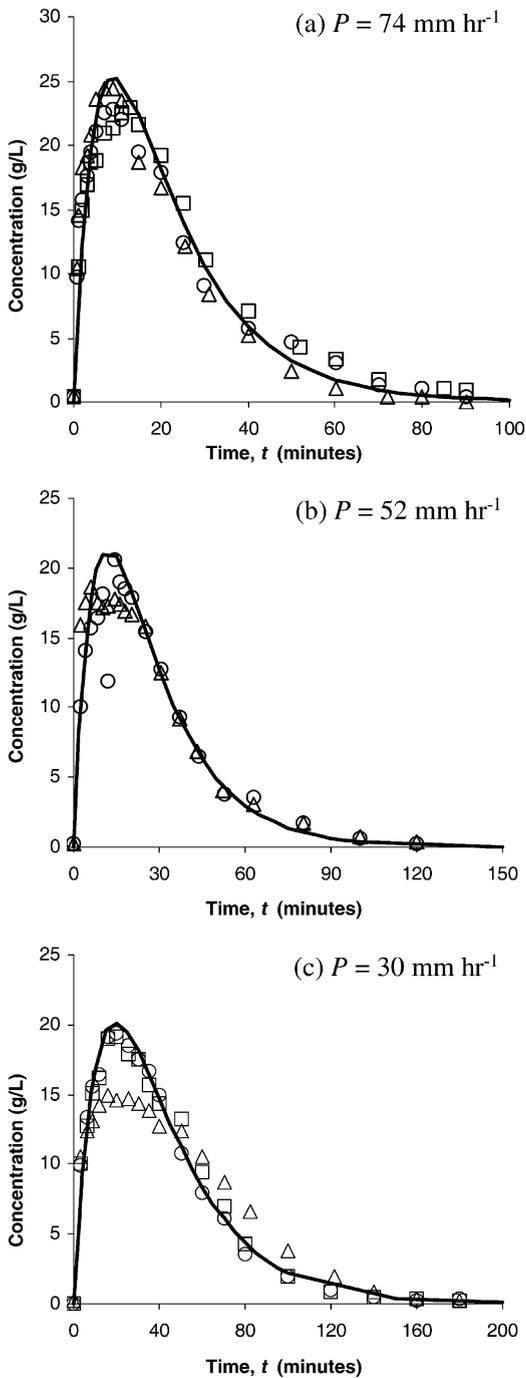


Fig. 2. Clay transport runoff with three different rain intensities (a) $P=74 \text{ mm h}^{-1}$, (b) $P=52 \text{ mm h}^{-1}$, (c) $P=30 \text{ mm h}^{-1}$. Symbols: data, circles, triangles, and squares correspond to experimental runs 1, 2, and 3, respectively, as shown in Table 1; lines: Rose model.

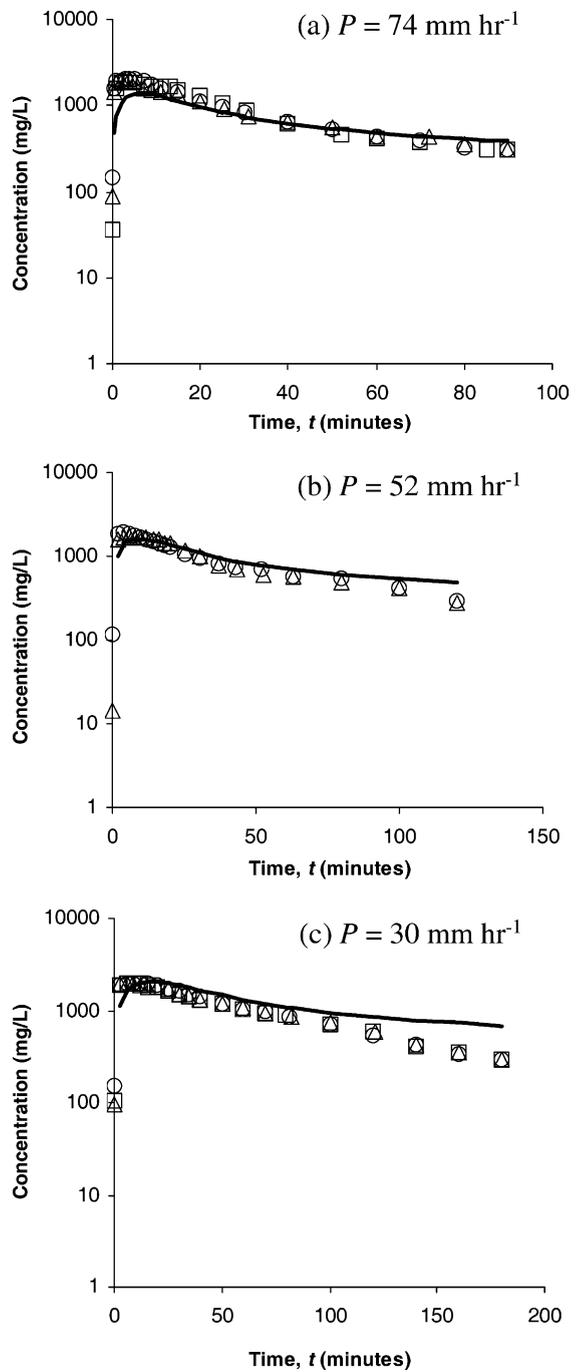


Fig. 3. Transport of Cl in runoff with three different rain intensities, (a) $P=74 \text{ mm h}^{-1}$, (b) $P=52 \text{ mm h}^{-1}$, (c) $P=30 \text{ mm h}^{-1}$. Symbols: data, circles, triangles, and squares correspond to experimental runs 1, 2, and 3, respectively, as shown in Table 1; lines: Rose model.

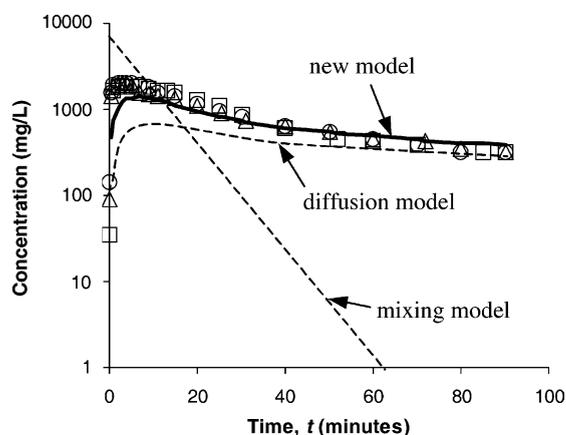


Fig. 4. Example comparison of different solute transport models with experimental data ($P=74 \text{ mm h}^{-1}$). Symbols: data, circles, triangles, and squares correspond to experimental runs 1, 2, and 3, respectively, as shown in Table 1; solid line: new model presented in this paper; dashed lines: other recognized models as labeled.

with $t > 5 \text{ min}$ were: 0.92 for $P=30 \text{ mm h}^{-1}$, 0.88 for $P=52 \text{ mm h}^{-1}$, and 0.98 for $P=74 \text{ mm h}^{-1}$. These generally high R^2 suggest that our model correctly describes the solute behavior for all but the earliest time period. Note: Gao et al. (2004), in a paper submitted for publication after this one, were able to get better results by restructuring the model so that diffusion transported chemicals into the ‘mixing-layer’ and raindrop impacts transported chemicals into overland flow; in essence the revised model assumed the processes occurred in series as apposed to the parallel arrangement presumed in this study.

By way of comparison, we applied a ‘mixing-layer’ model and a ‘diffusion-like’ model to the data from one of our experiments and neither performed well (Fig. 4). For the mixing model we used the observed ‘shield’ depth (4–7 mm) as the theoretical mixing depth (Table 1). Indeed, the ‘shield’ depth is the maximum penetration of the raindrops and conceptually matches the theoretical description of the mixing layer (Steenhuis et al., 1994, 1980; Zhang et al., 1997, 1999; Ahuja and Lehman, 1983). However, the observation that the chemical concentrations remain elevated after a long time indicates chemical transport from below the shield, or mixing layer, which violates a basic assumption in the

mixing-layer theory. Thus poor agreement between the mixing-layer model and our data is not unexpected. Note also, that the ‘mixing-layer’ model can be fitted to the data only by changing both the mixing-depth and initial concentration and, thus, it is not obvious how to interpret the parameters. The ‘diffusion-like’ model that we used is from Wallach et al. (1988) and is essentially the same as our model, Eq. (5), without the raindrop detachment term, e_s . Predictably, this model under-estimates the solute concentration. Wallach et al. (1988) originally overcame this deficiency by adding a term that accounted for dynamic soil–water exchanges during the development of a steady runoff depth. In these experiments our runoff depth is constant so this term cannot be meaningfully added. Other researchers have arbitrarily increased the diffusion coefficient to fit data without providing a straightforward justification or protocol for doing so. Note, however, that the Wallach et al. (1988) model agrees reasonably well with the experimental data for long time periods, indicating that prolonged movement of solutes between soil–water and overland flow is predominantly controlled by diffusion in these experiments.

6. Conclusions

Our results show that raindrop controls on solute transport can be directly incorporated into a solute transport model. By including raindrop-controlled solute transfer, based on the conceptual Rose model, with diffusion processes, as determined by Wallach et al. (1988), we developed a model that agreed relatively well with experimental data. Our results show that the mechanisms driving the rapid, early solute transfer from the soil into runoff are not well represented in models and we hypothesize that a rain-impact soil–water pumping mechanism may be an important control not currently considered in any models. As a final comparison, we showed that a popular ‘mixing-layer’ model and ‘diffusion’ model were unable to meaningfully fit our data, i.e., without seemingly arbitrary calibration. This paper suggests a new, physically-based solute transfer model for which all parameters can be directly measured. See also Gao et al. (2004), which expands on this work.

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