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Dynamic Nonequilibrium During Unsaturated Water Flow

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Abstract. Traditionally, the water retention characteristic is measured under equilibrium or steady-state flow conditions, and it is assumed that the so-determined properties can be used to model transient water flow under field conditions. However, comparison of retention data obtained with the conventional method and using transient methods indicate the presence of "dynamic effects" which show up as differences between static and dynamic hydraulic functions. This paper reviews previous evidence of dynamic effects, illustrates the presence of dynamic effects during a multistep outflow/inflow experiment, and discusses an experiment specifically designed to quantify dynamic effects. Finally, a comparison of simulations with the classical Richards model and a two-phase flow model show that dynamic effects are to a considerable part caused by non-negligible resistances of air flow.

INTRODUCTION

Hydraulic properties of porous media are generally assumed to be of a static nature, i.e., to depend only on the distribution and geometric arrangement of pores in the porous medium and on the wetting/drying history (hysteresis), provided the solid matrix is rigid, fluid properties are constant, and external conditions do not change with time. This implies that the relationship between water content, θ, and matric potential, ψ, during monotonous draining or imbibition processes is not affected by the dynamics of water flow, and that $\theta(\psi)$ is expressed by a unique characteristic. Traditionally, the water retention characteristic is measured in the laboratory under equilibrium conditions, and it is generally assumed that the so-determined properties can be used to model transient water flow under field conditions. However, comparison of drainage retention data obtained with dynamic methods and using the conventional approach indicate that more water will be held at a given suction in the dynamic case. According to *Klute* [1986], "the reasons for and implication of this observation continue to be a subject of investigation." If dynamic nonequilibrium occurs between the water content and the water potential during transient water flow, the presently used standard simulation technique for describing saturated/unsaturated flow cannot be reliably applied under transient flow conditions.

The aim of this paper is to investigate dynamic effects during transient flow experiments, which are generally used to identify hydraulic parameters by inverse simulation. Contrary to most investigations in the past, we use undisturbed columns and investigate both the imbibition and the drainage process. Furthermore, we use smooth pressure changes at the column boundary instead of the commonly applied stepwise changes. The paper is organized in three parts. First we briefly review past experimental evidence of dynamic nonequilibrium during unsaturated water flow. Then we present an experiment involving a series of outflow/inflow processes applied to soil samples at different flow rates, but otherwise identical conditions. An evaluation of the experiment by inverse simulation allows us to assess the importance of dynamic effects on flow processes under natural conditions. In the final part of the paper, we simulate our experiment with a two-phase model, and show that dynamic nonequilibrium may be caused by limited air permeability at high water saturation.

EVIDENCE OF DYNAMIC NONEQUILIBRIUM IN THE LITERATURE

Investigations by various authors in the past indicate that hydraulic properties may depend not only on the wetting and drying history, but also on the dynamics of water flow [*Davidson et al.*, 1966; *Topp et al.*, 1967; *Smiles et al.*, 1971; *Vachaud et al.*, 1972; *Stauffer*, 1977; *Lennartz*, 1992; *Plagge*, 1991; *Plagge et al.*, 1999; *Kneale* 1985]. Without relating the observed phenomena to any specific process concept, we may categorize the observations in a general way as *dynamic nonequilibrium*. Dynamic nonequilibrium during water flow may be important because of the implication that hydraulic functions measured under static equilibrium conditions may not be applicable to dynamic water flow simulations.

In imbibition and drainage studies, *Davidson et al.* [1966] found that equilibrium water contents apparently were affected by the dynamics of the wetting history. More water was removed from samples when one single large pressure decrease was imposed rather than a sequence of small decreases. In contrast, more water was sorbed by the soils when a series of small pressure steps was applied in the imbibition process. *Topp et al.* [1967] compared water retention characteristics of vertical sand columns under static equilibrium conditions, under steady-state flow, and under transient flow conditions. They found for transient experiments that, at a given potential, water contents during drainage were significantly higher than water contents determined under static equilibrium or steady-state conditions. *Smiles et al.* [1971] performed experiments where a series of imbibition/drainage cycles was applied to horizontal sand columns by imposing stepwise changes in the water pressure at one column end. They found that the retention characteristic was a unique function during imbibition, but not during drainage. For a given initial water content, which had to be greater than some critical value, the retention characteristic depended on the size of the imposed pressure step, and on the time required to approach static equilibrium. No deviations from atmospheric air pressure have been observed in the soil during any of these experiments. Using a similar experimental setup, *Vachaud et al.* [1972] confirmed these results for vertical soil columns.

Stauffer [1977] investigated drainage in vertical columns packed with sand. During dynamic flow he measured higher water contents at a given pressure as compared to the static $\theta(\psi)$ relationship. Again, no deviations in the gas phase pressure from atmospheric air pressure were observed. Stauffer showed that the dynamic process could be simulated with the Richards equation by using a dynamic retention characteristic, with the difference to the static characteristic depending upon the rate of change of the local water content with time. All of the above-mentioned experiments indicate that the dynamic effect depends on the size of the pressure changes, being larger for large changes. Some indication also exists that dynamic effects are more important for soils with a wide pore-size distribution.

Plagge [1991] confirmed these findings with evaporation experiments on silty soils. He measured matric pressures and water contents with micro-tensiometers and TDR probes which were installed at multiple depths in the columns. Dependent upon the distance from the boundary of the soil column, the locally measured retention curves differed considerably. Plagge could show that these differences were systematic and reproducible, and that they could not be attributed to shortcomings in the packing technique or to measurement errors. *Lennartz* [1992] investigated dynamic nonequilibrium systematically with evaporation experiments on repacked soil samples of four different substrates. He found clear differences between static and dynamic $\theta(\psi)$ characteristics, but could not identify a unique and simple trend.

Recently, *Wildenschild and Hopmans* [1999] investigated the rate-dependence of unsaturated hydraulic properties for three disturbed soils in short laboratory columns by performing onestep and multistep outflow experiments. In situ retention characteristics were obtained from measured tensiometric pressures in the soil column and the average water content. They also found clear effects: water contents at a given tensiometric pressure were higher for fast drainage with large pressure drops. These findings were recently further confirmed experimentally by *Plagge et al.* [1999].

Based on thermodynamic considerations, *Hassanizadeh and Gray* [1993] postulated the existence of a dynamic component during unsaturated water flow. Their theoretical analysis yields an approximate capillary pressure equation with a dynamic term that depends linearly on the rate of change in water saturation:

$$p_a - p_w = p_c - \frac{1}{L}\frac{\partial \Theta}{\partial t} \tag{1}$$

where p_w is the macroscopic pressure in the water phase, p_a is the pressure in the air phase, p_c is the capillary pressure, L is a non-negative material coefficient, Θ expresses the soil water saturation, and t indicates time. Equation (1) states that the pressure difference between air and water is larger than the capillary pressure under drainage conditions, and smaller when imbibition occurs.

ILLUSTRATION OF DYNAMIC NONEQUILIBRIUM

The purpose of this section is to illustrate how dynamic nonequilibrium shows up in experiments which primarily focus on parameter identification by inverse modeling [*Durner et al*., 1996; *Schultze and Durner*, 1996]. Multistep experiments were performed on 15 cm long soil samples of an undisturbed sandy soil. The cores were instrumented with tensiometers and TDR probes. Details of the soil, the experimental device, and sample handling will be given in the next section. More information on the multistep procedure, and on inverse modeling, is given in *Durner et al.* [1999] and *Zurmühl and Durner* [1998]. Figure 1 shows a comparison of observed and simulated cumulative outflow (right), and the pressure at the lower boundary as well as tensiometer readings at two depths (left), for a typical experiment. A stepwise change of water pressure was applied at the bottom of the column, starting from saturation (+15.7 cm), down to –150 cm, and going back again to saturation. Both the drainage and wetting cycle took place within 96 hours. An equilibration time of 48 hours was allowed between drainage and imbibition. From Fig. 1 it is evident that, after a pressure step, tensiometer readings reached the new equilibrium levels relatively quickly, whereas outflow or inflow of water continued for periods of 24 hours or longer [*Schultze and Durner,* 1998]. The observed outflow dynamics cannot be simulated with the Richards equation and a single retention characteristic. We note that failure of the model to describe the shape of the outflow data could not be attributed to limited flexibility of the hydraulic model [*Zurmühl and Durner,* 1998]. Apparently, the most significant deviations between model and observation occurred in the moisture range near the air entry point. The phenomenon is not limited to the drainage branch (left side offigure), but occurs also during imbibition. This is contrary to observations of *Smiles et al*. [1971] who similarly investigated both imbibition and drainage, and found that observed dynamic effects occurred only under drainage conditions.

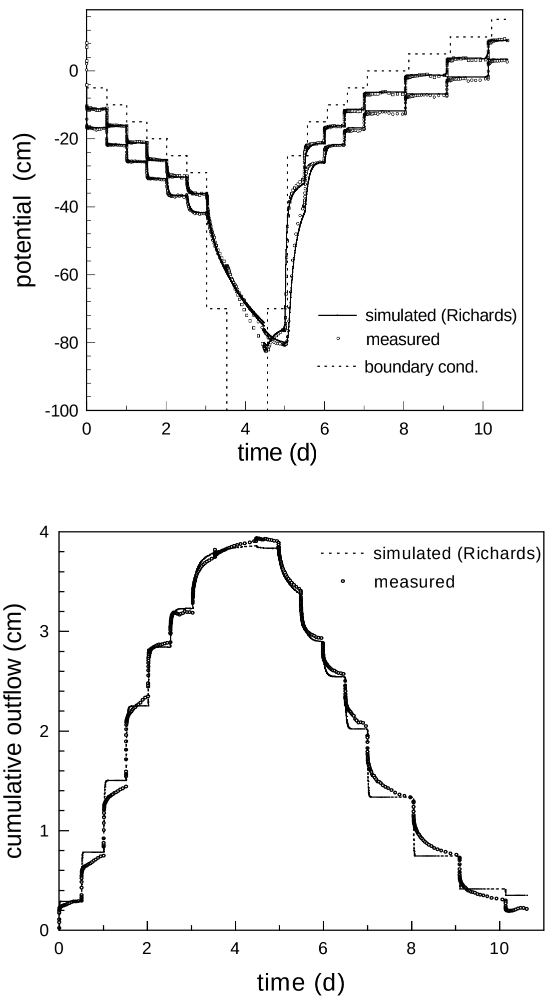

Fig. 1. Measured and simulated outflow and backflow during a multistep experiment on undisturbed Forchheim sand. Left: Tensiometric potential (measured), matric potential (simulated) and pressure at the lower boundary. Right: Cumulative water flow.

A different illustration of dynamic nonequilibrium is obtained if the tensiometric pressures which have been measured in two depths are plotted versus water contents, which have been simultaneously measured at the same depths (Fig. 2). Apparently, the pressure steps applied at the bottom boundary are somehow reflected in the in situ retention characteristics at both depths. For comparison, the van Genuchten retention curve which has been obtained by the inverse simulation of the Multistep experiment is also depicted. The deviations between the ideal smooth curve and the measurements are most pronounced in the relatively narrow pressure region of –15 cm to –35 cm for the drainage curve. For the imbibition curve, the influence of the pressure steps is limited to the pressure range of –25 cm to –5 cm. Figure 2 also indicates a systematic shift between the retention characteristics that were measured at the upper and the lower level, particularly for the imbibition part of the curve. This observation is in qualitative agreement with the findings of *Plagge et al.* [1999], although we cannot precisely judge the significance of the effect since our soil columns were undisturbed and local heterogeneity and layering could have impacted the results.

Based on the above observations, two hypotheses can be formulated:

1. Depending upon the rate of changes in the boundary pressure, the simulation may require different retention and conductivity curves. Inverse simulation of transient

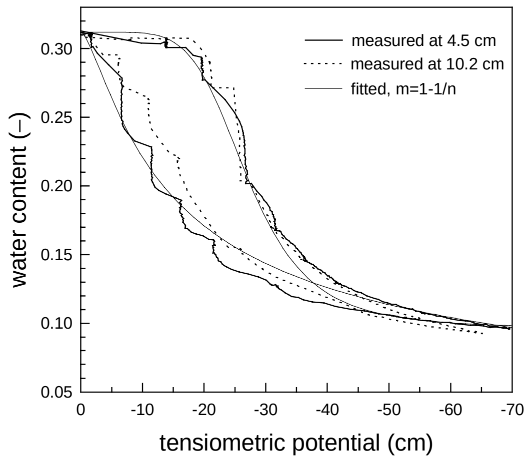

Fig. 2. In-situ measured retention curves of a sample of Forchheim sand during a multistep drainage and imbibition experiment.

experiments will therefore yield different results if boundary pressures change at a different rate.

2. Air movement near water saturation may be restricted because of limited gas phase permeability. Experiments on soil columns with and without lateral venting should therefore lead to different results.

We investigated the validity of these hypotheses by repeating a specific outflow/inflow scenario on a soil sample at different time scales.

MATERIAL AND METHODS

Soil and Experimental Device

We used a fully automated, computer-controlled device to perform inflow/outflow experiments with 1000 cm^3 soil samples (ID=9.4 cm, height=15.7 cm). Undisturbed soil samples were sampled by pushing guided plexiglass cylinders into a sandy forest soil (Typic Distrochept) near Forchheim, Bavaria, Germany. The texture is 95% sand, 1% silt, and 4% clay [*Deschauer*, 1995]. The samples were taken from the Bvs horizon at 15-40 cm depth. Since ease of access of air may influence water flow, we used soil columns with closed casings and perforated walls. The hole size of the perforation was 1.5 mm which enables lateral air access at capillary pressures below –2 cm.

The cores were placed on a glass sintered porous plate and mounted in the experimental apparatus. At distances of 4.5 cm and 10.2 cm distance from the top, pressure transducer tensiometers (tip diameter 6 mm, length 15 mm) and TDR probes (length 8 cm, rod distance 2 cm, coated rods, Type IMCO P2) were installed. Measurement accuracy for the tensiometers was 0.5 cm (absolute) and 0.2 cm (relative), and for the water content 2% (absolute) and 0.5 % (relative). The TDR calibration curves had to be re-calibrated according to the mass balance obtained by the outflow curves.

In a typical experiment, water pressure at the lower end of the soil column is changed from saturation to an unsaturated state, and back. Inflow and outflow of water, water pressures below the porous plate, and tensiometric pressures and water contents in two depths were continuously recorded. Figure 3 shows a schematic of the measurement device.

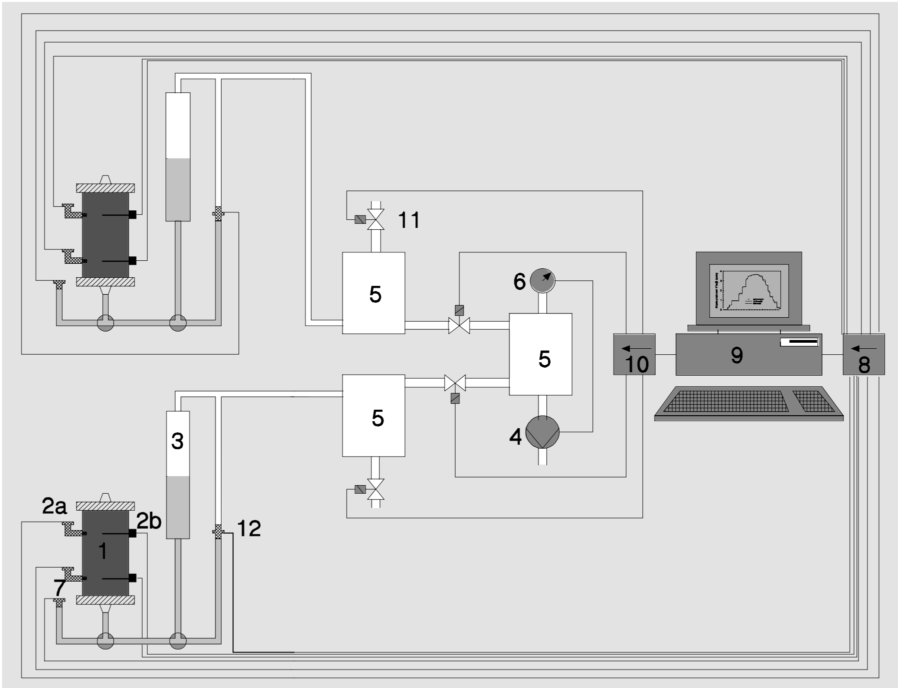

Fig. 3. Schematic of the experimental device: (1) soil core, (2a) microtensiometer with pressure transducer, (2b) TDR, (3) burette, (4) vacuum pump, (5) storage bin, (6) manometer, (7) pressure transducer, (8) multiplexer and analog/digital converter, (9) computer, (10) digital/analog converter, (11) magnetic valve, and (12) difference pressure transducer.

Boundary Conditions

To investigate how the rate of change in the pressure affects the parameter inverse simulation results, we used the "continuous" outflow/inflow experimental approach of *Durner et al*. [1996]. In this approach, the water pressure at the bottom of the soil column is changed smoothly from full saturation to an unsaturated state, and back. The top of the column has a no-flow boundary condition for water, but permits free access of air. After saturating the soil for 48 hours with a 0.001 $CaSO_4$ solution, the saturated conductivity was measured during constant gradient flow from bottom to top. Microbial activity was inhibited by adding 40 µM $AgNO_3$ to the fluid. The experiment consisted of four cycles of drainage and imbibition (Fig. 4). Initially, the soil was saturated at its top boundary, with the pressure at the lower bound being maintained at $\psi_l = 15.7$ cm. During each cycle, the pressure was decreased linearly from $\psi_l = +15.7$ cm to $\psi_l = -60$ cm, where it remained for a redistribution period of at least 24 hours. The pressure was then linearly increased. Another redistribution period finishes one cycle. The total time for the first cycle was 432 hours. The experiment was repeated three times, increasing the speed of the drainage and imbibition process by a factor of four each time. This process accelerated the drainage and imbibition rates from the first to the fourth cycle by a factor of 64. To make sure that no changes in the soil's pore system occurred during the measurements, the standard multistep experiment was repeated at the end of the entire measurement series, and was found to produce identical flow behavior.

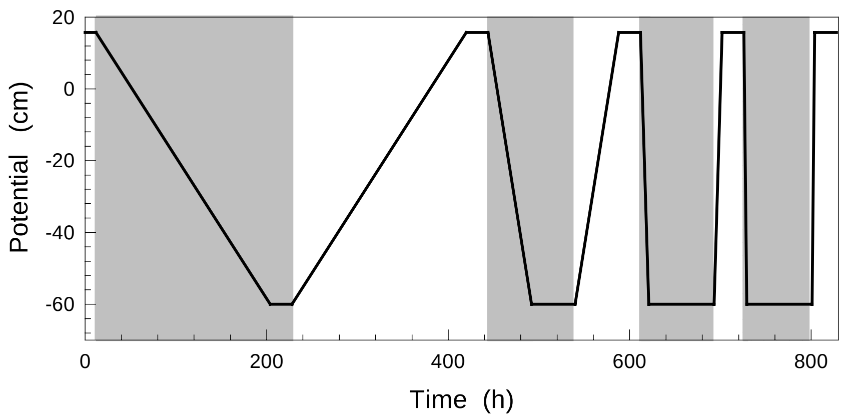

Fig. 4. Lower boundary condition of the outflow/inflow experiment with smooth continuous pressure changes. The experiment consisted of four cycles. The drainage and redistribution phase of each cycle is shaded. The rate of the pressure changes is first slow [cycle 1]. The experiment is then repeated three times with increasing flow rates [cycle 2 to 4]. In order to start each drainage or imbibition cycle with comparable initial conditions, the redistribution phases were extended in cycles 2 to 4.

RESULTS AND DISCUSSION

Inverse Optimization

The experiments were evaluated by inverse simulation of the flow process using the Richards equation as the flow model, and the van Genuchten/Mualem parameterization for the hydraulic properties. Details on the numerical treatment of the inverse procedure are given in *Zurmühl and Durner* [1998] and *Durner et al.* [1999]. In the objective function we included cumulative flow and tensiometric measurements. The parameters θ_s and K_s were fixed at their independently measured values, whereas the parameter m was set to $m = 1-1/n$. The remaining inverse problem was used to optimize the parameters α, m, θ_r, and τ. By optimizing the hydraulic parameters separately for each drainage and imbibition cycle, we could evaluate whether or not changes in the hydraulic parameters occurred. The water content measurements using the TDR probes, which were also available, were not included in the optimization, but used subsequently for validation (next section).

Figure 5 shows the results of the inverse optimization for the outflow branches. The results of the imbibition branches are similar, but bear a slightly greater uncertainty because the initial conditions are not as well defined as for the drainage branches. We see for both the perforated-wall sample (left) and the closed-wall sample (right) that the change in the rate of the outflow process did not affect the optimized water retention characteristics. However, the optimized conductivity curves differ systematically, with the near-saturation conductivity being higher for the faster experiments. In Fig. 5, some independent measurements of the saturated and unsaturated conductivity are also depicted. These measurements were obtained on the same samples at the end of the experimental series by steady-state irrigation from a needle irrigation head until unit-gradient conditions were established. It is obvious that the fast outflow experiments gave results that are closer to the independently measured values. Similar results were obtained for the imbibition branches (not shown). Bearing in mind that the sensitivity of any outflow experiment is very low with respect to the hydraulic conductivity function near saturation, this low sensitivity is apparently slightly increased for the faster drainage experiment, which gave results which are closer to the values obtained with the flux measurements.

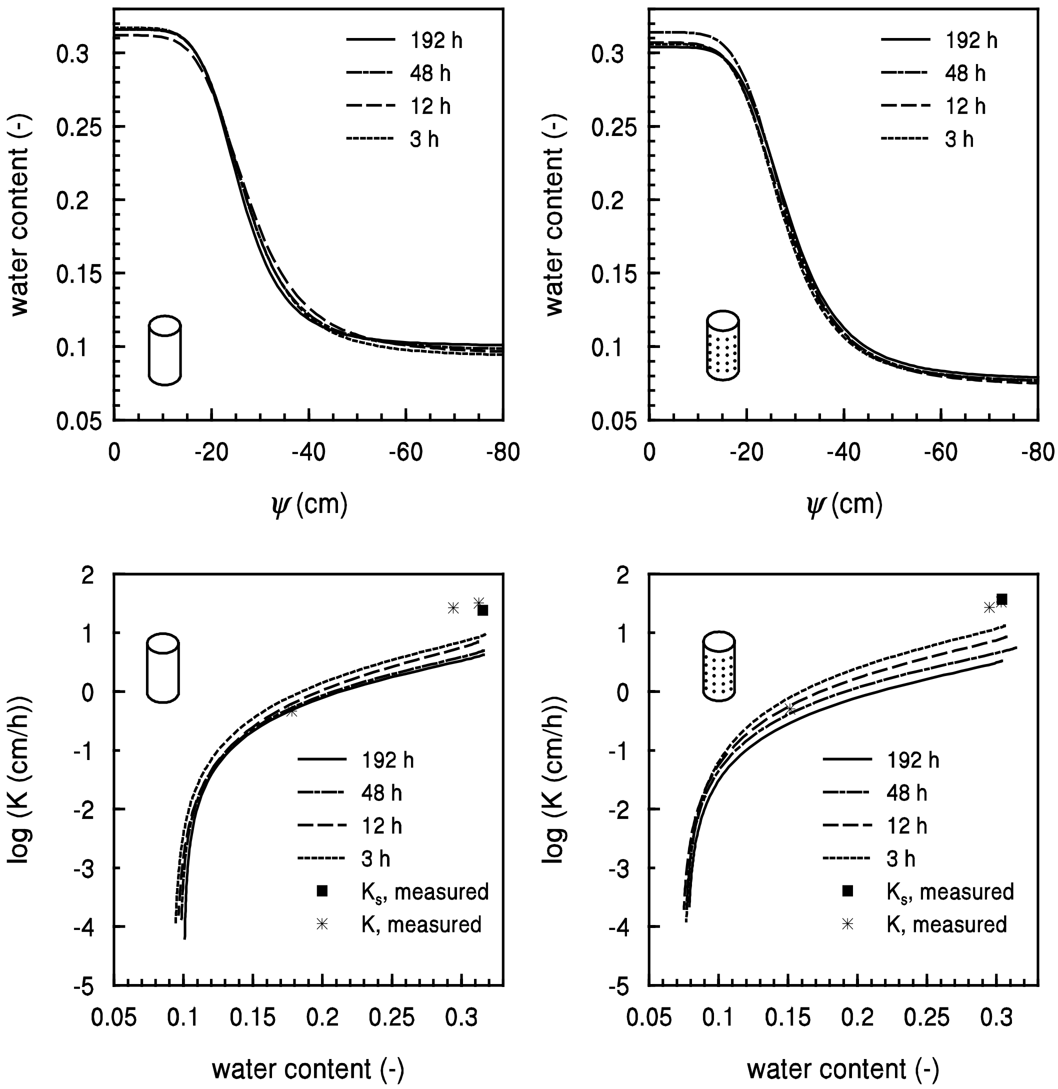

Fig. 5. Hydraulic properties of Forchheim Sand as obtained by inverse simulation of a continuous outflow experiment at four different time scales. Top: optimized retention curves of the closed-wall (left) and open-wall (right) variants. Bottom: associated conductivity curves.

Summarizing the inverse evaluation results: (1) we did not detect any dynamic effects with both the perforated and open-wall soil samples, and (2) inverse modeling of drainage experiments which were performed within three hours, allowed a unique identification of the hydraulic properties of 15 cm long soil columns. We will compare these results now to the directly measured retention curves. This is a validation of the inverse method.

Plotting the tensiometric and water content measurements against each other, without any smoothing of the raw data, leads to in-situ measured retention characteristics for each cycle and for each depth, giving a total of 16 retention curves for each soil sample. Again we restrict ourselves here to results from the drainage process. Furthermore, the results for the slowest drainage cycle of the perforated wall sample are missing due to failure in the data acquisition.

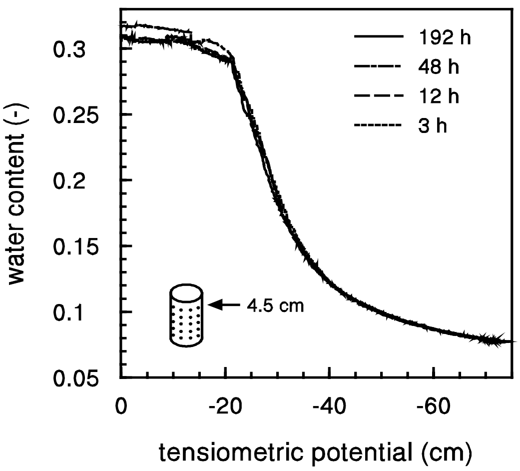

Fig. 6. Influence of drainage speed on in-situ retention curves for Forchheim sand, obtained by directly measuring tensiometric pressures and water contents during a *Continuous* outflow experiment. Soil column with open walls.

Figure 6 shows that despite the very different rates of drainage, the retention curves of the soil columns with perforated walls were almost identical. We were not able to perfectly reproduce the same initial saturation conditions for the soils, but the initial saturation differences should have leveled out during the drainage process. However, the results of the same experiment for the closed samples was different (Fig. 7). Contrary to the perforated-wall columns, the retention characteristics show now a small but significant shift towards higher water contents at a given potential for the fast drainage. The effect appears to be small, but can reach up to 10% water content (~30% of saturation) at a given tensiometric pressure. In Fig. 7, this most sensitive region is indicated by a dotted line.

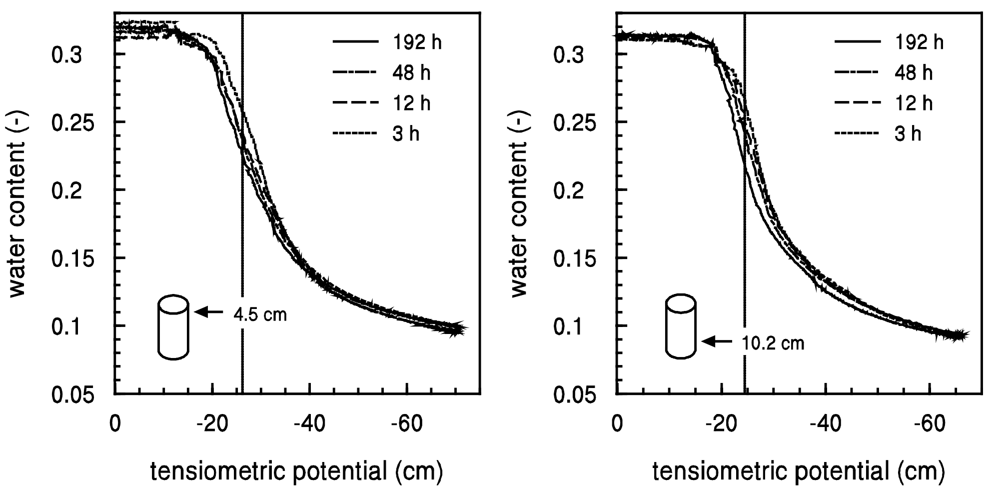

Fig. 7. Influence of drainage rate on in-situ retention curves for Forchheim sand, obtained by directly measuring tensiometric pressures and water contents during continuous outflow experiment. Soil column with closed walls.

MODELING DYNAMIC NONEQUILIBRIUM

Formally, any transient water flow process in porous media must be regarded as a two-phase flow problem since changes in water saturation must be compensated for by the entry or exit of air. For many or most applications, this approach can be simplified by assuming that the resistance to air flow is negligible, which means that no pressure gradients will build up in the gas phase, and hence that the gas phase pressure in the system at any point and time is equal to atmospheric pressure. However, if our aim is to use drainage and imbibition processes for identification of hydraulic properties, we must be aware of these effects, and should seek to isolate them in order to be able to quantify their importance for making standard simulations of water flow. The most decisive important factor causing deviations from atmospheric air pressure in a soil is the loss of gas-phase continuity. Measurements of gas-phase permeability indicate that the air phase does not become continuous during drainage until a significant amount of water has left the pore system, and an "emergence point" saturation is reached. The gas permeability then suddenly jumps to a finite value [*Corey and Brooks*, 1999]. Measurements on repacked sands show that the emergence point occurs at a water saturation of about 50% to 70% [*Fischer*, 1995]. At higher saturation, transfer of gas takes place by effective diffusion. Fischer's results have been confirmed by our own measurements on undisturbed Forchheim sand (Fig. 8).

To test whether the above two-phase flow hypothesis leads to results which can explain our nonequilibrium observations, we used a numerical model, MUSIC, which simulates the one-dimensional coupled transport of gases, water and heat [*Ippisch*, 1997]. Water flow is described by Darcy's law, where the total head, h, contains an additional component for the pneumatic pressure

$$h = \psi_m - z + \frac{p_w - p_a}{\rho_w g} \tag{2}$$

in which ψ_m is the matric head, ρ_w is the density of water, g is the gravitational acceleration, and where the other coefficients are as defined before. Gas flow is simulated by means of a convective-diffusive transport equation involving the species O_2, CO_2, N_2, and H_2O. The gas phase pressure is calculated from temperature and the gas density using

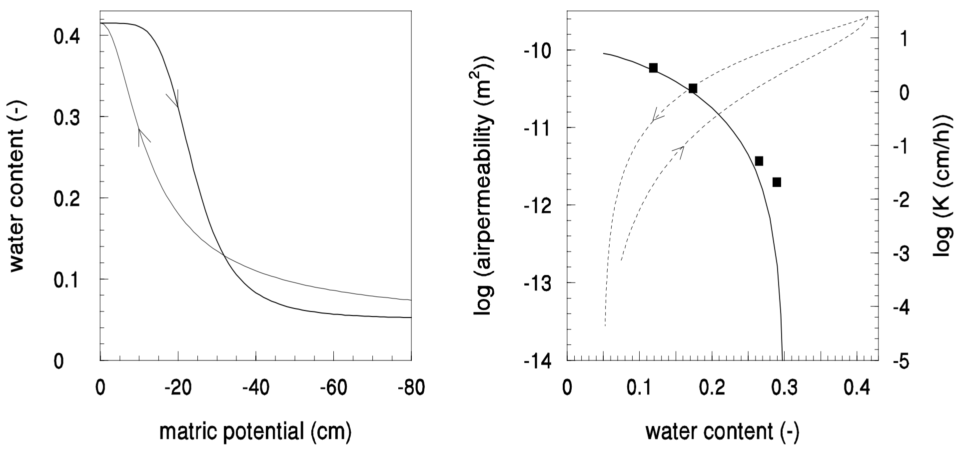

Fig. 8. Constitutive relationships used in the multistep drainage simulation with MUSIC. Left: Retention function. Right: air permeability function (solid), measured values of the air permeability (squares), and the hydraulic conductivity function (dashed).

the ideal gas law. As long as the gas phase is continuous, almost all flow takes place by convection. When the gas phase becomes discontinuous, a pressure difference builds up which causes the gas flow to become proportional to a saturation-dependent effective diffusion term. For diffusive gas transport, a composition-dependent diffusion coefficient is used and multiplied with a tortuosity coefficient. The diffusion coefficient of the gas species is bounded by a minimum value corresponding to the diffusion coefficient in pure water. Transport of dissolved gases in the water itself is not explicitly considered. This should cause some errors, particularly for CO_2 at high pH values. The above system leads to a system of four coupled partial differential equations, which are solved numerically. Details about the model and the numerical solution are given in *Ippisch* [1997].

Using MUSIC, we simulated a multistep drainage experiment. The hydraulic relationships, obtained by inverse simulation of the data in Fig. 1, are shown in Fig. 8 (left). The air permeability for different water saturations was independently determined on a parallel sample. The measured values were fitted with the air permeability function of *Fischer* [1996] (Fig. 8, right). Boundary conditions were identical to those depicted in Fig. 1. As initial condition, we assumed saturation at the bottom of the column and hydrostatic equilibrium within the column. Figure 9 shows a comparison of the simulation results, obtained by the two-phase Richards models. The Richards equation overestimated the initial outflow (Fig. 9, right). Also, a similar smoothing in the shape of the outflow curve was obtained as in our experiments (Fig. 10). A plot of the simulated water pressures versus water contents gave results that were very similar to the observations in Fig. 2 (not further shown here).

Figure 9 (left) compares simulated tensiometric potentials (which is the absolute pressure in the water phase) and matric potentials (pressure difference between water and air phase) at the top and the bottom of the column. The top exhibited a continuous air phase connection from the beginning of the experiment. Accordingly, the air phase pressure remained atmospheric when the water pressure at the bottom of the column was lowered, i.e., the matric and tensiometric potentials were equal. At the bottom, the pressure of the gas phase is initially also at atmospheric pressure. However, after applying suction, the air-phase pressure near the bottom decreased almost in the same way as the water pressure. Accordingly, the matric potential remained close to its initial value. Not until the fourth pressure step did the forces become high enough to desaturate the soil down to the bottom plate. The air phase at that time (t=1.5 d) became continuous throughout the

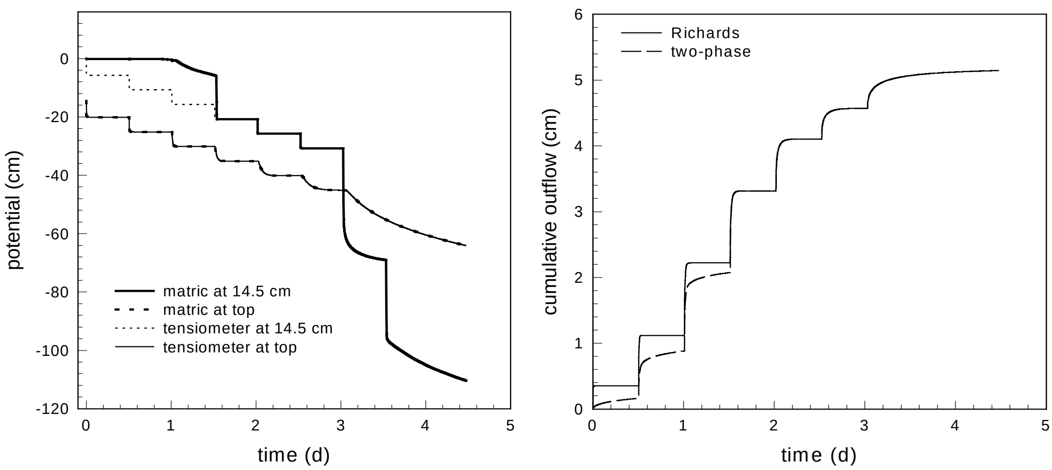

Fig. 9. Simulation of a multistep drainage experiment with the Richards and a two-phase flow model. Left: tensiometric and matric potentials. Right: umulative outflow.

sample, and the air phase pressure increased back to atmospheric levels. This explains in a qualitative way the phenomenon that was illustrated in Figs. 1 and 2. Since the pressure actually being measured by tensiometers does not represent the matric potential, a plot of water content versus tensiometric pressure will yield an incorrect retention curve. This may explain some of the observations discussed in the review section of this paper.

Finally, we illustrate some characteristic differences between the Richards and two-phase flow simulations by discussing several details of the multistep imbibition process. Initially, the soil was in equilibrium with the lower boundary potential at ψ_l = –150 cm. Next, a sequence of 9 pressure steps within 6 days was applied, which did lead to a final pressure ψ_l = +15.7 cm. The constitutive relationships were as in the previous example. First, we focus on the cumulative flow across the lower boundary. Figure 10 shows how the simulation results reflect the behavior that was observed experimentally. Initially, the water content of the soil was quite low, and the air phase is continuous. During that phase the two models produced identical results. After approximately 3.5 days, the fifth pressure step occurs in which the lower boundary pressure drops from pressure ψ_l = 0 cm to ψ_l = +5 cm. From then on, air phase continuity is lost and there is an increasing difference between the two simulations.

Figure 11 (left) shows simulated depth profiles of the total water head and matric head at the end of the fifth pressure step (*t*=4.6 days). At this time the gas phase was continuous down to a depth of approximately 6 cm. Movement of air above this depth took place purely by convection (with a gradient too small to be measurable). Below this depth, the gas phase became discontinuous, and air escape occurred entirely by effective diffusion (Fig. 11, right). Microscopically, the transport of air is a three-dimensional process, and the loss of air phase continuity extends over a transition zone. The air will be gradually organized in air pockets, which are separated by thin water bridges. For the one-dimensional simulation, this is reflected by a high value of the effective diffusion coefficient. Deeper in the column, water saturation is higher and diffusion becomes very slow. At full saturation, the diffusion process is reduced to that of diffusion in pure water. Due to the very low mobility of air at high water contents, the saturation process itself becomes extremely slow towards full saturation. This is illustrated in Fig. 12 which shows the water saturation depth profile at the end of the simulation (*t*= 6.1 d, ψ_l = +15.7 cm).

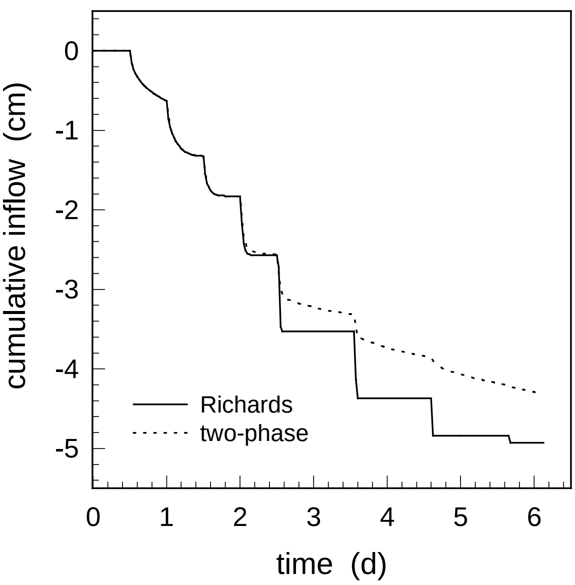

Fig. 10. Simulation of cumulative inflow during multistep imbibition.

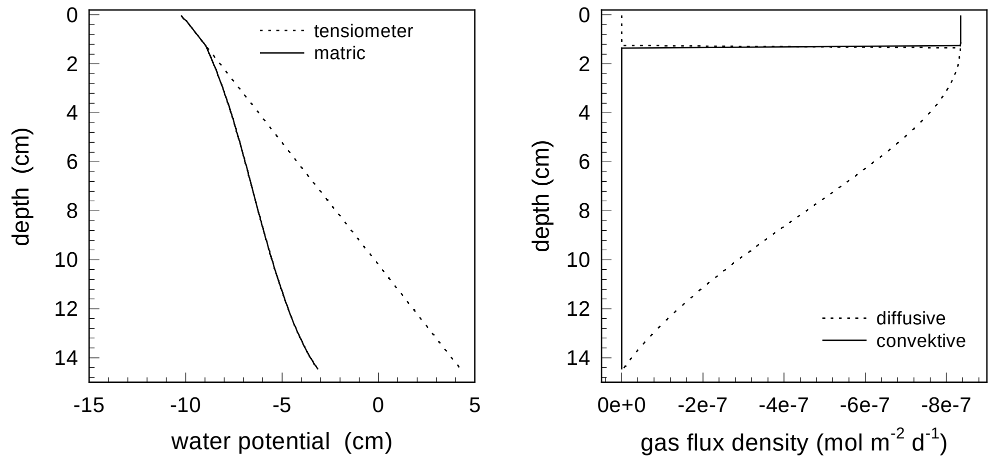

Fig. 11. Left: Simulated depth profiles of matric potential and water phase pressure (=tensiometric potential) at *t*=4.6 days. Right: Gas-flux density at *t*=4.6 days.

If the Richards equation is used as the process model, the soil is now fully saturated. The two-phase flow simulation, however, shows a somewhat surprising result. Due to entrapped air, the saturation is far lower than the pore volume. Saturation is highest at the bottom, and lowest close to the top, as may be expected for a flow process which goes from bottom to top. However, saturation close to the soil surface increases again to 100%. This reflects the fact that right at the atmospheric boundary, it is again easy for air to escape. Once saturation at the top is near 100%, this small saturation zone acts as a impeding layer for further escape of air.

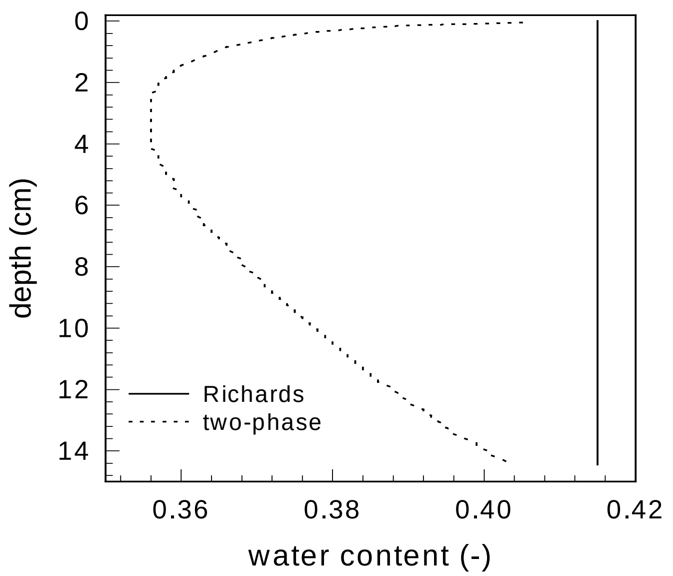

Fig. 12. Simulated depth profile of water saturation after 6.1 days. Dashed line: Richards simulation; Solid line: Two-phase simulation.

CONCLUSION

Experiments on soil columns with open and with closed walls, and simulations with a one-phase and a two-phase flow model indicate that the standard use of the classic Richards equation for water flow becomes invalid under transient conditions if changes from full water saturation to unsaturated conditions (and vice versa) occur. For packed as well as for undisturbed soils, the air phase can loose its continuity already at 50% to 70% water saturation, with air movement then being limited to effective diffusion. Under transient conditions, the gas-phase pressure will deviate from atmospheric pressure. This invalidates the assumption of a unique relation between water phase pressure and water content, which is intrinsic to the Richards equation. Under these conditions, tensiometric measurements can no longer be interpreted as matric potential measurements [*White et al.*, 1970)], and solutions of the Richards equation with time-invariant hydraulic functions (which is presently the standard technique for simulating saturated/unsaturated water flow) cannot be reliably applied to transient flow. It appears remarkable that this hypothesis was already formulated in the early 1960s [*Nielsen et al.*, 1962], but that its consequences had not yet been fully investigated.

We believe that a considerable portion of the dynamic effects is caused by non-negligible resistance to air flow. As a consequence, accurate parameter identification based on transient flow experiments is not possible without explicitly considering this process in the simulations. Furthermore, the phemomenon is interfering with the description of hysteresis in the hydraulic properties; without separating the two processes it is not possible to investigate hysteresis in a reliable manner [*Schultze et al*., 1996]. The two-phase flow hypothesis explains the systematic shift which has been observed in water retention curves measured in the laboratory at different depths of an homogeneous soil sample [*Plagge*, 1991]. Two-phase flow may also be a primary cause for frequently observed discrepancies between static hydraulic properties, as determined in the laboratory and field measurements where dynamic flow conditions prevail. Some researchers have taken great care in measuring air-phase pressures in their experiments [*Smiles et al.*, 1971, *Vachaud et al*., 1972, *Stauffer*, 1977], and found no or negligible pressure changes. Apparently, these findings are contradictory to our hypothesis. However, we believe that there is a severe measurement problem with respect to obtaining the pressure of a non-continuous phase of residual air. An external sensor, brought into contact with some part of the porous medium, would need a quasi perfect measuring characteristic, since the sensor likely sees only a very small pocket of air. Even a minute volume change in the measurement process will lead immediately to a strong pressure drop since the pressure of the air pocket is not buffered by flow from a larger air reservoir. According to *Corey and Brooks* [1999], “there is no known method for directly measuring the pressure of an entrapped nonwetting fluid in the interior of porous media”.

In experiments where changes in boundary conditions occur in a shock-wise manner, e.g., the classical onestep outflow experiment, still other phenomena such as instable air/water fronts and thermodynamic energy changes due to local reorganization of water with respective consequences for the distribution of air–water interfaces, may play a role [*Hassanizadeh and Gray*, 1993]. These effects may be referred to as true dynamic nonequilibrium, as opposed to apparent dynamic nonequilibrium caused by two-phase flow. In order to fully understand water flow in porous media, including hysteresis, two-phase flow, and true dynamic effects, we must be able to quantify and isolate these processes in the analysis. This requires a profound knowledge of the saturation dependence of the air phase permeability for different soils. Thus, future research should aim at detailed measurements of air-phase permeability. Furthermore, the importance of nonequilibrium effects in simulations of field scale water flow under natural conditions is not known presently, and should be investigated in the future.

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