We measured management-induced changes in water retention and hydraulic conductivity for two soil management systems, mixed crop farming and pasture farming. The results were incorporated into an existing model describing the evolution of the pore size distribution as the soil management changes from mixed crop farming to pasture.

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Abbreviations: PSD, pore size distribution.

Compared with pasture or forests, arable cropping can have large impacts on soil properties due to frequent soil disturbance, compaction, greater use of agrochemicals and higher removal of plant biomass (Sparling et al., 2000). Land management practice must be adapted, therefore, to ensure both economic crop production and soil fertility. Mixed pasture—crop—pasture rotations (each of 2–6-yr duration) are common practice in New Zealand because the pasture enables soil structure building and nutrient replenishment. It is generally accepted that such management practices modify the soil structure and pore size distribution (PSD). In particular, the saturated and near-saturated soil hydraulic properties are very sensitive to these management-induced changes (e.g., Ahuja et al., 2006).

The storage and supply of water, nutrients, and O2 in the root zone is controlled by the geometry of the soil pore space and therefore by the soil structure. Hence, soil structure parameters such as bulk density, macroporosity, and saturated and near-saturated hydraulic conductivity are accepted indicators to assess the effect of different soil and crop management practices on a soil’s physical state (e.g., Tebrügge and Düring, 1999; Ball et al., 2007). Many studies have dealt with the change in soil structure caused by different soil and crop management practices (e.g., Bodner et al., 2008; Daraghmeh et al., 2008; Hu et al., 2009, Krümmelbein et al., 2009). Nevertheless, incorporating these measurement results into mathematical equations that quantify soil structure changes is rarely done (Roger-Estrade et al., 2009). A promising approach to capturing the change in soil pore geometry was developed by Or et al. (2000) and Leij et al. (2002a,b). They developed a stochastic model with physically based coefficients to predict changes in the PSD as a function of time and pore size. Leij et al. (2002a) concluded that the lack of soil structural and hydrologic data may limit the application of the PSD model.

Quantifying management-induced changes in soil structure and incorporating these results into mathematical functions describing temporal changes in the PSD may improve our
capacity to assess the overall impacts of different soil and crop management practices on soil physical properties and on the water balance. Furthermore, the ability to measure and predict management-induced changes in soil structure would enable evaluation of how new land use and crop management systems may maintain or enhance soil health. The objective of this study was therefore twofold: (i) characterization of the alteration in soil structure due to a crop–pasture rotation and the corresponding changes in the soil properties that control the fate of water and nutrients and the development of plant root systems; and (ii) incorporation of the measurement results into an existing model that describes the dynamics (i.e., evolution with time) of the PSD resulting from different management practices (Leij et al., 2002a,b).

Materials and Methods

Study Site

Measurements were conducted at Lincoln University’s organic cropping farm, on the Canterbury Plains, South Island, New Zealand (43°38′ S, 172°27′ E, 10 m above sea level). The annual rainfall is 667 mm, supplemented by irrigation due to 870 mm of annual evapotranspiration. The soil is a well-drained Templeton fine sandy loam, formed in 50 to 100 cm of fine-textured alluvium (New Zealand soil classification: Immature Pallic Soil; World Reference Base, Haplic Cambisol). In the Canterbury region, such soils are used predominantly for grazed pasture, with some mixed cropping. The two sites sampled were in adjacent fields on the same soil type (unpublished New Zealand Soil Bureau soil map, scale 1:6670, 1971), with a soil pit excavated at each site to confirm that they had the same soil type. A detailed list of the crops grown at the sites since 1999 is given in Table 1. One site had been under extensively grazed pasture for 2 yr after being cropped for 2 yr. The adjacent site had been cropped for 2 yr after being in grazed pasture for 2 yr. Note that at the cropped site, seedbed preparation was some weeks before the infiltrometer measurements. Tillage involved one pass of the plow followed by power harrows, then the site was bed formed and sowed. The maximum tillage depth was 18 cm. The pasture site was established by no-till direct drilling into the stubble of the previous crop. The pasture crop was predominantly perennial ryegrass (Lolium perenne L.), with red clover (Trifolium pratense L.) and buck’s-horn plantain (Plantago coronopus L.), and managed by sheep and beef cattle grazing (T. Chamberlain, personal communication, 2010). We assumed that the soil structure under the cropped treatment represented an initial stage and the soil structure under the pasture a final stage of a mixed pasture–cropping rotation cycle.

Measurements

In this study, both hood and disk infiltrometers were used to characterize the impact of the crop–pasture rotation management on saturated and near-saturated hydraulic conductivity at 10 locations for the cropped site and 11 locations for the pasture site. The hood infiltrometer enables measurement of the hydraulic conductivity including flow in macropores from saturation down to the bubble point of the soil without the need for a contact disk and contact layer. For measurements at pressure heads beyond the soil’s bubble point, a standard 124-mm-diameter disk (instead of a hood) was connected to the Mariotte water supply. Detailed information on the method was given in Schwärzel and Punzel (2007). Hood infiltrometer (HI) measurements were conducted at pressure supply heads of 0, −2, and −4 cm, followed by disk infiltrometer measurements (over exactly the same area measured by the HI) at pressure supply heads of −8 and −12 cm. In some cases at the cropped site, HI measurements at pressure heads of −4 cm were not possible due to the soil’s low bubble point. In those cases, disk infiltrometer measurements started earlier, at −4 cm. For the disk experiments, we removed all vegetation from the infiltration surface, laid a nylon guard cloth on the soil, and prepared a 10-mm-thick contact layer using dry SphériGlass no. 2227 glass spheres (Potters Ballotini GmbH, Kirchheimbolanden, Germany). The soil was visually dry before starting the infiltration experiments.

Analysis of the hood and disk infiltrometer measurements was based on Wooding’s (1968) solution for infiltration from a circular source with a constant pressure head at the soil surface. If the unsaturated hydraulic conductivity $K(h_0)$ is given by an exponential function (Gardner, 1958)

$$K(h_0) = K_s \exp(\alpha_{GE} h_0) \quad h_0 \leq 0$$

where $K_s$ [L T$^{-1}$] is the saturated hydraulic conductivity and $\alpha_{GE}$ [L$^{-1}$] is the slope of the ln$K$ vs. pressure head ($h$) [L] curve, then the steady-state flow rate $Q$ [L$^3$ T$^{-1}$] is given by

$$Q = \pi b^2 K(h_0) + \frac{4h}{\alpha_{GE}} K(h_0)$$

Table 1. Crop rotation history at Lincoln University’s organic cropping farm at Canterbury, New Zealand. The soil is a well-drained Templeton fine sandy loam, and the soil type according to the World Reference Base for Soil Resources is a Haplic Cambisol.

<table>
<thead>
<tr>
<th>Site</th>
<th>Crop history†</th>
</tr>
</thead>
<tbody>
<tr>
<td>A3</td>
<td>pasture</td>
</tr>
<tr>
<td>A5</td>
<td>cropped</td>
</tr>
</tbody>
</table>

where \( b \) [L] is the radius of the infiltration surface, \( h_0 \) [L] is the applied pressure head, and \( K(h_0) \) is the unsaturated hydraulic conductivity at pressure head \( h_0 \). Equation [2] can be solved for \( K(h_0) \) using multiple pressure heads for a given hood or disk radius (Ankeny et al., 1991) whereby Eq. [1] and [2] are applied piecewise such that \( \alpha_{GE} \) is a constant in the interval between two successively applied pressure heads (Reynolds and Elrick, 1991).

Following all infiltration experiments, we extracted 21 (11 from the pasture site and 10 from the cropped site) undisturbed soil cores (volume of the cores was 420 cm\(^3\), from beneath the positions where the infiltration had been measured. These cores were used to determine water retention curves (desorption) and bulk densities. The dewatering process was performed using ceramic plates connected to a hanging water column down to a pressure of –100 cm. A pressure cell (Soilmoisture Equipment Corp., Santa Barbara, CA) was used below this pressure. The following pressure heads were applied: log10\( |h| = 1.0, 1.5, 2.0, 2.5, 3.0, \) and 4.2, where \( b \) is in centimeters. At log10\( |h| = 4.2 \), repacked small cores (8-cm\(^3\) volume) were measured using soil sampled from beneath the positions where the infiltration had been measured. For each sample, the C and N contents were also measured (Vario EL III, Elementar Analysensysteme GmbH, Hanau, Germany). In addition, disturbed soil samples from the 0- to 7- and 20- to 30-cm depths from each site were collected. The soil samples were sieved through a 2-mm sieve and then used for particle size analysis, and measurements of pH and C and N contents (Vario EL III), and particle density (using a multipycnometer, Quantachrome Instruments, Boynton Beach, FL). To analyze the particle size distribution, the sieve method was applied for soil particles <2000 \( \mu \)m; for particles <63 \( \mu \)m, the pipette method was used. It can be seen from Table 2 that the soil texture at the depth of 0- to 7-cm of both fields on the same soil type, these differences are probably attributable to differences in management history. Although the farm has had a long history of cropping (>40 yr), however, there is limited historical data on the management practices and soil C levels. It is interesting to note that the lower amount of C at the 0- to 7-cm depth under the pasture seems to be contradictory to the generally suggested and often confirmed assumption that conversion of a cropped soil to pasture leads to an increase of soil organic C stocks (e.g., Haynes and Francis, 1990; Aslam et al., 1999; Persson et al., 2008).

### Pore Space Evolution Parameterization

For each location, the calculated hydraulic conductivities, \( K(h_0) \), and measured soil water retention data, \( \theta(h) \), were simultaneously fitted to the lognormal distribution model of Kosugi (1996). The effective degree of water saturation, \( S_e \), is given by

\[
S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} = F_n \left[ \frac{\ln(h/h_m)}{\sigma_{Ko}} \right] \quad [3]
\]

where \( \theta_s \) and \( \theta_r \) [L] denote the saturated and residual water contents, \( h_m \) [L] is the pressure head at \( S_e = 0.5 \), \( \sigma_{Ko} \) is a dimensionless parameter representing the width of the pore size distribution, and \( F_n \) is the normal distribution function, defined as

\[
F_n(x) = \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{x} \exp \left( -\frac{u^2}{2} \right) du \quad [4]
\]

where \( u \) is the variable of integration. The hydraulic conductivity, \( K \), is given by

\[
K = K_0 \left[ F_n \left[ \frac{\ln(h/h_m)}{\sigma_{Ko}} \right] \right] ^{0.5} \left[ F_n \left[ \frac{\ln(h/h_m)}{\sigma_{Ko}} + \sigma_{Ko} \right] \right] ^{2} \quad [5]
\]

### Scaling

To facilitate comparisons between the crop and pasture sites and to answer the question how these management practices affect soil structure, the scaling approach was applied. Scaling enables description of the spatial variation of soil hydraulic properties by a set of scaling factors relating the soil water retention and hydraulic conductivity data at each location to a representative mean. Several methods have been developed to derive scaling factors and the corresponding representative mean (see reviews by Tillotson and Nielsen, 1984; Vereecken et al., 2007). In our study, scaling factors were estimated by minimizing the residual sum of square differences between the data and the scaled reference curves. The scaling factors \( \delta_{e,i} \) (for the water

<table>
<thead>
<tr>
<th>Site</th>
<th>Treatment</th>
<th>Depth</th>
<th>Particle size distribution</th>
<th>Organic C/N</th>
<th>C/N ratio</th>
<th>D_p</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>ID A3</td>
<td>pasture</td>
<td>0–7</td>
<td>Sand: 28.5 Silt: 51.9 Clay: 19.6</td>
<td>2.214</td>
<td>0.199</td>
<td>11.1</td>
<td>2.58</td>
</tr>
<tr>
<td></td>
<td></td>
<td>20–30</td>
<td>Sand: 29.7 Silt: 51.9 Clay: 18.4</td>
<td>0.620</td>
<td>0.076</td>
<td>8.2</td>
<td>2.68</td>
</tr>
<tr>
<td>A5</td>
<td>cropped</td>
<td>0–7</td>
<td>Sand: 26.2 Silt: 51.9 Clay: 21.9</td>
<td>2.943</td>
<td>0.238</td>
<td>12.4</td>
<td>2.58</td>
</tr>
<tr>
<td></td>
<td></td>
<td>20–30</td>
<td>Sand: 34.2 Silt: 46.6 Clay: 19.3</td>
<td>0.583</td>
<td>0.069</td>
<td>8.4</td>
<td>2.68</td>
</tr>
</tbody>
</table>

† Sand = 2.0– 0.063 mm, silt = 0.063– 0.002 mm, and clay = < 0.002 mm. 
† Particle density.
The reference curves are given by

\[ h_i (S_c) = \delta_{b_i} h (S_c) \quad \text{(6)} \]

\[ K_i (b) = \delta_{K_i} K^* (b) \quad \text{(7)} \]

The reference curves are given by

\[ S^*_c = F_n \left[ \frac{\ln \left( \frac{h_i}{h_m^*} \right)}{\sigma_{Ko}^*} \right] \]

\[ K^* = K_i^* \left[ F_n \left[ \frac{\ln \left( \frac{h_i}{h_m^*} \right)}{\sigma_{Ko}^*} \right] \right]^{0.5} \]

\[ \times \left[ F_n \left[ \frac{\ln \left( \frac{h_i}{h_m^*} \right)}{\sigma_{Ko}^*} + \sigma_{Ko}^* \right] \right]^2 \quad \text{(9)} \]

A nonlinear least squares optimization procedure provided in Microsoft Excel was applied to minimize the residual sum of squared differences (SS\text{total}) between the data and the scaled reference curves for \( i \) locations, defined as

\[ SS_{\text{total}} = SS_b + SS_K \]

\[ = W_1 \left[ \sum_{i,p} \left( S^*_{c,i} \delta_{b_i} h (p) - S_{c,i}^* \right)^2 \right] + W_2 \left[ \sum_{i,p} \left[ \log K^* \delta_{K_i} h (p) \right] - \log \left( K^* \delta_{K_i} h (p) \right) \right]^2 \quad \text{(10)} \]

where \( p \) represents discrete measurements of the water retention and hydraulic conductivity at each location. The relative saturation, \( S^*_{c,i} \delta_{b_i} h (p) \), and the hydraulic conductivity, \( K^* \delta_{K_i} h (p) \), were obtained by substituting Eq. [6] and [7] into the reference soil water (Eq. [8]) and conductivity (Eq. [9]) functions. The values of the parameters of the reference curves \( (h_m^*, \sigma_{Ko}^*, \sigma_{Ko}^*, \delta_{b_i}^*) \) and the scaling factors \( \delta_{b_i} \) and \( \delta_{K_i} \) were then optimized to minimize SS\text{total}: To allow for equal weighting between retention and conductivity data, the weighting factors \( W_1 \) and \( W_2 \) were defined as proposed by Tuli et al. (2001). They defined \( W_1 \) and \( W_2 \) by the inverse of the standard deviation of the observed water retention and hydraulic conductivities, computed from all data within each field. In addition, we fixed the geometric mean of the scaling factors at unity, as a normalization condition (Kosugi and Hopmans, 1998).

**Pedotransfer Functions**

The neural network model ROSETTA (Schaap et al., 2001) and the approach of Wöstjen et al. (1999) were used to predict the parameters of the van Genuchten equation for describing the water retention curve using the percentage of sand, silt, and clay, the amount of organic matter, and the bulk density as predictors. Furthermore, the approach of Zacharias and Wessolek (2007), which estimates the parameter of the van Genuchten equation using only the amount of sand, silt, and clay and the bulk density, was applied.

**Pore Size Quantification**

For each field (the cropping and pasture sites), a set of scaling factors and their respective reference curves were computed as outlined above. From the parameterized curves, \( K(h_0) \) values were calculated for the pressure heads, \( h_0 \), at which the infiltrometer measurements were made. Once the \( K(h_0) \) values were calculated, the representative mean radius \( r_{\Delta h_0} \) for two consecutive pressure heads, \( \Delta h_0 \), was determined as (Moret and Arrúe, 2007a)

\[ r_{\Delta h_0} = \frac{\sigma_{\text{surface}} \Delta K}{\rho g M_0} = 7.50 \frac{\Delta K}{\Delta M_0} \quad \text{(11)} \]

where \( \sigma_{\text{surface}} = 73.49 \text{ g s}^{-2} \) is the air–pore water interfacial surface tension, \( \rho = 9.991 \times 10^{-4} \text{ g mm}^{-3} \) is the density of water, \( g = 9810 \text{ mm s}^{-2} \) is the acceleration due to gravity, \( \Delta K \text{ (mm s}^{-1}) \) is the difference in hydraulic conductivity between two applied pressure heads, and \( \Delta M_0 \text{ (mm s}^{-1}) \) is the difference in matrix flux potential between two applied pressure heads (see Fig. 1). The matrix flux potential is the area beneath the hydraulic conductivity function \( K(h_0) \) within chosen limits (Gardner, 1958):

\[ \Delta M_0 = \int_{h_0}^{h_0+\Delta h_0} K(h) \text{d}h \]

\[ \Delta K_0 = \int_{h_0}^{h_0+\Delta h_0} K_0 \text{d}h \]

Fig. 1. Relationships between the matrix flux potential (\( M_0 \)), the tension flux potential (\( M_{\text{t}} \)), and the residual flux potential (\( M_{\text{r}} \)) (modified from Reynolds et al., 1995). Included are hypothetical \( K(h_0) \) vs. \( h_0 \) data points and a spline fitted through them, where \( K(h_0) \) values are the hydraulic conductivities measured by tension infiltrometer (TI) at pressure heads \( h_0 = 0, -2, -4, -8, \) and \(-12 \text{ cm} \); \( h_0 \) is the minimum \( h_0 \) at which TI measurements were conducted; and \( h_b \) is the background pressure head in the soil at the time of the TI measurement. In these examples, the following values were found. At \( h_0 = -4 \text{ cm}: K(h_0) = 0.010 \text{ mm s}^{-1}, M_0 = 0.473 \text{ mm}^2 \text{s}^{-1}, \) and \( M_{\text{r}} = 0.124 \text{ mm}^2 \text{s}^{-1} \). At \( h_0 = -8 \text{ cm} \): \( K(h_0) = 0.006 \text{ mm s}^{-1}, M_0 = 0.172 \text{ mm}^2 \text{s}^{-1}, \) and \( M_{\text{r}} = 0.124 \text{ mm}^2 \text{s}^{-1} \). Using Eq. [11], the representative mean radius for these two consecutive pressure heads is: \( r_{\Delta h_0} = 7.50 \frac{\Delta K}{\Delta M_0} = 0.100 \text{ mm} \).
\[ M_0 = \int_{h_0}^{h} K(h) \, db \]  \hspace{1cm} \text{[12]}

The representative mean radius, \( r_{Δh_0} \), defines a sort of pore index of “water conductivity” that relates to the flow impedance for a pressure head range corresponding to a specific “pore size” (Moret and Arrúe, 2007a). Calculation of \( r_{Δh_0} \) requires estimates of \( M_0 \). These were obtained as described by Reynolds et al. (1995), who split the \( M_0 \) integral into “tension” and “residual” flux potential components, \( M_n \) and \( M_r \), respectively:

\[ M_0 = M_n + M_r \]  \hspace{1cm} \text{[13]}

with

\[ M_n = \int_{b_n}^{b_0} K(b) \, db \quad b_n \leq b \leq b_0 \]  \hspace{1cm} \text{[14]}

and

\[ M_r = \int_{b_1}^{b_n} K(b) \, db \quad b_1 \leq b \leq b_n \]  \hspace{1cm} \text{[15]}

where \( b_0 \) and \( b_n \) are the maximum and minimum pressure heads at which the infiltrometer measurements were made and \( b_1 \) is the background pressure head in the soil at the time of measurement. Following the proposal of Reynolds et al. (1995), the values of \( M_n \) were then determined by numerically integrating under the splines that were fitted through the \( K(b_0) \) vs. \( b_0 \) data. The values of \( M_r \) were determined as

\[ M_r = \frac{Q_n}{\pi b^2} \frac{K_n}{4b} \]  \hspace{1cm} \text{[16]}

where \( Q_n = Q_0 \) and \( K_n = K_0 \) at \( b_n = b_0 \) (Reynolds et al., 1995). Using the original data, \( M_r \) was calculated for each location and then averaged for each site (see Fig. 1).

After calculation of the representative mean radius, the number of pores for two consecutive pressure heads, \( N_0 \) [L\(^{-2}\)], was estimated using Poiseuille’s law for flow in a capillary tube (Reynolds et al., 1995):

\[ N_0 = \frac{8\mu \Delta K}{\rho g \pi \left( r_{Δh_0} \right)^4} \]  \hspace{1cm} \text{[17]}

where \( \mu = 1.404 \times 10^{-3} \text{ g mm}^{-1} \text{ s}^{-1} \) is the dynamic viscosity of water, and \( \Delta K \) is the difference in hydraulic conductivity between two applied pressure heads (\text{mm s}^{-1}). Finally, the effective water-conducting porosity, \( \Phi_e \) (%), for two successively applied pressure heads was determined by (Watson and Luxmoore, 1986)

\[ \Phi_e = N_0 \pi r_{Δh_0}^2 \times 10^{-3} \]  \hspace{1cm} \text{[18]}

The macro- and mesoporosity were derived from the measurements of the infiltration as well as soil water retention. In this study, the amount of macropores corresponds to pores that drain at pressure heads \(|b| < 4 \text{ cm} \) and the amount of mesopores to pores draining at \(|b| \) between 4 and 12 cm. Furthermore, the amount of plant-available water and the air capacity were derived using the measured soil water retention curves (see Fig. 2).

**Change of Pore Size Distribution**

The reference water retention data for each different crop and soil management practice was used to adapt the pore space evolution model of Leij et al. (2002a). In doing so, we described the evolution of the PSD from a mixed cropping farm soil (cropped treatment) to a pasture farm soil (pasture treatment). In the approach of Leij et al. (2002a), the PSD is described with the Fokker–Planck equation (Or et al., 2000):

\[ \frac{\partial f}{\partial t} = \frac{\partial}{\partial r} \left[ D(r, t) \frac{\partial f}{\partial r} \right] - \frac{\partial}{\partial r} \left[ V(r, t) f \right] - Z(t) f \]  \hspace{1cm} \text{[19]}

where \( f \) is the PSD or “frequency” [L\(^{-1}\)] as a function of time \( t \) [T] and pore radius \( r \) [L], \( D \) is a dispersion coefficient [L\(^2\) T\(^{-1}\)], \( V \) is the drift coefficient [L T\(^{-1}\)], and \( Z \) is a degradation coefficient [T\(^{-1}\)]. The drift coefficient describes the rate at which the mean change in pore radius evolves with time for a given pore size class (Or et al., 2000). The dispersion coefficient represents the rate at which the variance of changes in pore radii evolves with time (Or et al., 2000). The degradation coefficient \( Z \) is a first-order pore degradation factor representing instantaneous pore loss. Solving of Eq. [19] requires a set of initial and boundary conditions:

\[ f(r, 0) = f_0(r), \quad 0 < r < \infty \]  \hspace{1cm} \text{[20]}

![Fig. 2. Calculation of the plant-available water capacity (AWC), air capacity (AC), and amount of macropores (pores that drain at \(|b| < 4 \text{ cm} \) or \(\log|h| < 0.6\) and mesopores (pores draining at \(|b| = 4–12 \text{ cm} \) or \(\log|h| = 0.60–1.08\)) using the water retention curve; FC = field capacity, corresponding to the water content at \(\log|h| = 2.0\); PWP = permanent wilting point, corresponding to the water content at \(\log|h| = 4.2\).](Image)
\[ Vf - D \frac{\partial f}{\partial r} = 0, \quad 0 = r, \ t > 0 \]  \hspace{1cm} (21)

\[ \frac{\partial f}{\partial r} = 0, \quad r \to \infty, \ t > 0 \]  \hspace{1cm} (22)

The initial distribution \( f_0 \) of the PSD is described by (Kosugi, 1996)

\[ f_0(r) = \frac{\phi_0}{r \sigma_{Ko} \sqrt{2\pi}} \exp\left[ - \frac{\ln(r/r_m)^2}{2\sigma_{Ko}^2} \right] \]  \hspace{1cm} (23)

with

\[ \int_0^\infty f_0(r) dr = \phi_0, \quad 0 < r < \infty \]

where \( r_m \) is the initial median pore radius or geometric mean \([L]\), \( \sigma_{Ko} \) is the standard deviation of the log-transformed pore radius, and \( \phi_0 \) is the total (initial) porosity that determines the maximum of the cumulative distribution as defined by Eq. (23). Leij et al. (2002a) assumed that \( \phi_0 \) is equal to the difference between the saturated and residual water contents, \( \theta_s - \theta_r \). Furthermore, they related the dispersion \( (D) \) and drift \( (V) \) coefficients in the same way as for solute transport:

\[ \lambda = \frac{D(t)}{V(t)} \]  \hspace{1cm} (24)

where \( \lambda \) is the dispersivity \([L]\). The drift coefficient will normally be negative because the pore radius is prone to decrease after soil tillage. Leij et al. (2002a) obtained the following solution for the PSD:

\[ f(r, t) = \exp\left[ \int_0^T Z(t, r) d\tau \int_0^\infty f_0(\zeta) \right] \times \frac{1}{\sqrt{4\pi T \lambda}} \exp\left[ -\frac{(r - \zeta - T)^2}{4T \lambda} \right] \]  \hspace{1cm} (25)

\[ + \exp\left[ -\frac{r}{\lambda} \left( r + \zeta - T \right)^2 \right] + \frac{1}{2\lambda} \exp\left[ -\frac{r}{\lambda} \right] \text{erfc}\left[ \frac{r + \zeta - T}{\sqrt{4T \lambda}} \right] \int_0^\infty d\zeta \]

where \( \tau \) and \( \zeta \) are dummy integration variables, while the cumulative drift term, \( T[L] \), is defined by

\[ T(t) = -\int_0^t V(\tau) d\tau \]  \hspace{1cm} (26)

Leij et al. (2002a) described the drift term as follows:

\[ V(t) = \frac{d}{dt} \langle r \rangle = a \left( 1 - \frac{\langle r \rangle}{b} \right) \langle r \rangle \]  \hspace{1cm} (27)

with

\[ \langle r \rangle = \frac{b \langle r_0 \rangle}{\langle r_0 \rangle + (b - \langle r_0 \rangle) \exp(-at)} \]

where \( \langle r_0 \rangle \) is the mean pore size at the initial stage, \( \langle r \rangle \) is the mean pore size at the final stage, and \( a[T^{-1}] \) and \( b[L] \) are empirical coefficients. Note that the difference between \( \langle r \rangle \) and \( \langle r_0 \rangle \) corresponds to \( T \).

Leij et al. (2002a) determined the initial and final mean pore size from the moments of the actual PSD (Eq. [24]) using the definitions for mean according to Aitchison and Brown (1963):

\[ M_1 = \langle r \rangle = \phi_m \exp\left( \frac{\sigma_{Ko}^2}{2} \right) \]

The degradation coefficient \( Z(t, r) [T^{-1}] \) can be defined as (Or et al., 2000)

\[ Z(t, r) = \lim_{\Delta t \to 0} \frac{1}{\Delta t n} \sum_{i=1}^{n} \frac{m_i}{m} \]

where \( n \) is the total number of pores of radius \( r \) and \( m \) corresponds to the pores of radius \( r \) that are closed during the time interval \( \Delta t \).

\section*{Results and Discussion}

\subsection*{Soil Hydraulic Properties}

A compilation of soil physical data is given in Table 3. Even though the soil texture of both fields is similar (see Table 2), significant differences in soil properties between the topsoils of both treatments were detected.

The lower bulk density and higher porosity of the topsoil under the cropped treatment reflects the loosening effect of tillage a few weeks before the measurements were conducted. Similar results were reported by Schwartz et al. (2003). They compared the hydraulic properties of native grassland, re-established grassland, and recently tilled cropped soils. Lower bulk densities associated with higher porosities after tillage operations have also been observed by other researchers (e.g., Francis and Knight, 1993; Tebrügge and Düring, 1999; Moret and Arrúe, 2007b). Changes in total porosity are generally related to alterations in the soil air and water storage (e.g., Schjønning and Rasmussen, 2000; Kay and VandenBygaart, 2002; Lipiec et al., 2006). These findings are supported by our study. The decrease of porosity under the pasture treatment was accompanied by a decrease in air capacity and an increase in plant-available water capacity in comparison to the cropped soil (Table 3). This increase of plant-available water capacity under pasture was due to a higher field capacity and lower permanent wilting point (Table 3). The lower amount of hygroscopic water (water stored at the permanent wilting point and unavailable to plants) of the soil under pasture can be attributed to the slightly smaller amounts of clay and soil organic C, the surfaces of which can adsorb more water than other soil particles.
(see Tables 2 and 4). Our explanation is supported by Rawls et al. (2003), who investigated the influence on soil water retention of the proportion of textural components and the amount of C. On the other hand, management-induced changes in bulk density and soil structure may have partially compensated for the impact of clay and soil organic C on the water content at field capacity. This was already suggested by Reynolds et al. (2002), who investigated the impact of different crop and soil management regimes on indicators of good soil physical quality. In that study, the indicators field capacity, permanent wilting point, and plant-available water capacity showed no consistent differences among different management practices. Similar results were reported by Schjønning and Rasmussen (2000).

The unscaled and scaled water retention and hydraulic conductivity data are presented in Fig. 3. Additionally, the original measured data are listed in Table 4.

It can be seen that the measured conductivities at and near saturation under cropping were always larger than under pasture. Furthermore, the variability of the data is larger under cropping than under pasture. For example, the coefficient of variation of...
The parameters

The scaled water retention and conductivity values are well

Table 5 shows the parameter values obtained from the fit of the

pressure head exceeded 31 cm (see Table 4). Moreover, due to the

larger variability of conductivity at and near saturation indicates

a greater small-scale variability of macropores (corresponding to

pore radii that will be emptied when \( |h_0| < 4 \) cm) compared with

the variability of mesopores.

The separate scaling of water retention and hydraulic conductivity

produced two sets of scaling factors for each treatment (see Table 5). In accordance with other studies (e.g., Clausnitzer et al., 1992; Kosugi and Hopmans, 1998), a lognormal distribution, rather than a normal distribution, describes the distribution of the scale factors (not shown). Scaling of similar media assumes that the factors obtained from scaling water retention and hydraulic conductivity should be identical or at least highly correlated. Nevertheless, in this study, the scaling factors were only weakly correlated (−0.41 for the pasture soil and −0.29 for the cropped soil). The weak correlation between the scaling factors might be attributed to differences in the measurement range between the water retention and hydraulic conductivity data. The water retention data were scaled across six pressure heads ranging from 10 to 1000 cm; the conductivity data were scaled across five pressure heads ranging from 0 to 12 cm. Therefore, the scaling factors obtained for the water retention consider its variability from near-saturated to dry. The high correlation between the logarithmic scaling factors of water retention and the amount of plant-available water supports this conclusion (correlation factor: −0.78 for the pasture soil and −0.84 for the cropped soil). In contrast to the water retention, the scaling factors for hydraulic conductivity describe the variability of the pasture soil. Differences in water retention as a result of different land management practices were also found by Schwartz et al. (2003). In contrast to our study, the influence of tillage on water retention was mainly confined within the region from saturation up to 100 cm pressure head. Ahuja et al. (1998) stated, on the basis of a review of the literature, that tillage generally increases the porosity as well as the water retention in the pressure head range from about 60 to 330 cm.

Despite the differences in \( \sigma_{K_o} \) between treatments, the shape of the hydraulic conductivity functions is quite similar but offset along the \( y \) axis. Our differences in saturated and near-saturated hydraulic conductivities between treatments were mainly due to tillage at the cropped site a few weeks before our measurements. Tillage is known to create a loose, fragmented, macropore-rich soil structure. Our observed high infiltration rates under cropping could indicate that water flow took place between the aggregates rather than through the soil matrix. This suggestion is supported by Lipiec et al. (2006), who studied how different tillage methods (ranging from conventional tillage to no-till) affect porosity and water infiltration. They derived aerial porosity from resin-impregnated blocks and stained porosity (flow-active porosity) from horizontal sections of column samples taken after infiltration of a methylene blue solution. Lipiec et al. (2006) found that the soil under conventional tillage had the greatest areal porosity and stained porosity within the plow layer. This resulted in the highest infiltration throughout 3 h of water application compared with other tillage methods. They concluded that the soil pore system under conventional tillage with a higher contribution of large, flow-active pores enhances infiltration and water storage capacity.

The measured hydraulic conductivity ranged between 29 (at saturation) and 55\% (at \( |h_0| = 12 \) cm) under pasture and between 41 (at saturation) and 54\% (at \( |h_0| = 12 \) cm) under cropping. The larger variability of conductivity at and near saturation indicates a greater small-scale variability of macropores (corresponding to pore radii that will be emptied when \( |h_0| < 4 \) cm) compared with the variability of mesopores.

Table 5 shows the parameter values obtained from the fit of the soil hydraulic functions (Eq. [3] and [5]) to the measured data. The parameters \( \theta_s, \theta_r \) and \( K_s \) tended to be higher for the cropped treatment than for the pasture. By contrast, the fitted values of \( h_{m_a} \) were substantially smaller for the cropped treatment than for the pasture. Significant differences between the treatments were also detected in the fitted values of \( \sigma_{K_o} \); the range of \( \sigma_{K_o} \) values was significantly larger for the cropped treatment than for the pasture.

The scaled water retention and conductivity values are well coalesced to a narrow band around the reference curves (Fig. 3). It can be seen that the reference water retention functions of the treatments differ significantly (see also Table 6). The cropped topsoil drained a considerable amount at pressure heads <31 cm, in contrast to the topsoil, which released little water until the pressure head exceeded 31 cm (see Table 4). Moreover, due to the smaller value of \( \sigma_{K_o} \), the water retention curve of the cropped soil is steeper at the inflection point than the corresponding curve of

Fig. 3. Unscaled (above) and scaled (below) hydraulic conductivity data (left side) as well as soil water retention data (right side) of the Templeton sandy loam under 2 yr of pasture (red circles) or tillage and cropping (blue triangles).
of conductivity near saturation. This is supported by the high correlation between the near-saturated hydraulic conductivities ($|h_0| \geq 2$ cm) and the logarithmic scaling factors (correlation factors vary between $-0.90$ and $-0.97$).

**Pedotransfer Functions**
Different management practices result in different hydraulic functions even though the basic soil physical data (texture, bulk density, C content) are quite similar. Pedotransfer functions (PTFs) are often used to estimate soil hydraulic parameters from readily available soil data. Commonly available PTFs are generally not suitable, however, to predict the change in soil hydraulic functions caused by different crop and soil management practices, at least for the soils of this study. This can be seen in Fig. 4, where the estimated water retention curves for the soils under pasture and cropping are contrasted with the measured data. Due to the same soil texture but different bulk densities and C contents, the estimated curves of each PTF approach differ substantially between treatments only.
in the wet pressure head range. For drier soil, the course or shape of the estimated curves becomes very similar but slightly shifted along the y axis.

It was not the aim of this study to evaluate the accuracy of the PTFs, but the comparison between measured and predicted water retention data indicates that the soil management-induced change in hydraulic properties cannot be predicted by static PTFs. More process-oriented models are needed to describe the dynamics of the pore space as a result of soil management.

**Pore Size Quantification**

We suggested above that the differences in water retention and hydraulic conductivity between the pasture and cropped soils mainly reflect the creation of a loose and macropore-rich soil structure due to tillage a couple of weeks before our measurements. Equation [17] clarifies that the hydraulic conductivity depends not only on the pore size of water-conducting pores but also on the corresponding number of pores. A comparison of this relation for the pasture and cropped sites in this study is presented in Fig. 5. The estimated representative mean pore radii varied between 0.08 and 0.26 mm, as found by other researchers (e.g., Reynolds et al., 1995; Angulo-Jaramillo et al., 1997; Moret and Arrúe, 2007a). The largest representative mean pore radius was detected under the cropped treatment. Furthermore, the macropore abundance of the cropped soil was significantly larger than that of the pasture soil, which is attributed to seedbed preparation a couple of weeks before the infiltrometer measurements. As a result, the saturated and near-saturated conductivities of the cropped topsoil were up to four times larger than those of the pasture soil. This is in accordance with the above-mentioned study of Lipiec et al. (2006) and the results reported by Moret and Arrúe (2007a). The latter observed significantly higher values of the representative mean pore radius for macropores under no-till but significantly lower near-saturated hydraulic conductivity under no-till than under conventional tillage due to the lower abundance of water-transmitting macro- and mesopores.

Our results are also in agreement with those reported by Francis and Kemp (1990), who used micromorphological techniques to investigate the influence of short-term changes in land management on soil structure. They found that after only 2 yr under pasture, the soil was significantly denser than the cropped soil, although a considerable number of vertically oriented biopores had formed. In contrast, the reference cropped soil had voids that were often horizontally oriented. Other studies showed that the earthworm population under no-till and grassland was greater than under tillage (e.g., Francis and Knight, 1993; Aslam et al., 1999). This is also true for the pasture soil in the present study. Visual assessment of the earthworm population during the infiltrometer measurements and soil sampling indicated a high abundance of earthworms under the pasture, but no earthworms were detected in the cropped soil. A higher amount of biopores is often related to higher saturated and near-saturated hydraulic conductivities (e.g., Bodhinayake et al., 2004; Buczko et al., 2006). As outlined above, however, the higher conductivities under cropping compared with pasture can be attributed to loosening of the cropped soil before our measurements.

Using the results of the infiltrometer measurements, the effective water-conductive porosities were calculated according to Eq. [18]. The water-conducting macroporosity was 0.047% (v/v) for the cropped soil and 0.012% (v/v) for the pasture soil. For the water-conducting mesoporosity, values of 0.075% (v/v) (cropped)
and 0.016% (v/v) (pasture) were estimated. These values of the hydraulically active porosities are of the same order of magnitude as reported by other researchers (e.g., Bodhinayake et al., 2004; Buczko et al., 2006; Moret and Arrúe, 2007a). In contrast to the hydraulically active porosity, the total porosity relates to the water storage capacity (Reynolds et al., 1995). Based on the fitted water retention curves, the total macro- and mesoporosities were calculated as the difference between the water content at 0 and 4 cm of pressure head and at 4 and 12 cm pressure head. For the cropped soil, the total macroporosity was 0.43% (v/v) and the total mesoporosity 1.16% (v/v). The corresponding values for the pasture soil were 0.21 and 0.50% (v/v). As in previous studies (e.g., Bodhinayake et al., 2004; Buczko et al., 2006; Moret and Arrúe, 2007a), the water-conducting macro- and mesoporosities were about 10 times smaller than the corresponding total macro- and mesoporosities. This could indicate that the connectivity of large voids is generally low. Dye staining tests of other researchers (e.g., Weiler and Flühler, 2004; Lipiec et al., 2006; Cey and Rudolph, 2009) may support this suggestion.

**Change in Pore Size Distribution**

Based on the reference water retention curves, probability density functions (Eq. [23]) of the PSD for both treatments were determined (Fig. 6). It was shown above that after 2 yr of pasture, the total porosity decreased and the bulk density increased compared with the cropped site. It can be seen from Fig. 6 that the topsoil under pasture had a more balanced PSD than the cropped topsoil. The loss of porosity under pasture was accompanied by a reduction in the mean pore size and an increase in the variance of the PSD. Compared with the cropped soil, we observed a shift in the PSD toward smaller pores, highlighting that the abundance of large pores created by tillage is temporally unstable when the soil is in a pasture phase. Such a collapse of large pores may be due to soil settlement, partial slaking on wetting, shrink–swell cycles, freeze–thaw cycles, cattle treading, and biological activity (e.g., Schwartz et al., 2003; Bodner et al., 2008; Daraghmeh et al., 2008; Drewry et al., 2008; Hu et al., 2009). From a review of New Zealand research, Hewitt and Shepherd (1997) predicted that our soil type should have a high vulnerability to structural collapse due to the low content of clay, organic matter, and short-range-order oxyhydroxides of Al and Fe. They also identified our soil as having a high tendency for slaking and dispersion.

As discussed above, tillage created a loose, fragmented, macropore-rich soil structure; however, our measurements indicate also that this created soil structure is unstable. We observed almost 80% fewer large macropores (pores that drain at |h₀| < 2 cm) in the pasture soil compared with the cropped soil (see Fig. 5). This can be partly interpreted as a loss of aggregate stability due to cropping, resulting in the collapse of interaggregate pores. Previous studies have shown that such a collapse of large pores takes place in the early stages after tillage. Moret and Arrúe (2007b) conducted infiltrometer measurements after primary tillage but before any post-tillage rainfall events had occurred and compared their results with those conducted after primary tillage but following a period of intermittent rainfall events. They observed a significant decline in saturated and near-saturated hydraulic conductivity 1 to 2 mo after primary tillage because of soil reconsolidation by post-tillage rain and associated wetting and drying cycles. Hu et al. (2009) performed infiltrometer tests under soybean [Glycine max (L.) Merr.] at four times from May to August. The field was moldboard plowed just before being planted in early May. Their results showed that the saturated and near-saturated hydraulic conductivities generally decreased by a factor of 1.3 to 2.2 from spring (May) to summer (August). Interestingly, with time the contribution of macropores (>0.5 mm in diameter) and mesopores (0.5–0.1 mm) to the total water flow decreased while the contribution of micropores (<0.1 mm) increased. Hu et al. (2009) suggested that this phenomenon can be attributed to the loss of large pores as a result of changes in the arrangement of soil particles and pore structure. To characterize the instantaneous mortality of pores per unit time, an empirical model similar to the model proposed by Hara (1984) was used:

\[ Z(t) = d \exp(ct), \quad c < 0 \]  

where \( Z(t) \) is the mortality of large macropores (pores that drain at |h₀| < 2 cm) at time \( t \) per unit time due to the collapse of the pores and \( c \) and \( d \) are empirical coefficients. We assumed that a high number of large pores will be closed in the early stages after tillage (cf. Or et al., 2000; Moret and Arrúe, 2007b; Hu et al., 2009). Furthermore, the gradual closure of the smallest interaggregate pores will be neglected. Figure 7a presents the assumed degradation dynamics of large macropores after tillage calculated.
The combined use of hood and disk infiltrometer experiments in conjunction with the laboratory measurement of soil water retention enabled the quantification of management-induced changes in soil structure. Our study verified that the conversion of a cropped soil to pasture within a rotation improves the soil structure. In comparison with the cropped soil, the topsoil under pasture had a more balanced PSD and a higher capacity to store plant-available water. Due to loosening of the soil structure by tillage before our measurements, the saturated and near-saturated hydraulic conductivities and the amount of flow-active pores of the cropped topsoil were considerably higher than those of the pasture topsoil. The measured high infiltration rates under the cropped soil could indicate that water flow took place between the aggregates rather than through the soil matrix. Based on the observed lower hydraulic conductivities and the smaller number of flow-active pores of the pasture soil compared with the cropped soil, it can be concluded that the macropore-rich soil structure created by tillage is unstable.

The adapted pore size evolution model was suitable for predicting management-induced changes in soil structure and therefore in soil properties that control the fate of water and nutrients, the development of plant root systems, as well as soil erodibility. A reasonable agreement between the measured and predicted PSD was obtained only when an assumed time-dependent degradation of pores was incorporated in the model, however. Further research should therefore focus more on the temporal changes in the soil hydraulic properties. Repeated infiltrometer measurements in the early stages after tillage would be an ideal way to do that. Our ability to measure and model management-induced changes in the PSD may improve our capacity to evaluate how new land-use systems may maintain or enhance soil health.

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