A hillslope-scale experiment to measure lateral saturated hydraulic conductivity

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[1] One of the most challenging parameters in hillslope- and watershed-scale, distributed, hydrologic models is the lateral saturated hydraulic conductivity (K_s). In this paper, we present a methodology to determine the hillslope-scale lateral K_s above a moderately deep sloping restrictive layer in an 18 × 35 m hillslope plot using perched water level measurements and drain tile outflow data. The hillslope-scale lateral K_s was compared to small-scale K_s measured with small soil cores and the Guelph permeameter. Our results show that small-scale K_s measurements underestimate the actual hillslope-scale K_s. The hillslope-scale K_s measurements were 13.7, 4.1, and 3.2 larger than small soil core measurements in the A, B, and E horizons, respectively. We argue that the gap between small-scale and hillslope-scale K_s within the same porous medium is foremost a measurement problem. Data analysis provided the K_s distribution with depth, showing a sharp decrease in K_s within the first 0.1 m of the soil and an exponential decline in K_s below 0.1 m. The distribution of K_s with depth was best described by a double-exponential relationship. Overall, results indicate the importance of macroporosity, perhaps of biological origin, in determining K_s at a hillslope scale. *INDEX TERMS:* 1894 Hydrology: Instruments and techniques; 1860 Hydrology: Runoff and streamflow; 1829 Hydrology: Groundwater hydrology; 1831 Hydrology: Groundwater quality; *KEYWORDS:* distributed modeling, drain tiles, fragipan soils, macropores, scaling, through flow

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1. Introduction

[2] Hydrologists and watershed managers need to develop models that accurately simulate the generation of runoff and erosion within a landscape to assess and recommend effective and appropriate pollutant management techniques [*Dunne*, 1983]. Distributed or semi-distributed parameter hydrologic models, which operate in a geographic information system (GIS), present an attractive option (e.g., TOPMODEL [*Beven et al.*, 1995], DHSVM [*Wigmosta et al.*, 1994], SMR [*Frankenberger et al.*, 1999], TOPOG [*Vertessy et al.*, 1993], THALES [*Grayson et al.*, 1992a], and SHE [*Bathurst and O'Connell*, 1992]). However, the limitation to this option has been data availability and inability to assess the accuracy of these models with data other than streamflow. In many applications, information on hydraulic properties is unavailable and as a result these properties become calibration parameters. The lateral saturated hydraulic conductivity (K_s) or the transmissivity profile are examples of such parameters.

[3] Methods to measure K_s range from the laboratory-based "constant head" and "falling head" permeameters using soil cores extracted from the field [*Klute and Dirksen*, 1986] to field-based Guelph permeameters [*Reynolds et al.*, 1983; *Gallichand et al.*, 1990], disk permeameters [*Perroux and White*, 1988], ring infiltrometers, auger hole and piezometer pumping tests [*United States Bureau of Reclamation*, 1993]. Despite the availability of this wide range of tools to measure K_s, many modelers continue to treat K_s as a calibration parameter.

[4] One reason why K_s is often obtained by calibration is because the calibrated K_s values in process-based hydrologic models tend to be larger than most laboratory and field-derived K_s values [*Grayson et al.*, 1992b; *Chappell et al.*, 1998; *Blain and Milly*, 1991]. *Beven et al.* [1995] pointed out that calibration of the transmissivity parameter used in TOPMODEL to measured streamflow "often yields very high values." *Beven* [1997], citing *Binley and Beven* [1991], suggested that the high, calibrated transmissivity values used in TOPMODEL might be realistic since the "more normal" K_s values are not able to mimic the recession curves from long slopes. *Grayson et al.* [1992a] demonstrated that the THALES model was not able to adequately simulate the base flow coming out of the Wagga Wagga catchment in New South Wales, Australia, without using an average K_s which was 10 times larger than the average measurements made using 83 disc permeameter measurements. *Wigmosta et al.* [1994] found that the "effective" lateral K_s determined by calibration was 100 times the vertical K_s using the DHSVM model on the Middle Fork Flathead River basin in northwestern Montana. *Vertessy et al.* [1993] found that streamflow could be most accurately simulated using the

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TOPOG_Yield model when using a lateral K_s value which was nearly 10 times larger than the mean vertical K_s measured using constant head well permeameter experiments following the design of *Bell and Schofield* [1990]. *Davis et al.* [1999] revisited this analysis and showed that using vertical K_s values taken from measurements made on large (0.22 m $\times$ 0.30 m) soil cores, which were one to three orders of magnitude larger than small core measurements, resulted in better model predictions. *Hopmans et al.* [2002] stated that small-scale laboratory soil cores could be used to estimate large-scale effective soil hydraulic properties with some kind of fitting along the way. These studies strongly suggest that use of small-scale K_s values in distributed models operating at the watershed or landscape scale remains a challenge.

[5] When the calibrated lateral K_s is greater than the small-scale K_s, either the measurements made in the watershed do not adequately describe the true lateral K_s at the modeling scale or the basic hydrologic processes within the model are not adequately described. In most watershed applications, sufficient distributed measurements seldom are made to rule out either explanation. *Grayson et al.* [1992a, 1992b] demonstrated how two fundamentally different models could be fitted to streamflow resulting in the ''right results for the wrong reasons'' [*Klêmes*, 1986]. Without distributed measurements (e.g., runoff depth, perched water depth) throughout the watershed it is difficult to interpret the physical meaning of a calibrated K_s parameter. The calibrated K_s value may be more a function of the error in other hydrologic parameters or processes in the model such as evapotranspiration, percolation, or soil depth than reality [*Beven et al.*, 1995]. *Sherlock et al.* [2000] demonstrated using an uncertainty analysis that the interpretation of flow pathways in hillslope-based field experiments can be drastically different based on errors in measured or assumed K_s. They concluded that ''saturated hydraulic conductivity is the critical hillslope hydrological parameter, and there is an urgent need to address the issues regarding its measurement further.''

[6] The application of many distributed hydrologic models requires that the magnitude of K_s and its distribution with depth be specified [*Wigmosta et al.*, 1994; *Boll et al.*, 1998; *Beven and Kirkby*, 1979; *Vertessy et al.*, 1993]. *Ambroise et al.* [1996a, 1996b] demonstrated that shape (exponential, parabolic, or linear) of the transmissivity profile, which reflects the shape of the K_s profile, has a significant effect on the prediction of saturated areas throughout a watershed. In the absence of a technique to measure the transmissivity profile, recession data have been used to indicate the shape of the transmissivity profile. *Ambroise et al.* [1996a, p. 2143] stated,

> any quasi-physical subcatchment-scale parameterization must be dependent on effective transmissivity values, yet it is not clear whether any adequate techniques exist for the direct measurement of such values except for large scale (and expensive) tracer experiments. Even if K_s measurements are available, they have usually been measured only within the root zone and from small samples or from vertical rather than lateral fluxes.

[7] To account for natural heterogeneity within a landscape the K_s should ideally be measured at the same scale as the desired application. *Grayson et al.* [1992b, p. 2660] stated,

> the scale at which homogeneity is assumed in models is larger than that of field measurements such as those for hydraulic conductivity but smaller than that represented by outflow hydrographs. Therefore, measured parameter values do not integrate the response of the ''elemental'' area, and there is an inconsistency in scale between that used in measurement of field variables and the way in which they are applied in models.

Hopmans et al. [2002] noted ''significant progress in the understanding of fundamental flow processes in heterogeneous soils is possible only if scale-appropriate measurement technologies are available''.

[8] The experiments that seem to have a great potential for characterizing the hydrologic properties of larger areas are controlled trench and drain tile experiments. *Whipkey* [1965] was able to measure hillslope-scale lateral flow and calculate K_s using an isolated hilllslope trench experiment. The K_s for a soil horizon was determined using Darcy's law for an entire saturated layer through measurements of the wetted cross-sectional area, the length of the saturated area, and change in hydraulic head over the entire saturated length. This experiment along with the experiments of *Dunne and Black* [1970], who also applied this technique, and *Hewlett and Hibbert* [1963] enlightened many hydrologists to the importance of subsurface lateral flow in hillslope hydrology. Around the same period of time, *Hoffman and Schwab* [1964] presented a method to calculate K_s and drainable porosity using drain tile outflow and drawdown measurements for the purpose of predicting drain spacing. In a similar technique, *Skaggs* [1976] presented a method, which was later used by *Rogers et al.* [1985], to determine the K_s/drainable porosity ratio from water table drawdown and rise events which satisfy the boundary conditions of the Boussinesq equation. For sloping ground, *Parlange et al.* [1989] used a drain tile experiment to describe the hydraulic properties of a 35 $\times$ 20 m hillslope plot in New York State. Although they did not explicitly determine the lateral K_s for the plot, their methodology is well-suited to measure the large-scale lateral K_s and the transmissivity profile.

[9] In this paper, we address the following objectives: to measure the lateral K_s at the hillslope-scale, to derive the distribution of K_s with depth, and to compare the measured hillslope-scale K_s to small-scale K_s and those reported in the local Soil Survey report. We adopted the methodology of *Parlange et al.* [1989] and used the solution of *Childs* [1971] for groundwater flow over a sloping impermeable bed to measure the lateral K_s at the hillslope-scale and to develop the relationship of the lateral K_s as a function of depth below the soil surface.

2. Materials and Methods

2.1. Site Description

[10] The research site is located 8 km north of Troy, ID in the eastern Palouse region. A perched water table develops above a fragipan located at ~0.65 m during the winter season. The Latah County Soil Survey [*Barker*, 1981] classifies the soil type as a Santa Series (coarse-silty, mixed, superactive, frigid Vitrandic Fragixeralfs (Soil Survey Division, official soil series descriptions, available at http://soils.usda.gov/technical/classification/osd/index.html)),

which consists of moderately well-drained soils with a moderately deep profile extending to a fragipan. This soil is classified as a Luvisol using the World Reference Base for Soil Resources. Soil horizons include A, Bw (B is substituted for Bw in the remainder of the text), E, and a Btxb (or fragipan). The A horizon is a yellowish brown silt loam that is dark brown when moist with a subangular blocky structure that extends to an average depth of 0.38 m. The Bw horizon is a brown silt loam with a blocky structure and occurs between average depths of 0.38 to 0.68 m. The E horizon is a pale brown silt loam with a massive structure that is slightly hard and occurs between depths of 0.68 to 0.86 m. The Btxb horizon is a yellowish brown silt loam and silty clay loam soil with a coarse prismatic and medium angular blocky structure that is very hard, firm, and brittle and occurs at an average depth of 0.86 m. The A, Bw, and E horizons formed in Holocene silty sediments, while the underlying fragipan (Btxb horizon) is part of a buried late Pleistocene soil [*McDaniel et al.*, 2001]. Soil horizon thickness was documented when installing the shallow observation wells (see below). The soil is composed on average of 8% sand, 74% silt, and 18% clay [*McDaniel et al.*, 2001]. Field observations at the site indicate the presence of back-filled channels of tree roots and animal burrows. These macropores are most abundant in A horizons. Mean annual precipitation for the eastern Palouse ranges from 500 mm in the west to over 830 mm in the east. More than 60% of the annual precipitation occurs from November to April with low-intensity rainfall or snowmelt. The topography was measured with a detailed site survey.

[11] Prior to the 1960s the field was forested with native tree species including Grand fir, Douglas fir, and Ponderosa pine. The land was cleared and active farming took place using a grain/legume rotation between the mid-1960s to 1994 when it was seeded with perennial grasses and enrolled in the Conservation Reserve Program.

2.2. Plot Design

[12] A hillslope plot (~35 m × 18 m) having an average slope of 20% and aspect of 345° was isolated using tile lines and plastic sheeting following a methodology similar to *Parlange et al.* [1989]. The length of the plot was selected to represent a typical flow length in a mapping unit (30 m) in GIS modeling applications. An upper tile line was installed on the fragipan to divert incoming lateral subsurface flow and a lower tile line was installed to collect and divert the outgoing lateral flow to an automated tipping bucket (Figure 1). A datalogger recorded the bucket tips every 15 minutes. All sides of the plot were lined with plastic sheeting and incoming surface runoff was diverted using galvanized sheet metal borders. Outgoing surface runoff was measured using a trough draining to another automated tipping bucket.

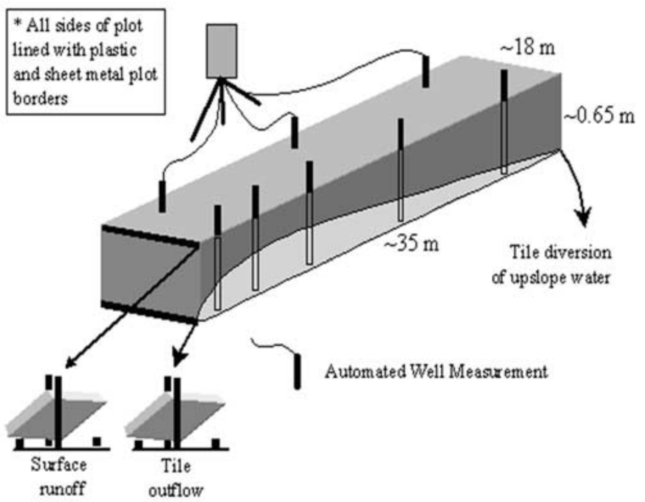

Figure 1. Hillslope plot layout and instrumentation.

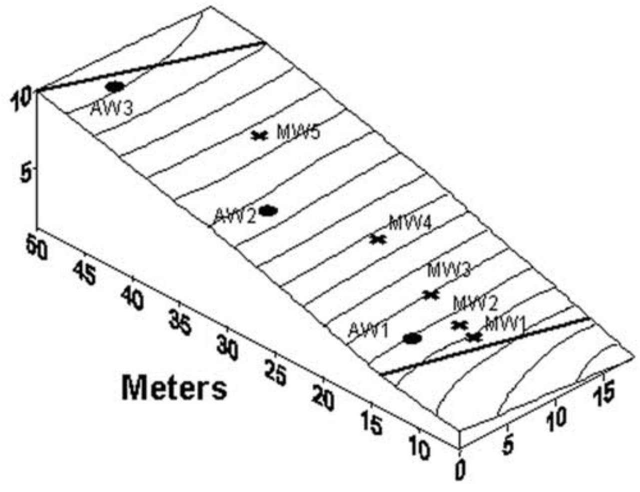

Figure 2. Contour map of the hillslope plot. Wells identified with a circle were read hourly. Wells identified with a cross were read during site visits.

[13] Three automated shallow observation wells extending down to the fragipan (AW1-AW3) were installed in the plot. The depth to the fragipan recorded at points AW1, AW2, and AW3 was 0.64, 0.74, and 0.62 m, respectively. Each well was screened by 0.45 m above the top of the fragipan. These wells were equipped with pressure transducers (model PX26-005DV, Omega Technology), which recorded hourly perched water table height above the fragipan using a Campbell Scientific CR10x datalogger (Figure 2). The shape of the water table profile was monitored manually during site visits (about weekly) using five shallow observation wells (MW1-MW5) spaced at distances of 0.5, 2.0, 4.9, 10.9, and 22.8 m above and perpendicular to the lower tile line (Figure 2).

[14] A separate experiment was conducted to ensure that the single automated well AW1 could properly account for the saturated thickness along the lower width of the plot. Measurements at three temporary wells and at three soil moisture sensors installed along the width of the tile line indicated a uniform water table that varied on average by less than 0.05 m (data not shown).

2.3. Drainage of Sloping Land

[15] Tile drains installed on sloping land are referred to as interceptor drains. Interceptor drains are used to reduce the saturation excess runoff downslope of the drain. *Childs* [1971] presented a solution to groundwater flow to a ditch or tile line over a sloping impermeable bed. Assuming flow lines are approximately parallel to the bed, following the Dupuit-Forchheimer approximation, the flow of water per unit width to the drain can be written in terms of the lateral

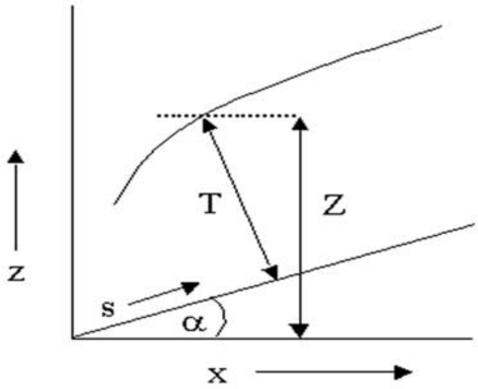

Figure 3. Conceptual model of lateral saturated flow to a tile [*Childs*, 1971].

K_s and the absolute slope of the water table by means of Darcy's law. A conceptual model of lateral flow to the tile drain is shown in Figure 3. The bed slope is at an angle α, the thickness of the saturated layer over the bed is T, and the distance along the bed is defined as s.

[16] By letting x be an arbitrary datum and Z the distance from the datum to the water table, the flow per unit width to the tile or ditch (q) can be written as:

$$q = -K_{s,avg} T \left(\frac{dZ}{ds} \right) = -K_{s,avg} T \left(\frac{dT}{ds} \cos \alpha + \sin \alpha \right) \quad (1)$$

[17] In the present study, q, T, and dZ/ds were measured directly or determined using simple geometry, so that equation (1) can be solved directly for the average lateral saturated hydraulic conductivity ($K_{s,avg}$) for the saturated soil thickness.

[18] We distinguish specifically between $K_{s,avg}$ and the lateral saturated hydraulic conductivity $K_s(z)$ at a specific depth z below the soil surface. The lateral $K_{s,avg}$ value determined by equation (1) is an average value for the portion of the soil profile that is saturated. By definition $K_{s,avg}$ determined by equation (1) can be related to $K_s(z)$ by the following equation where the numerator is the transmissivity of the saturated layer and the denominator is the saturated thickness:

$$K_{s,avg}(h) = \frac{\int_h^{D_s} K_s(z) dz}{(D_s - h)} \quad (2)$$

h is the depth to the water table from the soil surface, D_s is the depth from the soil surface to the hydraulically restricting layer such that $T = D_s - h$, z is a positive variable which increases with depth below the soil surface.

[19] Tile outflow (q) was measured between 26 March 2002 to 14 April 2002. The thickness of the water table (T) was measured with well AW1 located 2.3 m above the downslope tile. The term dT/ds cosα + sinα in equation (1) (hence referred to as "gradient term") was first determined from water level measurements made in wells MW1-MW5 and then related to automated measurements at well AW1. Mass balance calculations for the hillslope plot indicated that there was less than 0.001 m of water lost through vertical percolation through the fragipan layer. No hysteresis patterns were observed in the data.

3. Results and Discussion

3.1. Tile Outflow

[20] The analysis period was characterized by five days of snowmelt, followed by ten days of drainage and ending with five days of low-intensity rainfall (Figure 4a). Snowmelt was predicted using an hourly mass and energy balance snowmelt model [*Brooks*, 2003].

[21] Water levels in well AW1 were highly correlated with tile outflow (r^2 = 0.88) as shown in Figures 4b and 5.

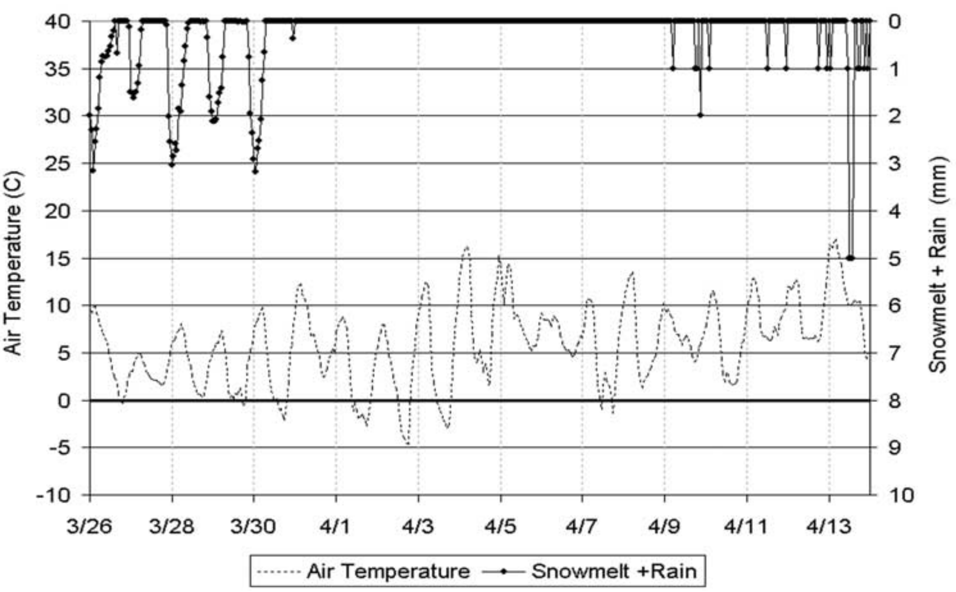

Figure 4a. Measured air temperature (°C) and the summation of predicted snowmelt and rain for the period between 26 March 2002 and 13 April 2002.

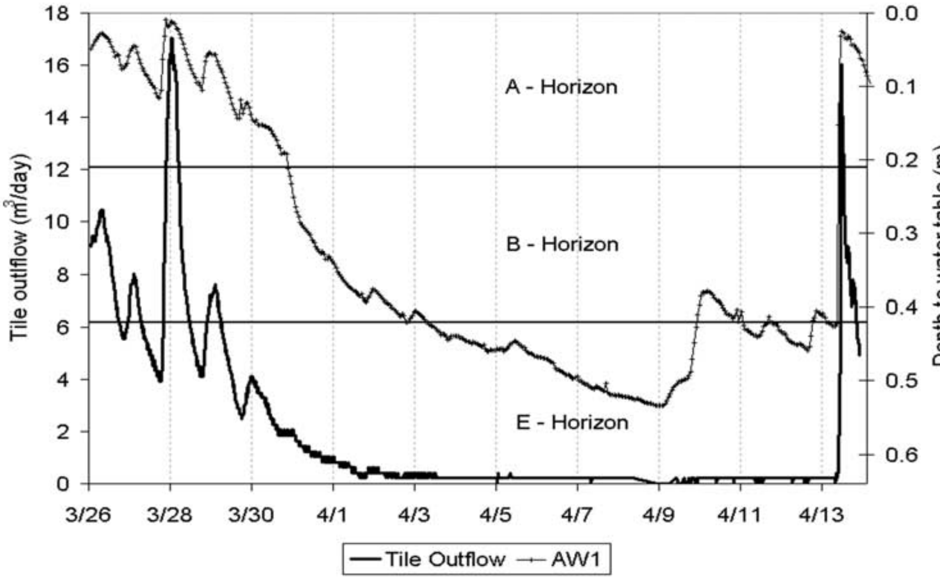

Figure 4b. Measured tile outflow (m^3/d) and the depth to the water table (m) at well AW1 for the period between 26 March 2002 and 13 April 2002.

In a similar hillslope trench study *Weyman* [1973] found that the correlation between subsurface lateral flow and water table height at the seepage face was 0.835. Water level fluctuations in wells AW2 and AW3 also correlated well with tile outflow, but the saturated thickness at these wells was smaller during the time period considered, resulting in very few K_s values in the A horizon.

3.2. Hydraulic Gradient Measurements

[22] Water table thickness increased with distance above the lower tile line for the first 5 m, and decreased from 5 m to 35 m based on measurements in wells MW1–MW5. This is illustrated with measurements from four selected dates in Figure 6. Hence the effective range of influence of the upper tile line was approximately 30 m.

[23] In the first 5 m, the water table slope was equal to the land slope when the water table was below the A horizon, and sloped toward the downslope tile line when the water table was in the A horizon. Furthermore, when the water table height in the A horizon increased, the slope of the water table close to the downslope tile line also increased, as can be seen in Figure 6 for measurements on 7 March 2002 and 26 March 2002. These observations suggest that drainage in the A horizon was more rapid than in the B and E horizons, and that within the A horizon drainage rates were greater near the surface than near the A-B horizon interface.

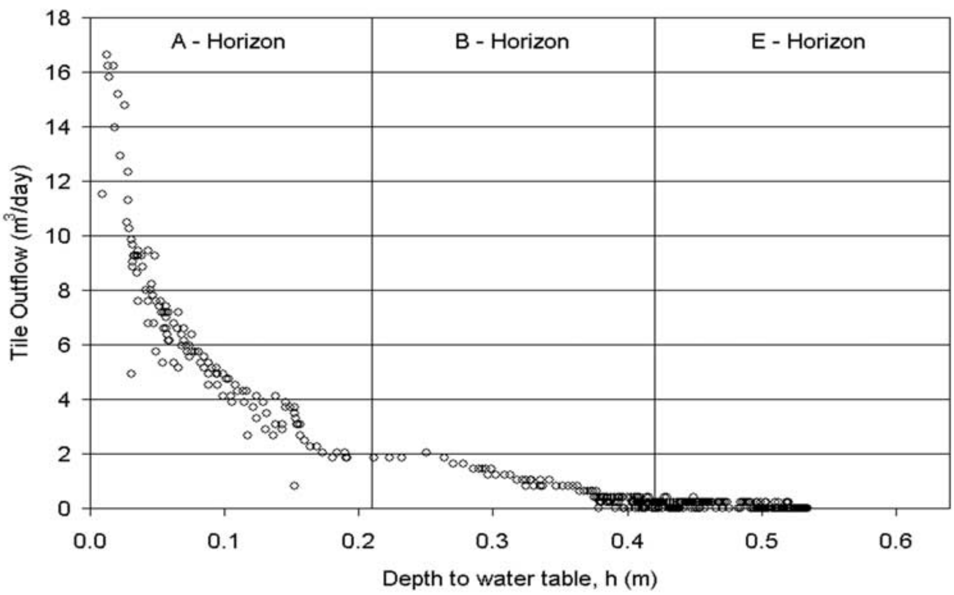

Figure 5. Measured tile outflow (m^3/d) versus the depth to the water table (m) at well AW1 for the period between 26 March 2002 and 13 April 2002.

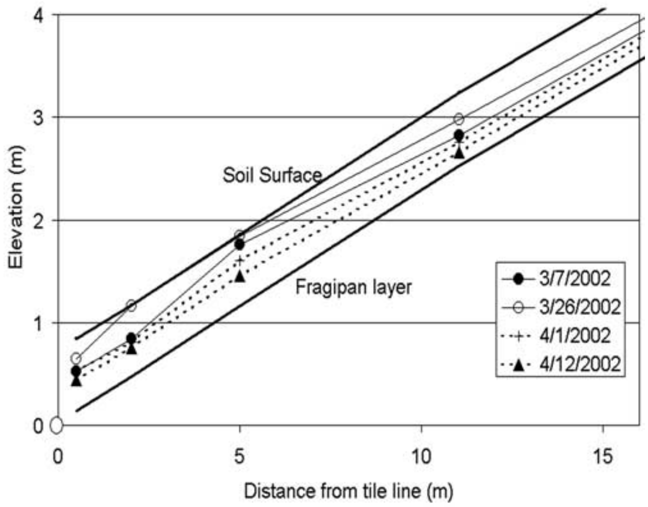

Figure 6. Water table profile upslope of the tile line for four measurement dates.

[24] The effects of increasing hydraulic gradients with depth are illustrated in Figure 7 where the gradient term in equation (1) for all observations in MW1–MW2 is plotted versus depth to the water table (h) in well AW1, which was located 2.3 m above the lower tile. The slope of the water table is approximately equal to the land slope until the water table at AW1 increases to within 0.18 m of the soil surface, which is close to the A-B interface. The slope of the water table then increases linearly from the land slope to 32% when the water table is at the surface at AW1. This relationship was used in the analysis to predict the hydraulic gradient term in equation (1) using automated well measurements at AW1.

3.3. Lateral Saturated Hydraulic Conductivity

[25] We determined the lateral $K_{s,avg}$ for a given saturated thickness from hourly data using equation (1) and plotted $K_{s,avg}$ versus depth to the water table as shown in Figure 8. In equation (1), the lateral flow (q) was measured with the tipping bucket, the thickness of the saturated layer (T) was measured at well AW1 (Figure 2), and the slope of the water table was determined from well measurements at AW1 following the relationship presented in Figure 7. Also shown in Figure 8 is the $K_{s,avg}$ of 0.8 m/d (dotted line) as reported in the Latah County Soil Survey report [*Barker*, 1981]. Clearly, in the A horizon and part of the B horizon, the hillslope-scale $K_{s,avg}$ was much greater than the reported value.

[26] Two additional observations further illustrate the rapid flow in our experimental plot. On 12 March 2002, we observed the largest tile flow rate of 24.3 m^3/d. Unfortunately, owing to a datalogger malfunction, measurements in well AW1 were unavailable, so we are not sure about the saturated thickness. By assuming that the entire soil profile was saturated at this time the $K_{s,avg}$ for this event would have been 7.3 m/d. During an earlier study by *Barndt* [2000], the front of a plug of potassium bromide traveled 7 m in 9 hours when the perched water table was 0.15 m below the soil surface, equivalent to a pore velocity of 18.7 m/d. The bulk bromide concentration was measured after 58 hours resulting in a mean pore velocity of 2.9 m/d. During that time, the perched water table had receded to a depth of 0.3 m below the soil surface. Rapid flow in this soil most likely is attributed to macropores as discussed below.

[27] Using the double-exponential function described by equation (3) as the $K_s(z)$ function in equation (2) provided the best fit of $K_{s,avg}(h)$ to the observed data as depicted by the solid line in Figure 8 (R^2 = 0.94, root mean square error = 0.39 m/day).

$$K_s(z) = 11\exp(-5.5z) + 90\exp(-50z) \tag{3}$$

The first exponential term represents the slow exponential trend seen at depths below 0.1 m while the second exponential term is required to fit the rapid decline in the near surface K_s values. Below 0.1 m, the contribution of the second term becomes negligible. The uncertainty in the gradient term in equation (2) leads to an average percent error in the K_s relationship of 47% at a 95% confidence level. The linear "stair-step" pattern seen at low-flow rates in Figure 8 is caused by the minimum resolution of the tipping bucket flow measurement device. Figure 9 shows the actual $K_s(z)$ function with the contribution of each exponential term with depth below the soil surface, and the comparison with the Soil Survey K_s of 0.8 m/d. Several activities may have generated near surface macropores explaining the large K_s and its sharp decline within the top 0.1 m of the soil surface. First, the thick rooting layer associated with the perennial grass cover has porosity greater than the native soil. Second, after snow has melted, channels and pathways are visible created by foraging small rodents. Third, raw organic matter at the soil surface also may have contributed to stronger structure near the soil surface. Fourth, despite their absence for nearly ten years, tillage implements typical for the region, which tend to break up the soil surface only within the first 0.1 m, may have caused preferential pathways.

[28] The shape of $K_s(z)$ function below 0.1 m agrees with the assumption used in TOPMODEL [*Beven and Kirkby*, 1979] that the transmissivity decreases exponentially with depth below the soil surface. A single exponential fit, however, did not account for the large K_s values near the

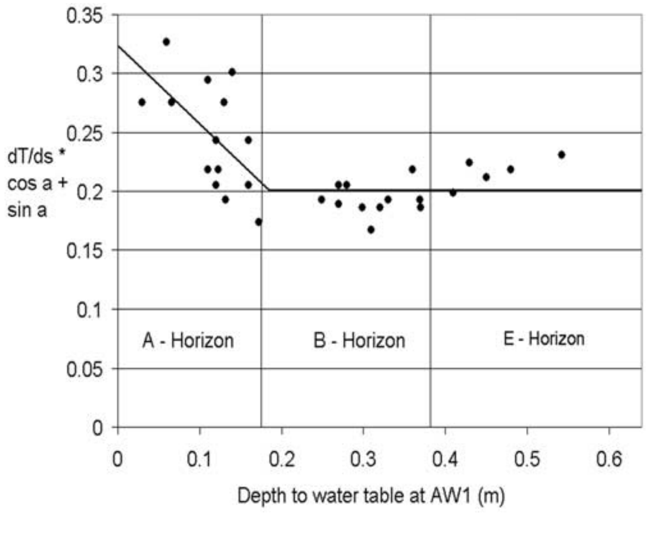

Figure 7. Gradient term in equation (1) versus depth to water table (m) at well AW1.

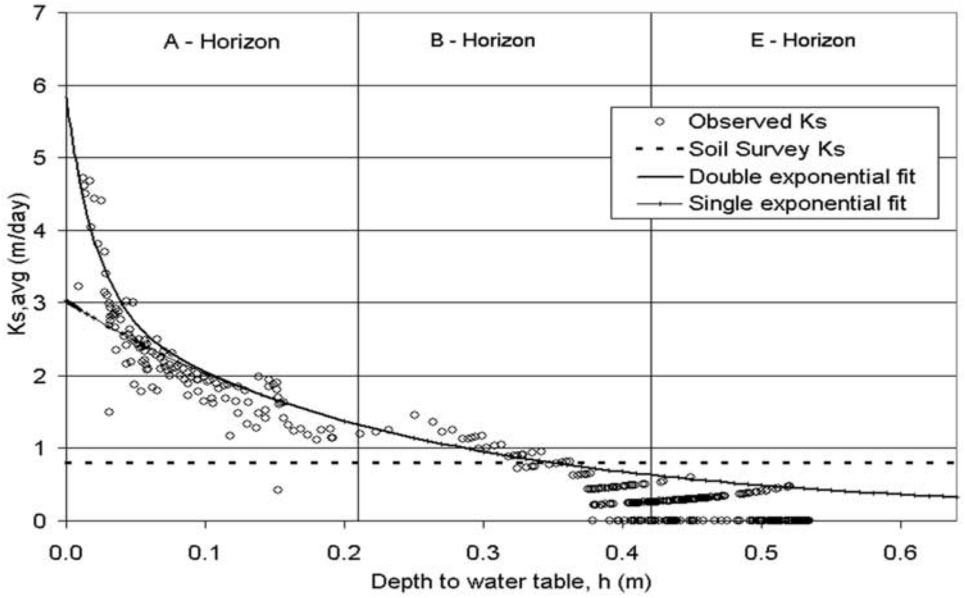

Figure 8. Effective lateral $K_{s,avg}$ (m/d) versus depth to the water table (m) at well AW1.

soil surface in this study. Recently, *Ambroise et al.* [1996a, 1996b] developed $K_s(z)$ relationships for TOPMODEL to replicate parabolic and linear transmissivity profiles. Similarly, *Iorgulescu and Musy* [1997] presented modifications to TOPMODEL to replicate a power law vertical profile of K_s, which most closely resembles our double-exponential fit.

3.4. Comparison of Hillslope-Scale K_s to Small-Scale K_s

[29] We now compare our hillslope-scale K_s values to small-scale K_s values determined at the site by *Young* [1998] and *Regan* [2000] and to K_s values reported in the Latah County Soil Survey [*Barker*, 1981]. Both *Young* [1998] and *Regan* [2000] conducted their studies in the same field within 100 m of the hillslope plot. *Young* [1998] extracted 3.4×10^{-4} m^3 vertical soil cores from each soil horizon and measured K_s using constant and falling head methods (Table 1). *Regan* [2000] used a Guelph permeameter, which was stated to have a sample volume of $\sim 1.2 \times 10^{-4}$ m^3, to measure the K_s in each soil horizon (Table 2). The Soil Survey report [*Barker*, 1981] lists the K_s in the top 0.66 m of the soil between 0.37 m/d and 1.2 m/d. The Soil Survey K_s values are in agreement with K_s values measured by *Young* [1998]. Interestingly, the soil core K_s measurements by *Young* [1998] were three to four times larger than the measurements made by *Regan* [2000] using the Guelph permeameter, assuming the sample volume of 1.2×10^{-4} m^3 is correct. This difference agrees with the findings of other studies. *Dorsey et al.* [1990] found the Guelph permeameter provided significantly lower values

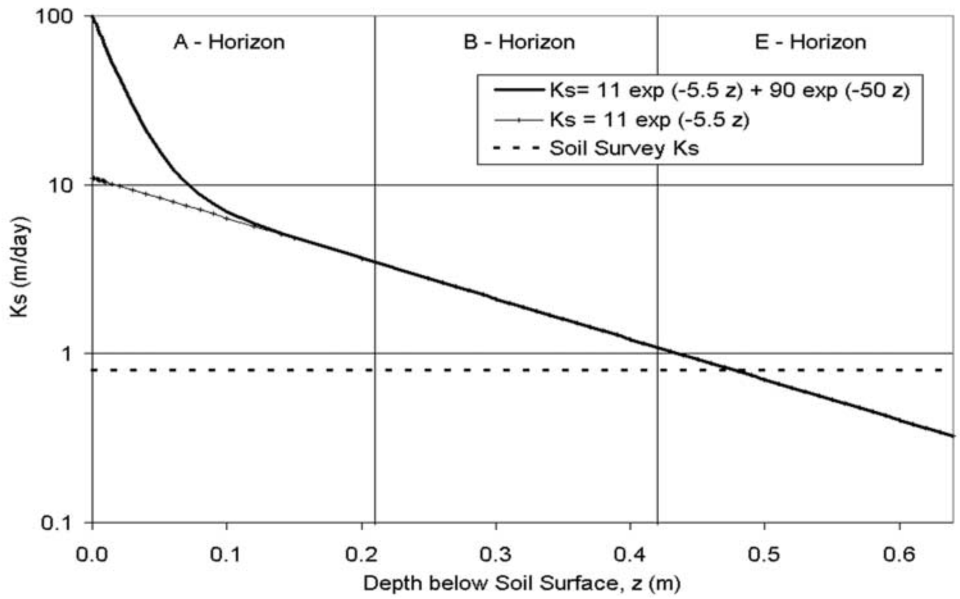

Figure 9. Actual lateral K_s (m/d) versus depth below the soil surface (m).

Table 1. Saturated Hydraulic Conductivity K_s Using Constant Head and Falling Head Permeameters[a]

Soil Horizon	Minimum K_s, m/d	Maximum K_s, m/d	Geomean, m/d	Number of Samples
A	0.49	2.95	1.11	7
B	0.049	1.16	0.42	14
E	0.083	0.53	0.17	7
Fragipan	0	0.03	NA	7

[a]See *Young* [1998].

of K_s than using pumping test, auger hole, and velocity permeameter methods. *Mohanty et al.* [1994] found that the Guelph permeameter also provided significantly lower K_s in comparison with undisturbed soil cores, velocity permeameter, disk permeameter, and double-tube methods. Maximum variability was found in undisturbed soil cores due to open-ended macropores. Differences in measured K_s were attributed to the size of the sample volume. *Paige and Hillel* [1993] showed that the K_s measured using a Guelph permeameter was one to three orders of magnitude less than K_s measured by soil cores.

[30] For this study, a K_s value for each horizon was determined by applying equation (2) to each soil horizon: 15.1, 2.1, and 0.6 m/d for the A, B, and E horizons, respectively, and 5.8 m/d for the entire soil thickness. Hillslope-scale K_s values were between one and two orders of magnitude larger than the small-scale K_s values (Table 3). The average hillslope-scale K_s was 7.7, 37.0, and 7.3 times larger than the average K_s listed by *Young* [1998], *Regan* [2000], and *Barker* [1981], respectively. The magnitude and distribution of the lateral K_s with depth agrees reasonably well with *Boll et al.* [1998] who found a lateral K_s that was 10, 5, and 2 times larger than vertical soil core measurements made in the A, B, and E soil horizons, respectively, through model calibration.

[31] The average K_s for the entire soil layer determined in this experiment, 5.8 m/d, is similar in magnitude to the K_s values determined by *Whipkey* [1965], 6.9 m/d, and *Dunne and Black* [1970], 8.2 m/d, in their untilled forest soil horizons using hillslope trench studies.

3.5. Limitations of Small-Scale Measurements of K_s

[32] Our results show that K_s increased with measurement scale, from the Guelph permeameter ($\sim 1.2 \times 10^{-4}$ m^3) to small cores (3.4×10^{-4} m^3) to the hillslope plot. These results confirm what many hydrologists have determined through model calibration of K_s, that small-scale K_s measurements are not representative of hillslope-scale K_s [*Grayson et al.*, 1992a; *Wigmosta et al.*, 1994; *Vertessy et al.*, 1993; *Boll et al.*, 1998]. In all these cases small-scale K_s values consistently underestimated large-scale K_s values.

[33] We discuss three possible reasons why small-scale K_s measurements do not represent large-scale K_s values, including those in our study.

[34] 1. Resources generally are not devoted to acquire a sufficient number of samples to statistically calculate a mean hillslope-scale K_s. It is commonly acknowledged that in most natural landscapes the variation in K_s measurements ranges over one order of magnitude [*Nielsen et al.*, 1973]. This variation is a product of the natural heterogeneity within a hillslope which in the case of the K_s at the hillslope-scale is caused primarily by macropore flow [*Bouma*, 1981; *Beven and Germann*, 1982; *Steenhuis and Muck*, 1988; *Flury et al.*, 1994]. The difficulty is trying to physically characterize the extent and continuity of macropores within a landscape at the scale of a single modeling element. If the local K_s varies by one order of magnitude, the number of samples required to determine the true mean can exceed 100. In the studies of *Young* [1998] and *Regan* [2000] this was certainly the case (Tables 1 and 2). However, even if the statistically correct number of samples was taken, the small-scale K_s measurements may still not represent the large-scale values.

[35] 2. Lack of adequate sampling is compounded by physical limitations inherent in small-scale sampling, when soils have weak structures or large macropores, such as in our hillslope plot. In regard to the extraction and subsequent measurement of K_s on soil cores, *Young* [1998] stated, "In some cases there was a high degree of variation found between cores from the same horizon. In those cases an additional core was run to improve the average. Cores that were observed to have weaknesses such as fractures or gaps along the rim, were not included in the calculation of the mean." We can only speculate over the bias in the average K_s values by *Young* [1998] due to the omission of cores representing these exceptional heterogeneities. *Sobieraj et al.* [2001] showed that pedotransfer functions, which predict K_s based on soil texture and bulk density could not predict storm flow using a distributed hydrologic model. They explained that the pedotransfer functions failed to include soil samples that yielded high K_s values which were thought to be biased by macropores [*Rawls et al.*, 1998]. Clearly, it is impossible to quantify K_s if a soil core contains a large macropore similar to the rodent burrows and tree-root channels found at our research site because all applied water would immediately pass through the core. Hypothetically, if two samples out of 100 have a K_s value of 1000 and the 98 remaining samples have a K_s value of 1, the geometric mean for this hypothetical case would only be 1.1. Thus, while isolated macropores physically are difficult to include in the data set, their contribution to the average value can be minimal.

[36] 3. Even if we were to overcome the lack of sampling numbers and above bias in sampling techniques, the small-scale mean still may not represent the large-scale mean because the connectivity of the macropores at the larger scale is unknown [*Beven and Germann*, 1982]. *Peterson and Dixon* [1971], as described by *Beven and Germann* [1982], found that flow through a single macropore representing 0.002% of the total plot area accounted for 40% of the total infiltration. In this case since the macropore was "active" and well connected to the

Table 2. In Situ Saturated Hydraulic Conductivity K_s Using the Guelph Permeameter[a]

Soil Horizon	Minimum K_s, m/d	Maximum K_s, m/d	Geomean, m/d	Number of Samples
A	0.085	2.86	0.31	15
B	0.014	0.275	0.122	17
E	0.005	0.517	0.042	10
Fragipan	0.0001	0.071	0.009	14

[a]See *Regan* [2000].

Table 3. Comparison of Lateral Saturated Hydraulic Conductivity K_s Values From the Hillslope Plot to Other Methods

Description	A Horizon	B Horizon	E Horizon	All Horizons
Lateral K_s from hillslope plot, m/d	15.1	2.1	0.6	5.8
Ratio of lateral K_s from hillslope plot to *Young* [1998] vertical K_s	13.7	4.1	3.2	7.7
Ratio of lateral K_s from hillslope plot to *Regan* [2000] Guelph permeameter K_s	48.7	16.8	15.1	37.0
Ratio of lateral K_s from hillslope plot to soil survey K_s, m/d	NA[a]	NA[a]	NA[a]	7.3

[a]The Latah County soil survey treats the A, B, and E soil horizons as one soil layer.

deeper soil layers, the geometric mean of 100 samples taken from this plot would not be an appropriate statistic to characterize this flow. The true mean for the plot should be weighted by the amount of water that actively passes through each sample. Another good example of the importance of connectivity is the study by *Sharrat* [2001] in the Prairie Pothole Region of North America where surface depressions fill with spring runoff, and remain ponded for weeks and then drain rapidly. *Sharratt* [2001] reported that the water table below a 2000 m^2 pond rose by 1 m within a 24-hour period despite a 0.85 m thick layer of concrete frozen soil beneath the pond. Five unfrozen ‘‘conduits’’ beneath the pond, occupying an area of 10 m^2 (or 0.5% of the total area), drained the entire pond. A geometric mean K_s based on small-scale measurements from this site would not be able to quantify the movement of water recorded in this study. Without large-scale K_s measurements it is difficult to determine to what degree an active macropore contributes to the water movement. In summary, the gap between hillslope-scale K_s and small-scale K_s within a porous medium appears to be a measurement problem rather than a scaling problem. The magnitude of this gap will depend on the extent of macropores in the porous medium.

[37] As described above, the method to determine the lateral K_s ideally should measure a volume comparable to the basic modeling element [*Grayson et al.*, 1992b]. Our study did not determine the optimal sampling size between the small-cores and the hillslope plot, similar to the minimum representative elementary volume (REV) discussed by *Lauren et al.* [1988], if it even exists. An intermediate size that incorporates the natural heterogeneity of the hillslope may generate K_s values comparable to large-scale methods [*Davis et al.*, 1999; *Mendoza and Steenhuis*, 2002]. As pointed out by *Beven and German* [1982] the REV for soils with macropores may be much larger than for the soil matrix or micropores. In the study of *Sharratt* [2001], the REV of the thawed preferential flow paths most likely is the entire 2000 m^2 area since the majority of flow took place in only five locations.

[38] The hillslope-scale experiment in this study overcomes the pitfalls of small-scale sampling and is recommended in lieu of model calibration. For purposes of distributed modeling in heterogeneous media, however, a more economical method may be needed for routine measurement of the lateral K_s. Our methodology may be simplified by eliminating the full isolation of the plot and the upper tile line. At a minimum, however, the method requires a perched water level measurement above of a tile line of known length, the slope of the water table near the tile line, and a measurement of tile outflow. Additional measurements of the thickness of the perched layer along the tile line add greater confidence in the results. Furthermore, while this method was designed to work with a single interceptor drain on a restricting layer, it is possible to apply this design to more complex situations such as a tile line located above the restricting layer or on relatively flat ground where a solution similar to that presented by *Skaggs* [1976] and *Hoffman and Schwabb* [1964] can be applied. In such cases, however, the accuracy of the approach may be limited by the ability to describe the pressure and flow field directing water to the tile line.

4. Conclusions

[39] In this paper, we described a methodology to measure the lateral K_s at the hillslope scale using a 35 × 18 m plot, and derive the K_s distribution with depth. Our results show that hillslope-scale K_s was between one and two orders of magnitude larger than small-scale K_s measurements (small soil cores and Guelph permeameter) and those in the Soil Survey report. The difference between hillslope- and small-scale values was greatest in the A horizon. The relationship of the lateral K_s as a function of depth was best described by a double-exponential function. In this relationship, the lateral K_s decreased sharply from 101 m/d to 6 m/d within the top 0.1 m and then followed an exponential decline with depth below 0.1 m. Field observations indicate that the high lateral K_s in the top 0.1 m was caused by macropores. The gap between hillslope-scale K_s and small-scale K_s within a heterogeneous porous medium appears to be a measurement problem rather than a scaling problem. The hillslope-scale experiment in this study overcomes the pitfalls of small-scale sampling and is recommended in lieu of model calibration. Further research should refine this experiment for more routine use to support distributed hydrological modeling.

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