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# A mathematical model for variation in water-retention curves among sandy soils

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**Abstract:** Equations were developed to predict soil matric potential as a function of soil water content, texture and bulk density in sandy soils. The equations were based on the additivity hypothesis - that water-retention of a whole soil depends on the proportions of several particle size fractions, each with fixed water-retention characteristics. The new model is an advancement over previously published models in that it embodies three basic properties of water-retention curves: a) matric potential is zero at saturation water content, b) matric potential approaches -∞ as water content approaches zero, and c) volumetric water content in dry soil is proportional to bulk density. Values of model parameters were taken from the literature, or estimated by fitting model predictions to data for sandy soils with low organic matter content. Most of the variation in water-release curves in the calibration data was explained by texture, with negligible effects of bulk density and sand particle size. The model predicted that variation in clay content among soils within the sand and loamy sand textural classes had substantial effects on water-retention curves. An understanding of how variation in texture among sandy soils contributes to matric potential is necessary for interpreting biological activity in arid environments.

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**Key words:** Antarctic soils, biological activity, bulk density, soil texture, soil matric potential, soil water, water-release curve

## Introduction

Biological activity in soil varies with water content, soil texture and structure, temperature, energy and nutrient resources, and species composition (Coleman *et al*. 2004). The effect of water is usually expressed as a function of matric potential (Griffin 1981), although other measures such as the fraction of water-filled pore space (Linn & Doran 1984) and volumetric water (West *et al*. 1992) have been employed. Soil water potential has several components including matric, osmotic, pressure and gravitational potentials. Matric potential, resulting from physical forces between water and soil particles, is the most important determinant of biological activity under most natural conditions, except in salty soils where the osmotic component also must be considered (Griffin 1981).

A variety of mathematical equations has been used to describe the relationship between soil water content and matric potential. Many published equations are semi-empirical, with some parameters related to physically measurable soil properties and other parameters (fitted parameters) adjusted to achieve a good fit to data (Clapp & Hornberger 1978, McCuen *et al*. 1981, Vereecken *et al*. 1989). Water-retention equations (WREs) have been adapted to soils with different bulk properties (e.g. texture, bulk density, organic matter content) by either using regression analysis to relate parameter values to soil properties, or specifying different parameter values for different textural classes. Models of the second sort implicitly assume that variation within a textural class is unimportant. The empirical nature of most WREs is reflected in their failure to represent one or more fundamental characteristics of water-retention data, for example that water content equals the saturation level at zero matric potential.

We required a WRE for use in studies of biological activity in soils of the McMurdo Dry Valleys of Antarctica. Most of the soil water data available from the Dry Valleys (e.g. Vishniac 1993, Freckman & Virginia 1997, Treonis *et al*. 1999, Courtright *et al*. 2001, Porazinska *et al*. 2002, Parsons *et al*. 2004) are reported as gravimetric water, which should be converted to matric potential for the most accurate prediction of organism activity (Griffin 1981, Killham 1994, Hunt *et al*. 2001). The Dry Valleys are cold deserts with an annual precipitation less than 10 cm (Fountain *et al*. 1999); thus, a useful equation must perform well in dry soil. Because there are few moisture release data available for Antarctic soils, we required a theoretically sound equation to minimize the number of parameters that must be estimated from data. Dry Valley soils generally are sandy, poorly weathered and poorly



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structured (Campbell & Claridge 1987), with extremely low organic matter contents (0.02–0.08%, Campbell 2003) and with poorly weathered clay minerals (Pastor & Bockheim 1980). Soils with distinctive properties from geographical regions such as the Dry Valleys may require special WREs (Wösten *et al*. 2001). The objective of the present research was to develop a WRE that is applicable to sandy soils with low organic matter content, requires a minimal number of parameters estimated from data, requires minimal information on texture (sand, silt and clay only), and correctly represents the fundamental characteristics of water-retention curves.

## Literature

We evaluated several WREs that have been applied to sandy soils (here denoting USDA sand and loamy sand textural classes). The first equation is a simple power curve (Clapp & Hornberger 1978):

$$\theta = \theta_s \left(\frac{\psi_s}{\psi}\right)^{1/b}, \quad (1)$$

where θ is volumetric water content ($cm^3$ water per $cm^3$ dry soil), $\theta_s$ is water content at saturation, ψ is matric potential (kPa), $\psi_s$ is matric potential at saturation (air entry potential) and b is a "fitting parameter" (adjusted to achieve a fit to data). Clapp & Hornberger (1978) used a modified version of Eq. (1) to represent variation in water-retention data among soil textural classes by assigning values for parameters $\theta_s$, $\psi_s$ and b for each of eleven textural classes, for a total of more than 33 fitted parameters. The second equation is a version of the van Genuchten (1980) equation recommended by Vereecken *et al*. (1989):

$$\theta = \theta_r + \frac{\theta_s\text{-}\theta_r}{1 + (\alpha \cdot \psi)^{\eta}}, \quad (2)$$

where $\theta_r$ is residual (minimal) volumetric water ($cm^3$ water per $cm^3$ dry soil), α and η are parameters, and other variables are defined as above. Vereecken *et al*. (1989) fitted Eq. (2) to data for different soils by deriving regression equations for the four parameters ($\theta_s$, $\theta_r$, α and η) as functions of bulk density, sand, silt, clay and organic matter content, with a total of 15 fitted coefficients. The third equation (Brooks & Corey 1964) is:

$$\theta = \theta_r + (\theta_s\text{-}\theta_r) \cdot \left(\frac{\psi_s}{\psi}\right)^{1/\lambda}, \quad (3)$$

where λ is a parameter, and other variables are defined as above. Equation (3) is similar to Eq. (1), but with the addition of a parameter for residual water. McCuen *et al*. (1981) estimated values for the four parameters of Eq. (3) ($\theta_s$, $\theta_r$, λ and $\psi_s$) for each of eleven textural classes, for a total of 44 fitted parameters. The fourth equation, based on the "additivity hypothesis" (Zeiliguer *et al*. 2000), is:

$$\theta = \frac{\theta_s\text{-}(f_t \cdot w_t + f_c \cdot w_c) \cdot bd}{\exp\left[\left(\frac{\psi}{h_s}\right)^{m_s}\right]} + \frac{f_t \cdot w_t \cdot bd}{\exp\left[\left(\frac{\psi}{h_t}\right)^{m_t}\right]} + \frac{f_c \cdot w_c \cdot bd}{\exp\left[\left(\frac{\psi}{h_c}\right)^{m_c}\right]} \quad (4)$$

Equation (4) differs fundamentally from Eqs (1–3) in that some soil properties (sand, silt and clay fractions; bulk density) are incorporated directly into the structure of the equation, rather than being related statistically via fitting parameters. The three terms on the right-hand side of Eq. (4) represent water associated with the sand, silt and clay fractions as a function of matric potential ψ. Variables $f_t$ and $f_c$ are the mass fractions of silt and clay, respectively, in dry soil, and bd is bulk density (g dry soil per $cm^3$). Operator exp(x) denotes $e^x$, where e is the base of natural logarithms. There are eight fitting parameters: $w_t$ and $w_c$ are gravimetric water (g water per g dry soil) in silt and clay, respectively, at saturation, and $m_s$, $h_s$, $m_t$, $h_t$, $m_c$ and $h_c$ (with s, t and c denoting sand, silt or clay) determine the shape of the curves relating water content to ψ for the three particle size fractions.

Parameters in models based on Eqs (1) & (3) took distinct values for each textural class (in the analyses of Clapp & Hornberger (1978) and McCuen *et al*. (1981), and thus these models did not predict variation of water-release curves within classes. Models based on Eqs (2) & (4) allowed for continuous variation between and within textural classes. The four equations differ in their performance at extreme water contents. For example, Eqs (1) & (3) assign a matric potential ψ less than zero ($\psi_s$) at saturation water content $\theta_s$, while Eqs (2) & (4) assign a value of zero, as appropriate from the definition of water saturation. As ψ approaches -∞, volumetric water θ approaches a non-zero residual level ($\theta_r$) according to Eqs (2) & (3), but approaches zero according to Eqs (1) & (4).

Assuming saturation water content equals total porosity, $\theta_s$ is related to soil bulk density bd according to

$$\theta_s = 1\text{-bd/sg}, \quad (5)$$

where sg is the particle density of solid soil constituents. The equating of porosity and saturated water content has been applied across a range of soil textures (Cosby *et al*. 1984). $\theta_s$ in Eq. (4) is calculated using Eq. (5). For Eq. (3), McCuen *et al*. (1981) calculated porosity from bulk density and particle density, presumably according to Eq. (5). For Eq. (2) (Vereecken *et al*. 1989), $\theta_s$ was predicted from a multiple regression equation in which $\theta_s$ declined linearly with increasing bd. For Eq. (1), $\theta_s$ was estimated for each water-release curve to maximize fit to data, without reference to bd (Clapp & Hornberger 1978). Assuming that

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$\theta_s$ declines with increasing bd, then models based on Eqs (1) & (3) predict that θ decreases as bd increases in dry as well as in wet soil, for a given texture. To the contrary, θ in dry soil (for a given texture and matric potential) should be directly proportional to bd, assuming that water in dry soil is mainly adsorbed to particle surfaces and held at interstices between adjacent particles, because there are more soil particles per unit volume at higher bd. Such a relationship is supported empirically (Siegel-Issem *et al*. 2005). In the Vereecken *et al*. (1989) application of Eq. (2), θ increases with bd at intermediate water contents, but the effect of bd is too large (greater than proportionality) at intermediate water contents, and the effect disappears in dry soil as θ approaches $\theta_r$. Several other models in the literature incorporate effects of bd in addition to that in Eq. (5). Campbell (1985) expressed $\psi_s$ in Eq. (1) as an increasing function of bulk density, but the resulting model fails to predict that θ increases with bd in dry soil. Winfield (2005) applied a WRE (Rossi & Nimmo 1994) with three different functional forms for different regions of the water-release curve. Parameters of the WRE were predicted from regression equations as functions of bulk density and particle-size distributions. Because the equations are complicated, we evaluated them numerically for the sandiest set of soils (the third dataset in Winfield's table 3). Again, θ in dry soil failed to increase with bd.

## Materials and methods

### *Data employed*

We found only two studies which provided texture and water-release data for Antarctic Dry Valley soils, and which employed standard particle size definitions. The first is an unpublished part of a study of nematode activity and abundance in McMurdo Dry Valleys (Treonis *et al*. 1999, 2000, 2002). Soil collected from the upper 10 cm near Lake Hoare, Taylor Valley, was sieved to remove particles larger than 2 mm. Soil structure was not preserved by this sampling procedure, but Dry Valley soils generally have no cohesive structure due to high sand and low organic content, and frequent freeze-thaw cycles (Campbell & Claridge 1987). A tension plate apparatus (Klute 1986) was used to manipulate soil water content over a range of high potentials (0, -1, -2.5, -5, -8.8 kPa). Soil was packed into soil-retaining rings and wetted to saturation on ceramic plates so that the curve obtained represents the drying curve. After 24 h equilibration at saturation, suction was applied. Soil moisture was determined gravimetrically (24 h at 105°C) after at least 24 h equilibration at each suction level. Soil texture (95.7% sand, 1.8% silt, 2.5% clay) was determined by the hydrometer method (Gee & Bauder 1986). Bulk density (1.67) was determined by packing dry soil into a container of known volume. The second study (Ugolini 1963) employed soils from a profile (2–6, 6–15, 15–25, 25–30, and 30–37 cm) in the Wright Valley, Antarctica. Samples were sieved to remove particles greater than 2 mm, leached to remove salts, and exposed to atmospheres with relative humidity regulated to either 98.2% (-2499 kPa, assuming tests were at 25°C) or 50.0% (-95 361 kPa).

The above two studies from the Dry Valleys cover the wet and dry ends of the water-release curve. A third study (Rivers & Shipp 1978) was not of Dry Valley soils, but employed samples "from soil profiles developed in materials deposited by glacial melt waters" in North Dakota. Samples were taken from below the surface horizons to minimize organic matter. Five particle sizes within the sand category were given, and average sand diameter ranged from 0.14 to 0.55 mm. A pressure plate apparatus was used to set potentials to -5.0, -6.7, -10. and -1500 kPa, which bridge the gap between the first two datasets. Bulk

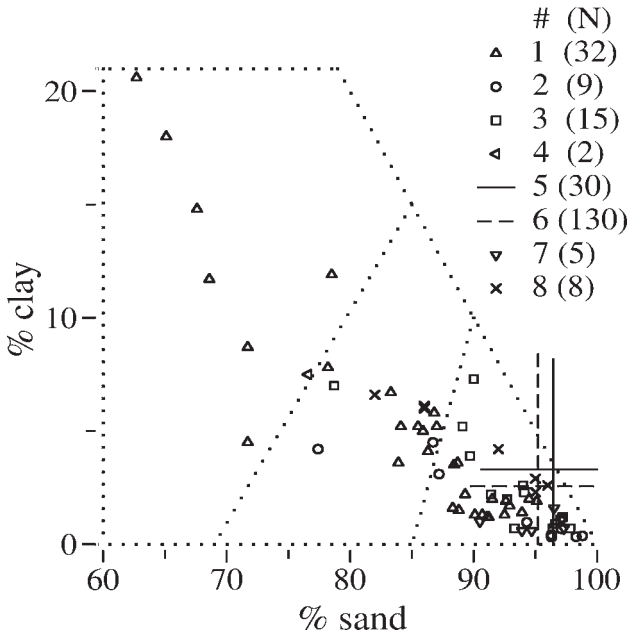



**Fig. 1.** Soil textures of samples from Antarctic ice-free areas and of samples used for model development. Three regions of the graph separated by dotted lines are for sandy loam, loamy sand, and sand textures (USDA definitions), left to right. The legend gives source of data (#) and sample size (N). Sources are #1 (Claridge & Campbell 1968), #2 (Claridge 1965), #3 (Pastor & Bockheim 1980), #4 (Ugolini & Anderson 1973), #5 (Parsons *et al*. 2004), #6 (Freckman & Virginia 1997) and #7 (Ugolini 1963), and #8 (Rivers & Shipp 1978, used for model development). Data from sources 5 & 6 are presented as ranges including the mean ± two standard deviations, which should include about 95% of observations. Sample locations included the Dry Valleys (sources #2–7), ice-free ground in the Shackleton Glacier Region, Antarctica (source #1), and North American glacial deposits (source #8). Antarctic samples were of variable thickness, but all were from above 13 cm depth. All samples were sieved to exclude particles larger than 2 mm.

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densities were determined from field samples, and varied from 1.47 to 1.59.

We were unable to fit Eq. (4) to all three datasets using a single set of parameters, because the samples at -2500 kPa (Ugolini 1963) had too high a water content in relation to their clay content compared to the samples at -1500 kPa (Treonis unpublished). The explanation for the discrepancy may lie in the use of different methods for manipulating soil water content suction plates by Treonis, pressure plates by Rivers & Shipp (1978), and control of atmospheric humidity by Ugolini (1963). Only the humidity method is sensitive to soluble salts, which increase water content for a given total (matric + osmotic) potential. Data for leached soils only were used from the Ugolini study, but he did not describe the leaching method or present evidence of its efficacy. Thus soil salinity cannot be ruled out as an explanation for high water contents (relative to soil clay) in the Ugolini data. The suction plate method (Treonis data), in contrast, produced low soil water contents relative to matric potential and clay content. Entrapped air (Sakaguchi *et al*. 2005) is a possible explanation for anomalously low θ values at high potentials. We selected the Rivers & Shipp (1978) data for model development because 1) these samples shared properties - glacial origin, low organic matter, high sand (82–96%) and low clay contents (2.3–6.6%) - with most Dry Valley soils, 2) these data appeared to suffer from fewer technical problems than the Dry Valley samples, and 3) the data included the widest range of soil textures and matric potentials.

Figure 1 shows the distribution of soil textures of the samples used for model development, and of published data from the McMurdo Dry Valleys and other Antarctic ice-free areas. Textures were about 90% sand class, 6% loamy sand and 3% sandy loam. The Rivers & Shipp (1978) samples used for model development approximated the distribution of Antarctic soil textures, but differed in the absence of sandy loam soils, most of which were from ice free areas other than the McMurdo Dry Valleys, and in having slightly greater ratios of clay to sand.

### *Parameter estimation*

Volumetric water θ was taken as the dependent variable in Eqs (1–4) to facilitate comparison among equations. However, for purposes of fitting the equations to data, we treated matric potential ψ as dependent variable (a function of θ, soil properties, and parameters) because in the field θ is more easily and reliably measured than ψ. Equations (1–3) were algebraically inverted to solve for ψ. Equation (4) was solved numerically for ψ in order to treat it as the dependent variable.

Some of the parameters in Eq. (4) were taken from the Zeiliguer *et al*. (2000) model, and others (fitted parameters identified below) were estimated to minimize E, the sum of squared errors in the natural logarithm of potential:

$$E = \Sigma[\ln(-\psi_m + 0.1) - \ln(-\psi_d + 0.1)]^2, \qquad (6)$$

where $\psi_m$ is matric potential predicted by Eq. (4) as a function of observed θ, $\psi_d$ is observed matric potential, and the summation is over 31 observations (eight soil samples times four potentials minus one missing observation). The logarithm transformation was chosen on the assumption that measurement errors of potential were proportional to the absolute value of potential. The added constant (0.1) prevents undefined logarithms in case $\psi = 0$. The dependent variable ($-\psi_m + 0.1$) varied over almost five orders of magnitude. Minimization of E was accomplished by augmented SIMPLEX optimization (Hunt *et al*. 2003).

Several alternative models also were fitted to the data. Equation (2) was fitted by retaining the original regression equations for parameters ($\theta_r$, $\theta_s$, α and η) as a function of texture, bulk density and carbon content (Vereecken *et al*. 1989, first entries in their table 7), but adjusting each parameter value by a multiplier estimated to minimize error according to Eq. (6). A modified version of Eq. (2) was fitted to the data by setting $\theta_r$ to zero and estimating multipliers for the remaining three parameters to minimize error. We modified Eq. (3) to predict continuous variation in water-release curves within sandy soils by using the Vereecken equations for parameters $\theta_r$ and $\theta_s$ (see above), and estimated parameters $\psi_s$ and λ to minimize error. Another modified version of Eq. (3) (equivalent to Eq. (1)) was fitted by setting $\theta_r$ to zero, and estimating the remaining parameters as explained above.

## Results

### *Development of new water-release equation*

Equation (4) is derived from a model (Zeiliguer *et al*. 2000) based on the additivity hypothesis, according to which experimentally determined water-release curves for nine individual soil particle size fractions (0–0.001, 0.001–0.005, . . . , 0.5–1, 1–3 mm) contribute to the release curve of a whole soil in proportion to the mass of each size fraction in the whole soil. We could not apply the Zeiliguer *et al*. (2000) model directly because most texture data available for Dry Valley soils distinguish only three particle size fractions. Thus we simplified the Zeiliguer model by reducing the number of soil fractions from nine to three (sand, silt and clay). In Eq. (4), the "h" parameters ($h_s$, $h_t$ and $h_c$ for sand, silt and clay, respectively) are the matric potentials at which water associated with a size fraction ≈ 36.8% (1/e) of its saturation value. The "m" parameters ($m_s$, $m_t$ and $m_c$) determine how abruptly water associated with a size fraction changes with ψ, when ψ equals the corresponding "h" parameter.

Saturation water ($\theta_s$) in Eq. (4) was predicted from bulk density bd according to Eq. (5), using a value of 2.75 g $cm^{-3}$

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for particle density sg. Vereecken *et al.* (1989), treating the whole spectrum of soil textures, included an effect of clay on $\theta_s$, but according to their model the clay effect would be insignificant in sandy soils. The value usually assumed for sg (e.g. Cosby *et al.* 1984) is 2.65, that of quartz. Campbell & Claridge (1987) reported orthoclase, oligoclase, biotite, quartz and hornblende as common minerals in Dry Valley soils. The particle density of these minerals ranges from 2.55 to 3.45. We found that the fit of the model to data was not very sensitive to the value of sg, and arbitrarily assigned an intermediate value of 2.75.

Parameter $w_t$ in Eq. (4) (g water per g silt at saturation) was set to 0.30, the average for two silt fractions of the corresponding parameter ($W_{i,0}$, with i indicating particle size fraction) in the Zeiliguer *et al.* (2000) model. Parameter $w_c$ was estimated to achieve the best fit to data, and its value (0.92 g water per g dry clay) was within the range (0.31–0.92) of Zeiliguer model parameter $W_{i,0}$ for their two clay size fractions. The Zeiliguer model parameters $W_{i,0}$ for five sand size fractions varied from 0.24 to 0.27 g water per g dry sand, but there is no corresponding parameter in Eq. (4) (see discussion below).

Zeiliguer *et al.* (2000) employed a double Weibull equation with seven parameters for the release curve of each of nine particle size fractions. We reduced the number of parameters by using a single Weibull equation for each of three size fractions (see Eq. (4)). We eliminated the Zeiliguer parameter for residual water (by setting $\theta_r = 0$), on the assumption that soil water will approach zero if matric potential falls low enough. The single Weibull "h" parameters (as in Eq. (4) for each of the nine particle size fractions in Zeiliguer *et al.* (2000) were estimated by numerically evaluating their double Weibull. The resulting h values (kPa) for their seven sand and silt fractions were regressed on the mid point d (mm) of the particle size range:

$$\ln(-h) = -0.161 - 1.193 \cdot \ln(d). \quad (7)$$

The form of Eq. (7) is theoretically appropriate (cf. eq. (4) in Haverkamp & Parlange 1986). Equation 7 accounted for 98.5% ($P < .0001$) of the variation in ln(-h), and was used to estimate $h_s$ for the sand fraction in Eq. (4), with d equal to mean particle size for the sand fraction. For a nominal sand diameter of 0.28 mm, $h_s$ takes a value of -3.9 kPa. Values of "h" for silt ($h_t = -230$ kPa) and clay ($h_c = -4250$ kPa) were estimated to achieve the best fit to data. The relative values of "h" parameters ensure that sand-associated water is lost first as a soil dries, followed by water associated with silt and clay fractions. The "m" parameters of Eq. (4) could not be directly estimated from corresponding Zeiliguer model parameters, because smaller values of "m" are required to achieve a smooth water-release curve when using fewer particle size fractions (three in Eq. (4) vs. nine in the Zeiliguer model). Thus $m_t$ for silt and $m_c$ for clay were assigned values of 1.0 and 0.5 (non-dimensional), respectively. Parameter $m_d$ for sand was estimated (1.35) to achieve the best fit to the data.

Zeiliguer *et al.* (2000) adjusted their nine $W_{i,0}$ parameters (i = 1,9; five for sand fractions, plus two each for silt and clay fractions) so that the observed capillary water at saturation in a whole soil ($\theta_s$) equalled the sum of saturated capillary water associated with each size fraction, weighted by the dry weight fraction $f_i$ of each size class in the whole soil:

$$\theta_s = \sum_{i=1}^{n} (fi \cdot W'_{i,0}) \quad (8)$$

where $W'_{i,0}$ is the adjusted value and $n = 9$. This adjustment was done keeping the ratios of $W'_{i,0}$ parameters among the nine size fractions the same as the ratios of the $W_{i,0}$ parameters determined for individual fractions. In other words, the amounts of saturated water associated with the unmixed size fractions were assumed to keep the same ratios after being mixed together in whole soil. We tried this adjustment method in an earlier version of Eq. (4) which included a parameter $w_s$ for capillary water associated with sand at saturation; that is, $w_s$, $w_t$, and $w_c$ were adjusted similarly to Eq. (8) with $n = 3$. The problem with this method of adjusting $w_s$, $w_t$, and $w_c$, was that $\theta$ in dry soil inappropriately decreased with increasing bulk density. Zeiliguer *et al.* (2000) pointed out that their method (Eq. (8)) of adjusting pore spaces associated with different particle size fractions neglects the likelihood that finer soil particles occupy spaces between larger particles

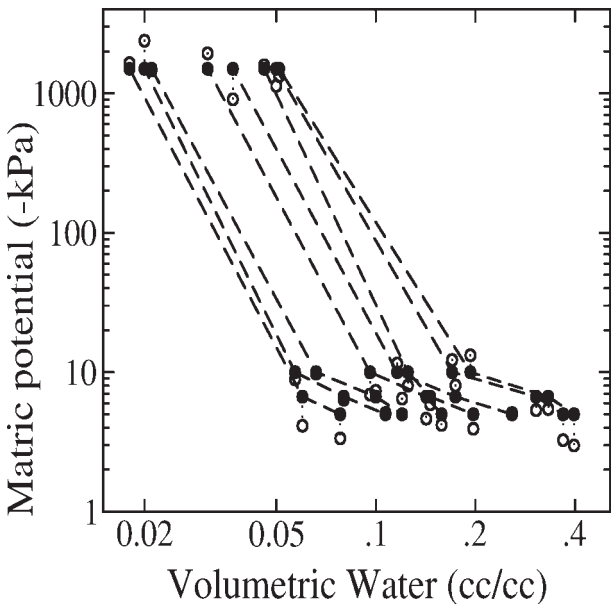



**Fig. 2.** Water-release curves for the eight soil samples (Rivers & Shipp 1978) used in model calibration. Data (filled symbols) for a given soil sample at different potentials are connected by dashed lines. Predictions of the model (Eqs (4, 5 & 7)) are denoted by open symbols, which are connected to the corresponding data points by dotted lines.

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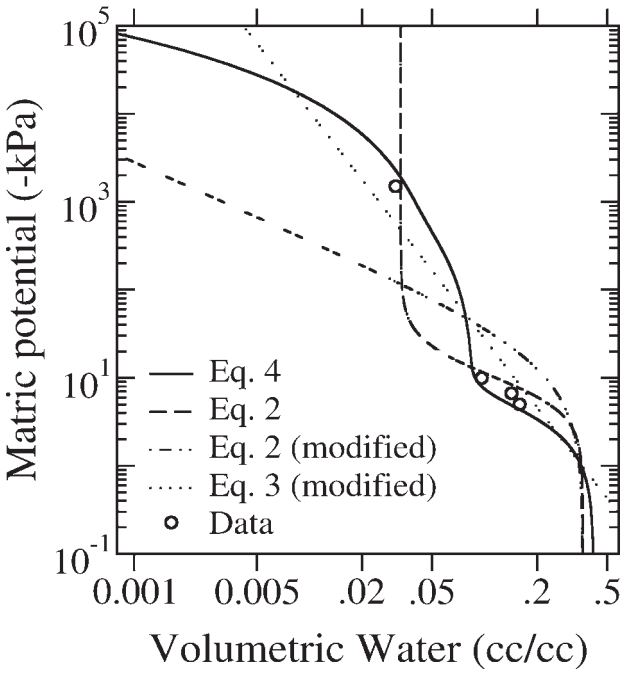



**Fig. 3.** Water-release curves (lines) for Eqs (2–4) and for equations modified by setting $\theta_r$ to zero (see text). Open circles are data for one of the soil samples of Rivers & Shipp (1978), included to indicate the range of matric potential in the data.

and thus reduce the relative contribution of larger pore spaces. We represented the packing of small particles among larger particles by assuming that large pore space (numerator of first term of Eq. (4)) equals total pore space $\theta_s$ minus pore space associated with silt ($f_t \cdot w_t \cdot bd$) and clay ($f_c \cdot w_c \cdot bd$). With this assumption and the use of Eq. (5) for $\theta_s$, Eq. (4) correctly predicts that $\theta$ in dry soil increases proportionately with increasing bulk density. We are not aware of any other water-release models with this property.

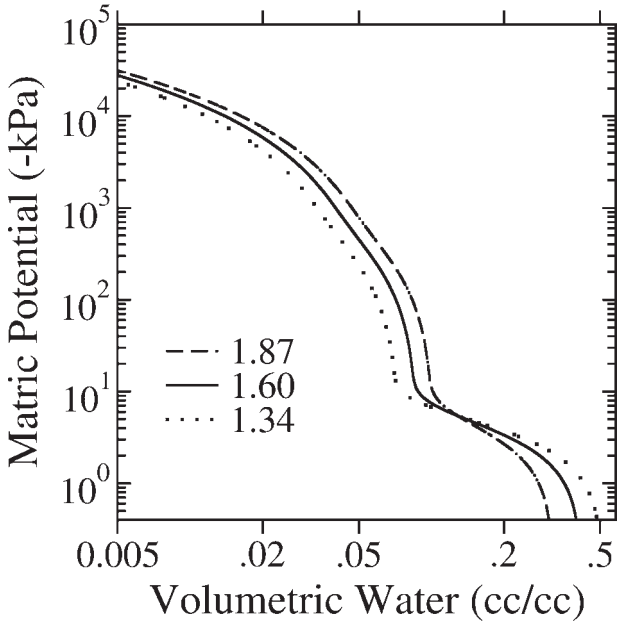



**Fig. 4.** The predicted effect (Eqs (4, 5 & 7)) of bulk density bd (g dry soil $cm^{-3}$) on the water-release curve.

### *Fit of models to data*

Our model accounted for 98.7% ($R^2 = 0.987$) of the variation in $\ln(-\psi_d + 0.1)$. Figure 2 shows water-release curves and model errors. Most of the variation among these water-release curves is related to clay content, which increases left-to-right among the curves in the figure. Alternative models with $\theta_r \neq 0$ (Eqs (2) & (3)) also fit the data well, with $R^2$ of 97.7% and 99.4% respectively, but $\theta_r$ values ranged from 0.03 to 0.06, well above the minimal values of 0.1–0.3% reported by Ugolini (1963) for Dry Valley soils. When $\theta_r$ was set to zero and the equations refitted to data, $R^2$ declined to 82% for Eq. (2) and to 87% for Eq. (3), and both models consistently overestimated matric potential at an observed potential of -1500 kPa. Figure 3 compares the shapes of equations, and illustrates how setting $\theta_r$ to zero degrades the fit of Eqs (2) & (3).

To determine the importance of soil properties for the fit of Eq. (4) to data, variation among soils was eliminated by replacing property values for each soil sample with the mean value over all eight samples. When texture of every soil was replaced with mean texture (90.5% sand, 4.4% clay), $R^2$ declined from 98.7% to 70%. Similarly, $R^2$ was reduced to 98.5% when using mean bulk density (1.54 g $cm^{-3}$), to 96.1% using average sand diameter (0.28 mm), and to 70% when all soil properties simultaneously were set to their means. The residual 70% of variation accounted for by the model with constant soil properties results from within sample effects of variable water content on potential.

### *Effect of bulk density*

Figure 4 shows the predicted effects of bulk density in a hypothetical soil with 90.5% sand, 4.4% clay, and average sand diameter of 0.28 mm. At high matric potentials near saturation, $\theta$ decreases with increasing bd in accordance with Eq. (5). At lower potentials ($<$ -10 kPa), ($\theta$ is directly proportional to bd. The crossing point for any two curves of different bd is about -5.4 kPa, not far from the approximate range of -0.1 to -5 kPa given for the crossing point by Siegel-Issem *et al*. (2005) for three sandy loams and a silt loam. Matric potential for a given volumetric water may vary with bd by a factor of ten or more over the range of bd indicated in Fig. 4. In contrast, matric potential for a given gravimetric water (g water per g dry soil) and soil texture is virtually independent of bd in dry soil (Siegel-Issem *et al*. 2005). This can be seen from Eq. (4) by setting the first term for sand-associated water to zero,

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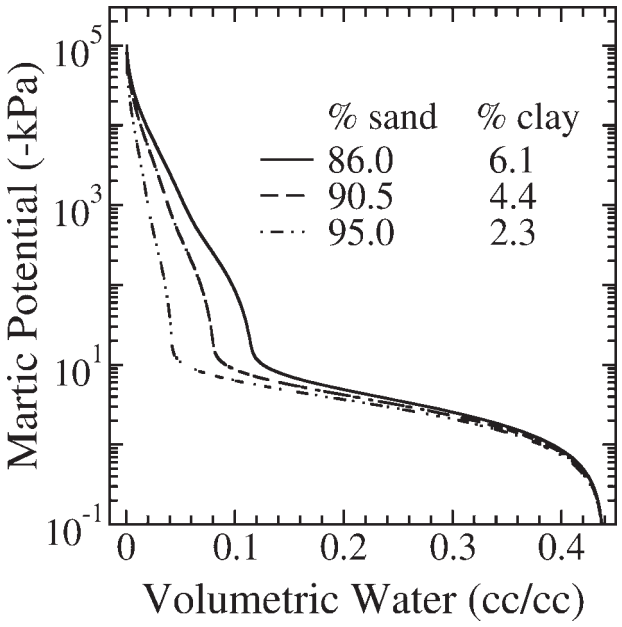



**Fig. 5.** The predicted effect (Eqs (4, 5 & 7)) of texture on the water-release curve.

and dividing both sides of the equation by bd (gravimetric water = θ/bd). Thus it is unnecessary to know bd to estimate matric potential from gravimetric water in dry soil.

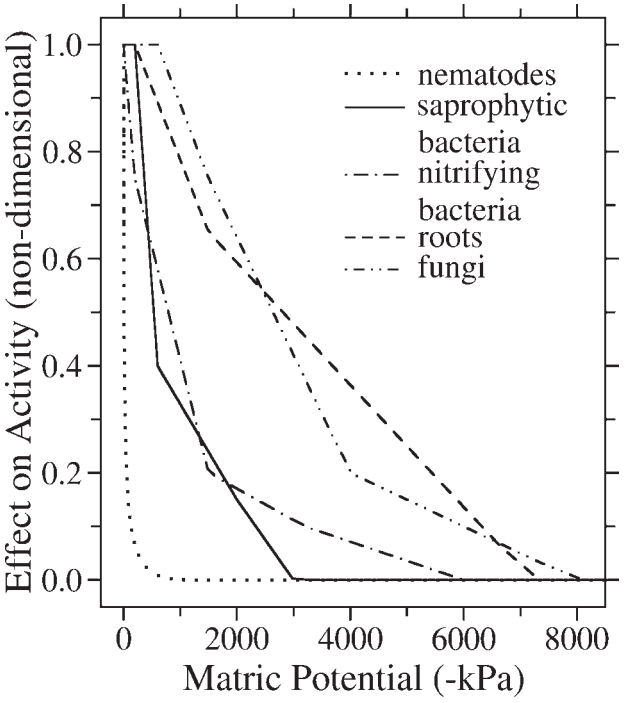



**Fig. 6.** Effects of water potential on activities of selected soil organisms as defined in two simulation models. The curve for nematodes is from Hunt *et al*. (2001). Other curves are from McGill *et al*. (1981). The plant is a drought-resistant warm season (C4 photosynthetic pathway) perennial grass.

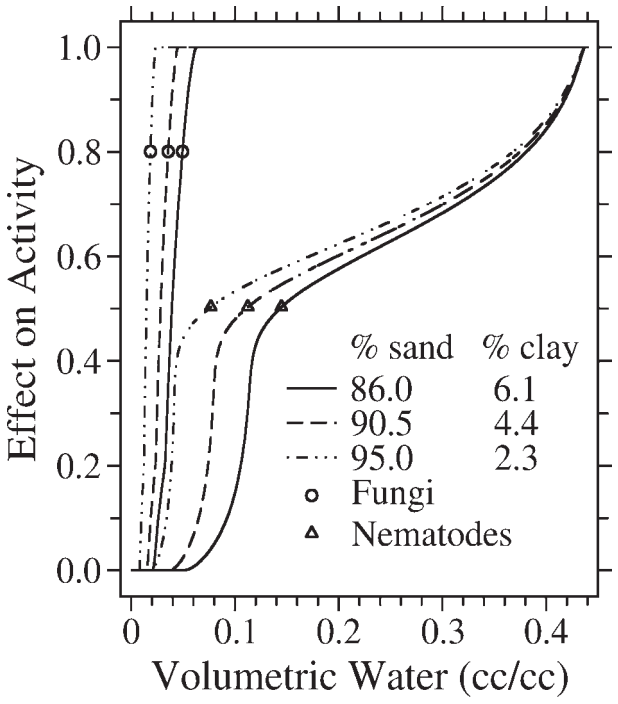



**Fig. 7.** The effects (non-dimensional reduction factors) of texture and volumetric water on biological activity, predicted using the water-release curves in Fig. 5 along with the effects of matric potential on organism activity in Fig. 6.

*Effects of texture*

Figure 5 shows the predicted effects of texture over the range of textures in the calibration data (cf. Fig. 2), but extrapolated over a wider range of matric potential and water content. Water contents were very similar at high matric potentials, because all three soils were assigned the same bulk density (1.54). However, at 5% volumetric water, potentials varied from a low of about -1500 to a high of about -10 kPa, a difference that would affect biological activity. Figure 6 shows curves predicting the effects of matric potential on activity of various soil organisms. Figure 7 combines results from Figs 5 & 6 to show the joint effects of texture and volumetric water on activity of nematodes and fungi. Nematodes were little affected by texture until water fell below about 0.15. A range of 1–5% gravimetric water is typical for McMurdo Dry Valley soils (Cameron & Conrow 1969), and this corresponds to about 1.5–8% volumetric water. At the wetter end of this range, nematode activity is predicted to vary from about 5% to 50% of maximum depending on texture. At a volumetric water content of 4%, nematode activity would vary from zero to 25% of maximum depending on texture. The sizes of these effects are probably an underestimate because clay contents of Dry Valley soils may be lower than the minimum of 2.3% used in Fig. 7. The predicted effects of texture on fungal activity are as great as those on nematodes, but the

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curves are displaced towards drier soils because fungi are more desiccation tolerant than nematodes.

## Discussion

Parameters for residual water employed in some water-release equations have been given values (volumetric water, $cm^3\ cm^{-3}$) of 0.015 (Vereecken *et al*. 1989), 0.016 (McCuen *et al*. 1981), and .05–.06 (summarized by El-Kadi 1985). These values are an order of magnitude greater than the smallest values in the data of Ugolini (1963) for McMurdo Dry Valley soils. Zeiliguer *et al*. (2000) reported zero residual water for all but the finest of their sand fractions. Residual water has been defined neither conceptually nor as a measurable soil property (Wösten *et al*. 2001), and the inclusion of residual water in water-release equations limits their use to matric potentials greater than the lowest potential available in the data used for equation development (van Genuchten 1980, Rossi & Nimmo 1994), typically -1500 kPa.

The advantage of the proposed WRE over more empirical WREs is that it performs well at both extreme high and low matric potentials, and correctly predicts the effect of bulk density in dry soil. The additivity hypothesis should in principle be applicable to non-sandy soil textures (Zeiliguer *et al*. 2000). However, for non-sandy soils, the simple assumption implicit in Eq. (4) about packing of particle sizes might have to be replaced with a more complex treatment that allows for interaction between silt and clay fractions.

To apply the model confidently to McMurdo Dry Valley soils, it ideally should be tested against water-release data from the Dry Valleys over a range of sandy textures and water contents including the low values typical in the field. There is a great need for such data. Nevertheless, the model was developed specifically to include features necessary for application to the Dry Valleys. The lowest clay content in the calibration data was 2.3%. Application to the Dry Valleys would require extrapolation to clay contents as low as 0.3%. Such extrapolation seems reasonable, given that the model is based on a more detailed model incorporating information for pure sand fractions. Variation in bulk density had little effect on the fit of the model to data because bd in the calibration data varied only from 1.5 to 1.6. However, the effect of bd in the model is conceptually sound, and the model correctly represented the effects of bd over a wider range (0.8 to 1.7) in the results of Siegel-Issem *et al*. (2005). Thus the model should be applicable over the range of bd present in the Dry Valleys. Average sand diameter varied appreciably (0.14 to 0.55 mm) in the calibration data, but had little effect on the fit of model to data over the range of potentials employed. Thus sand diameter can probably be set to a constant (0.28 mm) for application to other sandy soils.

Model predictions of nematode activity agree in general with observations (Treonis *et al*. 2000) that the percentage of active (uncoiled) nematodes in McMurdo Dry Valley soils was consistently near its lower values in soils drier than 2% gravimetric water, consistently in a high range in soils wetter than 8% water, and relatively most variable in soils with 3–8% water. However, model predictions and data are not perfectly comparable, for three reasons. First, the effect of matric potential on activity in the model was based on a mix of temperate species, which might be less desiccation tolerant than Dry Valley nematodes (Treonis & Wall 2005). Second, nematode activity is affected by pore size distribution (a function of texture and bulk density), independently of the effect of matric potential (Hunt *et al*. 2001). The pore size effect would cause 17% greater activity in sand than loamy sand textures at a bulk density of 1.6 and thus slightly increase the size of the texture effect in Fig. 7. Third, model predictions do not account for the effects of osmotic potential, which also affects nematode activity in the Dry Valleys (Treonis *et al*. 2000). Many Dry Valley soils are salty, with some sites too salty to support nematodes (Wall & Virginia 1999). Osmotic and matric potentials may have different effects on organisms (Griffin 1981). One approach to estimating osmotic potential would be to use the water-release curve to estimate matric potential, which is then subtracted from total (matric plus osmotic) potential estimated by thermocouple psychrometry. This approach would rely on an accurate estimate of matric potential as a function of gravimetric water, which estimate is not reliable without information on soil texture.

Antarctic Dry Valley soils are frozen for most of the year and undergo repeated freeze thaw cycles in the summer (Thompson *et al*. 1971). The change in freezing point $\Delta T$ of soil water in unsaturated soil with zero ice pressure is given by (eq. (34) in Buchan 1991): $\Delta T = 0.00082 \cdot (\psi + \pi)$, where matric potential $\psi$ and osmotic potential $\pi$ are both in kPa. At a typical low volumetric water content of 1.5% and an intermediate texture for Dry Valley soils, matric potential would be about $-10^4$ kPa (Fig. 5), which yields a freezing point depression of 8.2 C. Thus Dry Valley soils may contain liquid water at temperatures well below 0°C, and there is a broad temperature range over which the soil contains a mixture of liquid water and ice. The significance of this phenomenon for organism activity and survival is unknown.

Among studies relating organism abundance and activity to water content in Dry Valley soils (Vishniac 1993, Freckman & Virginia 1997, Treonis *et al*. 1999, Courtright *et al*. 2001, Porazinska *et al*. 2002, Parsons *et al*. 2004), none have considered soil texture as a factor interacting with water content. These authors may have reasoned that Dry Valley soils will not vary significantly in their water-release curves, because the soils are mostly of sand texture with very low organic matter content, and because clay content does not vary much in absolute terms, with most

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soils between 0.3 and 8% clay. However, our results suggest that the large relative variation in clay content among these sandy soils (greater than the 16 fold difference in clay between average clay and average sand textured soils) causes important differences in water-release curves over the range of matric potentials affecting organism activity. Taking soil texture into account should advance our understanding of the biological effects of gravimetric water in sandy soils.

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