A time series approach to inferring groundwater recharge using the water table fluctuation method

Russell S. Crosbie,[1] Philip Binning,[2] and Jetse D. Kalma

School of Engineering, University of Newcastle, Callaghan, New South Wales, Australia

Received 5 February 2004; revised 19 October 2004; accepted 10 November 2004; published 18 January 2005.

[1] The water table fluctuation method for determining recharge from precipitation and water table measurements was originally developed on an event basis. Here a new multievent time series approach is presented for inferring groundwater recharge from long-term water table and precipitation records. Additional new features are the incorporation of a variable specific yield based upon the soil moisture retention curve, proper accounting for the Lisse effect on the water table, and the incorporation of aquifer drainage so that recharge can be detected even if the water table does not rise. A methodology for filtering noise and non-rainfall-related water table fluctuations is also presented. The model has been applied to 2 years of field data collected in the Tomago sand beds near Newcastle, Australia. It is shown that gross recharge estimates are very sensitive to time step size and specific yield. Properly accounting for the Lisse effect is also important to determining recharge.

Citation: Crosbie, R. S., P. Binning, and J. D. Kalma (2005), A time series approach to inferring groundwater recharge using the water table fluctuation method, *Water Resour. Res.*, *41*, W01008, doi:10.1029/2004WR003077.

1. Introduction

[2] Time series approaches have been widely used to forecast the water level in wells [*Rennolls et al.*, 1980; *Graham and Tankersley*, 1993; *Bierkens et al.*, 2001]. However, relatively few studies have employed this methodology to infer recharge. Examples include that of *Viswanathan* [1984], who extracted recharge estimates from the coefficients used to model water level fluctuations due to rainfall. *Larocque et al.* [1998] used correlation and spectral analysis to characterize the hydrologic processes in a karstic aquifer. *Lee and Lee* [2000] undertook a similar study in a fractured rock aquifer to identify the recharge mechanisms.

[3] *Healy and Cook* [2002] presented an excellent review of the theory and application of the water table fluctuation method for estimating recharge. They identified the following limitations in the water table fluctuation method: If rainfall is of a low intensity and long duration then recharge may be underestimated; the diurnal fluctuations in the groundwater level due to evapotranspiration need to be taken into account; the Lisse effect can lead to greatly overestimated recharge calculations; there are many difficulties in defining and evaluating the specific yield; they also suggested that it was important that the uncertainty associated with recharge calculations be quantified. The present study overcomes many of these limitations through an improved evaluation of specific yield and by addressing problems in determining the change in water level. The new method is presented in a time series framework and can be applied to long-term records of precipitation and water table elevation.

[4] The definition of recharge used for this study is based upon that used by *Sophocleous* [1991]: water that percolates into the lower limits of the vadose zone and reaches the water table. This definition of recharge includes water that does not necessarily cause the water table to rise.

[5] The water table fluctuation method is based upon the assumption that water level rises are caused by rainfall recharging the aquifer. If the water level rise is known and the specific yield is known then recharge can be inferred:

$$R = \Delta h \times S_y, \tag{1}$$

where R is recharge, Δh is change in water table height and S_y is specific yield.

[6] Estimating Δh can be problematic because rainfall is not the only factor that can cause the water table to rise. All other causes of the water table rising need to be filtered out to prevent an overestimation of the recharge.

[7] The Lisse effect is one type of water level rise that is not due to the addition of water to the water table [*Heliotis and DeWitt*, 1987]. High-intensity rainfall can trap air in the unsaturated zone between the wetting front and the water table. The increased gas pressures in the unsaturated zone can cause the water table to rise [*Heliotis and DeWitt*, 1987]. The Lisse effect causes a very rapid water level rise that is greater than would be expected considering the depth of infiltrating rainfall, and it is also characterized by a rapid dissipation of the water level rise. Proving that a water level rise is caused by the Lisse effect and not recharge is also difficult. *Healy and Cook* [2002] suggested that it would be necessary to check if the subsurface soils were fully saturated or determine if the

[1]Now at Orange Agricultural Institute, Orange, Australia.

[2]Now at Institute of Environment and Resources, Technical University of Denmark, Lyngby, Denmark.

0043-1397/05/2004WR003077

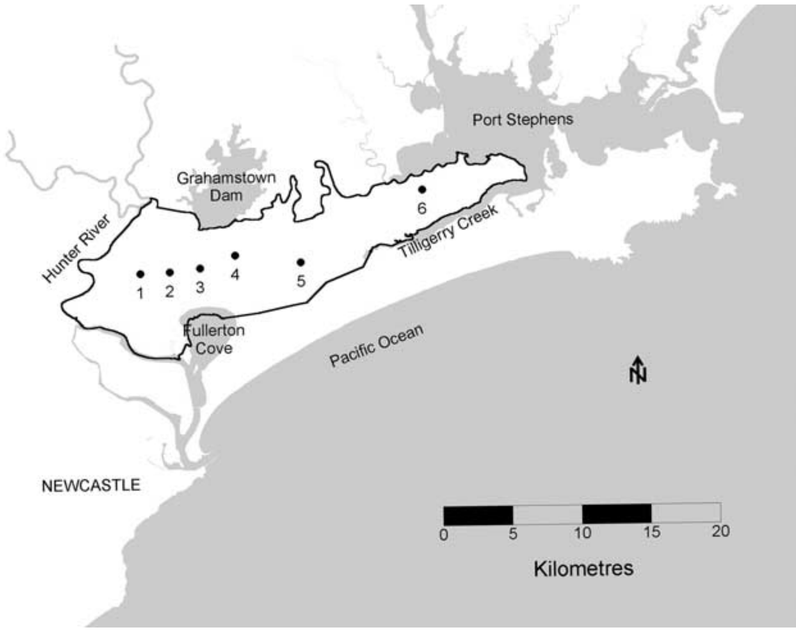

Figure 1. Location of the six field sites within the Tomago sand beds.

gas-phase pressures in the unsaturated zone were greater than atmospheric pressure. Time domain reflectometry or neutron probes could be used to measure the moisture content of the soil but their accuracy may not be sufficient to determine if the soil was fully saturated [*Healy and Cook*, 2002]. The pressure of the gas phase of the unsaturated zone could be measured using a small diameter gas piezometer, however this would become a vent allowing the increased pressures to escape [*Healy and Cook*, 2002]. A reliable method for detecting and removing the Lisse effect from a water level signal has not been reported in the literature, as far as the authors have been able to ascertain.

[8] Other causes of water level rises that are not caused by recharge include periodic signals: ocean tides [*Townley*, 1995], earth tides [*Bredehoeft*, 1967] or atmospheric tides [*Palumbo*, 1998]. If the periodic signal is measured or calculated then it can be removed using a transfer function [*Igarashi and Wakita*, 1991]. An alternative if the signal has not been measured or calculated is a prediction error filter [*Masterlark et al.*, 1999].

[9] There are many methods available for estimating the specific yield; unfortunately they often give inconsistent results [*Nwankwor et al.*, 1984]. If a constant value is assumed for the specific yield then the recharge calculated can be greatly overestimated [*Childs*, 1960; *Sophocleous*, 1985]. Even though a definition of a depth-dependant specific yield has existed for 30 years [*Duke*, 1972], it has not been routinely used in the water table fluctuation method for estimating recharge.

[10] The method presented in this paper is an improvement over previous work because recharge is inferred from the fluctuating water table and rainfall in a time series framework with: a specific yield that varies with depth; consideration of the recession limb of the hydrograph accounting for drainage away from the water table; and the filtering of the water level signal to remove water level rises caused by the Lisse effect and periodic water level fluctuations caused by evapotranspiration and atmospheric tides.

2. Site Description

[11] An investigation is underway to determine the sustainable limits of groundwater extraction for the Tomago sand beds near Newcastle, Australia. The sand beds provide about 25% of the potable water for the Newcastle region. Groundwater recharge remains one of the principal uncertainties in the water balance at Tomago.

[12] The Tomago sand beds are an unconsolidated, unconfined aquifer consisting of aeolian deposits of fine sand. Sand layers are identified by color, in order of increasing depth: upper light sand (ULS); dark sand (DS); lower light sand (GS); and grey sand (GS). The aquifer has a surface areal extent of 152 km^2, an average thickness of 18 m, an average depth to water of about 2 m, and a hydraulic conductivity of 23 m/d [*Crosbie*, 2003]. The aquifer is replenished by rainfall and discharges to the Hunter River to the West, Tilligerry Creek to the South and Port Stephens to the Northeast [*Woolley et al.*, 1995]. Figure 1 shows the location of these discharge points as well as six field sites used in this study.

[13] The field sites were chosen to cover the aquifer geographically as well as sample the different vegetation types. SK3536, SK5498 and 40a are all in forested areas, 284 and SK1708 are in an area of heath and SK6451 is in a mining revegetation area. The chosen sites are away from pumping infrastructure and areas with steep hydraulic gradients to minimize the effects of lateral flow.

[14] Each field site is instrumented with a well containing a pressure transducer and a rain gauge. The pressure transducers are Solinst Leveloggers LTs that are recording at a frequency of 5 min with a resolution of 1 mm. The rain gauges are Hydrological Services model TB3/0.2, these record every 5 min on a TinyTag data logger with a

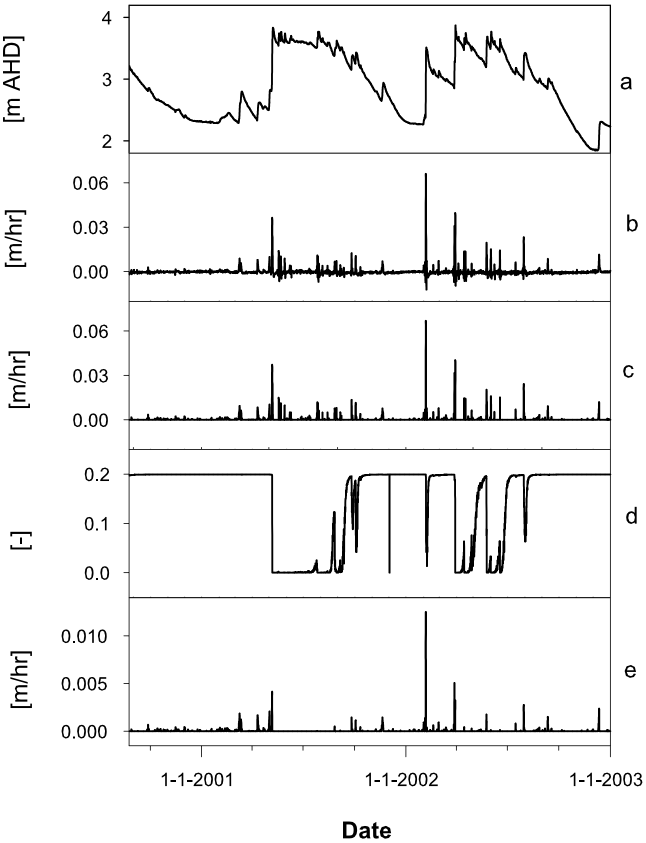

Figure 2. Recharge calculations at SK6451. (a) Water level signal (mAHD). (b) Differenced water level signal (m/hr). (c) Water level rises contributing to recharge (m/hr). (d) Apparent specific yield. (e) Recharge signal (m/hr).

resolution of 0.2 mm. The data recorded was retrieved from the field sites monthly.

3. Methodology

[15] The methodology used for calculating recharge using a time series approach is given in this section. Firstly the model is presented and then the justification is given.

3.1. Time Series Model

[16] A time series approach is used rather than an event based approach [*Delin et al.*, 2000] because we wish to obtain recharge for use in aquifer water balance calculations. In this context an event based approach is less useful than an approach that derives recharge over a longer historical time series. There are two main processes involved to infer recharge: calculating Δh; then finding the correct value of S_y to multiply it by.

[17] The time series model devised is shown in equation (2) and is based upon that shown in equation (1):

$$R_t = \begin{cases} [(h_t - h_{t-1}) + D\Delta t]S_{ya} & \text{if} \begin{cases} [(h_t - h_{t-1}) + D(z)\Delta t] > 0 \\ \sum\limits_{t>t'>t-\alpha} P_{t'} > 0 \\ h_t < h_{t+\beta} + D(z)\Delta t \end{cases} \\ 0 \text{ otherwise} \end{cases} \quad (2)$$

where R_t is recharge at time t, h_t is water level at time t, D(z) is the drainage rate (which accounts for how far the water level would have fallen had recharge not occurred), S_{ya} is apparent specific yield (defined later), P_t is precipitation at time t, α is a precipitation time parameter and β is a Lisse time parameter. This shows that recharge can be inferred from the water level rise multiplied by the apparent specific yield, provided that the change in water level plus the drainage over the time period is greater than 0; there has been rainfall in a certain time period (α) before the water level rise; and the water level is not falling at a rate greater than the drainage rate within time β, such falls are indicative of the Lisse effect.

[18] The steps involved in inferring recharge from the water table response to rainfall are illustrated in Figure 2 for site SK6451. These are as follows: (1) apply a low-pass fast Fourier transform (FFT) filter to the water level signal to remove the diurnal and semidiurnal fluctuations and the noise from the pressure transducer (Figure 2a); (2) sample the filtered signal hourly and difference once to reveal the change in water level signal (Figure 2b); (3) add the drainage term; (4) remove all negative terms from the change in water level signal; (5) remove all terms that are not preceded by rainfall in a specified amount of time (α); (6) remove all terms that are caused by the Lisse effect (Figure 2c); (7) evaluate S_{ya} for each term in the water level signal (Figure 2d); and (8) multiply the signal created in step 6 by the S_{ya} signal; this reveals the hourly recharge signal (Figure 2e). The recharge signal was then aggregated to monthly recharge.

3.2. Water Level Changes

[19] Not all water level changes are caused by rainfall infiltrating to become recharge. Other causes of water table fluctuations are due to: ocean tides; earth tides; barometric pressure changes; increases in gas pressure in the unsaturated zone; pumping; and lateral flow. These effects need to be avoided, minimized or removed from the water level signal.

[20] The effect of ocean tides, pumping and lateral flow were minimized or avoided by the careful selection of the field sites. The sites chosen are all several kilometers inland, upstream of pumping infrastructure and in areas of low hydraulic gradient. Unconfined aquifers are generally assumed to be insensitive to changes in barometric pressure, Tomago is no exception [*Freeze and Cherry*, 1979].

[21] The water level signal from Tomago is contaminated by periodic water level fluctuations and by noise from the pressure transducer. Figure 3 shows the power spectrum of the water level signal at SK6451, peaks in spectral power can be seen at the diurnal and semi diurnal frequencies. The periodic signals and the noise from the pressure transducer were removed using a low-pass FFT filter [*Vauterin and Van Camp*, 2000]. To remove the frequencies in the tidal band a notch filter was applied but better results were found using a low-pass filter with a cut-off frequency of 0.93 cycles per day (cpd) and a bandwidth of 0.1 cpd.

[22] The combined effects of direct evaporation from the water table, transpiration by phreatophytes and lateral flow to the aquifer boundaries, mean that between recharge events the water level is falling. These effects have been grouped together in a drainage (D) term. The change in water level used for the calculation of recharge needs to have this drainage term added to account for the rise that would have occurred in the water level had the drainage not occurred.

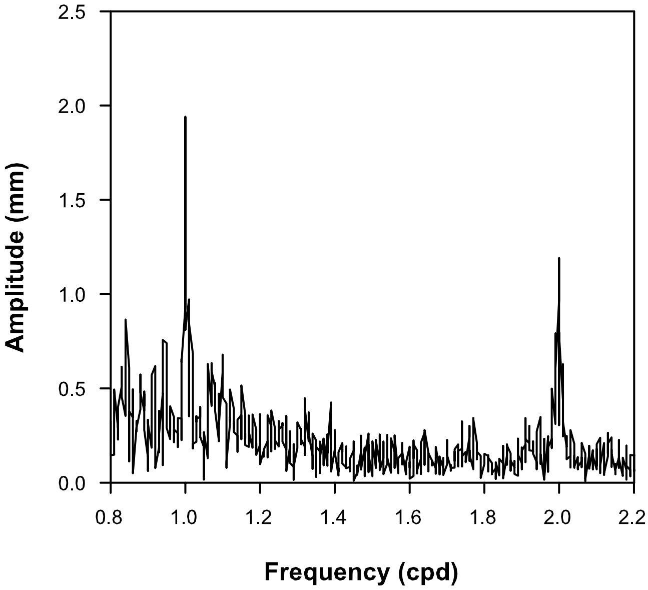

Figure 3. Tidal band of power spectrum of water level signal from site SK6451.

[23] The drainage rate was calculated from the falling water table. In periods of no rainfall the filtered water level signal was analyzed to find how far the water table fell in one day. Figure 4 shows the drainage rate at 284, this figure shows that the closer the water table is to the surface the greater the drainage rate and the greater the variability in the drainage rate. The scatter in this data is due to the way it is defined. The drainage rate includes all the water level decreases, these include lateral flow, evaporation, transpiration and pumping. The impact of pumping has been minimized by careful selection of field sites. Lateral flow is head-dependent and does not display much scatter. However both evaporation and transpiration are more complex than can be described by a simple drainage term. Despite this, a simple drainage model can be justified by noting that its affect on the model is only minor. A linear trend line has been fitted through the means of the classes shown in Figure 4 to fit a function to the drainage rate. This is a head-dependent relationship as would be expected from Darcy's law, the higher the water level, the greater the drainage rate. The drainage rate at Tomago is on the order of 1–2 cm/day.

[24] Some restrictions were placed on the water level rises: for the rises to be included in the model, they must occur within a certain time (α) after the rainfall (p) had ceased and not be caused by the Lisse effect [*Heliotis and DeWitt*, 1987]. This time was determined from the partial cross correlation [*Masters*, 1995] between the differenced water level signal and the rainfall signal, which for SK6451 is shown in Figure 5. The figure shows that the water level responds quickly to rainfall and after about 15 hours the effect of the precipitation event is no longer significant. The time series model assumes that the soil moisture distribution is at equilibrium and Figure 5 shows that this is attained after about 15 hours at SK6451.

[25] Infiltrating rainfall can trap air in the unsaturated zone and cause the Lisse effect. As the infiltration moves through the soil profile, the air pressure is increased and the water table rises to compensate for the increase in pressure in the unsaturated zone [*Heliotis and DeWitt*, 1987]. This is not recharge, as the infiltrating water has not yet reached the water table. Figure 6 shows an example of the Lisse effect upon the water level at SK3536. A period of high intensity (5 mm in 5 min) rainfall has caused the water level to rise disproportionately to the quantity of rainfall. After about 2 days, one third of the water table rise has disappeared and two thirds still remain. The water level rise that remains is due to recharge; the water level rise that dissipates is due to the Lisse effect. In a conventional application of the water table fluctuation method for estimating recharge (equation (1)), the event shown in Figure 6 would have 24.5 mm of recharge from 26.2 mm of rainfall. This is calculated by a water table rise of 150 mm multiplied by a specific yield of 0.163. For the model presented here, the recharge is considerably lower due to the Lisse effect and has been calculated to be 16.0 mm.

[26] The Lisse effect can also be seen in the partial cross correlations shown in Figure 5. Site SK6451 is affected by the Lisse effect as can be seen from the negative partial cross correlation but the site 40a is not. The reason that 40a is not affected by the Lisse effect is due to the depth from the surface to the water table. The Lisse effect only occurs in areas with <1–1.3 m between the surface and the water table [*Weeks*, 2002].

[27] The Lisse effect is detected in the recharge model from the recession limb of the hydrograph. If, some time after the current time (β), the water table has fallen further than would be predicted by the drainage rate, then the Lisse effect has occurred. The water level rise caused by the Lisse effect is then removed from the calculation of recharge.

[28] Another cause of disproportionate rise of the water table is the Reverse Wieringermeer effect [*Gillham*, 1984]. This is caused when the capillary fringe intersects the surface, the soil is fully saturated but the water table is still below the surface. The addition of a very small amount of

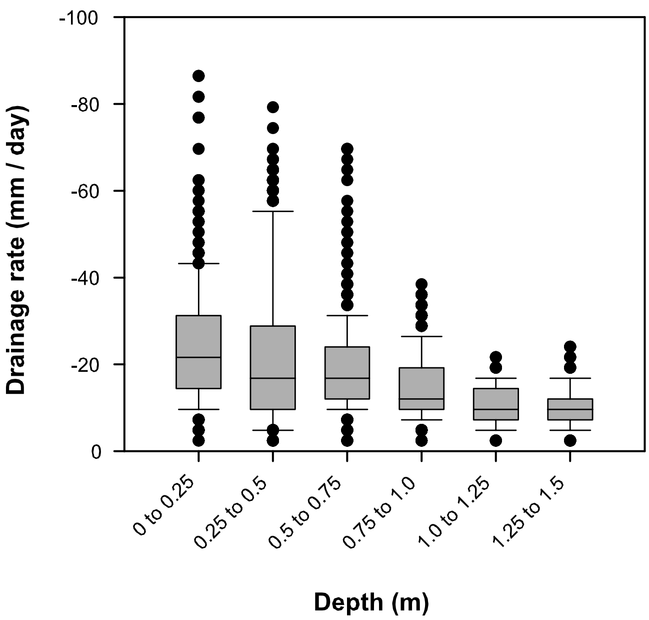

Figure 4. Box plots of water level decreases at 284 binned into 0.25 m classes by depth to water table.

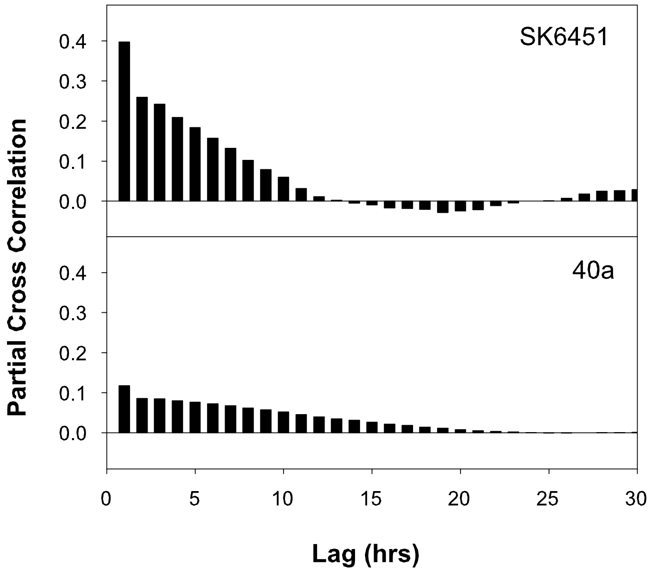

Figure 5. Partial cross correlation of differenced water level and rainfall at sites SK6451 and 40a.

water (>1 mm) will cause the conversion of capillary water to phreatic water and the water table will rise to the surface [*Gillham*, 1984; *Heliotis and DeWitt*, 1987]. This does not affect the calculation of recharge using this methodology because the definition of specific yield used implicitly removes this effect from the recharge calculation process (see next section).

3.3. Specific Yield

[29] A commonly used definition of specific yield (S_y) is: "volume of water that an unconfined aquifer releases from storage per unit surface area of aquifer per unit decline in the water table" [*Freeze and Cherry*, 1979, p. 61]. This definition can be expressed as

$$S_y = \theta_s - \theta_r, \tag{3}$$

where θ_s is the saturated moisture content and θ_r is the residual moisture content.

[30] The problem with this definition of S_y is that it neglects the store of water held above the water table by capillary forces. *Childs* [1960] showed that the specific yield is not a constant and is related to the depth from the surface to the water table. At great depth S_y approaches that of the definition of *Freeze and Cherry* [1979] but when the water table approaches the ground surface S_y approaches 0.

[31] *Duke* [1972] provided the solution to this variable specific yield problem. If it can be assumed that: the porous medium is uniform; there is an initial static equilibrium profile above the water table; and upon change in water table depth, the static equilibrium profile is reattained instantaneously; then "The apparent specific yield (S_{ya}) is equal to the change in soil water storage per unit area per unit change in water table depth" [*Duke*, 1972, p. 689]. This is illustrated in Figure 7:

$$S_{ya} = \frac{1}{\Delta z}\left(\int_{z_1}^{H} \theta_2 dz - \int_{z_1}^{H} \theta_1 \, dz\right), \tag{4}$$

where θ_1 and θ_2 are the moisture distributions as a function of depth z at times t and t + 1 respectively. At those times the water table is at z_1 and z_2 respectively with $\Delta z = z_2 - z_1$. The surface is at z = H. If the aquifer material is homogeneous from the water table to the surface and the curves θ_1 and θ_2 are identical apart from being displaced by Δz then equation (4) becomes [*Duke*, 1972]

$$S_{ya} = \theta_s - \theta(H - z_1), \tag{5}$$

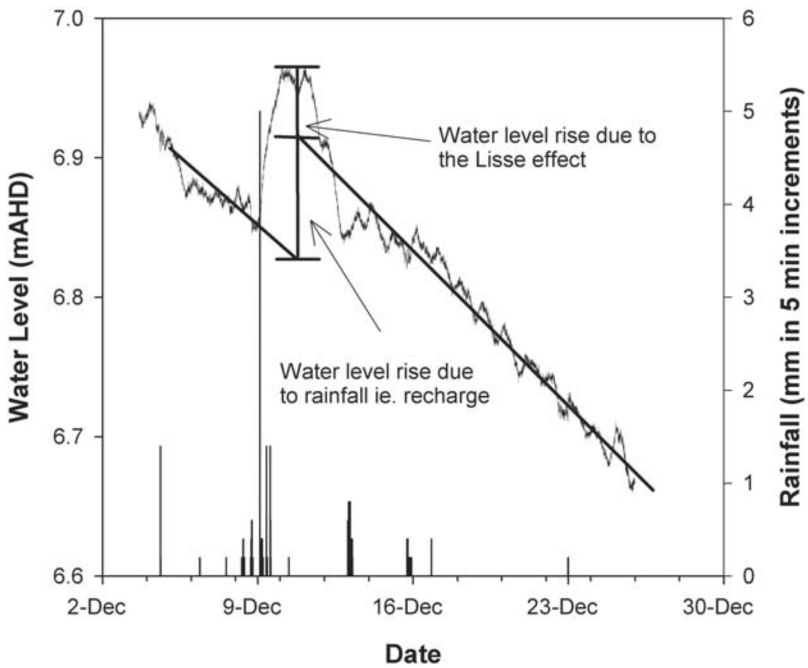

Figure 6. Water level signal at SK3536 in December 2001 showing the Lisse effect.

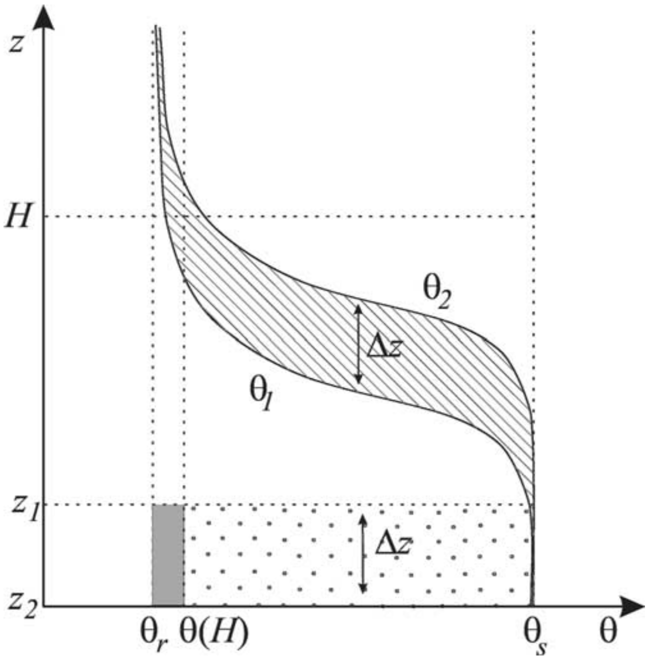

Figure 7. Definition of depth-dependant specific yield for a homogeneous material.

where θ_s is the saturated moisture content and $\theta(H - z_1)$ is the average moisture content in a layer of thickness dz at the soil surface. As shown by *Childs* [1960], equation (5) describes the area shown in Figure 7 by the cross hatching below the ground surface (H) divided by dz which is equivalent to the area shown by the dots divided by dz. As $H - z_1$ approaches ∞ then equation (5) will become equivalent to equation (3), and S_{ya} will be equal to the ultimate specific yield (S_{yu}) as which is defined to be equal to the classical definition of specific yield given by equation (3).

[32] In a layered soil (5) does not apply and a more general solution needs to be found. Figure 8 shows the case of a sand layer overlying a clay layer. The surface is now at H_2, H_1 is the interface between the sand and the clay and the other symbols are as they were in Figure 7. Denote by $\hat{\theta}(z)$ and $\theta(z)$ respectively the moisture content function of the upper and lower materials respectively. Then,

$$S_{ya} = \frac{1}{\Delta z}\left(\int_{z_1}^{H_1} \theta_2 dz - \int_{z_1}^{H_1} \theta_1\, dz\right) + \frac{1}{\Delta z}\left(\int_{H_1}^{H_2} \hat{\theta}_2 dz - \int_{H_1}^{H_2} \hat{\theta}_1\, dz\right). \tag{6}$$

This can then be evaluated to be

$$S_{ya} = \theta_s - \theta(H_1 - z_1) + \hat{\theta}(H_1 - z_1) - \hat{\theta}(H_2 - z_1). \tag{7}$$

This result can be generalized to arbitrarily many layers. Note that if $\theta = \hat{\theta}$ then the above result degenerates to the homogeneous result.

[33] The *van Genuchten* [1980] equation describes the moisture content as a function of head h:

$$\theta(h) = \theta_r + \frac{(\theta_s - \theta_r)}{\left[1 + (\alpha h)^n\right]^{1-\frac{1}{n}}}, \tag{8}$$

where α and n are soil specific parameters. Figure 5 shows that approximately 15 hours after a rainfall event the soil moisture comes to equilibrium. At equilibrium, the pressure distribution in the soil is assumed to be hydrostatic and equation (8) can be substituted into equation (5) to derive an equation for the S_{ya}:

$$S_{ya} = S_{yu} - \frac{S_{yu}}{\left[1 + \left(\alpha\left(\frac{z_i + z_f}{2}\right)\right)^n\right]^{1-\frac{1}{n}}}, \tag{9}$$

where z_i is the initial depth to water and z_f is the final depth to water. In equation (9) when the depth to water

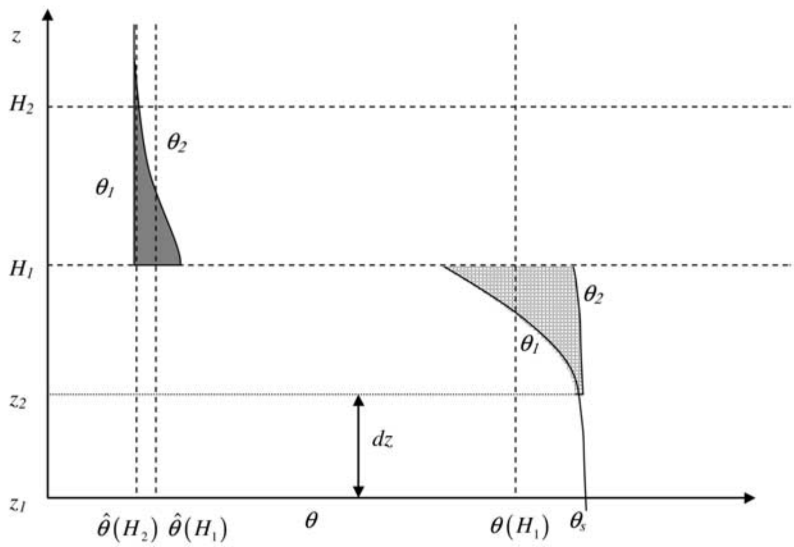

Figure 8. Definition of depth-dependant specific yield for a heterogeneous material.

Table 1. Ultimate Specific Yield Determination at Each Field Site

Site	Soil	Tempe Cell	Water Table Response to Rainfall	Pump Test
40a		0.363	0.177	0.136
SK6451		0.289	0.199	
SK5498		0.230	0.230	0.065
SK3536	ULS1		0.120	
	ULS2	0.260	0.163	
	DS		0.193	
SK1708		0.233	0.166	
284		0.299	0.169	

approaches 0, S_{ya} approaches 0, and when the depth to water is great S_{ya} approaches the ultimate specific yield (S_{yu}).

3.4. Determination of Ultimate Specific Yield

[34] While the historical estimates of S_{yu} at Tomago mirror those reported by *Nwankwor et al.* [1984], the laboratory estimates are 2–3 times higher than those estimated by pump test. In this case it is impossible to determine if this is due to the temporal and spatial scales involved, the failure of the theory used in the pump test analysis or the failure to properly implement the theory in the pump test analysis. What is clear is that the best available estimates of S_{yu} need to be used.

[35] Three methods were used to determine S_{yu}: the soil moisture retention curve was measured in the laboratory using a Tempe pressure cell; S_{yu} was inferred from the water table response to rainfall; and, historical pump tests have been reanalyzed using the lessons learnt from *Moench* [1994].

[36] The drying curve of the soil moisture retention curve was measured using a Tempe pressure cell and then the measured data was fitted to equation (8) using a least squares methodology. The fitted values of θ_s and θ_r were used to estimate S_{yu} (shown in Table 1).

[37] S_{yu} was inferred from the water table's response to rainfall based upon the methodology presented by *Armstrong and Narayan* [1998]. On an event basis the water table's response to rainfall was recorded and then plotted. The slope of the line being the specific yield, and the intercept on the *y axis* being the threshold rainfall to make the water table rise. Storm events were only included if they were of sufficient depth that S_{ya} approached S_{yu} and if the initial and final depth of the water table rise were within the same soil strata. The results of this process are also shown in Table 1.

[38] A series of historical pump tests [*Hunter District Water Board*, 1982; *Hamilton*, 1992] at two locations have been reanalyzed. Previous work analyzed the observation well data separately and the results were averaged. Here the methodology recommended by *Moench* [1994] is applied. The results are shown in Table 1.

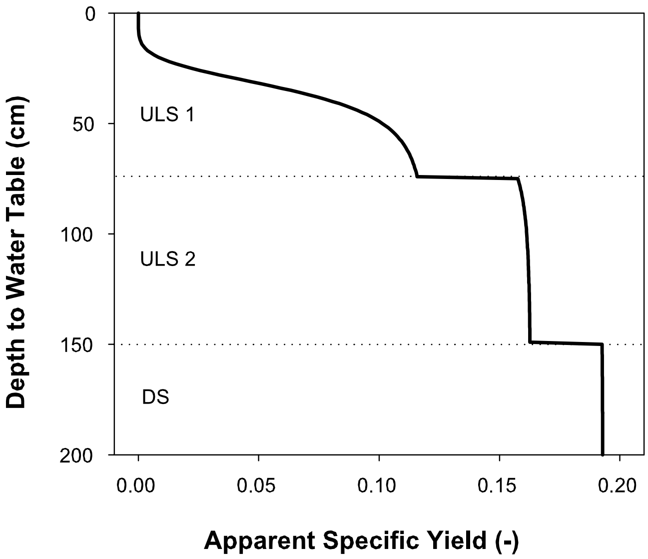

Figure 9. Apparent specific yield function used at SK3536, showing three soil layers.

[39] At field sites SK5498 and 40a three different estimates of S_{yu} have been made. In order to determine which value should be used, each is incorporated into the time series model (equation (2)) and the results interpreted (Table 2).

Table 2. Using the Three Estimates of S_{yu} to Estimate Recharge as a Proportion of the Rainfall Using the Time Series Model at Two Field Sites

Method	Site 40a, %	Site SK5498, %
Tempe cell	119	65
Water table response to rainfall	58	65
Pump test	44	17

[40] The results at field site SK5498 show that the S_{yu} determined by the pump test results in an unacceptably low estimate of recharge given that the net recharge determined by chloride mass balance was 18% [*Crosbie*, 2003]. The S_{yu} estimates from the Tempe cell and the rainfall-WT response seem reasonable. For field site 40a the S_{yu} estimate from the Tempe cell is clearly too high and the result from the pump test and the rainfall-WT response produced a recharge estimate that seems reasonable.

[41] The results of this test of the different methods of estimating S_{yu} show that the rainfall – WT response seems to provide the best estimates for these two locations. This may be due to the calculation of S_{yu} using the same temporal and spatial scale as it is to be applied in the time series model.

[42] Figure 9 shows the apparent specific yield as a function of depth for the field site at SK3536. There are three layers in the soil shown in this figure, each having a different S_{yu}. This shows that S_{ya} is effectively equal to 0 for a depth up to 15 cm and that S_{ya} is effectively equal to S_{yu} by a depth of 100 cm.

4. Results

[43] Figure 10 shows the monthly recharge calculated for each field site. Recharge between field sites is reasonably consistent in most months due to the consistency of rainfall between sites, however May 2001 is very different. In this month sites SK1708, SK5498 and 40a had much greater recharge than at sites 284, SK6451 and SK3536. This is due to the water table reaching the surface at these three sites

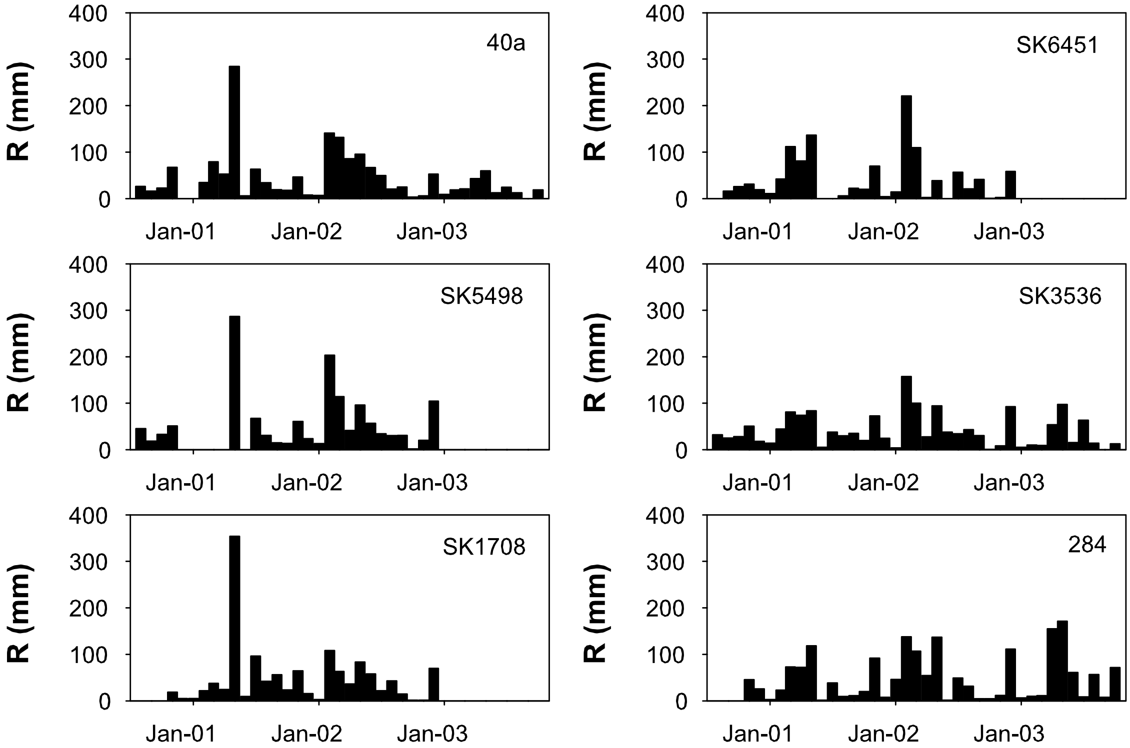

Figure 10. Monthly recharge at each field site.

preventing further recharge whereas SK1708, SK5498 and 40a had a deeper unsaturated zone that was capable of accepting more recharge into storage.

5. Discussion

[44] The sensitivity of the model to its input parameters was determined at the field site 284 over the time period 23 October 2000 to 5 June 2002. The parameters tested were the S_{yu} estimates, the time step chosen, the precipitation time parameter (α) and the Lisse time parameter (β). Table 3 shows the results of the sensitivity analysis.

[45] The estimate of recharge shows a proportionate response from a change in S_{yu}. This is as expected considering equation (1). This means that if there is a 10% error in estimating S_{yu}, there will be a corresponding 10% error in the estimate of recharge. If S_{ya} is assumed to be constant there is a substantial increase in the recharge, from 44.9% to 75.5%. This confirms the findings of *Childs* [1960] and *Sophocleous* [1985] who suggested that using a constant S_y will greatly over estimate recharge.

[46] The model appears to be very sensitive to the time step. The recharge estimate converges with decreasing time step. For large time steps differences are significant, for example a 24 hour time step results in a recharge estimate that is a third less than that found with a 1 hour time step. This is because of the loss of resolution with increasing time, the longer the time step, the greater the chance of missing the peaks and troughs in the water level signal. A one hour time step was used for this study.

[47] The Lisse time parameter, β, determines how long after the recharge event the water level is tested to see if the Lisse effect has occurred. The water level signal at 284 does not show the sharp peaks that some of the other field sites have and so is not expected to be particularly sensitive to this parameter. The recharge estimate without accounting for the Lisse effect is 49.3%, whereas if it is considered the recharge is calculated to be 44.9%. This is not a great difference considering some of the other field sites; SK5498 shows a decrease from 78.2% to 53.1% for the same time periods.

[48] The precipitation time parameter, α, is used to help exclude water level rises that are not caused by rainfall. There must be rainfall in the preceding time period for the water level rise to be included in the model. This parameter is evaluated from the partial cross correlation between the differenced water level signal and the rainfall signal. Table 3 shows that recharge increases with increasing α. This is as expected as the longer the time the more water level rises that are included as recharge. The model is not as sensitive

Table 3. Sensitivity Analysis of Recharge Estimates to Model Input Parameters at 284[a]

S_{yu} (R, %)	Time Step, hours (R, %)	β, hours (R, %)	α, hours (R, %)
0.146 (34.5)	0.5 (45.1)	0 (49.3)	5 (37.0)
0.189 (44.9)	1 (44.9)	6 (48.5)	10 (42.1)
0.233 (55.2)	2 (44.3)	12 (47.1)	15 (44.0)
0.299 (70.9)	3 (43.9)	24 (45.0)	20 (44.9)
0.189[b] (75.5)	6 (42.1)	36 (44.9)	25 (45.3)
	12 (37.8)	48 (44.2)	30 (45.7)
	24 (30.5)	60 (44.0)	50 (47.6)

[a]Baseline parameters are S_{yu} = 0.189, time step = 1 hour, Lisse time β = 36 hours, and P time α = 20 hours.

[b]This value of S_y is constant, i.e., S_{ya} does not decrease to 0 at the surface.

to this parameter as it is to either the specific yield or the time step.

6. Conclusions

[49] This paper has shown that it is possible to estimate the groundwater recharge from piezometric head data in aquifers where the water table is close to the surface. The time series methodology presented here is an advance on previous approaches because it employs a variable apparent specific yield and is capable of detecting and removing the Lisse effect using a function for the hydrograph recession curve. The specific yield is determined from the pressure saturation curve and depth to water table assuming that the soil pressure distribution is at equilibrium. For the aquifer analyzed in this study, soil water pressures have been shown to reach equilibrium in about 15 hours and so the equilibrium assumption in the model is reasonable. The other major assumption of the model is that all contributions to aquifer drainage can be lumped in a single term. This assumption was shown to be reasonable for the case study considered, but should be carefully evaluated before the method is applied elsewhere. The time series method has the advantage of not requiring extensive field instrumentation. However, it does rely very heavily on being able to estimate the apparent specific yield function and it does require water level data with a high temporal resolution.

[50] For the field site under investigation the model developed is very sensitive to the specific yield and the time step and some field sites were more sensitive to the Lisse effect than others. It was shown that if a constant specific yield were used instead of the apparent specific yield then the recharge estimate would be nearly 70% greater. If a time step of 24 hours were to be used instead of 1 hour then the recharge estimate would be one third less. Without accounting for the Lisse effect the recharge estimate could be up to 50% greater.

[51] **Acknowledgment.** The authors wish to acknowledge the Australian Research Council, Hunter Water Corporation and the former NSW Department of Land and Water Conservation for the financial support that made this project possible.

References

Armstrong, D., and K. Narayan (1998), *Using Groundwater Responses to Infer Recharge*, CSIRO Publ., Collingwood, Victoria, Australia.
Bierkens, M. F. P., M. Knotters, and T. Hoogland (2001), Space-time modeling of water table depth using a regionalized time series model and the Kalman filter, *Water Resour. Res.*, *37*, 1277–1290.
Bredehoeft, J. D. (1967), Response of well-aquifer systems to Earth tides, *J. Geophys. Res.*, *72*, 3075–3087.
Childs, E. C. (1960), The nonsteady state of the water table in drained land, *J. Geophys. Res.*, *65*, 780–782.
Crosbie, R. S. (2003), The regional scaling of groundwater recharge, Ph.D. thesis, Univ. of Newcastle, Callaghan, NSW, Australia.
Delin, G. N., R. W. Healy, M. K. Landon, and J. K. Bohlke (2000), Effects of topography and soil properties on recharge at two sites in an agricultural field, *J. Am. Water Resour. Assoc.*, *36*, 1401–1416.
Duke, H. R. (1972), Capillary properties of soils—Influence upon specific yield, *Trans. ASAE*, *15*, 688–691.
Freeze, R. A., and J. A. Cherry (1979), *Groundwater*, Prentice-Hall, Upper Saddle River, N. J.
Gillham, R. W. (1984), The capillary fringe and its effect on water-table response, *J. Hydrol.*, *67*, 307–324.
Graham, W. D., and C. D. Tankersley (1993), Forecasting piezometric head levels in the Floridian aquifer: A Kalman filtering approach, *Water Resour. Res.*, *29*, 3791–3800.
Hamilton, S. (1992), Hunter Water Corporation pumping test analysis bore 3, report, NSW Dep. of Water Resour., Sydney.
Healy, R. W., and P. G. Cook (2002), Using groundwater levels to estimate recharge, *Hydrogeol. J.*, *10*, 91–109.
Heliotis, F. D., and C. B. DeWitt (1987), Rapid water table responses to rainfall in a northern peatland ecosystem, *Water Resour. Bull.*, *23*, 1011–1016.
Hunter District Water Board (1982), Pumping tests at station no. 2 Tomago, report, Newcastle, Australia.
Igarashi, G., and H. Wakita (1991), Tidal responses and earthquake-related changes in the water level of deep wells, *J. Geophys. Res.*, *96*, 4269–4278.
Larocque, M., A. Mangin, M. Razack, and O. Banton (1998), Contribution of correlation and spectral analyses to the regional study of a large karst aquifer (Charente, France), *J. Hydrol.*, *205*, 217–231.
Lee, J. Y., and K. K. Lee (2000), Use of hydrologic time series data for identification of recharge mechanism in a fractured bedrock aquifer system, *J. Hydrol.*, *229*, 190–201.
Masterlark, T. L., H. F. Wang, L. S. Chan, and Y. T. Che (1999), Coseismic fluid-pressure response estimated from prediction-error filtering of tidal-band loading, *Bull. Seismol. Soc. Am.*, *89*, 1439–1446.
Masters, T. (1995), *Neural, Novel and Hybrid Algorithms for Time Series Prediction*, John Wiley, Hoboken, N. J.
Moench, A. F. (1994), Specific yield as determined by type-curve analysis of aquifer-test data, *Ground Water*, *32*, 949–957.
Nwankwor, G. I., J. A. Cherry, and R. W. Gillham (1984), A comparative study of specific yield determinations for a shallow sand aquifer, *Ground Water*, *22*, 764–772.
Palumbo, A. (1998), Atmospheric tides, *J. Atmos. Sol. Terr. Phys.*, *60*, 287–297.
Rennolls, K., R. Carnell, and V. Tee (1980), A descriptive model of the relationship between rainfall and soil water table, *J. Hydrol.*, *47*, 103–114.
Sophocleous, M. (1985), The role of specific yield in groundwater recharge estimations: A numerical study, *Ground Water*, *23*, 52–58.
Sophocleous, M. (1991), Combining the soil water balance and water level fluctuation method to estimate natural groundwater recharge: Practical aspects, *J. Hydrol.*, *124*, 229–241.
Townley, L. R. (1995), The response of aquifers to periodic forcing, *Adv. Water Resour.*, *18*, 125–146.
van Genuchten, M. T. (1980), A closed–form equation for predicting the hydraulic conductivity of unsaturated soils, *Soil Sci. Soc. Am. J.*, *44*, 892–898.
Vauterin, P., and M. Van Camp (2000), TSoft, report, R. Observ. of Belgium, Bruxelles.
Viswanathan, M. N. (1984), Recharge characteristics of an unconfined aquifer from the rainfall–water table relationship, *J. Hydrol.*, *70*, 233–250.
Weeks, E. P. (2002), The Lisse effect revisited, *Ground Water*, *40*, 652–656.
Woolley, D., T. Mount, and J. Gill (1995), Technical review of the Tomago/Tomaree/Stockton groundwater, report, NSW Dep. of Water Resour., Sydney.

P. Binning, Institute of Environment and Resources, Technical University of Denmark, Lyngby, DK-2800, Denmark. (pjb@er.dtu.dk)
R. S. Crosbie, Orange Agricultural Institute, NSW Department of Primary Industries, Forest Road, Orange, 2800, Australia. (russell.crosbie@agric.nsw.gov.au)
J. D. Kalma, School of Engineering, University of Newcastle, University Drive, Callaghan, NSW 2308, Australia. (cejk@alinga.newcastle.edu.au)