Propagation of pressure change through thick clay sequences: an example from Liverpool Plains, NSW, Australia

W. A. Timms · R. I. Acworth

Abstract In-situ hydraulic conductivity and specific storage measurements are derived from an analysis of pore-water pressure changes in a nest of piezometers installed in a 40-m-thick succession of smectitic clay on the Liverpool Plains of northern New South Wales, Australia. The cumulative response to the rainfall events that typically occurs during winter or early spring is propagated through the clay with measurable loss of amplitude and increasing phase lag. Five major rainfall events occurred over the four years of detailed monitoring. The phase lag at the base of the clay varied between 49 and 72 days. Barometric efficiency (BE) measurements for the clay sequence (BE = 0.07) and the underlying confined aquifer (BE = 0.10) were used, with a known porosity of 0.567, to derive specific storage values of 3.7×10^{-5} and 6.8×10^{-6} m^{-1} respectively. Vertical hydraulic conductivity (K_v) of the clay sequence derived from observed amplitude and phase changes, resulted in an average value of 2.8×10^{-9} m/s. These in-situ-derived values indicate that previous estimates of vertical hydraulic conductivity of the clays, made on core samples, are unrealistically high. The instantaneous response to individual rainfall events transmitted through the clay succession (tidal efficiency of 0.93) is also described.

Résumé Des mesures in-situ de conductivité hydraulique et d'emmagasinement spécifique sont obtenues à partir de l'analyse de la variation des pressions interstitielles telles que mesurées au droit d'un nid de piézométres interceptant une épaisseur de 40 m d'argile smectique dans les plaines de Liverpool, région nord du New South Wales, Australie. La réponse cumulative aux événements pluviométriques qui surviennent typiquement durant l'hiver ou le début du printemps, se propage dans les argiles suite à une diminution d'amplitude (des pressions interstitielles) et à un accroissement du pas de temps. Cinq événements pluviométriques sont survenus lors d'un suivi détaillé effectué sur quatre (4) années. Le pas de temps à la base de l'argile a varié de 49 à 72 jours. Les mesures d'efficience barométrique (EB) de l'assemblage argileux (BE = 0.07) et de la nappe aquifére confinée sous-jacente (BE = 1.10) ont été utilisées, avec une porosité connue de 0,567, afin d'obtenir des valeurs d'emmagasinement spécifique de respectivement 3.7×10^{-5} et 6.6×10^{-6} m^{-1}. La conductivité hydraulique verticale (Ky) de l'assemblage argileux a été déterminée à partir de la mesure des variations d'amplitudes ainsi que des pas de temps, permettant la détermination d'une valeur moyenne de 2.8×10^{-9} m/s. Ces valeurs obtenues in-situ montrent que les estimations de conductivité hydraulique verticale des argiles telles qu'obtenues à partir d'échantillons non remaniés sont surestimées. La réponse instantanée à des événements pluviométriques individuels transmis au travers de la succession d'argile (facteur d'efficacité de marée de 0.93) est également décrite.

Resumen Se han derivado mediciones in-situ de conductividad hidráulica y almacenamiento específico a partir de un análisis de cambios de presión intersticial en una red de piezómetros instalados en una secuencia gruesa de 40 m de arcilla esmectítica en las Planicies Liverpool del norte de Nueva Gales del Sur, Australia. La respuesta acumulativa de los eventos de lluvia que típicamente ocurren en invierno y principio de primavera se propaga a través de la arcilla mediante pérdida de amplitud y un incremento en retraso de fase. Durante los cuatro años de monitoreo detallado ocurrieron cuatro eventos de lluvia principales. El retraso de fase en la base de la arcilla varió de 49 a 72 días. Las mediciones de eficiencia barométrica (BE) para la secuencia arcillosa (BE = 0.07) y el acuífero confinado subyacente (BE = 0.1) se utilizaron, con una porosidad conocida de 0.567, para derivar valores de almacenamiento específico de 3.7×10^{-5} y 6.8×10^{-6} m^{-1}, respectivamente. La conductividad hidráulica vertical (Kv) de la secuencia arcillosa derivada de cambios observados en amplitud y fase dio por resultado un valor promedio de 2.8×10^{-9} m/s. Estos valores derivados in-situ indican que los estimados previos de conductividad hidráulica vertical de las arcillas,

Received: 23 August 2003 / Accepted: 21 December 2004
Published online: 14 June 2005

W. A. Timms · R. I. Acworth (✉)
Water Research Laboratory, School of Civil and Environmental Engineering, University of New South Wales,
Manly Vale, NSW, 2093, Australia
e-mail: i.acworth@unsw.edu.au
Tel.: +61-2-9949-4488
Fax: +61-2-9949-4188

DOI 10.1007/s10040-005-0436-7

hechos en muestras de núcleo, son muy altos y poco confiables. También se describe la respuesta instantánea de eventos de lluvia individuales transmitidos a través de la secuencia arcillosa (eficiencia de marea de 0.93).

Keywords Water level · Specific storage · Hydraulic conductivity

Introduction

Clay sequences are important components of groundwater systems, both limiting vertical groundwater flux and acting as a source and sink for salt and other contaminants. However, groundwater resource investigations have typically focused on the sand and gravel sequences that form high yielding aquifers, while processes within adjacent clay aquitards are generally ignored. The lack of piezometer measurements in aquitard systems is unfortunate as the opportunity to estimate aquitard storage and hydraulic conductivity is lost. Furthermore, the lack of understanding of pressure propagation through aquitard systems has led some observers to interpret water level rise in clay piezometers as due to the advective transport of rainfall recharge rather than as a response to an applied load. There are clearly major implications for resource analysis in aquifer-aquitard systems if an inappropriate conceptual model is employed.

An understanding of the mechanisms that cause hydraulic head variation in clay is critical for the successful implementation of groundwater management strategies on the Liverpool Plains in northern New South Wales, Australia. Large-scale irrigation of cotton has been developed that exploits reserves of low salinity groundwater in Tertiary sands and gravels. Smectitic clays, that frequently contain high salt levels, overlie the sand and gravel aquifers and careful management of the groundwater resource is therefore required. A program of investigation including the installation and monitoring of a nest of five piezometers has been carried out at a site on the Liverpool Plains. Values of specific storage and hydraulic conductivity for the clay sequence were derived by analysis of the barometric efficiency and time lag response of the systems. The theory of pressure transmission through clays is briefly reviewed to provide a basis for the analysis and discussion that follow.

Pressure transmission through clays

Within laterally extensive aquitards, the flow of groundwater can be assumed to be essentially vertical and the general differential equation for pressure transients, where P_{ex} is the excess pressure (Domenico and Schwartz 1997) and K_z is the vertical hydraulic conductivity, can be written as:

$$K_z \partial^2 P_{ex}/\partial z^2 = \rho g(\phi\beta + \beta_p)\partial P_{ex}/\partial t - \rho g \beta_p \partial\sigma/\partial t \quad (1)$$

Equation (1) is the normal time-dependent equation for groundwater with an additional term on the right hand side that represents the pressure response to changes in stress (pressure) with time. The coefficients before the first term on the right hand side of Eq. (1) describe the specific storage of the medium (S_s with dimensions l^{-1}) and depend upon on the density of water (ρ), the porosity (φ), the compressibility of water (β) and the compressibility of the formation (β_p).

$$S_s = \rho g(\phi\beta + \beta_p) \quad (2)$$

Dividing Eq. (1) by S_s and recognizing the hydraulic diffusivity $D=K_z/S_s$ gives

$$D\, \partial^2 P_{ex}/\partial z^2 = \partial P_{ex}/\partial t - \beta_p/(\phi\beta + \beta_p)\partial\sigma/\partial t \quad (3)$$

The term $\beta_p/(\varphi\beta+\beta_p)$ represents the tidal efficiency (TE) first described by Jacob in 1940. It is that fraction of the total specific storage that relates to the compressibility of the formation. It is the fraction of the external load change (in an undrained system) that is carried by the water as a change in pressure. The term is also sometimes referred to as a loading efficiency (Van der Kamp and Gale 1983; Neuzil 2003) and denoted by γ.

For sands and gravels, γ typically has values of >0.1 but much less than 1.0. Clay aquitards have a drained compressibility much greater than water and have γ values of almost 1.0. For this reason, the impact of the second term in Eq. (1) is frequently ignored for sands but can be significant for clays.

The specific storage (Eq. 2) is the sum of two terms. The first term represents the water released from storage as water expands ($\rho g\varphi\beta$) and the second term represents the water released by the compression of the matrix ($\rho g\beta_p$). If TE represents the proportion of S_s produced by the matrix, then $\varphi\beta/(\varphi\beta+\beta_p)$ must represent that proportion of S_s due to the expansion of water. This term is the barometric efficiency (BE) of Jacob (1940) and clearly

$$\mathrm{BE} + \mathrm{TE} = 1 \quad (4)$$

The terms BE and TE can be thought of as partitioning coefficients that distribute the imposed stress depending upon the ratio of β to β_p.

In geotechnical engineering it is frequently the case that consolidation of weak soils is a concern. Equation (1) can be modified based upon the assumption that the stress, once applied, does not change with time and therefore $\partial\sigma/\partial t=0$ and that $\beta_p \gg \beta$ for drained compressibilities, allowing terms involving β to be dropped. These modifications to Eq. (1) give:

$$K_z \partial^2 P_{ex}/\partial z^2 = \rho g \beta_p\, \partial P_{ex}/\partial t \quad (5)$$

Terzaghi and Peck (1948) introduced the concept of a coefficient of consolidation C_v to obtain their $1-D$

 DOI 10.1007/s10040-005-0436-7

consolidation equation where $C_v = K_z/\rho g \beta_p$. This leads to a further representation:

$$\partial^2 P_{ex}/\partial z^2 = 1/C_v \partial P_{ex}/\partial t \qquad (6)$$

Note that although C_v is also referred to as a hydraulic diffusivity in texts dealing with geotechnical engineering, it is not the same quantity initially defined here because that proportion of water released by the expansion of water itself has been ignored. From a historical perspective, it is interesting to note that Jacob made the opposite assumption in his initial derivation, namely that the compressibility of the matrix could be ignored and this lead him to underestimate the groundwater yield of formations in the US A (Domenico and Schwartz 1997). In effect, one group working with clays made the assumption that the partitioning coefficient TE = 1, while the other group working with sandstones made the assumption that partitioning coefficient BE = 1 when, in reality, Eq. (4) holds.

Response to loading events

The implications of the above for interpreting pore water pressure response in a piezometer to various loading events will now be examined in more detail. Some loads, such as atmospheric pressure, P_{atm}, act on both the formation and the water in a piezometer at the same time, while other loads, such as additional weight of rain, floods or tides act only on the formation and not on the water level in the piezometer. This difference leads to important variations in the observed response. The total vertical stress (pressure) (σ) acting on a horizontal plane at any depth can be resolved into two components: the effective stress supported by grain-to-grain contact in the medium (σ_e) and the water pressure in the pore space (P_w). This stress distribution is simply represented as:

$$\sigma = \sigma_e + P_w \qquad (7)$$

Naturally occurring changes in the stress distribution can be used to determine the compressibility of the formation. Atmospheric pressure, P_{atm}, has an inverse relationship with pore water pressures, P_w, observed as a change in hydraulic head, h, in a piezometer. This relationship can be used to derive a value of the barometric efficiency:

$$\mathrm{BE} = \Delta h \rho g / \Delta P_{atm} \qquad (8)$$

Another relationship for BE can be derived using Eq. (4) by substituting values for TE leads to:

$$\mathrm{BE} = \phi\beta/(\phi\beta + \beta_p) \qquad (9)$$

Equation. (9) can be solved for β_p if suitable values of BE and porosity are available. The constant of compressibility for water β is 4.6×10^{-10} Pa^{-1}.

Transmission of pressure

The transmission of pressure changes to a specific depth within a matrix depends on the frequency of pressure changes in addition to the hydraulic diffusivity of the matrix. The fluctuation is transmitted downward into the aquitard and is delayed and damped out as it moves (Van der Kamp and Maathius 1991). If a sinusoidal fluctuation is assumed, as would approximately result from the annual wetting and drying of a soil in a semi-arid climate, the amplitude of the wave diminishes with depth according to Keller et al. (1989):

$$a(z) = a_0 \exp[-z\pi/C] \qquad (10)$$

where $a(z)$ is the amplitude of the wave at depth z, and a_0 is the amplitude of the wave at the top of the aquitard. The relative penetration depth, C, is defined by Eq. (11) , where T is the period of the sinusoidal pressure change and D is hydraulic diffusivity (K_z/S_s):

$$C = (\pi T D)^{1/2} \qquad (11)$$

From Eq. (10) , at a depth z equal to C, the amplitude of the wave equals only 4% of the initial amplitude and there is a phase lag δ of π radians (or half the period of fluctuation). From this relationship it is evident that the relative penetration depth of short-term pressure changes (e.g., atmospheric pressure) is much less than for long term pressure changes (e.g., seasonal loading by soil moisture).

The relative amplitude and phase lags of pressure responses in piezometers at specific depths may be used to derive hydraulic diffusivity. Given S_s derived from atmospheric pressure response, K_v may then be determined over a specific depth range, z, using Keller et al. (1989).

$$K_v = S_s\pi/T(z/\delta)^2 \qquad (12)$$

where the phase lag δ is expressed in radians.

Alternatively, the change in amplitude of a pressure wave transmitted to a specific depth can be used to calculate K_v. Substitution of Eq. (11) into Eq. (10) and solving for K_z gives

$$K_v = z^2 S_s\pi/T\ [\ln(a_0/a_z)]^{-2} \qquad (13)$$

Where a_0 is the initial amplitude and a_z is the amplitude at a specific depth (amplitude equivalent to half the peak height measured from trough to trough). This approach was adopted by Boldt-Leppin and Hendry (2003) to determine K_v for clay rich till using a vertically separated dataset that had been filtered and decomposed using harmonic analysis.

 DOI 10.1007/s10040-005-0436-7

Field and laboratory methods

Description of study site and properties of strata

The Liverpool Plains have been formed by fluvial deposition of extensive silty clay deposits derived from the weathering of Tertiary basalts on the Liverpool Ranges to the south (Fig. 1a). Gravel filled channels occur widely beneath these silt-clays and form an extensive, high yielding, aquifer with good quality water, developed in valleys eroded into the Tertiary land surface. The clays are dominated by smectite, which has a high cation exchange capacity. The cation ratios for calcium, magnesium and sodium were 40:40:20. The Plains are very flat and surface drainage channels are often discontinuous and ill defined. The top 2 m of silty clay has weathered to a self-mulching soil (black cotton soil) that has exceptionally high agricultural productivity (Fig. 1b). The dominant mode of deposition in this unit was fluvial although there is evidence for the development of saline lakes and evaporites at various stages in the past 30,000 years (Timms and Acworth 2002).

The Yarramanbah nested piezometer site is located in the constricted outlet of the Yarramanbah Creek subcatchment, on the southern edge of the Liverpool Plains (Figs. 1a, b). The saturated zone at this site is generally less than a metre or two below ground surface (Timms et al. 2001). Uniform silty clays were identified at four cored boreholes to 10 m depth at a site approximately 2,000 m farther up the valley at Claremount (Fig. 1a) (Acworth and Beasley 1998). The mean clay content was 69.8% (number = 95; standard deviation = 10.4) and the mean silt content was 25.6% (number = 95; standard deviation = 37.2).

Conductivity cone penetration testing (CCPT) to a depth of 16 m, at 14 locations approximately 100 m apart extending along the unsealed road (Fig. 1a) from the piezometer site at Yarramanbah (shown by a cross on Fig. 1a) to the southern edge of the valley close to Connamara, indicated that the thick clay sequence was laterally extensive (Weisner and Acworth 1999). A stiff to very stiff clay, lightly overconsolidated below 3 m, was identified on the basis of the CCPT data. Shallow (<10 m depth) piezometers installed in the clay along the unsealed road to the south of the Yarramanbah site have fluid electrical conductivities (EC) of >20,000 μS/cm and an electrical induction log in a borehole 1,200 m to the south has a maximum bulk EC of >1,000 mS/m. These observations indicate that the clay contains significant quantities of salt. The continuity of the clay was established by an electrical image (Acworth 1999) collected at the site. A base electrode separation of 5 m was used with two cables each containing 25 cores. The image line length was then 245 m.

Drilling and piezometer installation

In order to monitor groundwater levels and quality variability with depth at the Yarramanbah site, a series of piezometers was initially installed above and below an existing piezometer 30061 (intake screen 50–55 m depth) by the Department of Water Resources. Four piezometers were installed in three holes drilled by rotary mud rig in February 1997. In anticipation of groundwater in the deep alluvium having a flowing artesian head, piezometers 40822-3 (87 m depth) and 40822-4 (108 m depth) were installed in separate holes, each with triple 2-m-thick bentonite seals precluding flow around each 60 mm Class 12 PVC pipe. Dual piezometers 40822-1 and 40822-2 were installed in a third hole screened between 15 and 17 m and between 28 and 30 m depth within the clay sequence. As a precaution against annular leakage, dual bentonite seals were placed between the intake screens. Construction details are shown in Table 1.

The top of each of these piezometers was completed with steel casing 1.2 m above ground to prevent ingress of flood water. A thick cement pad eliminated leakage of surface water along the clay-casing interface, a common design problem for piezometers in swelling smectite clay (Timms et al. 2001). Piezometers 408221 and 30061 are seen in the foreground in Fig. 1b. Flood debris can be seen hanging on the fence from floods in 1998. The floods overtopped the older piezometer 30061.

Cuttings returned by the rotary drill rig were catalogued in 1 m intervals for the clay sequence, and a bulk sample from each interval was retained for laboratory analysis. While analysis of physical parameters and detailed porewater chemistry was not possible on these disturbed samples, some information was gained from chloride analysis.

Downhole geophysical logging

Gamma ray activity and bulk electrical conductivity (EC_a) were obtained using GEONICS EM-39 geophysical logging equipment (McNeil 1986). The gamma ray activity was run at a speed of 0.015 m/s using a time constant of 12 s. Data were sampled at 25 mm depth increments.

Characteristics of sediment samples

Duplicate clay samples obtained within 0.1 m of the ground surface were analysed for cation exchange and carbon content. The analyses were carried out by Enviro-Test Labs, Saskatoon, Canada (J. Hendry, University of Saskatchewan, personal communication). Auger cuttings from 1 m intervals were recovered during drilling for laboratory analysis of soluble chloride at the Department of Land and Water Conservation (DLWC) Gunnedah Research Centre (Fig. 1a). Soils were dried at 105°C and prepared as 1:5 solutions for titration against silver nitrate.

Water-level monitoring

Groundwater levels in the five piezometers were automatically logged on an hourly or 6-hourly basis between February 1997 and April 2001. These data were verified and augmented by manual groundwater level measurements.

A dual channel logger (HYDROKIT), fitted with a DRUCK pressure transducer was deployed above the screen intake of piezometer 30061. These pressure

DOI 10.1007/s10040-005-0436-7

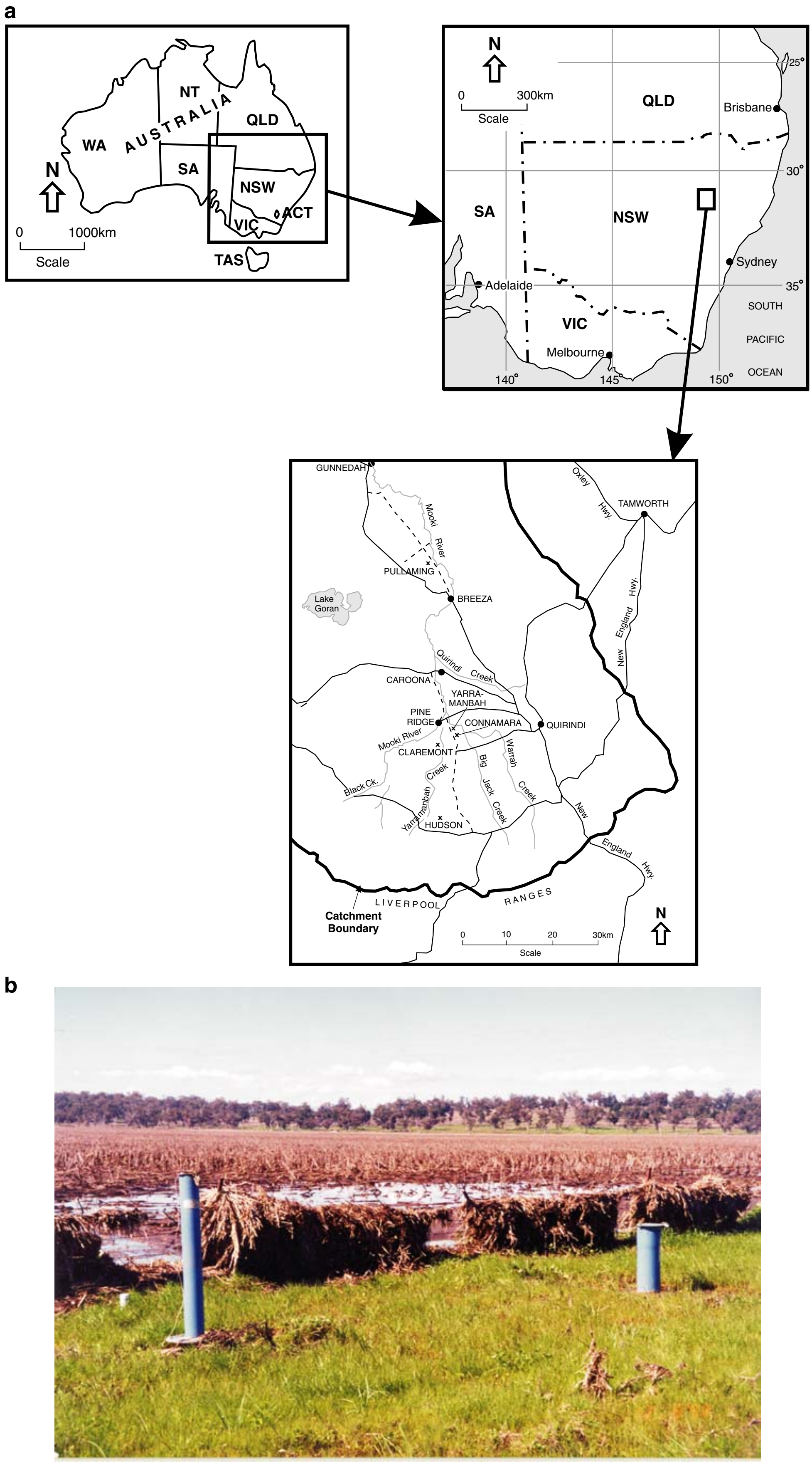

Fig. 1 **a** Location map showing the position of the Liverpool Plains within the Murray-Darling Basin. The location of investigations at Yarramanbah, Connamarra, Claremont and Pullaming are indicated by crosses. **b** Plate showing piezometers 40822-1 and 30061 in the foreground. Note the flood debris attached to the fence from the 1998 floods when standing water remained for several weeks at this site. The ridge line in the distance is approximately 2 km away

DOI 10.1007/s10040-005-0436-7

Table 1 Piezometer installation details for the Yarramanbah site. The ground elevation at the site is 312 m ASL

Piezometer	Installation date	Screen intake (m bgl)	Total depth (m bgl)	Pipe height (m above ground)	Pipe diam (mm)	Lithology
40822-1	2-May-97	15–17	19	1.18	50.8	Alluvial clay
40822-2	2-May-97	28–30	30	1.18	50.8	Alluvial clay
30061	1970's	50–55	55	0.72	63.5	Alluvial gravel
40822-3	2-Apr-97	80–86	87	1.22	63.5	Alluvial gravel
40822-4	27-Jan-97	102–108	108	1.17	63.5	Basalt bedrock

transducers were vented to the surface, so groundwater levels are differentiated from atmospheric pressures. Piezometer series 40822 were each equipped with a WESDATA electrical capacitance groundwater level logger. Though not as precise as loggers equipped with pressure transducers (S. Heidenreich, personal communication, Department of Infrastructure Planning and Natural Resources, New South Wales), the water level drift of these units was minimised by regular calibration against observed levels measured with a measuring tape.

Atmospheric pressure was monitored and recorded automatically by the second channel of the HYDROKIT logger. Data were collected for two periods. Between May 1998 and February 1999 half-hourly atmospheric pressure was recorded at the Claremont site (Fig. 1a). Later the equipment was moved to the Pullaming site (Fig. 1a) where it recorded hourly atmospheric pressure between November 1999 and March 2000. Daily rainfall data was collected by New South Wales Agriculture at the Connamara site (Fig. 1a).

Groundwater salinity

Water samples were taken on nine occasions from the nested piezometers and from a shallow piezometer installed by hand augering. A KENT fluid EC probe, positioned near the screen depth in bore 30061, was connected to the HYDROKIT logger between July 1998 and November 2000. This EC probe was calibrated against in situ groundwater on deployment, and the results verified on two occasions by sampling with a GRUNDFOS submersible pump. Groundwater salinity values were used to determine the density of the water column within the piezometers and to correct the observed environmental heads h_s, to reference fresh water heads, using:

$$h = (1 - \alpha)h_s - \alpha z \qquad (14)$$

where h is the reference hydraulic head, α is the ratio of densities $(\rho_s - \rho)/\rho$, with ρ_s the density of the water column and ρ the fresh water density. The elevation of the piezometer intake above datum is z. The fluid density was estimated from the fluid EC using the relationship given by Stuyfzand (1993).

Analysis of hydraulic properties

The observed data were analysed to derive the matrix properties BE, S_s, D, and the penetration depth C using the equations defined above. The vertical hydraulic gradient between the clay sequence and the confined aquifer was determined on the basis of corrected water heads at each observation point and the volume flux calculated using Darcy's Law.

Results

Electrical imaging of strata

The results of the electrical imaging are shown in Fig. 2. The piezometer nest is located at the center of the image. A laterally continuous zone of high bulk electrical conductivity (>650 mS/m) is seen above 10 m depth, with conductivities falling to <250 mS/m below 30 m depth.

Borehole geophysics

The results of natural gamma-ray and bulk electrical conductivity (EC) logging of the deepest piezometer (40822-4) are shown in Fig. 3. Logging of drill samples indicated black organic-rich clay within the top 1 m contrasted with brown clay between 1 and 8 m depth, and light brown clay between 8 and 39 m depth respectively. The proportion of lithic fragments increased between 33 and 39 m depth. The

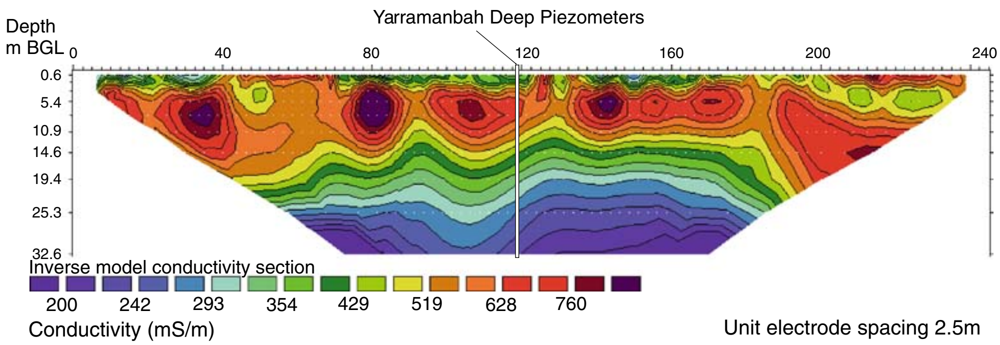

Fig. 2 Electrical image through the Yarramanbah site showing the high values of bulk electrical conductivity associated with the smectite dominated clays containing variable quantities of salt

DOI 10.1007/s10040-005-0436-7

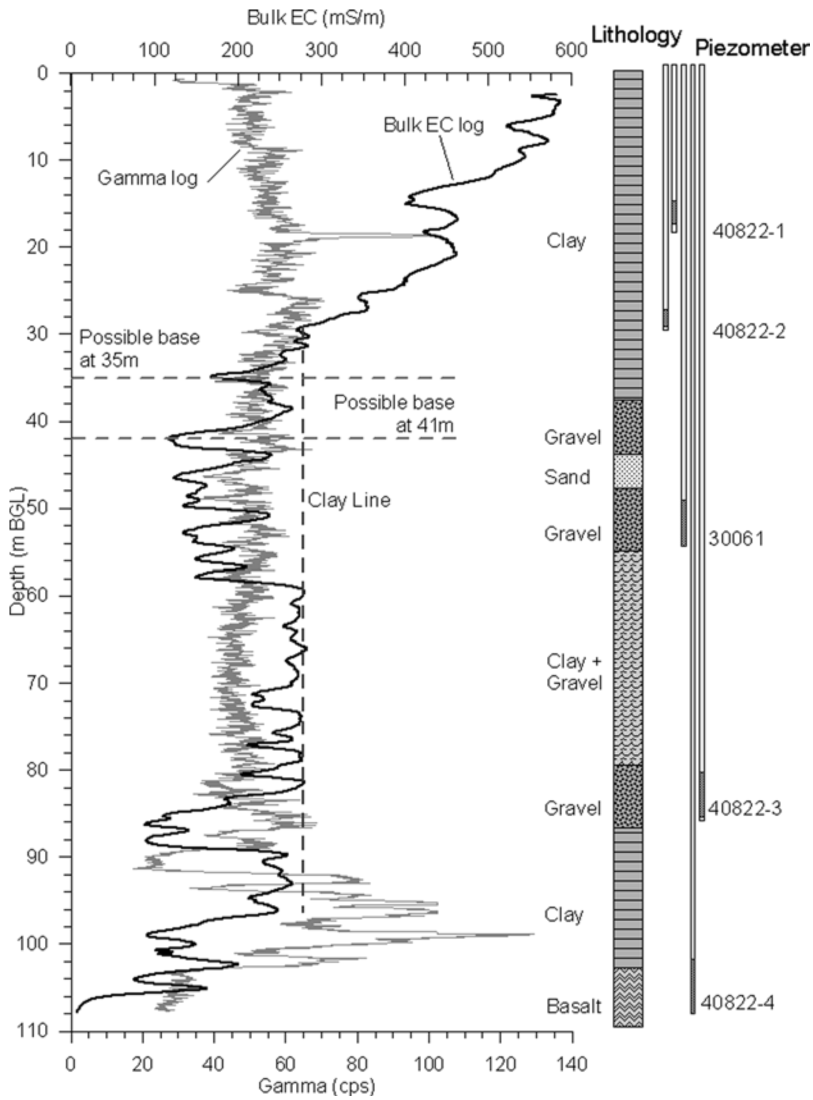

Fig. 3 Vertical profile at the Yarramanbah site with geophysical downhole logs of natural gamma and bulk EC (EM39) shown with the piezometer depths

geophysical logs show the change from brown clay to light brown clay at 8.6 m with the gamma activity increasing and the bulk EC decreasing. The brown clay has the higher bulk electrical conductivity and is the same unit indicated on the electrical image. A sharp gamma anomaly occurs at 18.6 m. This is the same depth as an inferred evaporite layer at the Pullaming site (Fig. 1a) to the north (Timms and Acworth. 2002). A further gamma-ray anomaly occurs at 26 m with a marked increase in activity.

The base of the clay was logged at 39 m. There is a bulk EC minimum at 41 m but it could be argued that the major change in response begins at 35 m. There is a continuous decrease in bulk EC from 600 mS/m close to the surface, to 120 mS/m at 41 m. The bulk EC log shows indications of layers comprising different salt loads. Piezometers 40822-1 and 40822-2 are clearly completed within the clay aquitard. For the purposes of this paper, the base of the aquitard has been taken as the first occurrence of sands indicated on the gamma log at 39 m.

The geological log derived from inspection of drill returns indicated fine well-sorted gravel comprised of basalt and chert fragments between 39 and 44 m with poorly sorted gravel between 48 and 54 m. Between 41 and 58 m a sequence of alternating low and high conductivity values indicates layers of sand/gravel with interlayered clays. There is a particularly clear sequence of clay interlayers below 58 m with a bulk EC value of approximately 280 mS/m indicating a deeper aquitard unit beyond the focus of this paper. The intercalated sand/gravel and clay layers between 41 and 58 m form the confined aquifer monitored by Bore 30061.

Analysis of clay sequence

Analysis of the clay sequence was limited to samples obtained near the surface and cuttings from rotary drill and hand auger installations. Soluble chloride content was relatively low in the top 1 m (4.7 meq/100 g), reached a maximum of 27 meq/100 g between 1 and 2 m and declined to levels below detection at 29–30 m. The linear decrease of soluble chloride with depth correlated well with the bulk EC log (Fig. 3).

 DOI 10.1007/s10040-005-0436-7

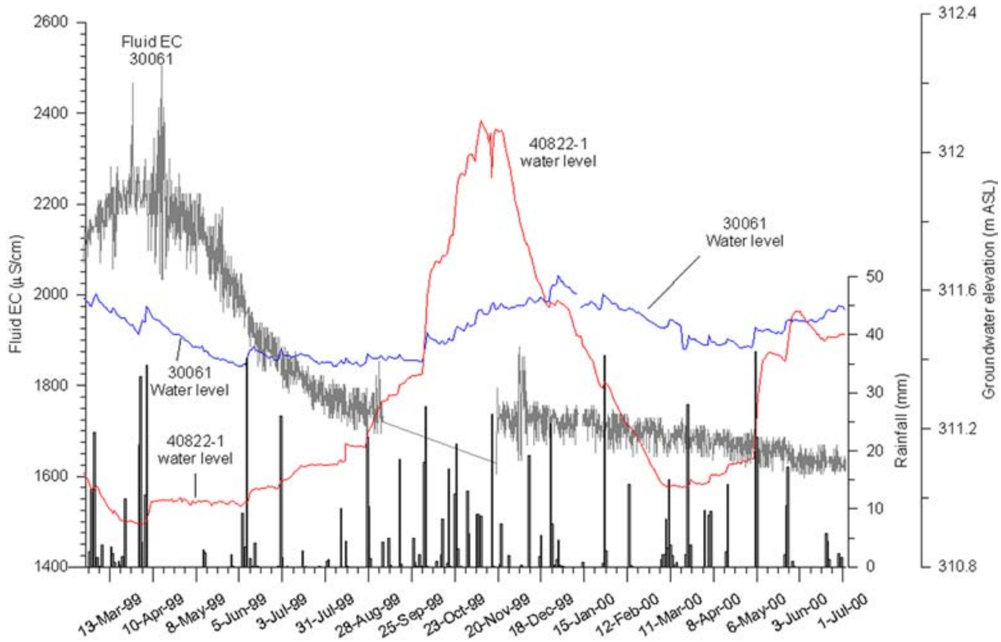

Fig. 4 Fluid electrical conductivity, daily rainfall and water levels for piezometers 40822-1 and 30061 for the period March 1999 to July 2000. The major ticks on the x-axis refer to the dates shown. Minor ticks represent weekly intervals

Cation exchange properties and carbon content were determined for the top of the clay sequence (J. Hendry, University of Saskatchewan, personal communication). The cation exchange capacity was 67.1 meq/100 g. The exchange sites on the clay were dominated by Ca and Mg ions (34.5 and 37.9 meq/100 g respectively) with a small proportion occupied by Na and K ions (1.5 and 1.3 meq/100 g respectively). Total carbon (TC) was 1.4%, of which 1.3% was organic (TOC) and 0.07% inorganic (TIC). Analysis results from a duplicate sample showed little variation.

Groundwater salinity

Groundwater salinity was highest near the surface and gradually decreased with depth through the clay aquitard, correlating with bulk EC estimated from the electrical image (Fig. 2) and the borehole logging (Fig. 3). Small temporal variations were observed in the clay sequence (40822-1: mean 11,910 μS/cm, standard deviation = 613, number = 6; 40822-2: mean = 10,550 μS/cm, standard deviation = 527, number = 5). Groundwater salinity within the confined aquifer was relatively low (mean = 2,079 μS/cm, standard deviation = 520, number = 9).

Temporal variability was considered to be a valid feature since standard deviations of this data set were larger than the estimated measurement error. To confirm this, fluid EC measurements in the confined aquifer (30061) were recorded every four hours between Aug-98 and Nov-00. Part of this data set is shown in Fig. 4, with daily rainfall and the groundwater levels for piezometers 40822-1 and 30061. Fluid EC decreased rapidly from more than 3,600 μS/cm in August 1998, several days after a flood peak (not shown in Fig. 4) when water spilled down inside the piezometer from the surface, to 2,200 μS/cm by the end of October 1998. After this time there has been a general steady decline with variations unrelated to the observed change in water levels.

Groundwater hydrographs and time lags

A detailed hydrograph, based on hourly and 6-hourly measurements in five piezometers between November 1996 and March 2001, is presented in Fig. 5. Five significant peaks in water level are shown. There was some uncertainty regarding measurement dates during the 1st and 2nd peaks, and the 5th peak was obscured by anomalous data from piezometer 30061. A brief period of anomalous data was also recorded for this piezometer after the 3rd peak.

The timing and magnitude of five major groundwater level peaks, relative to the available rainfall record, were analysed in detail (Table 2). Major rainfall events occurred episodically with an interval of 234–425 days (average 341 days) and, with one exception, occurred in the spring. In response to these events, groundwater pressures within the clay sequence typically increased by a metre within 6–102 days. Smaller pressure increases, of 0.17–0.49 m, were observed in the confined aquifer, with a time lag of between 49 and 72 days after peak levels in 40822-1 (Table 2). Figure 6 shows the data for piezometers 40822-1 and 30061 (exaggerated scale) for peaks 3 and 4. The time lag for peak 4 is 49 days (Table 2 and Fig. 6).

There are also immediate (zero phase) responses to individual rainfall events recorded in both piezometers and clearly seen in Fig. 6. The amplitude of the response varies between rainfall events and more detailed monitoring is required to resolve the variation.

Elastic storage estimates determined from the barometric efficiency

The barometric efficiency of piezometers 40822-2 and 30061 was determined by application of Eq. (8). The data for barometric pressure were plotted against water level measurements made at the same time. The scatter plot that results should contain repeated straight line segments that have a slope representing barometric efficiency (Eq. 8).

 DOI 10.1007/s10040-005-0436-7

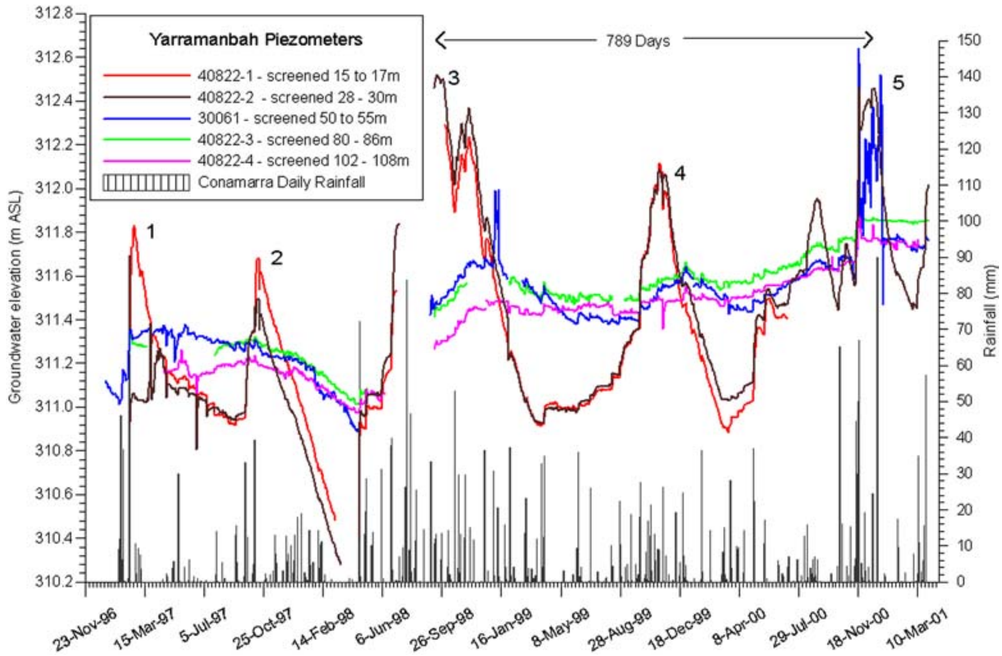

Fig. 5 Groundwater hydrograph for the nested piezometers at the Yarramanbah site, November 1996 to March 2001, based on 3-hourly data. Daily rainfall for the Connamara site is also shown. Major ticks refer to the dates shown, minor ticks show weekly intervals

Table 2 Major rainfall events related to peak groundwater levels at the Yarramanbah site, 1997–2000

Peak no. (Fig. 5)	Date that upward trend begins	Total rainfall during upward trend of 40822-1 hydrograph (m)	Peak to peak time interval (days)	Size of 40822-1 response Δh (m)	Time lag to reach 40822-1 maximum response (days)	Size of 30061 response Δh (m)	Lag between 40822-1 and 30061 peaks (days)
1	23-Feb-97	0.051	–	0.66	10	–	–
2	15–Oct-97	0.110	234	0.67	26	–	–
3	13-Sep-98	0.435	333	1.43	102	0.49	72
4	12-Nov-99	0.292	420	1.20	95	0.27	49
5	19-Nov-00	0.256	369	0.90	6	0.19	67

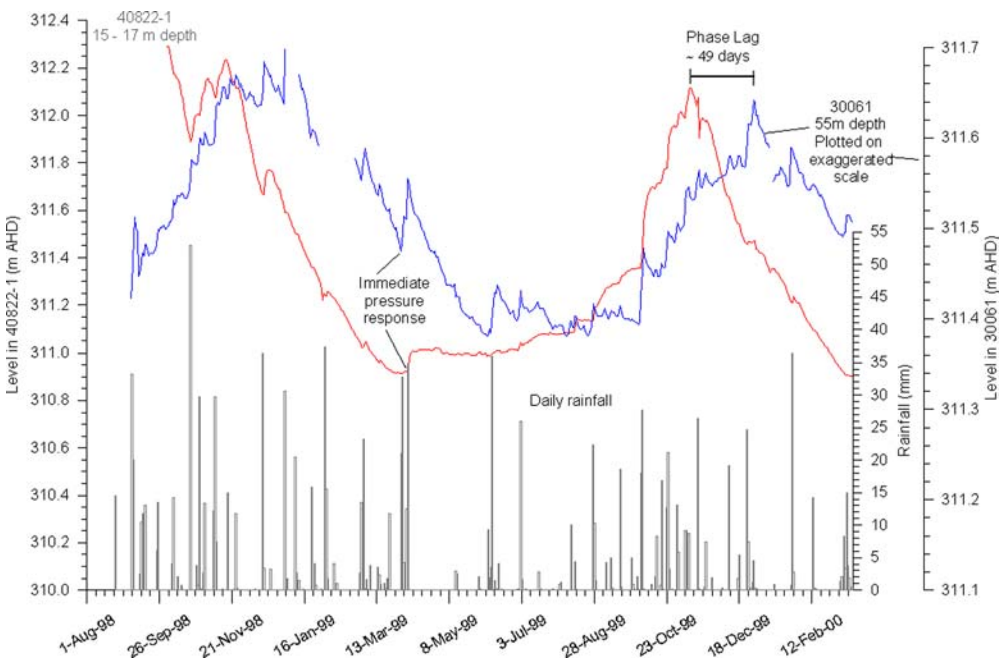

Fig. 6 Relationship between groundwater peaks within the clay sequence (40822-1, 15 m depth) and the confined aquifer (30061, 55 m depth, vertically exaggerated) illustrating time lags and response to major wet periods. Major ticks refer to the dates shown, minor ticks refer to weekly intervals

 DOI 10.1007/s10040-005-0436-7

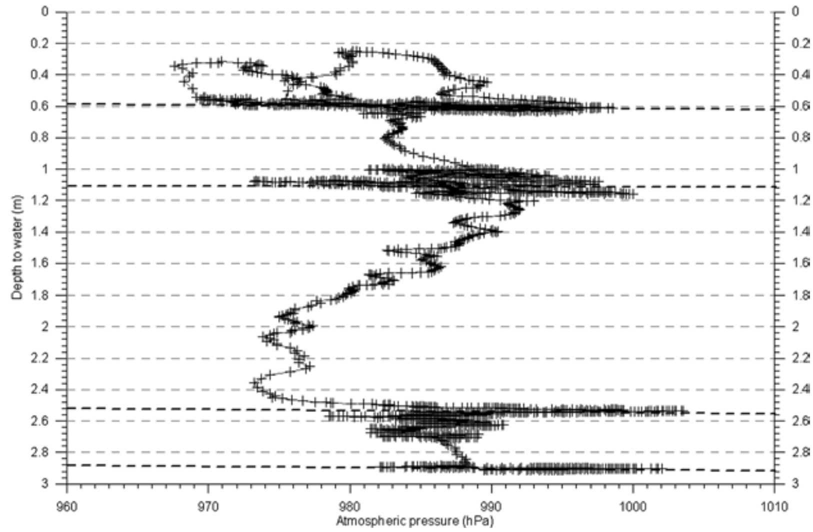

Fig. 7 Relationship between atmospheric pressure and water level in a shallow piezometer at Connamara. The family of straight dashed lines have a slope that represents the barometric efficiency of the piezometer. Deviations from the straight line segments represent recharge or evapotranspiration

Transverse departures from the straight lines would represent changes of water level not related to variation in barometric pressure (rainfall or transpiration). This approach is demonstrated with data measured every 30 min for a 10 m deep piezometer at the Claremont site (Fig. 1a). The results are presented in Fig. 7 and clearly show the relationship indicated by Eq. (8). The data for the piezometers at Yarramanbah were only measured every 6 h and the relationship is therefore less clear, however the scatter plot (not presented) shows evidence of similar straight line segments with an almost identical slope. This procedure makes maximum use of all the available data to produce a single value of barometric efficiency that can be considered to be representative of the formation.

The slope of the scatter plot data for the aquitard gave a barometric efficiency of 0.07. The BE for the aquifer was higher at 0.10. The porosity of the samples of silty clay from the cores (Acworth and Beasley 1998) was 0.567 (number = 15; standard deviation = 0.036). Assuming a temperature of 20°C and a β value of 4.58×10^{-10} ms^2/kg, Eq. (9) can be used to determine β_p. The clay compressibility (β_p) was determined to be 3.5×10^{-9} m s^2/kg. Equation (2) was then used to determine a value of 3.7×10^{-5} m^{-1} for S_s of the clay sequence (40822-1).

The S_s of the confined aquifer (30061) was lower. If a porosity of 0.15 is assumed for this poorly sorted material, then β_p of 6.2×10^{-10} m s^2/kg is given by Eq. (9) and a S_s value of 6.8×10^{-6} m^{-1} is produced by application of Eq. (2).

Hydraulic conductivity estimates

The K_v of the clay sequence was determined from observed amplitude and phase changes given the S_s value derived previously. Data for all five peaks is presented in Table 2, however, data for peak 4 has been used to demonstrate the analysis as this peak is clearly defined. A phase lag of 49 days (0.84 radians) was observed for the pressure fluctuation to be transmitted between 17 and 35 m depth (base of the intake screen in piezometer 40822-1 and the top of the confined aquifer respectively). The period (T) of this wave was calculated as half the time between peaks 3 and 5 (394.5 days). Substituting these values into Eq. (12), together with the estimated S_s value (3.7×10^{-5} m^{-1}) and the phase lag, gives a K_v value of 1.6×10^{-9} m/s.

In a similar manner, K_v was calculated from the changed amplitude observed at depth. An amplitude of 0.60 m (a_0) was observed at 17 m depth for peak 4 (Table 2 and Fig. 6). This had decreased to 0.16 m (a_z) inferred for the top of the confined aquifer at 35 m. The time lag for this difference to propagate was 49 days (Table 2). Based on the S_s value given above, substitution of values into Eq. (13) gives a K_v of 4.0×10^{-9} m/s for an assumed clay thickness of 18 m.

The average value of K_v from these two estimates is 2.8×10^{-9} m/s.

Penetration depth of an annual pressure fluctuation

Hydraulic diffusivity ($D=K_z/S_s$), based on an S_s value of 3.7×10^{-5} m^{-1} and the K_v of 2.8×10^{-9} yields a D value of 7.6×10^{-5} m^2/s. The penetration depth of an annual pore pressure fluctuation through the clay sequence was then determined by substituting the above parameters in Eq. (11). A penetration depth (4% of original amplitude based on Eq. (10) of 90 m was calculated for the derived value of hydraulic diffusivity. It is of interest to note that if a higher value of K_v is used, such as 8.7×10^{-6} m/s (derived from a simple average of the 15 falling head test results reported by Acworth and Beasley 1998), then a C value of 5017 m is calculated. This is clearly inappropriate given the observation that the deepest piezometer at the site shows a very subdued response to the approximately annual cycle of loading in the overlying clay (Fig. 5).

Hydraulic gradient and groundwater flux

The hydraulic heads were corrected to reference freshwater heads using Eq. (14). The daily values of fluid EC were used

 DOI 10.1007/s10040-005-0436-7

to calculate the equivalent head in the deeper piezometer. Only five values of fluid EC were available for the shallow piezometer and the average of these five were used for the density correction. The correction for 36001 was minor (0.088 m) but an average correction for the clay piezometer was 0.180 m. There are clearly periods of time when the gradient between the piezometers reverses (Fig. 5) but the total mass flux over the period is close to zero.

Discussion

No crops are grown during the winter months in this part of the Liverpool Plains and the fields are left bare. Evapotranspiration is therefore very low and rain falling during this time is effective in increasing soil moisture content. With the exception of the first peak (Fig. 5, Table 2), hydrograph peaks in the clay all occur in the winter or spring months with a rapid decline as summer evapotranspiration occurs. The smectite-dominated soils store large quantities of water and the local farmers wait until soil moisture levels are sufficient before planting crops. The accumulated mass of rainfall from periods of wet weather raises the shallow water table in the clays, increasing the water pressure at depth.

There is a clear and marked difference between the response in the clay aquitards and that in the underlying aquifers. As the net mass flux between aquitard and aquifer has been shown to be insignificant then the response of the whole system is taken to indicate the movement of a pressure pulse through the ground originating in the seasonal wetting and drying of the soil profile. There are clear phase lags and amplitude changes that occur in the hydrographs between the top of the clay and the underlying aquifer (Fig. 5). While individual piezometers installed throughout the clay would be preferable, the pressure change in the aquifer at 55 m has been taken as indicative of the pressure change at the base of the clay at 35 m. The aquifers are thought to be isolated lenses of sand and gravel at this depth and do not normally receive recharge. For this reason, the pressure at this depth in considered to be representative of conditions at the base of the clay.

Estimated S_s values for the clay sequence were relatively low (3.7×10^{-5} m^{-1}) when compared to typical S_s values for medium-hard clay of between 6.8×10^{-4} and 1.3×10^{-3} m^{-1} (Domenico and Schwartz 1997). The relatively low S_s for this clay sequence may be attributed to the fact that the clay is slightly overconsolidated (T. Wiesner, Douglas Partners, personal communication). CPPT profiles at the Yarramanbah site detected 'stiff to very stiff' clay below 8 m depth (Wiesner and Acworth 1999). A hard silty clay was also detected towards the edge of the valley at 16 m, the maximum CPPT penetration depth, and possibly correlates with the gamma-ray anomaly at 18 m beneath the site and an evaporite layer at a similar site at Pullaming (Fig. 1a).

The low S_s values are supported in a review by Van der Kamp (2001) who concluded that reliable S_s values were best derived either from multi-piezometer tests or based on barometric efficiency and measured porosity. Van der Kamp found that S_s values derived from constant head piezometer tests and consolidation tests carried out on core recovered during drilling were generally several orders of magnitude higher than the alternative in situ measurements used in this analysis. The analysis of the data set presented here supports Van der Kamp's analysis.

Low permeability media are defined by a K_v of 10^{-8} m/s (Neuzil 1986) and compacted clay liners are required to have a $K_v<10^{-9}$ m/s to be considered as practically impermeable (Rowe et al. 1995). The estimated K_v values (average of 2.8×10^{-9} m/s) for the Yarramanbah site, fall within these definitions of low permeability media. It is important to note that the estimates are at least two orders of magnitude lower than previous measurements for saturated clay below the near-surface transitional fractured zone. For instance, K_v values in the order of 10^{-6} m/s were measured by falling head tests on disturbed core that was obtained from 6.9 to 9.1 m depth at the Claremont site (Acworth and Beasley 1998) 2 km to the south.

The K_v of unoxidized clay tills reported from North America have values in the order of 10^{-11} m/s, in contrast with K_v of 10^{-9} m/s for shallow fractured tills (Shaw and Hendry 1998; Keller et al. 1989). Flow velocity through such low permeability unoxidized clay till is limited to 0.5–0.8 m per thousand years. The Yarramanbah K_v data are of a similar magnitude to the fractured tills of Keller et al. (1989), yet there is evidence from the chemical analyses available that only diffusion has occurred. The mode of occurrence of the clay tills is significantly different from that of the Yarramanbah clay sequence and it is probable that the higher value of K_v does not represent fracturing but may be attributed to the structure of Ca-smectite clay.

A number of physical and chemical factors determine bulk clay permeability such as fracturing, the proportion of silt and sand, consolidation history of the deposit, cementing and exchangeable cations and porewater chemistry (Mitchell 1993). At natural water contents, K_v typically ranges between 10^{-7} and 10^{-10} m/s, and the permeability of smectite clays is typically lower than for illite and kaolinite. The relatively small particle size of smectite means that mechanical compaction can significantly reduce permeability (Rowe et al. 1995). In natural environments however, the permeability of clay increases with a decrease in exchangeable Na^+ (increased Ca^{2+} and Mg^{2+}) and an increase in pore water solute concentration (Moutier et al. 1998).

Swelling Ca-smectite clay is comprised of structural domains, defined as associations of several quasi-crystals each of which contains 5–10 clay platelets (Quirk and Alymore 1971). The relatively high permeability of such clay is attributed to heterogenous pore size distribution, made up of intra-domain pores (2–5 nm), inter-domain pores (1–2 μm) and transmission pores (>50 μm). For example, Choi and Oscarson (1996) observed a bimodal pore size distribution for Ca-smectite clay, with large pores that were absent in Na-smectite. Since permeability is proportional to the square of pore diameter, such a

 DOI 10.1007/s10040-005-0436-7

structure can significantly increase K_v. Based upon the available literature, the results obtained at the Yarramanbah site are not unusual and confirm the observation that the previous measurements of K_v, carried out on disturbed core material, may have been influenced by fracturing.

The fluid EC data from bore 30061 in the gravel at 55 m shows considerable variation on a day-to-day basis as well as a longer term decline. Surface water contaminated with salt entered the top of the bore during a flood in August 1998. The debris from this flood can be seen hanging on the fence close to the bore in Fig. 1b. Shortly after this time, the fluid EC logger was installed and initially recorded a value of 3,500 μS/cm. This decayed rapidly and then more gently to read 2,100 μS/cm at the beginning of March 1999 (Fig. 4). The lack of any relationship between the fluid EC change in the aquifer and rainfall is apparent. The reduction in fluid EC is considered to be due to diffusion of the contaminated mass of water into the aquifer.

An immediate response to individual rainfall events in both the clay and the aquifer is shown in Figs. 4 and 6. This response is not the result of advective transport through the clay as the hydraulic conductivity is too low. The synchronicity of the response between 40822-1 and 408221-2 (Fig. 5) and the lack of any response to the fluid EC beneath the clay (Fig. 4) is further evidence that advective transport is extremely slow and could not be the cause.

The probable explanation lies in Eq. (3) where the loading efficiency (γ) of the clay is described. The loading efficiency has been calculated from Eq. (4) to be 0.93. The response to the extra weight of the soil caused by the rainfall is transmitted instantaneously through the aquitard and down into the underlying aquifers. Close inspection of Fig. 5 reveals a small but immediate response to the extra load from rainfall in all the piezometers, including the aquifer screened between 102 and 108 m. The hydrographs therefore provide a good example of a loading response as described by Van der Kamp and Maathuis (1991) as well as the diffusion of a pressure change through the system caused by a changing boundary condition at the surface.

Summary and conclusions

A major difficulty in the interpretation of water level data in clays has been lack of appreciation of the significance of clay compressibility and the impact of this on water pressures. In this paper, detailed monitoring of a nest of piezometers installed in a thick clay sequence on the Liverpool Plains of NSW is described. The piezometers were monitored either hourly or 6-hourly over a 4.5-year period. Analysis of the data has provided better information on hydraulic mechanisms within the thick clay sequence, with significant implications for hydrogeological interpretation and resource management.

Pore water pressures near the top of a smectite clay sequence increase in response to major rainfall events of several days duration and propagate downwards through a 30 to 40-m-thick massive smectite clay sequence by diffusion. The presence of high salt loads towards the top of the clay, combined with low salinity water in the underlying aquifer, strongly indicate that very little advective transport occurs and the increase in water level in the aquifer is mainly a pressure response. The amplitude of pore water pressure (P_w) change decays with increasing depth through the clay and aquifer sequence. The typical response to a major wet period resulted in P_w of $+0.98$ m near the top of the clay sequence, and was later reflected by P_w+0.17 m within the confined aquifer. For five major rainfall periods, a time lag of 49–72 days elapsed between maximum pore water pressure near the top and close to the base of the clay sequence. The analysis of the pore-pressure changes through the clay has provided estimates of specific storage and hydraulic conductivity of 3.7×10^{-5} m^{-1} and 2.8×10^{-9} m/s respectfully. The value of hydraulic conductivity is lower than other estimates by several orders of magnitude but is in agreement with analyses of similar systems by Van der Kamp (2001).

The response to individual rainfall events occurs with no phase lag between the various piezometers and can be detected at 100 m depth at the base of the alluvial sequence. This response varies in amplitude and is the result of different loading efficiencies as predicted by analysis of the effective stress distribution. The pressure response due to different loads at the surface can be used to quantify evapotranspiration, as described by Bardsley and Campbell (1994) and Van der Kamp and Schmidt (1997), and further detailed monitoring is planned.

The long period of detailed monitoring has allowed the separation of loading effects from the diffusion of hydraulic head downwards through the succession that is the result of changing the elevation of the water table at the surface. The analysis has shown that at this site very little mass flux occurs vertically and this has provided an explanation for the observation of very poor quality old water in the clay that appears to respond rapidly to rainfall recharge.

Acknowledgements The first author is indebted to the early training support of CSIRO Land and Water, particularly Dr Joe Walker and Peter Richardson, who provided additional groundwater level loggers to augment UNSW equipment for an extended monitoring period. The NSW Department of Land and Water Conservation funded installation of the nested piezometer site and ongoing support was given by Dawit Berhane, David Dight and others of the Department's Gunnedah Research Centre. Roland Thomas provided technical assistance while a summer student at the UNSW Water Research Laboratory. Climate data was supplied by Rick Young (NSW Agriculture). Assistance was provided by Terry Wiesner of Douglas and Partners and Prof Jim Hendry, University of Saskatchewan, Canada, and is acknowledged with thanks. The comments of Dr. Garth Van der Kamp of the National Hydrology Institute, Canada, led to improvements in the manuscript. Dr Chris Neuzil (USGS) and a second reviewer are acknowledged for their constructive comments.

 DOI 10.1007/s10040-005-0436-7

References

Acworth RI (1999) Investigation of Dryland Salinity using the Electrical Image Method. Aust J Soil Res 37:623–36

Acworth RI, Beasley R (1998) Investigation of EM31 Anomalies at Yarramanbah/Pump Station Creek on the Liverpool Plains of New South Wales, WRL Res Rep No. 195

Bardsley WE, Campbell DI (1994) A new method for measuring near-surface moisture budgets in hydrological systems. J Hydrol 154:245–254

Boldt-Leppin BEJ, Hendry MJ (2003) Application of harmonic analysis of water levels to determine vertical hydraulic conductivities in clay-rich aquitards. Ground Water

Choi JW, Oscarson DW (1996) Diffusive transport through compacted Na- and Ca- bentonite. J Contam Hydrol 22:189–202

Domenico PA, Schwartz FW (1997) Physical and chemical hydrogeology, 2nd edn. Wiley, New York

Jacob CE (1940) On the flow of water in an elastic artesian aquifer. Trans Am Geophys Union 21:574–586

Keller CK, Van der Kamp G, Cherry JA (1989) A multi-scale study of the permeability of a thick clayey till. Water Resour Res 25:11

McNeill JD (1986) Geonics EM-39 borehole conductivity meter: theory of operation, Mississauga, Ontario, Canada., Geonics Ltd., Technical Note TN-20, p. 9

Mitchell JK (1993) Fundamentals of soil behaviour, 2nd edn. Wiley, New York

Moutier M, Shainberg I, Levy GJ (1998) Hydraulic gradient, aging and water quality effects on hydraulic conductivity of a vertisol. Soil Sci Am J 62:1488–1496

Neuzil CD (1986) Groundwater flow in low-permeability environments. Water Resour Res 22:8

Neuzil CD (2003) Hydromechanical coupling in geological processes. Hydrogeol J 111:41–83

Rowe RK, Quigley RM, Booker JR (1995) Clayey barrier systems for waste disposal facilities. E & FN Spon, London, 390 pp

Terzaghi K, Peck R (1948) Soil mechanics in engineering practice. Wiley, New York

Timms W, Acworth RI, Berhane D (2001) Shallow groundwater dynamics in smectite dominated clay on the Liverpool Plains of New South Wales. Aust J Soil Res 39:203–218

Timms WA, Acworth RI (2002) Induced leakage due to groundwater pumping and flood irrigation at the Pullaming Agricultural Field Station, Liverpool Plains, Water Research Laboratory Research Report No. 208 available at http://www.wrl.unsw.edu.au/research

Quirk JP, Aylmore AG (1971) Domains and quasi-crystalline regions in clay systems. Soil Sci Soc Am J 35:652–654

Shaw RJ, Hendry MJ (1998) Hydrogeology of a thick clay till and Cretaceous clay sequence, Saskatchewan, Canada. Can Geotech J 35:1041–1052

Stuyfzand PJ (1993) Hydrochemistry and hydrology of the coastal dune area of the Western Netherlands. KIWA, Afd. Onderzoek & Advies–I11

Van der Kamp G, Gale JE (1983) Theory of earth tide and barometric effects in porous formations with compressible grains. Water Resour Res 19:2

Van der Kamp G, Maathuis H (1991) Annual fluctuations of groundwater levels as a result of loading by surface moisture. J Hydrol 127:137–152

Van der Kamp G, Schmidt R (1997) Monitoring of total soil moisture on a scale of hectares using groundwater piezometers. Geophys Res Lett 24:6

Van der Kamp G (2001) Methods for determining the in situ hydraulic conductivity of shallow aquitards—an overview. Hydrogeol J 9:5–16

Wiesner T, Acworth RI (1999) Groundwater contamination investigation using CCPTs: Water 99 Joint Congress, 6–8 July, pp. 801–806

 DOI 10.1007/s10040-005-0436-7