Monitoring the moisture balance of a boreal aspen forest using a deep groundwater piezometer

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Abstract

This study evaluates the piezometric weighing lysimeter method of van der Kamp and Schmidt, based on two annual cycles of measurements in a mature boreal aspen forest in western Canada. The method is used to infer components of the moisture balance on a scale of hectares. The method is based on accurate measurement of groundwater pore pressure in thick clay formations, and depends on the principle that changes of mass load result in changes of groundwater pressure. The resulting data are similar to those obtained by conventional weighing lysimeters, but on a much larger scale and with no significant disturbance of the site. The groundwater pressure was measured at the 34.6 m depth, effectively integrating the mass loading over an area of about 10 ha.

Cumulative evapotranspiration and precipitation, as inferred from the groundwater pressure time series, showed good agreement with direct measurements. The piezometric and gauged estimates for cumulative precipitation agreed to within 5%. The piezometer also captured the main features of the diurnal and seasonal cycles of evapotranspiration. However, the estimation of evapotranspiration was complicated by the presence of a seasonal background cycle in the groundwater pressure time series, the result of a higher-than-ideal permeability at this site. An initial attempt to estimate the seasonal background cycle from the observed pressure changes during quiescent periods met with moderate success. Work is in progress to estimate the background cycle using a more physically-based approach. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

In an earlier paper, van der Kamp and Schmidt (1997) reported a novel approach for continuous measurement of the moisture balance on a scale of hectares, which we will call the piezometric weighing lysimeter method. This method requires an accurate measurement of groundwater pore pressure in thick, low-permeability, clay formations, and is based on the principle that changes of mass load cause changes in groundwater pressure. The resulting piezometer data are similar to those obtained by conventional weighing lysimeters, but on a much larger scale and with no significant hydrologic disturbance of the site. van der Kamp and Schmidt suggested that the use of deep piezometers to infer the moisture balance would be particularly well suited to studies of forest hydrology, because the scale of the measurement allows observations of average moisture conditions over an area...
of tens of hectares. Deep piezometers may provide a cost-effective alternative for continuous, long-term measurement of the moisture balance.

The piezometric weighing lysimeter method of van der Kamp and Schmidt built on the earlier research of van der Kamp and Maathuis (1991) and Bardsley and Campbell (1994). These studies showed that seasonal changes of groundwater pressure in highly permeable, extensive aquifers could be related to the moisture balance above the piezometer. However, the relationship was complicated by the aquifers’ high permeability, which allowed transient horizontal and vertical flows. Such transient flows are complex and difficult to quantify. Consequently, the observed groundwater pressure changes could not be unambiguously related to moisture conditions near the ground surface. van der Kamp and Schmidt overcame this difficulty by analyzing data from aquitards (i.e., low-permeability formations), where daily and seasonal flow transients were negligibly small. They clearly demonstrated the usefulness of the piezometric weighing lysimeter method, but did not quantitatively evaluate it against independent moisture balance data.

This study evaluates the piezometric weighing lysimeter in a mature boreal aspen forest in central Saskatchewan, Canada, by comparison with measured precipitation and evapotranspiration. It demonstrates the use of the piezometric weighing lysimeter to estimate the moisture balance and its components on a scale of hectares. It also points out some ongoing challenges that must be overcome before the method can be more broadly applied.

2. Theory

For a more detailed description of the piezometric weighing lysimeter concept, the reader is referred to van der Kamp and Schmidt (1997). The underlying theory is described in van der Kamp and Gale (1983) and Rojstaczer (1988). Here we summarise the salient points.

Within extensive horizontal aquitards, the flow of groundwater may be assumed to be mainly vertical, and the equation that governs pressure transients can be written:

$$\frac{D \partial^2 p}{\partial z^2} = \frac{\partial p}{\partial t} - \gamma \frac{\partial \sigma}{\partial t}$$  \hspace{1cm} (1)

where $D$ (m$^2$/s$^{-1}$) is the hydraulic diffusivity, $p$ (Pa) is the groundwater pore pressure, $z$ (m) is depth, $t$ (s) is time, $\gamma$ is the loading efficiency and $\sigma$ (Pa) is the total mass load on the formation. The mass load $\sigma$ is equal to the mass of the overlying solids and water, plus atmospheric pressure. ‘Groundwater pressure’ as used here is equivalent to ‘hydraulic head’ as used in the groundwater literature (e.g. Freeze and Cherry, 1979). We will express both $p$ and $\sigma$ in terms of the equivalent height of water, in units of m or mm of water.

The loading efficiency $\gamma$ represents the proportion of a change of load that is borne by the water in the formation. Clay-rich aquitards have high compressibility compared to water, resulting in loading efficiencies only slightly less than 1.0. The value of $\gamma$ for a particular formation can be determined from its response to barometric loading.

The term on the left side of Eq. (1), involving $D$, represents the transient flow of groundwater. Transient groundwater flow occurs in response to two forcings: changes in water table depth and changes in mass load. The transient flow term in Eq. (1) is largest at the water table, but damps rapidly with depth in low-permeability materials such as clays. The effective penetration depth of the transient wave may be estimated as four times the damping depth $z_d$ (m), which in turn depends on the hydraulic diffusivity of the formation and the duration of the transient. (At four damping depths, 98% of the surface fluctuation is damped out.) For a sinusoidal fluctuation with period $T$ (s), the damping depth is:

$$z_d = \left( \frac{DT}{\pi} \right)^{0.5}$$ \hspace{1cm} (2)

In practical terms, the interior of an unfractured clay formation, with a value for $D$ of 10$^{-6}$ m$^2$/s$^{-1}$, has negligible transient flow at depths of 12.7 m and greater from the top or bottom of the clay over the annual cycle, and at depths of 0.6 m and greater over the diurnal cycle.

An inference can be drawn from Eqs. (1) and (2) that is the basis of the piezometric weighing lysimeter method, that changes in $p$ in the interior of thick unfractured clay formations reflect only changes in mass load $\sigma$, over both diurnal and annual cycles. Under such conditions, the measured changes of groundwater...
pressure provide a direct measurement of changes in total mass load. These changes are predominantly due to changes of atmospheric pressure and total moisture, including snow, above the measuring point.

However, many clays and similar formations are fractured to considerable depth and consequently have higher permeability and values of \( D \) much greater than \( 10^{-6} \text{ m}^2 \text{ s}^{-1} \) (e.g., Keller et al., 1988). The occurrence of fractures at depth cannot be predicted a priori. For fractured clays, much larger thicknesses are needed to isolate the groundwater pressure in the interior of the clay from pressure changes that occur at the top or bottom surface of the clay formation.

Eq. (1) may be rewritten as:

\[
\frac{\partial \rho}{\partial t} = \gamma \left( \frac{\partial \rho_a}{\partial t} + \frac{\partial \rho_e}{\partial t} + \mathcal{P} + E + R \right) + \beta \tag{3}
\]

where \( \rho_a \) is atmospheric pressure (expressed as an equivalent depth of water, in mm), \( \rho_e \) is the load associated with the earth tide dilatation (expressed as an equivalent depth of water, in mm), \( \mathcal{P} \) (mm h\(^{-1}\)) is the rate of precipitation, \( E \) (mm h\(^{-1}\)) is the rate of evapotranspiration, \( R \) (mm h\(^{-1}\)) is the rate of runoff and \( \beta \) (mm h\(^{-1}\)) is \( D \frac{\partial^2 \rho}{\partial z^2} \), the transient flow term from Eq. (1). We will refer to \( \beta \) as the seasonal background rate of change. Note the sign convention for \( E \) and \( R \); water losses from surface and soil are expressed as negative values.

The deep groundwater pore pressure effectively integrates the mass load at a scale of hectares. The response of the piezometer to changes of load near the ground surface falls off with horizontal distance away from the piezometer (van der Kamp and Schmidt, 1997), as:

\[
\Delta \rho = \frac{2}{3} \sigma_a \gamma (1 + v_u) \left[ 1 - \left( 1 + \frac{v_u^2}{z_p} \right)^{-1/2} \right]^{-1/2} \tag{4}
\]

where \( \Delta \rho \) is the pressure that results from the load within a circular area of radius \( r \) (m), \( \sigma_a \) is the uniform load per unit area, \( v_u \) is the undrained Poisson’s ratio of the formation (slightly less that 0.5 for clays) and \( z_p \) (m) is the depth of the piezometer. For the piezometer at the aspen site, with \( z_p = 34.6 \text{ m} \), Eq. (4) predicts that 29, 55, 80 and 90% of the piezometric load come from within radii of \( z_p, 2z_p, 5z_p \) and 10\( z_p \) (35, 69, 171 and 346 m), respectively, of the piezometer bore hole. The equivalent areas are 0.38, 1.5, 9.4 and 38 ha.

### 3. Methodology

#### 3.1. Field installation and instrumentation

This study evaluates the use of a piezometric weighing lysimeter at the Boreal Ecosystem Research and Monitoring Sites (BERMS) southern Old Aspen site (Black et al., 1996). BERMS is the field follow-on to the Boreal Ecosystem Atmosphere Study (BOREAS) (Sellers et al., 1995). The Old Aspen site is located in Prince Albert National Park, Saskatchewan, Canada (53.7°N, 106.2°W). In autumn 1996, we installed a deep groundwater piezometer at 34.6 m depth. The piezometer was sited in a mature aspen stand, 1.3 km from the Old Aspen flux tower. We chose the piezometer site because of its proximity to an access road where a conventional well drilling rig could be deployed. The forest stand around the flux tower is nearly uniform over a radius of 3.0 km and more. The stand characteristics at the piezometer site (mature aspen (Populus tremuloides L., with scattered Populus balsamifera L.) overstory with hazel (Corylus cornula Marsh.) understory are similar to those at the flux tower over approximately 90% of the piezometer’s area of loading. The remaining 10% of the piezometer’s area of loading is young aspen. The subsurface geology above the piezometer consists of mixed sands, gravel and glacial till to a depth of 20 m, and clay-rich glacial till below 20 m. The water table fluctuates between 3 and 4 m below ground surface. As discussed in Sections 4, 4.1 and 4.4 later, the hydraulic properties of this formation are adequate but not optimal for this method. Since installation, the piezometer has operated continuously.

Fig. 1 illustrates the piezometric weighing lysimeter concept and installation. We used two high-resolution vibrating-wire pressure transducers to make duplicate measurements of groundwater pressure throughout 1997 and 1998, and added a third, higher resolution, vibrating-wire pressure transducer in April 1998 (Table 1). Transducer #1 was buried at 34.6 m depth in a 0.15 m diameter, sand-filled cavity that extended from 33.5 to 35.7 m depth, and measured absolute pressure. Transducers #2 and #3 were connected via an oil-filled tube to a buried intake in the same sand-filled cavity, and measured the pressure in the oil-filled tube. The use of the oil-filled connecting tube added an offset to the pressure measurement, enabling the use of a
finer range transducer. It also kept Transducers #2 and
#3 accessible, in an access collar at 9.6 m depth. The
three transducers were deep enough that diurnal and
seasonal temperature changes were small, so that the
transducer’s temperature dependence did not degrade
the pressure measurement. After installation, the bore
hole was back-filled with bentonite, a clay mineral
that expands when wetted, to reseal the formation.
The transducer installations and data logging methods
stabilized in March 1997, when the most usable data
begin. A Setra barometer, housed in a heated hut at
the flux tower, was used to correct the groundwater
pressure for changes in atmospheric pressure.

The vibrating wire pressure transducers were logged
using a CR10 data logger with an AVW4 interface
(Campbell Scientific Inc.). The pressures were sam-
ples every 30 s and the means were logged every
30 min. The data logger resolution was 0.4 mm of wa-
ter. The nominal sensor resolutions were 8.8 mm for
Transducers #1 and #2, and 0.4 mm for Transducer #3.
However, we found in practice that Transducer #2 out-
performed this specification, and tracked Transducer #3
to well within 1 mm over the diurnal cycle.

Data from the three transducers were merged to cre-
ate one continuous time series. When their data were
good, we used Transducer #2 before May 1998 and
Transducer #3 after May 1998. Data segments were
rejected (twice in 2 years) when there was an obvious
excursion in the piezometer signal. At times, espe-
cially in late autumn and winter, the piezometer data
became unusually noisy. This increased noise level in
winter has been encountered in similar installations
elsewhere (van der Kamp and Schmidt, 1997), and is
thought to result from changes in horizontal stresses
near the ground surface in response to freezing and
thawing.

Evapotranspiration was measured continuously by
eddy correlation at the flux tower. Instrumentation
and signal processing details are given in Black et al.
(1996) and Chen et al. (1999).

Precipitation was measured year round by an accu-
mulating gauge (Belfort 5915, with an Alter shield),
with motor oil added to prevent water losses by evap-
oration. Rainfall was measured from spring to autumn
by a tipping bucket rain gauge (Texas Electronics
TE525M). Data were logged every half-hour. Precip-
itation was measured during both years at the flux
tower, and beginning in August 1997 at the piezome-
ter. At both sites, the gauges were sited in clearings,
at approximately one tree height from the clearing
edge. This geometry maximized snow catch efficiency
(Goodison, personal communication). A few gaps ex-
ist in the 1997 precipitation data from the flux tower
because of gauge disruption by a curious bear.

3.2. Data analysis

Before analysing the time series of groundwater
pressure, we removed two well-defined loading signals

<table>
<thead>
<tr>
<th>Transducer</th>
<th>Model</th>
<th>Venting</th>
<th>Full scale range</th>
<th>Nominal resolution (mm H2O)</th>
<th>Measurement</th>
</tr>
</thead>
<tbody>
<tr>
<td>#1</td>
<td>Geokon 4500H</td>
<td>Unvented</td>
<td>35 (50)</td>
<td>8.8</td>
<td>Absolute</td>
</tr>
<tr>
<td>#2</td>
<td>Geokon 4500H</td>
<td>Vented</td>
<td>35 (50)</td>
<td>8.8</td>
<td>Relative</td>
</tr>
<tr>
<td>#3</td>
<td>Geokon 4580-2</td>
<td>Vented</td>
<td>3.5 (5)</td>
<td>0.35</td>
<td>Relative</td>
</tr>
</tbody>
</table>

*a* Pressure in metres of water.

*b* The values in parenthesis denote pressure in lb in $^{-2}$. 
Fig. 2. Atmospheric and tidal adjustments to measured groundwater pore pressure at 34.6 m depth in a mature aspen forest in central Saskatchewan, September 1998. All values are in relative units, and are plotted with offsets to enable visual comparison.

from the raw piezometer output $p_t$ (Pa) — atmospheric pressure $p_a$ (Pa) and the load associated with earth tides $p_e$ (Pa). The earth tide dilatation was computed using the EarthTid program (Harrison, 1971), which required only site location, elevation and time. The resultant, adjusted groundwater pore pressure, which we will denote as $p^*$, was calculated for Transducer #1 (unvented) as:

$$p^* = p_t - \gamma(p_a + p_e)$$  \hspace{1cm} (5a)

and for Transducers #2 and #3 (vented to the atmosphere) as:

$$p^* = p_t - (1 - \gamma)p_a - \gamma p_e$$ \hspace{1cm} (5b)

The venting of pressure transducers removes the differential between atmospheric and groundwater pressure, so that the transducers measure only ($\gamma$) of the atmospheric load. Fig. 2 shows the magnitude and impact of these adjustments for the vented pressure transducers. The correction for atmospheric pressure provided an independent means to estimate the formation’s loading efficiency $\gamma$, which was found to be 0.91 at the aspen site. Earth tides contributed a complex component to the groundwater pressure that varied between 1 and 5 mm peak to peak over the diurnal and lunar cycles.

We also simplified Eq. (3) as:

$$\frac{\partial p^*}{\partial t} = \gamma(P + E) + \beta$$ \hspace{1cm} (6)

which combines Eq. (3) with Eqs. (5a) and (5b) and assumes $R$ to be zero. Considering the sandy nature of the soil, the low relief and the undisturbed forest vegetation, it is unlikely that any significant runoff occurred during the study period. If runoff did occur, the assumption of no runoff would cause $P$ to be underestimated by $R$ for runoff that occurred during precipitation events. Runoff that continued after precipitation events would be incorporated into $\beta$.

We decomposed the time rate of change of the adjusted groundwater pressure ($\partial p^* / \partial t$) into components for $\beta$, $P$ and $E$ as follows. First, we estimated $\beta$ as ($\partial p^* / \partial t$) during periods when $P$ was zero (i.e., when there were no positive ‘jumps’ in the $p$ time series) and $E$ was near zero. These included precipitation-free periods in the cold season (late autumn, winter and early spring), and precipitation-free nights in the growing season. The cold season was defined by daily mean temperature below 0.0°C. We merged the cold and growing-season estimates of $\beta$ to produce a continuous time series, and filled gaps by linear interpolation. The values of $\beta$ were then smoothed as 21-day running means to remove noise.

Secondly, we identified $P$ events from jumps in the $p^*$ time series, using a high-pass Haar wavelet filter (coefficients: $-1$, $-1$, $-1$, $+1$, $+1$, $+1$). We used a threshold of 2 mm to identify precipitation events in the Haar transform. Thresholds of less than 2 mm picked up excessive noise. The 2 mm threshold excluded most values for $P$ of below 0.2 mm per half-hour. We assumed that $P$ added to the groundwater pressure, so that only positive jumps in $p^*$ were considered. We also assumed that $E$ was negligible during precipitation. When precipitation was detected, the value for $P$ was computed from Eq. (6) as $\gamma^{-1} (\partial p^*/\partial t - \beta)$.

Thirdly, we calculated $E$ from Eq. (6) as $\gamma^{-1} (\partial p^*/\partial t - \beta) - P$, $P$.

We will refer to the term $\gamma^{-1} (\partial p^*/\partial t - \beta)$ as the net change in the moisture balance above the piezometer, and to its cumulative value as the net moisture balance. The net moisture balance integrates both changes in water table depth and changes in soil moisture storage above the water table. This term is equivalent to the
4. Results and discussion

We first qualitatively analyse the time series of groundwater pressure, and then decompose it into components for the seasonal background cycle, net moisture balance, precipitation and evapotranspiration (Sections 4.1, 4.2, 4.3 and 4.4, respectively). The method used to decompose \( \frac{\partial p^*}{\partial t} \) into \( \beta \), \( P \) and \( E \) uses only minimal ancillary data; we used air temperature in the \( \beta \) analysis to differentiate between the cold and growing seasons. An alternate approach would be to make greater use of ancillary data in the decomposition. This would be particularly valuable in detecting runoff. Snowmelt runoff could be inferred if an independent estimate of snow water equivalent was available just prior to melt. Surface runoff could be calculated using independent measurements of precipitation.

Fig. 3 shows two annual cycles of groundwater pore pressure, with the influences of atmospheric loading and earth tides removed. Fig. 4 enlarges a 6-week period from Fig. 3, and shows cumulative, gauged precipitation \( P \) for the same period. Clearly evident in the time series are sharp jumps of groundwater pressure. These jumps are most frequent in spring and summer, and show the response of \( p^* \) to changes in mass loading by \( P \). Note the quantitative match between gauged \( P \) and the increases in \( p^* \) during precipitation (Fig. 4). We compare piezometric and gauged estimates of \( P \) in (Section 4.3).

During growing-season periods with no precipitation, \( p^* \) shows a distinct diurnal cycle, associated with daytime evapotranspiration \( E \) (Fig. 4; see, for instance, the period from 1 to 11 July). However, the diurnal \( E \) cycle is partially obscured by the presence of a longer-term (seasonal) background cycle in \( p^* \), which we have denoted as \( \beta \) in Eqs. (3) and (6). The presence of a seasonal background cycle is evident in the winters of 1996–1997 and 1997–1998 (Fig. 3), when \( p^* \) declined steadily despite the accumulation of the winter snow pack. It is evident in June 1997, when \( p^* \) rose during some precipitation-free periods, despite the loss of water by \( E \). It is also evident during many precipitation-free nights, when \( E \) was small but \( p^* \) changed significantly. The \( \beta \) time series is complex, seasonal and contains both positive and negative values.

We attribute the existence of a significant seasonal background cycle in \( p^* \) to a higher-than-ideal hydraulic permeability at this site, which does not fully isolate the piezometer from low-frequency transients in vertical flow. (Another possible cause of \( \beta \) is net
lateral inflow or outflow of groundwater in the shallow sands above the piezometer, but a divergence of this magnitude would indicate major groundwater discharge in the area, and no such discharges have been found.) Low-frequency, transient flows are initiated by seasonal water table fluctuations and variations in soil moisture, and are transmitted to the depth of the piezometer intake. The two effects partially offset each other. Increases in water table depth cause a transient downward flow of water and a positive value for $\beta$, whereas increases in mass load cause a transient upward flow and a negative value for $\beta$. The transient flows are miniscule in mass terms but significant in pressure terms. Although the transients damp with depth, the aspen site appears to be too permeable for seasonal transients to be fully damped above the piezometer.

The occurrence of seasonal flow transients at the piezometer depth may indicate the presence of fractures in the 15 m of clay-rich glacial till that separate the piezometer intake from the overlying sands and gravels. Fractures can increase the hydraulic diffusivity by an order of magnitude or more above that of an unfractured till. For instance, Keller et al. (1988) reported a hydraulic diffusivity $D$ of $5 \times 10^{-5}$ m$^2$ s$^{-2}$ for a similar, fractured till at 18 m depth below ground level. The corresponding daily, monthly and annual penetration depths ($4z_d$, Eq. (2)) are 4.7, 26 and 90 m, respectively. This hydraulic diffusivity would effectively isolate the piezometer from pressure fluctuations with periods of 10 days and less, but would transmit a significant fraction of longer-period fluctuations to the depth of the piezometer intake. However, we have some evidence that the value for $D$ at the aspen site is smaller than $5 \times 10^{-5}$ m$^2$ s$^{-2}$. The till at the piezometer site is more deeply buried than that in Keller et al., and fractures tend to close with depth. Moreover, the observed time series for $\beta$ showed a response time of weeks not days. Work is in progress to estimate the response time of $\beta$ to observed changes in water table depth and $P+E$. Happily for the analysis in this study, the flow transients associated with higher-frequency features in the $p*$ time series, such as $P$ and $E$, have much smaller damping depths and are fully damped above the piezometer.

To put the aspen site in perspective with other sites in earlier studies, it is an aquitard with low but not very low hydraulic permeability. It is more permeable than the aquitard reported in van der Kamp and Schmidt (1997), where seasonal transients were fully damped at the piezometer depth, but much less permeable than the aquifers reported in van der Kamp and Maathuis (1991) and Bardsley and Campbell (1994), where high-frequency transients blurred the hourly and daily features in the $p*$ time series.

The non-zero value for $\beta$ adds complexity to the piezometric data analysis. The $\beta$ time series must be estimated and removed from the $p*$ time series before $P$, $E$ and changes in soil moisture can be inferred. This result contrasts with the simpler analysis in van der Kamp and Schmidt (1997), where the formation was very impermeable and $\beta$ was effectively zero.

Fig. 3 shows two annual cycles of cumulative seasonal background and net moisture balance. Fig. 5 decomposes the net moisture balance into components for $P$ and $E$. We will discuss each of the terms in turn.

4.1. Seasonal background cycle

The seasonal background cycle $\beta$ (Fig. 3) contributed significantly to the seasonal cycle of $p*$, but integrated to near zero over the annual cycle. The value for $\beta$ was comparable in magnitude to $P$ and $E$ during some periods (compare Figs. 3 and 5). However, we encountered two difficulties in estimating $\beta$ that diminished our confidence in its estimates.
First, the β estimates were not always consistent from day to day; at 34.6 m depth in a low-permeability formation, day-to-day variations in β should be largely damped out. The noise in β, estimated as the root-mean-squared difference between its daily mean and its 21-day running mean, was 0.11 mm per day. Further, the cold- and growing-season estimates of β did not always match in the transitional seasons. When the match was poor, we gave priority to the cold-season values, which showed greater day-to-day consistency than the growing-season values.

Second, the assumption that E was negligible during periods with low E (i.e., during the cold season and at night during the growing season) led to a small but systematic underestimation of . This underestimation was a necessary compromise. We could not use independent estimates of cold-season or nighttime E to calculate β from Eq. (6) because we would in turn use the estimates of β to calculate E from Eq. (6). We therefore chose periods to estimate β when E was near zero. In reality, E was small but not zero during these periods. The measured, eddy-correlation latent heat fluxes had a mean of 9 W m$^{−2}$ (0.3 mm per day or 0.1 mm per night) during precipitation-free, summer nights, and a mean of 3 W m$^{−2}$ (0.1 mm per day) in winter. The implications for (and the net moisture balance) are significant. The use of a zero E assumption in estimating resulted in an 82 mm underestimation of cumulative for April 1997 to December 1998, which accounts in part for the downward drift of in Fig. 3.

The analysis shows that the aspen site is not ideal geologically for the piezometric weighing lysimeter method, at least in its present form. The method should be used with caution, particularly when estimating E. Clearly, another approach is needed to estimate β that is independent of the $p^*$ time series and does not require assumptions about E.

4.2. Net moisture balance

Once the seasonal background time series was known, we calculated the net moisture balance from the difference of groundwater pore pressure $p^*$ and the cumulative value for β (Fig. 3). The most striking features of the net moisture balance are the deep drawdown of soil moisture by evapotranspiration in June–September 1997 and May–June 1998, the recharge of soil moisture by precipitation in April–June 1997 and June–July 1998, the contrast in moisture following the 1997 and 1998 growing seasons, the gradual accumulation of the snowpack in winter 1997–1998, and the lack of water loss by snowmelt runoff in both springs.

Although we lack the data to fully validate these results, two comparisons show the promise of the piezometric weighing lysimeter method for estimating the net moisture balance, based on soil moisture measurements at the flux tower, and water table and snow survey measurements at the piezometer site. First, the inferred snow accumulation in winter 1997–1998, which reached a maximum of 60 mm snow water equivalent on 30 March 1998 (Fig. 3), is in close agreement with snow survey measurements at the piezometer site, which showed a maximum snow water equivalent of 55 mm on 25 March 1998.

Second, the net moisture increase of 138 mm from 1 November 1997 to 1 November 1998, as inferred from the piezometer data (Fig. 3), agrees in sign with a measured 52 mm increase in root-zone (0.0–1.2 m) soil moisture, and a measured 145 mm increase in water table height (which is equivalent to a 66 mm increase in water mass, given the bulk density of 1450 kg m$^{−3}$ and a volumetric water content at field capacity of 0.15). However, Fig. 3 may overestimate the net moisture increase from autumn 1997 to autumn 1998, if indeed the value for β has been underestimated (Section 4.1). The underestimation of β would account for approximately 48 mm of the observed 138 mm moisture increase between autumn 1997 and autumn 1998.

4.3. Precipitation

Fig. 6 compares cumulative $P$, as inferred from groundwater pore pressure, with direct $P$ measurements from an accumulating gauge at the piezometer site, from June to November 1998. The period included both rain- and snowfall events. Fig. 7 compares individual precipitation events for the same period. Cumulatively, the piezometric estimate for precipitation (466 mm) exceeded the gauged measurement (444 mm) by 5%. Considering the scale differences of the two measurements (i.e., several hectares (piezometer) versus point (gauge)), the overall agreement was
encouraging and the agreement between half-hour values (0.01±0.20 mm h\(^{-1}\)) was remarkable. The 5% difference between the two is consistent with previous reports that accumulating gauges under-measure precipitation (Yang et al., 1999).

There is some evidence in Fig. 7 that the agreement between piezometric and gauged precipitation may be poorer for precipitation events with accumulations of more than 20 mm, but we have too few data from large events to draw any definite conclusions at this time. The piezometer and gauges agreed to within 2% for events with accumulation of less than 20 mm (all rainfall), with total accumulations of 139 mm — piezometer, 135 mm — accumulating gauge and 137 mm — tipping bucket rain gauge. For the five (only) rainfall events with accumulation of more than 20 mm, however, the piezometric \(P\) total (265 mm) was 12% higher than the accumulating gauge (237 mm) and 17% higher than the tipping bucket rain gauge (227 mm). A second accumulating gauge at the flux tower, 1.3 km away, gave cumulative \(P\) for the six large events in Fig. 7 that agreed closely with the accumulating gauge at the piezometer site.

We do not show precipitation results from 1997, because of gaps in measured \(P\). When data were available from 1997, the agreement between piezometric and gauged \(P\) was similar to 1998. All of the gauged events in 1997 had accumulations of less than 20 mm.

However, two details in the analysis may bias the piezometric \(P\) estimates low: the assumption of no evapotranspiration or runoff during precipitation, and the use of a threshold to detect \(P\). The piezometric estimates should therefore underestimate \(P\), by the sum of \((E+R\text{ during precipitation}) + (P\text{ at intensities below the Haar wavelet detection threshold of 0.2 mm per half-hour})\). The first term, \(E+R\) during precipitation, is certainly significant but is difficult to quantify. As stated previously, we have evidence that \(R\) was negligible throughout the study. The best estimate of \(E\) during precipitation came from the eddy-correlation measurements, which gave a mean of 0.06 mm h\(^{-1}\). This sums to 33 mm over the study period. If we assume that this value does not include any evaporation from falling raindrops (an assumption that places a ceiling on the resulting bias in \(P\)), then the assumption that \(E+R\) was negligible during precipitation caused the piezometer to underestimate \(P\) by 33 mm. The assumption of zero \(E+R\) during precipitation makes the piezometric \(P\) estimate an ‘effective’ precipitation, i.e., the total precipitation minus any evaporation or runoff that occur during the precipitation event. It is the portion of \(P\) that remains on the canopy, on the surface and in the ground at the end of the precipitation event.

The second term, \(P\) at intensities below the detection threshold, was also small but significant. Between June and November 1998, the piezometer detected only 2 mm of \(P\) at intensities below 0.2 mm per half-hour, whereas the accumulating gauge mea-
sured 19 mm. Considering the magnitude of these two biases, it was surprising that the piezometric estimate of $P$ agreed so closely with the direct measurement. It is possible that both methods have underestimated $P$.

We conclude that piezometric weighing lysimeters show great promise for measuring precipitation. They provide an indirect but physically fundamental measurement of ‘effective’ precipitation. They may have particular value for measuring large, intense events, where they could serve as reference standards for the validation of other methods. Work is in progress to compare piezometric versus gauged $P$ for a larger sample, at several sites, and using standard reference ‘pit’ gauges.

The greatest challenge in using piezometric weighing lysimeters to measure precipitation will be for the measurement of long-duration, low-intensity precipitation events (e.g., drizzle and light snowfall), which do not cause detectable jumps in $p^*$. The difficulty in detecting light snowfall is exacerbated by increased noise in the piezometer signal during winter. We have not yet developed a method to isolate low-intensity precipitation from the $p^*$ time series.

4.4. Evapotranspiration

Fig. 8 compares piezometric estimates of cumulative $E$, from Eq. (6), with direct measurements by eddy correlation, over two annual cycles. The seasonal integrals were roughly parallel, with cumulative piezometric $E$ ($-808$ mm) exceeding measured $E$ ($-760$ mm) by 6%. The piezometric and eddy-correlation $E$ totals for the growing season agreed to within $32$ mm (8%) in 1997 and 15 mm (4%) in 1998. In both years, the piezometer underestimated $E$ in early spring relative to the eddy-correlation measurements, and overestimated $E$ in autumn. The discrepancy may be related to uncertainty in the precise seasonality of $\beta$. The value for $\beta$ was hard to estimate, particularly during the spring and fall transitions. Our attempts to remove daily noise in $\beta$ by smoothing $\beta$ as a 21-day running mean may have blurred some important details in the $\beta$ time series during the transitional seasons.

Fig. 8 may underestimate both piezometric and eddy-correlation estimates for $E$. If, as discussed in (Section 4.1), the cumulative value for $\beta$ has been underestimated by 82 mm, then the piezometric $E$ total in Fig. 8 should be increased in magnitude by $\gamma^{-1} \times (82 \text{mm}) = 90 \text{mm} \rightarrow -898 \text{mm}$. However, Fig. 8 may also underestimate eddy-correlation $E$, because the eddy-correlation sensible and latent heat fluxes ($H$ and $\lambda E$) sum on average to 10% less than the surface available energy (net radiation minus ground and storage heat fluxes). If we increase both $H$ and $\lambda E$ by 10% to close the surface energy balance while preserving the ratio of $H/\lambda E$, the eddy-correlation $E$ total increases in magnitude by 76 mm to $-836$ mm, 7% lower than the adjusted piezometric $E$ total.

Fig. 9 shows cumulative $E$ during leaf out in 1997. Both piezometric and eddy-correlation estimates of $E$ showed a distinct diurnal cycle, beginning at leaf out. Some noise is evident in the piezometric estimates of $E$, with a peak-to-peak magnitude of about 1 mm. Although the piezometric and eddy-correlation $E$ totals diverged by 13 mm during the 25 days in Fig. 9, we were encouraged by the sensitivity of the piezometer and its ability to detect the pre- to post-leafout transition in $E$. Note that Fig. 9 is taken from the spring transition, when, as discussed earlier, the piezometric and eddy-correlation estimates of $E$ showed the greatest differences.

Fig. 10 compares daily piezometric versus eddy-correlation totals for $E$ from Fig. 8, for the 1997 growing season. Daily piezometric and eddy-correlation $E$ agreed to within 0.2±1.3 mm per day. We attribute the differences between piezometric and eddy-correlation estimates of $E$ in Figs. 8–10.
The eddy correlation measurement is offset by $C_{5}$ mm to enable visual comparison.

in part to noise in $p^*$, and in part to uncertainty in the seasonal cycle of $\beta$. It is also possible that the differences are real. The piezometer and eddy-flux sites were 1.3 km apart, and the two sites, although similar, were not identical. In particular, the soil at the piezometer site was sandier than the soil at the tower-flux site, and a difference in soil moisture could have caused a difference in $E$.

We conclude that piezometric weighing lysimeters have promise for estimating $E$ at the stand or field scale, but that their simplest application may be limited to sites with very low permeability. We have made some progress in implementing the method at a non-ideal site. Nevertheless, we recommend caution in applying the current methodology to estimate $E$ at sites where non-ideal sub-surface geology permits a significant seasonal background cycle.

5. Summary and conclusions

This study has demonstrated the utility of the piezometric weighing lysimeter method for inferring the moisture balance at the stand or field scale, based on two annual cycles of measurements in a mature aspen forest in central Saskatchewan, Canada. The deep groundwater pressure is shown to respond to changes of mass load associated with precipitation and evapotranspiration. This study extends the earlier work of van der Kamp and Schmidt (1997) to sub-optimal conditions, and broadens the application of the method.

Analysis of the $p^*$ time series revealed that the aspen site was not ideal geologically for the piezometric weighing lysimeter method. The simplest application of the method is limited to sites with very low hydraulic permeability, where seasonal changes in water table depth are fully damped at the depth of the piezometer intake. This was not the case at the aspen site, where $p^*$ showed a significant seasonal background cycle. This did not invalidate the use of the piezometric weighing lysimeter method at this site, but complicated the analysis and increased the uncertainty in the piezometric estimates of $E$.

To overcome this difficulty, we developed a procedure to isolate the seasonal background cycle from the $p^*$ time series. The seasonal background rate of change $\beta$ was estimated as $\partial p^*/\partial t$ during periods with no precipitation and when evapotranspiration was near zero, i.e., at night and during the cold season. Once $\beta$ was known, $P$ events were identified from positive jumps in the $p^*$ time series, and $P$ was estimated as $\gamma^{-1}(\partial p^*/\partial t - \beta)$. Lastly, $E$ was calculated as $\gamma^{-1}(\partial p^*/\partial t - \beta) - P$.

Piezometric estimates of cumulative $P$ exceeded in situ measurements by 5%, with excellent agreement between individual half-hour values. The 5% discrepancy was consistent with earlier reports that accumulating gauges under-measure precipitation. We
conclude that piezometric weighing lysimeters have definite utility for measuring $P$. Their advantages over conventional precipitation gauges include the large (e.g., several hectare) area of their catchment, and their independence from the deleterious effects of wind and air flow on gauge catch efficiency.

Growing-season evapotranspiration, as inferred from the $p^*$ time series, coincided to within 8% of direct measurements by eddy correlation over two growing seasons. The ability of the piezometric weighing lysimeter to detect the pre- to post-leaf-out transition in the diurnal cycle of $E$ illustrated the method’s sensitivity. However, our confidence in the piezometric estimates of $E$ was limited by uncertainty in $\beta$. The value for $\beta$ was difficult to estimate, particularly during the transitional seasons. (Uncertainty in $\beta$ also affected the piezometric estimates of $P$, but the effect was negligible because $P$ was much larger in magnitude than $\beta$.) We conclude that the piezometric weighing lysimeter method has promise for estimating $E$, but that its simplest application is limited to sites with low hydraulic permeability, where $\beta$ is effectively zero.

Work is in progress to improve the estimates of $\beta$ using a more physically-based approach. We have installed a shallow piezometer below the water table, and are attempting to evaluate the damping and lag of the seasonal $\beta$ wave with depth. We are also assessing the use of the piezometric weighing lysimeter to estimate changes in soil moisture by comparison with in situ measurements of root-zone soil volumetric water content, and have begun a comparison of piezometric estimates of $P$ with a reference precipitation standard.

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