Constraining the annual groundwater contribution to the water balance of an agricultural floodplain using radon: The importance of floods

Jackie R. Webb[a,b*], Isaac R. Santos[a,b], Barbara Robson[c], Ben Macdonald[d], Luke Jeffrey[a,b], Damien T. Maher[a,b]

[a]School of Environment, Science, and Engineering, Southern Cross University, New South Wales, Australia

[b]National Marine Science Centre, Coffs Harbour, Southern Cross University, New South Wales, Australia

[c]CSIRO Land and Water, GPO Box 1666, Canberra, ACT 2601, Australia

[d]CSIRO Agriculture, GPO Box 1666, Canberra, ACT 2601, Australia

*Corresponding author Jackie Webb (j.webb.20@student.scu.edu.au). Ph: +61 468 730 184

Key points:

- Combined radon and groundwater level observations allowed the construction of a detailed groundwater budget.
- Flood recession periods contributed 72-76% to annual groundwater discharge.
- Groundwater discharge contributed 30-80% to total surface discharge.

Abstract

The water balance of drained floodplains is highly dynamic with complex groundwater-surface water interactions operating over varying spatial and temporal scales. Here, we hypothesise that the majority of groundwater discharge will follow flood events in a modified wetland. To test this hypothesis, we developed a detailed water balance that quantifies the contribution of groundwater discharge to the annual water budget of an extensively drained agricultural floodplain. A clear relationship between surface water radon measurements and groundwater level indicated alternating connection-disconnection dynamics between the drains and shallow groundwater. This relationship was used to develop a radon mass balance to quantitatively model groundwater discharge continuously throughout the year. Groundwater discharge varied by four orders of magnitude over the study period, with daily average rates ranging from 0 to 27,200 m^3 d^{-1}, peaking just a few hours after floods receded. Flood events occurred only 12% of the time yet contributed 72 to 76% of the total groundwater discharge. During flood recession periods, aerial groundwater discharge rates reached up to 325 cm d^{-1} which were some of the highest rates ever estimated. We proposed that the high drainage density of this site (12.4 km constructed drains km^{-2} catchment area) enhanced groundwater discharge during wet periods due to increased connectivity with the soil. Overall, groundwater discharge contributed 30-80% to the total surface water discharge. This study offers insight into the dynamic behaviour of groundwater within an extensively drained floodplain, and the importance of capturing flood events to quantify total groundwater contribution to floodplain water balances.

Key words: drainage density, water budget, flood pulse, groundwater-surface water, connectivity, radon, artificial drains, groundwater exchange

1. Introduction

Hydrological interactions between streams and the land provide important feedbacks to land surface hydrologic, biogeochemical, and ecosystem dynamics [*Shen et al.,* 2016]. Hydrological flow regimes in floodplains are characterised by baseflow, flow pulse (below bankfull), and flood pulse (above bankfull), which can result from different water sources including tributary inflow, overland flow, and groundwater discharge [*Tockner et al.,* 2000]. All of these water sources are ultimately driven by precipitation, geomorphology, and aquifer properties. Groundwater-surface water (GW-SW) exchange is particularly important in floodplains and wetlands as water tables are generally near the surface in these depressional landscapes [*Jeffrey et al.,* 2016; *Ludwig and Hession,* 2015]. Specifically, GW-SW connectivity often controls surface discharge within floodplains along with the duration of inundation [*Mitsch and Gosselink, 2000*]. A relatively unconfined groundwater flow often develops in these systems and is referred to as 'interactive' groundwater due to active exchanges with surface water [*Harvey et al.,* 2006]. Groundwater-surface water interactions can strongly influence benthic productivity in wetland streams, habitat heterogeneity, and surface water biogeochemistry, and can even stimulate organic carbon turnover [*Tockner et al.,* 2000; *Hunt et al.,* 2006; *Stegen et al.,* 2016].

Most detailed studies of groundwater-surface water interactions focus on timescales ranging from hours to months. Short term studies can provide high spatial and temporal resolution of groundwater-surface water interactions [*Burnett et al.,* 2003; *Santos et al.,* 2011; *Sadat-Noori et al.*, 2015; *Luo et al.,* 2016], and reveal any groundwater driven interactions between hydrologic dynamics and biogeochemical processes [*Fleckenstein et al.,* 2010; *Atkins et al.,* 2013; *Makings et al.,* 2014]. However, short term studies are limited in their ability to quantify the contribution of groundwater to the total water balance because of hydrological time lags and temporal bias in sampling. Long term groundwater studies ranging over annual to decadal time scales [*Harvey et al.,* 2006, *Wilcox et al.,* 2006; *Wolksi and Savenije*, 2006; *Krause et al.,* 2007a] offer insights into the composition of old and new groundwater, residence times, and provide a whole-of-system picture of groundwater contribution to the water balance. However, such long term studies may lack detailed information on temporal groundwater responses to rain events and therefore provide a more generalized view of groundwater discharge.

The temporal and spatial behaviour of groundwater discharge in floodplains is tightly controlled by pressure head gradients and mediated by physical soil and sediment characteristics governing hydraulic conductivity [*Bencala,* 1993; *Stanford and Ward,* 1993]. These properties can vary widely over small spatial scales, leading to large uncertainties in hydraulic conductivity that make mechanistic estimates of groundwater-surface water fluxes difficult to constrain [*Kalbus et al.,* 2006; *Johnston et al.,* 2009]. Furthermore, groundwater discharge is subject to both temporal and spatial variability which may not be captured by point measurements of groundwater seepage [*Peterson et al.,* 2010; *Wilson and Rocha*, 2016]. Geochemical tracers integrate complex and variable groundwater discharge pathways as they reflect the net groundwater discharge averaged over the reach length [*Schmidt and Schubert,* 2007; *Cook,* 2013; *Atkinson et al.,* 2015].

The advent of new automated analytical techniques has allowed the use of chemical tracers to estimate groundwater-surface water exchange over seasonal time scales. For example, radon has been used as a tracer to quantitatively assess groundwater flux in river/stream systems [*Genereux and Hemond,* 1990; *Santos et al.,* 2008; *Peterson et al.,* 2010]. Radon is present in high concentrations in groundwater due to the continuous decay of uranium in soils, making it ideal for estimates of groundwater input. The conservative and inert nature of radon gives it some advantage over other geochemical groundwater tracers commonly used such as conductivity, alkalinity, methane (CH_4), and stable isotopes of lead, strontium, and carbon, which can be affected by biological and chemical transformations [*Bullen and Kendall,* 1998; *Kalbus et al.,* 2006; *Santos et al.,* 2008]. Chloride (CI) is one of the most widely used conservative tracers for estimating groundwater discharge. Chloride can be used when groundwater and surface water have contrasting concentrations, but may be a less useful tracer coastal systems subject to salt deposition [*Cook,* 2013]. The stable isotopes of water (δ^2H and δ^{18}O), which are indicators of "new" and "old" water [*Kendall and Caldwell*, 1998], are another commonly used conservative tracer. However, the ability of these tracers to differentiate between groundwater flows and fresh surface water flows is limited by the degree of difference in isotopic composition between the two components [*Buttle,* 1994], which may be similar for shallow groundwater in floodplain landscapes (i.e. recharges with "new" water from surface water and precipitation). The short lived nature of radon relative to other radioactive isotope tracers (e.g., tritium) means that radon is sensitive to rapid exchanges between surface and groundwater. This makes it an ideal tracer for characterising GW-SW exchange, which can rapidly switch direction during flood events.

The role of flood events in wetland and floodplain groundwater dynamics is poorly understood. However, a number of short term studies have highlighted notable increases in groundwater discharge following flood events [*De Weys et al.,* 2011; *Gilfedder et al.,* 2015], and this is consistent with theory. Flood-stimulated groundwater pulses may have important implications to the overall water balance of wetlands and floodplains. Enhanced groundwater discharge of solutes has also been linked to severe acidification, deoxygenation, and high carbon dioxide in some rivers post-flood [*Santos et al.,* 2011; *Atkins et al.,* 2013].

Studies that integrate both the quantitative and qualitative advantage of assessing groundwater dynamics over different timescales are limited [*Wilson et al.,* 2015]. Here, we model a detailed time-series of surface water radon concentrations to quantify groundwater discharge and its contribution to the water budget of an artificially drained agricultural floodplain. Through modelling of radon concentrations, we estimate groundwater discharge to the surface waters at high temporal resolution over an annual cycle. We aim to capture periods of high groundwater discharge that may be overlooked in annual time scales due to limited temporal resolution or not captured during short term studies. We hypothesise that the majority of groundwater discharge will follow flood events and occur over short periods.

2. Methods

2.1 Study area and sampling approach

The study site is a sub-catchment of the McLeods Creek catchment (28°16′50′′S, 153°30′12′′E) which is situated in the Tweed Valley floodplain in northern New South Wales, Australia (Figure 1). McLeods Creek generally floods annually, however the frequency, intensity, and duration of floods vary between years. The site receives on average 1,600 mm rainfall annually and experiences a subtropical climate with monthly mean minimum and maximum temperatures of 8.6 and 29.5°C, respectively [*Bureau of Meteorology*, 2016]. Rainfall persists year round with ~830 mm of rainfall occurring during the hottest months of December-March when large isolated rainfall events are most frequent.

The sub-catchment is a typical Australian coastal lowland that contains unconsolidated estuarine sediments originating from the Holocene period, which due to extensive drainage has developed oxidised acid sulfate soils [*White et al.,* 1997]. The site is positioned at a low elevation of ~0.15 m AHD (Australian Height Datum, where 0 m AHD approximates mean sea level). Three soil horizons exist at the site; an organic topsoil (0 to 0.25 m), the oxidised sulfuric horizon (0.25 to 0.9 m), and an unconsolidated sulfidic estuarine clay layer (>1 m) comprising of pyrite deposits from the Holocene period [*Lin et al.,* 1998; *Smith et al.,* 2003]. Hydraulic conductivity has been measured between 0.74 and 5.54 m day^{-1} in the upper soil horizon (0 to 1.5 m) [*White et al.,* 1993; *Johnston et al.,* 2009]. Large releases of sulfuric acid and dissolved metals including aluminium, iron, zinc and manganese have been observed in surface water discharge after flood events at this site, and was speculated to be related to groundwater discharge [*Wilson et al.,* 1999; *Green et al.,* 2006; *Macdonald et al.,* 2007]. However, groundwater discharge has not been quantified.

The floodplain consists of an extensive artificial drainage network and has been used for sugarcane production for the past 40 years [*Smith et al.,* 2003]. The drainage network consists of main drains which have widths ranging between 1.0 to 3.5 m and depths of 0.6 to 1.2 m, and are the main conduit for surface water discharge out of the sub-catchment (Figure 1). Smaller field drains have a width of 1.5 m and depth of 0.5 m, and are only connected to the main drains during periods of major rainfall. An automatic pump at the catchment outlet (Figure 1) controls drain discharge from the site which expels water into the Tweed River. This is triggered when surface water levels go above -0.45 m AHD (0.57 m drain depth). Surface water infiltration from the Tweed River and bordering creeks are blocked by permanent flood gates and levees, allowing for water levels within the site to be maintained below sea level. The main sources of water feeding the drains are rainfall and groundwater. A small amount of tidal creek water is allowed to inflow through McLeods Creek floodgate to the main drain to buffer acidification during dry periods. Surface water actively flows within the drains only after major rainfall, and remains stagnant with intermittent discharge pulses between periods of rainfall.

The sub-catchment is hydrologically isolated from adjacent land and Tweed River by natural levees, artificial bund walls and bordering tributaries [*Green et al.,* 2006]. Local groundwater flow is shallow, as is typical of these coastal lowlands [*White et al., 2003*], and contained within the sub-catchment area that defines the water balance boundary in this study. The estuarine clay layer that exists at 1 m (-0.4 m AHD) below the surface and extends for a further 10 m has an hydraulic conductivity of 0.03 m d^{-1}, which effectively impedes any rainfall transmitted through the upper soil horizons, and infiltration of seawater into the drains [*White et al.,* 1993]. Groundwater levels are typically maintained at -0.5 m AHD by artificial pumping of drainage water.

The sampling approach consisted of a combination of continuous and discrete measurements of various hydrological parameters over the course of 11 months (336 days). Continuous measurements for drain surface depth and velocity, groundwater depth, evapotranspiration and precipitation were taken using a series of data loggers (see below). Surface water radon concentrations were measured approximately every two weeks from the outlet drain during individual field campaigns. Samples were obtained using specially designed eight litre bottles [*Stringer and Burnett,* 2004], and analysed using a radon-in-air closed loop method [*Lee and Kim,* 2006]. During one major flood event in late January 2015, a time series of continuous radon measurements was taken over a period of six days [*Webb et al.,* 2016]. Radon concentrations from this sampling period are included in this paper along with bi-weekly discrete radon samples to help constrain the flood response of radon in surface waters. The groundwater radon endmember was characterised by incubating six 1 kg samples of sediment from the two distinct soil layers below the surface (0.5 m and 1 m) for 21 days to obtain the "sediment equilibrated" radon concentration [*Corbett et al.,* 1998]. This approach has been used previously to estimate radon endmembers [*Schmidt et al.,* 2010; *Peterson et al.,* 2010]. The analytical uncertainty for radon samples ranged from 2-35% for surface water depending on concentration and 5-9% for groundwater endmember concentrations. Endmember error from natural variability was calculated from the standard error of the six sediment incubation measurements (supplementary material).

2.2 Precipitation, evapotranspiration, and evaporation

Daily precipitation and evapotranspiration were measured directly at a meteorological tower located onsite. Rain was measured using a Rimco tipping bucket rain gauge (RIM7499, Campbell Scientific Inc.) with a 20.3 cm diameter collecting funnel and 0.2 mm tip. Gaps in on site precipitation data due to instrument issues made up 16% of the total study period, and were filled from the Murwillumbah Bureau of Meteorology measurement site located ~10 km. Evapotranspiration was measured using an eddy covariance system to measure fluxes of water vapour. Three-dimensional wind speed (CSAT3 sonic anemometer, Campbell Scientific Inc.) and moisture density in air (LI-7500 open path CO_2/H_2O analyser, Licor Inc.) were sampled at 10 Hz and then computed to 30 min mean fluxes of latent heat (W m^{-2}). Half-hourly logger computed fluxes were processed following standard OzFlux QA/QC procedures (Eamus et al., 2013), including Webb-Pearman-Leuning (WPL) density correction for latent heat fluxes [*Webb et al.,* 1980] and removal of erroneous data (range test, spike removal, u* filtering). Gaps in latent heat flux data amounted to 24% following QA/QC procedures and were filled using a self-organising linear output (SOLO) artificial neural network model [*Eamus et al.,* 2013]. Final gap-filled, corrected latent heat fluxes (energy flux, W m^{-2}) were then converted to evapotranspiration rates (water flux) as mm d^{-1}. Water loss from ET and rainfall were up scaled to the total area of the catchment (~1,000,000 m^2). Surface water evaporation data was taken from daily Morton evaporation over shallow lakes obtained from a nearby station (~12 km) from the SILO climate database [*SILO*, 2016]. Total daily evaporation was then extrapolated to the surface water area within the sub-catchment. The instrument uncertainty associated with the rain, ET, and evaporation measurements were 1%, 5%, and 7% respectively.

2.3 Surface Discharge

Surface flow velocity and depth were measured in a pipe culvert positioned 100 m upstream of the pump using a Starflow ultrasonic Doppler flowmeter. Manufacturer reported accuracy for this instrument is ± 2%. Average depth and velocity were recorded over 30 minute intervals. The cross sectional area of the flooded portion of the pipe was calculated from the dimensions of the pipe culvert. Half hourly surface discharge was then calculated by multiplying the cross sectional area by velocity. Significant decreases in surface water depth between intervals determined when the pump was on and when surface discharge was calculated. A two month data gap in flowmeter data was filled from a second depth logger located in a drain 700 m upstream from the catchment outlet. Average velocity for water depth increments of 0.1 m was used to gap fill the velocity data gap. The model fit of predicted total discharge using this gap filling method to actual discharge measurements was within 5%.

2.4 Drain volume and surface area

Cross sectional drain profiles were measured for each drain within the study site. A total of 40 measurements were taken across the length of 26 individual drains. Drains were grouped into two categories, main drains which generally remain connected for the entire year, and field drains which are only connected after high rainfall (Figure 1). Small field drains were grouped into five sections relating to their position within the property. Channel cross sectional area was calculated as a function of (depth –varying) drain width and water height below the surface. For the large drains which had multiple profile measurements, depth along each measured width interval was averaged. A best fit second or third order polynomial equation was used to calculate changing width as a function of changing height. The maximum cross sectional area when the drains were full was then given by the antiderivative of the polynomial equation for width to determine the area under the curve.

Surface water depth data was obtained from one of two depth loggers in the drains, depending on which one was closer. Surface water depths of main drains was taken from one of the two closest depth loggers deployed in the field, and corrected for depth difference between a nearby drain containing a logger. Drain volumes were calculated individually for each large drain by multiplying the drain-specific cross sectional area equation by total drain length. Volumes for the small field drains were calculated by taking the average cross sectional area equation for a section, and multiplying by the sum of drain lengths within that section.

2.5 Hydraulic head

Hydraulic head was calculated as the vertical difference in depth between the groundwater level and drain surface water level. Surface water depth was measured in the outlet drain from a Starflow ultrasonic Doppler flowmeter, which was placed in the invert of a drainage pipe located directly upstream. Water depth was converted to AHD from the knowledge that baseline water levels are maintained to at -0.453 m AHD within the drains [*Green et al.,* 2006]. This is equivalent to a controlled surface water depth of 0.575 m above the invert of the pipe. Groundwater depth was measured from a CTD diver deployed 1.6 m below the surface inside a perforated PVC pipe which recorded water pressure. The relative accuracy of depth measurements based on manufacturer specifications was ±0.5%. Once corrected for atmospheric pressure (measured at the flux tower), groundwater depth was referenced to surface water depth by accounting for the difference in peak flood depths and converted to AHD.

2.6 Groundwater storage

Groundwater storage to a depth of 1 m over the annual cycle was calculated as the difference between groundwater depth between the start and end of time series. The following equation was used to calculate total storage:

$$S_{GW} = (\Delta D(m) \times P(v/v)) \times A(m^2)$$

Where S_{GW} is the storage term in m^3, ΔD is the difference in groundwater depth, P is the soil porosity, and A is the catchment area excluding drains and roads (assumed impermeable surface). Soil porosity was estimated from the soil moisture content (v/v) at saturation measured by five soil water content reflectometers (CS655, Campbell Scientific) connected to the onsite meteorology tower at depths ranging from 5 cm to 50 cm below surface. The natural variability in soil porosity was found to be ~14%, which was calculated as the standard deviation of three measurements made from the same depth. On a volume per volume basis, the maximum soil moisture content is equal to the porosity of the soil at saturation [*Vomocil,* 1965].

2.7 Radon mass balance

The flux of groundwater discharge was calculated using a radon mass balance technique. Two estimates of groundwater discharge were calculated to represent the minimum ($Q_{GW\text{-}min}$) and maximum ($Q_{GW\text{-}max}$) range based on two sets of extreme conditions [*Peterson et al.,* 2010; *Santos and Eyre,* 2011]. The minimum and maximum approaches used here accounts for the likely heterogeneity of groundwater input along the drain stretch and provide a range of possible groundwater discharge. An absolute estimate of the real groundwater discharge is not quantifiable using this approach as the exact amount of radon losses from decay and evasion is unknown. Instead, we report final groundwater discharge as a range using the minimum and maximum estimates, between which the real estimate lies [*Peterson et al.,* 2010].

The minimum estimate was derived from the assumption that groundwater input enters directly at the point of measurement in the stream. In this case, any losses of groundwater-derived radon during surface water transit (i.e. decay and evasion), are neglected from the mass balance. Non-groundwater derived sources of radon are also accounted for in the mass balance and include radium (^{226}Ra) decay, sediment diffusion, and hyporheic exchange. We first estimate radon excess (Rn_{ex}) by removing sources of radon other than groundwater from the observed surface water concentrations. In order to achieve that, the minimum radon concentration observed during the year was assumed to represent all radon sources other than groundwater discharge.

The minimum groundwater input was then derived from the ratio of surface water radon concentration (Rn_{ex}) to the groundwater endmember concentration (Rn_{GW}) multiplied by the total surface water volume (Q_{TOTAL}) fluxed out of the system [*Burnett et al.,* 2010]:

$$Q_{GW-min}\,(m^3 d^{-1}) = \left(\frac{Rn_{ex}(dpm\; m^{-3})}{Rn_{GW}(dpm\; m^{-3})}\right) \times Q_{TOTAL}(m^3\; d^{-1})$$

The maximum groundwater discharge estimate takes into account radon losses from the system that would occur if groundwater input was located at the most upstream section of the catchment relative to the sampling point. These losses include atmospheric evasion (F_{Rn}) and radon decay (λ) and results in higher groundwater discharge required to sustain the measured surface water radon concentration:

$$Q_{GW-max}\,(m^3 d^{-1}) = \frac{\left(\left(Rn_{ex}(dpm\; m^{-3}) \times Q_{TOTAL}(m^3\; d^{-1})\right) + F_{Rn}(dpm\; d^{-1}) + \lambda\,(dpm\; d^{-1})\right)}{Rn_{GW}(dpm\; m^{-3})}$$

Air-water flux estimates of radon (F, dpm m^{-2} d^{-1}) were calculated as follows:

$$F = k \propto (C_{(w)} - C_{(a)})$$

where k is the gas transfer velocity (m d^{-1}), ∝ is the Ostwald solubility coefficient of radon, $C_{(w)}$ is the radon in water concentration (dpm m^{-3}), and $C_{(a)}$ is the radon in air concentration (dpm m^{-3}). Due to the dynamic nature of surface hydrology in this system, two sets of k values were used based on the varying conditions of surface water discharge. During stagnant conditions (no drain discharge and surface inundation during floods), a transfer coefficient derived from Ongori et al., [2015] for radon transfer at the water-air surface under very low turbulence conditions was used. This represents a diffusive radon loss to the atmosphere that is primarily driven by the concentration gradient. Radon evasion during active periods of surface discharge (only when surface waters were constrained within the drains) was calculated from the *k* parameterisation of O'Connor and Dobbins [1957] using surface water velocity and depth as driving factors for turbulence.

2.8 Model uncertainties

Uncertainties in our groundwater discharge model originate from each component of the radon mass balance. The largest degree of uncertainty stems from the assumptions associated with the radon mass balance, where the exact location of groundwater entry points along the drain length is unknown. Because residence time of groundwater discharged to the drains are unknown, we cannot quantify the exact amount of atmospheric loss groundwater-derived radon is exposed to before measurement at the outlet of the catchment. This is a limitation associated with this approach, and consequently makes reporting a singular groundwater discharge value difficult. Our approach relying on extreme assumptions [Peterson et al., 2010] provides a realistic range for possible groundwater discharge. To account for model and endmember uncertainties, we applied lower and upper limits for both the minimum and maximum groundwater scenarios. The lower and upper limits were defined as the mean (μ) – one standard deviation (σ) and mean + one standard deviation, respectively. Error from the model was derived from the standard error of the equation coefficients and constants and the endmember error was based on the natural variability in radon concentrations between samples (Supplementary material). The error of the two terms were then propagated to achieve a combined lower and upper error for minimum and maximum scenarios.

A sensitivity analysis was carried out on the effect of different radon evasion scenarios from a range of empirically derived *k* models. This included estimates combining the wind speed driven *k* parameterisation of *Raymond and Cole* [2001] with depth and current velocity from *O'Connor and Dobbin's* [1957], applying current velocity and depth only [*O'Connor and Dobbins,* 1957], diffusive radon evasion driven by the concentration gradient characterised by the coefficient derived from *Ongori et al.,* [2015], and an extreme case of no evasion (assuming loss by decay only). Final estimates of annual groundwater discharge for each evasion scenario were compared with annual surface discharge to constrain realistic estimates of the maximum groundwater range estimate, but not the minimum groundwater discharge estimate that is not influenced by evasion.

3. Results

The floodplain displayed highly dynamic hydrology throughout the year. Direct rainfall was the only source of water into this floodplain and caused large fluctuations in surface and subsurface hydrology. Total annual rainfall was recorded at 1,740 mm which is greater than the mean annual rainfall of 1,600 mm for this region [*Bureau of Meteorology*, 2016]. Large episodic rainfall events between 40 and 220 mm over 48 hours were recorded eight times throughout the year, and caused inundation of the land surface (Figure 2). Approximately 70% of total annual rainfall fell during these eight events.

Evapotranspiration (ET) was highly seasonal and reflected the growing stage of sugarcane, with peak biomass growth occurring during the wetter warm months of January to April and reduced evapotranspiration during the cooler months of July to September (Figure 2). During flood events ET reduced significantly to <1.5 mm d^{-1}, which was then followed by a sharp increase shortly after the peak of the flood to some of the highest ET levels observed in the study (6.1 mm d^{-1}).

The effect of enhanced surface drainage through the extensive network of artificial drains and intermittent pumping could be seen through the highly erratic nature of surface and subsurface hydrology during rain events (Figure 2). Surface water depth was temporally variable and reached peak height five-six hours following significant rainfall. Changes in groundwater levels were also responsive to rainfall, and spanned over 1.5 m. Groundwater levels typically increased at the same rate as surface water levels during flood events with minimal delay, implying a connected shallow aquifer. However, water table recession occurred at a slower rate compared to surface water levels (Figure 2). Steep vertical hydraulic heads developed during these recession phases. As a result, the depth difference between groundwater and surface water became positive, ranging from 0.1-0.53 m during flood recession periods (Figure 2).

Variation in surface radon concentrations spanned an order of magnitude and demonstrated a direct relationship with the position of the water table (Figure 3). Radon concentrations increased rapidly between 0.3 to 0 m AHD and decreased back to baseline radon levels of ~ 4.8 dpm L^{-1} between 0 to -0.2 m AHD water table position. Based on the change in observed radon concentrations, the model predicted surface water disconnection with the groundwater between groundwater positions <-0.35 m AHD during non-flood conditions and >0.42 m AHD during floods. The highest measured surface radon concentrations of 149 dpm L^{-1} occurred at a receding groundwater position of 0.01 m AHD, and was in the same range to the average groundwater radon concentration of 146 ± 31 dpm L^{-1} (supplementary material).

Surface water radon simulations indicated 13 occasions when radon concentrations exceeded 100 dpm L^{-1} (Figure 4A), approaching the groundwater endmember. Daily volumetric groundwater discharge varied over five orders of magnitude throughout the hydrological year, with discharge during baseline conditions between 0-10 m^3 d^{-1} and discharge during receding flood periods between 100-20,000 m^3 d^{-1} (Figure 4B). Average baseline groundwater discharge remained elevated at around 35 m^3 day^{-1} during January to February when the water table was relatively higher. Throughout the study period, minimum aerial groundwater discharge rates ranged from 0 to 180 cm d^{-1} and maximum rates from 0.3 to 320 cm d^{-1} (Figure 4D).

Figure 5 demonstrates the cumulative trend in all water balance components over the year as a function of cumulative annual rainfall. The total outputs trend represents the sum of all quantified water loss terms which explained 78% of total annual rainfall. Generally, the contribution of the different water budget terms remained similar because episodic events control hydrology. All water balance components repeat the same response to rainfall illustrated by a stepwise trend, where large increases in cumulative rainfall are followed by sudden increase in water outputs (Figure 5). Evapotranspiration represented the largest loss for rainfall (46%), followed by total surface discharge (28%) and drain evaporation (4%). As more rainfall cumulates over time, the discrepancy between total water outputs and the unaccounted for flux increases. For the first half of total rainfall, surface discharge remains similar to evapotranspiration, where the contribution of surface discharge to water loss becomes equal when annual rainfall increases to 0.25 (Figure 5). The difference then widens with each major rain increase, analogous with the increasing discrepancy in the unaccounted for flux.

4. Discussion

4.1 Hydrology time series observations

Both surface water and groundwater were highly responsive to rainfall, demonstrating the effective connectivity of the drainage network with the floodplain. A more rapid response in land surface hydrology including shorter lag times is often associated with extensively drained floodplains [*Blann et al.,* 2009; *Levavasseur et al.,* 2012; *Shen et al.,* 2016]. This effect may also influence near surface groundwater hydrology, where any recharge following rain appears to be offset by groundwater discharge. The residence time of surface water in this floodplain varied greatly, with long residence times between periods of rain when the water was stagnant to extremely short residence times during flooding. Such a wide variability is typical of floodplain landscapes which naturally experience periods of disconnection and connection between channel, floodplain surface and groundwater [*Tockner et al.,* 1999; *Karim et al.,* 2016]. For example, simulations of a floodplain river segment found that residence time decreased exponentially from 55 to 10 days as surface discharge increased [*Helton et al.,* 2014]. By taking into account the average discharge rate when the automatic pump is in continuous operation, it would take approximately 21 hours to replace all water held within all drains when at capacity. Residence time of floodwaters following inundation and back to baseline levels was approximately 3.4 days. These short residence times during flood events are similar to floodwater residence times estimated from another agriculturally impacted floodplain which varied between eight hours to several days during a flood event [*Karim et al.,* 2013].

4.2 Radon-derived groundwater discharge estimates

Measured radon concentrations of up to 149 dpm L^{-1} in surface waters (similar to the measured groundwater endmember of 146 ± 31) suggest that when the groundwater table is -0.1 to 0.1 m AHD elevation groundwater discharge accounts for nearly 100% of surface waters. At 0 m AHD elevation the groundwater level is equivalent to ~20 cm below surface level (top of the drains). This finding that maximum surface radon concentrations coincide with periods of highest groundwater table elevation within the drains confirm that the groundwater-surface water connectivity is maximised due to the lateral drain surface area. The fact that radon concentrations decrease to levels indicative of non-groundwater radon sources once water levels decrease to -0.35 m AHD (equivalent to where baseline surface water levels sit) supports evidence for shallow lateral groundwater flow as the main source of groundwater in this system, and not upwelling from deeper aquifers. Although this relationship between radon and groundwater table position has not been demonstrated before, such a trend supports evidence for the transmissivity feedback mechanism, which is often revealed by strong correlations between groundwater table levels and stream discharge [*Laudon et al.,* 2004].

The resulting relationship between surface water ^{222}Rn concentrations and groundwater table position provide a theoretical basis for modelling groundwater discharge. However, the direction of this relationship should be viewed only from high groundwater levels returning to baseline levels (as indicated by the arrows). This is because the rising limb of the groundwater level occurs with a simultaneous rise in surface water levels and no hydraulic head develops. Therefore, the relationship predicting radon concentrations with depth is only applied during the receding phase of the hydrograph. A default concentration equivalent to dilution concentrations were applied during the rising limb. Applying the equation derived from Figure 3, our biweekly surface water radon observations were used to estimate daily concentrations. By simulating continuous surface radon concentrations throughout the year, a detailed radon mass balance was constructed to estimate groundwater discharge into the drains (Figure 3).

For the radon mass balance calculations, the surface water radon concentration supported by non-groundwater source (sediment diffusion, ^{226}Ra decay, and hyporheic exchange) was estimated at 1,962 dpm m^{-3}, which is the average concentration of the three lowest discrete radon measurements (supplementary material). The same assumption has been applied to other studies where the lowest values are used as an indicator of “background” radon/radium activity [*Peterson et al.,* 2008; *Schmidt et al.,* 2010]. This approach was supported by the known lower groundwater depth relative to surface water depth in the drains during the times of these low concentration measurements, and significantly higher radon concentrations observed with higher groundwater depths.

The modelled groundwater discharge rates produced by the relationship between measured radon concentrations and groundwater depth provides a detailed analysis of groundwater discharge dynamics over an entire year at a resolution often only achieved in short term tracer time series (Figure 4). The results from the groundwater model confirm our hypothesis that flood events drive a large portion of the groundwater flux into surface water. Variations in rainfall events significant enough to increase water table height triggered significant variation in groundwater discharge by over five orders of magnitude (Figure 4). The largest variation observed was during the floods of January and June, when the groundwater discharge rate ranged from 0-15 m^3 day^{-1} pre flood to 1,000-27,200 m^3 day^{-1} immediately post-flood (Figure 4B). During peak groundwater discharge after these flood events, the groundwater aerial flux ranged from 185-325 cm d^{-1}, which given the high hydraulic conductivity of 545 ± 275 cm d^{-1} of surface layers at this site [*Johnston et al.,* 2009] suggests the groundwater flux is maximised within the limits of flow through the soil. Although measurements of hydraulic conductivity are highly heterogeneous, this finding gives confidence in our model derived groundwater aerial flux estimates. This also highlights the usefulness of geochemical tracers to estimate groundwater discharge in systems with heterogeneous hydraulic conductivities due to cracks and macropores.

The rate of groundwater flux varies seasonally in many natural floodplains and wetlands, as factors such as sustained groundwater storage after rainfall and evapotranspiration generally control GW-SW interactions [*Ramberg et al.,* 2006; *Krause et al., 2007a*; *Saha et al., 2012*; *Ludwig et al., 2015*]. The distinct pulses of groundwater discharge observed at this site highlight the responsiveness of groundwater to individual rainfall events, which prevents the development of a clear seasonal pattern. The rapid return of groundwater levels to baseline conditions due to forced pumping ensures that there is no groundwater storage in the top profile of the soil, and hence no sustained groundwater input after rainfall in contrast to more natural drainage conditions [*Saha et al.,* 2012].

Many other studies have reported noticeable increases in groundwater discharge during wetter periods, usually based on one or two surveys over contrasting conditions [*Peterson et al.,* 2010; *Santos and Eyre,* 2011; *Santos et al.,* 2011; *Atkins et al.,* 2013; *Sadat-Noori et al.,* 2015; *Jeffrey et al.,* 2016]. The amplitude of variation between baseline and flood groundwater discharge rates in the aforementioned studies ranged between 2 and 145-fold. In our continuous study, the average groundwater discharge rates of 1,600 and 4,900 m^3 d^{-1} during periods of flooding (within 5-10 days after floods) were 136 to 230-fold greater than the average baseline rates (7 and 36 m^3 d^{-1}), making groundwater discharge in the present study the most sensitive yet to rainfall input. Such an extreme groundwater-rainfall response appears to be most pronounced in areas where localised shallow groundwater dominates subsurface flow into relatively small tributaries or drains. High groundwater discharge rates following rainfall in this study is likely enhanced due to two major hydrological alterations to floodplain drainage: (1) forced pumping of surface water against natural flood levels creating a steep hydraulic head (up to 0.55 m), and (2); an extensive and dense network of drainage canals which expand the surface area of groundwater discharge.

4.3 Uncertainties and challenges

The ratio between groundwater discharge and surface water discharge (Figure 4C) provides a quality control measure and reality check of our minimum and maximum range groundwater flux model. By definition, plausible groundwater discharge rates should not exceed surface discharge. Throughout the study period, minimum groundwater discharge estimates exceeded surface discharge by on average 5% during periods of peak groundwater flow (Figure 4C). In contrast, the average maximum groundwater discharge exceeded surface discharge by ~60% (excluding >200% periods) during these peak groundwater flow periods. These relatively minor occurrences of numerical instability are similar to the 2% uncertainty in discharge measurements and 21% error in the endmember concentrations. Since the highest surface water radon concentrations observed is equivalent to average groundwater radon concentrations makes the maximum groundwater model more likely to reach (and exceed) 100% of surface water discharge when these conditions occur. This also suggests that groundwater input likely occurs throughout the entire drainage area rather than at the most upstream site, resulting in a higher probability that the maximum estimate overestimates groundwater discharge in this system. The highest rates of estimated groundwater discharge (provided by the maximum range) exceeded surface water discharge by 200-380% during short periods, suggesting unrealistic groundwater discharge values estimated from our model. Such numerical instability is an artefact of the maximum mass balance model, which can occur because the groundwater flux is being solved by linearizing the first order derivative of the radon mass balance equation [*Frei and Gilfedder*, 2015]. However, the frequency of this occurrence was only 4% over the year, and occurred when rates of surface discharge was reduced during pumping of flood water at high water tables (highly unstable surface water hydrology). The coinciding rates of maximum groundwater discharge during these exceedance periods were in the range of 830-31,175 m^3 d^{-1}, which are realistic values compared the maximum surface discharge recorded at 35,336 m^3 d^{-1} during the study period.

The three main sources of error that were applied to the final groundwater discharge estimates include error from the radon simulations (model), error from the variability in the groundwater radon concentrations (endmember), and assumptions within the radon mass balance calculation (section 2.7). Uncertainty estimates of final groundwater discharge are complicated by the fact that model and endmember error are based on observational data whereas uncertainty associated with the radon mass balance technique is reported in the form of an upper (maximum) and lower range (minimum) in possible groundwater discharge. Therefore, an absolute value for final groundwater discharge is difficult to report when using the minimum and maximum approach. Since model and endmember errors are proportional to the magnitude of the groundwater flux, these errors were applied to both the minimum and maximum model. Results from the uncertainty analysis provided in Table 1 demonstrates how the model and endmember error are minor relative to the difference between minimum and maximum estimates. The upper and lower ranges in model, endmember and combined error vary by a factor of 1.4 to 1.8, 1.5, and 1.7 to 2.1 respectively, and the minimum and maximum range vary by a factor of 2.5 to 3.1. The resolution provided by the difference in the minimum and maximum range (factor of ~3) is comparable to groundwater discharge estimates based on short term high-frequency radon time series [*Peterson et al.,* 2008; *Peterson et al.,* 2010; *De Weys et al.,* 2011]. Over longer time scales (annual) groundwater discharge estimates from studies using water balance approaches have reported ranges anywhere from a factor of 2 to 24 [*Laczniak et al.,* 2001; *Zhu et al.,* 2007].

Another source of uncertainty within the radon mass balance calculations is assigning the correct *k* value to estimate radon evasion. A sensitivity analysis was carried out on different transfer velocity (*k*) models for radon to determine the variation in maximum groundwater discharge and constrain realistic values (Table 1). The *k* value for the first scenario one was empirically derived from the combined parameterisation of *Raymond and Cole* [2001] and *O'Connor and Dobbins* [1957] as a function of wind speed (*w,* at 10 m), current velocity (*c*), and depth (*d*). This yielded the highest flux estimate for radon evasion and resulted in annual groundwater discharge estimate exceeding total surface discharge by 311%, which is clearly unrealistic. Wind speed was discounted in the second scenario (current velocity and depth only), which reduced annual groundwater discharge by a factor of 3.6 and provided a more realistic calculation of 87% of total surface discharge (Table 2). The reasoning behind excluding wind is based on the sheltered site characteristics of the drains. This has been demonstrated in other similar canal sites where the high banks caused low evasion of radon from the surface nearly equivalent to diffusion [*Burnett et al.,* 2010]. Wind speed was measured above the canopy of the vegetation and walls of the drains and is likely not representative of wind speeds (if any) surface waters were exposed to. The effect of current velocity, although intermittent, is likely an important component of our evasion calculations at our site due to the narrow and shallow nature of the drains. Hence, radon evasion driven by a diffusive concentration gradient only was used between periods of surface water discharge.

4.4 Impact of drainage density on water balance

Increased hydrological connectivity due to extension of the drainage network via artificial drains is widespread in agricultural and urban landscapes [*Levavasseur et al.,* 2012; *Kaushal et al.,* 2014]. An increase in drainage density can impact catchment water budgets by altering the rates and pathways of water movement through the landscape [*Blann et al.,* 2009]. Such extensions of the natural drainage network typically increase peak surface discharge rates, accelerate surface runoff, and reduce surface storage [*Krause et al.,* 2007b; *Blann et al.,* 2009; *Schottler et al.,* 2014]. In these hydrologically modified systems, flood events are generally more intense and return to baseline conditions rapidly, consistent with the surface hydrology results found in this study.

The effect of drainage density on groundwater discharge rates has seldom been discussed before. Model simulations of groundwater dynamics in a lowland-floodplain have predicted that higher drainage density increases surface water-groundwater interactions and reduces groundwater recharge [*Krause et al.,* 2007b]. When aerial groundwater discharge rates from the current study were compared with other inland aquatic landscapes, two patterns were revealed; 1) groundwater input is greater under wet conditions, and 2) sites with the largest reported groundwater fluxes also have the highest drainage density (Table 3). Here, we compare with a limited number of studies that have used the radon mass balance technique to compute groundwater fluxes reported as a normalised aerial flux (cm d^{-1}), and focused on lowland aquatic landscapes such as wetlands, floodplains, and estuaries rather than streams with steep topographic relief. Groundwater discharge is highly variable between sites as it is strongly governed by localised catchment properties, including soil transmissivity and surface water flow dynamics [*Krause and Bronstert,* 2007]. However, drainage density may be an important underlying factor defining lowland groundwater dynamics (Table 3).

Unique to our study is the substantial increase in drainage density from 2.1 to 12.4 during flood conditions, when an increase in hydrologic connectivity facilitates comparatively larger groundwater input (Table 3). This drainage density is much larger than previous studies and may explain the higher groundwater discharge rates. A conceptual model of surface water and groundwater flood response was developed to illustrate how increased drainage modifies the surface water and groundwater flows (Figure 6). The shortened yet amplified hydrograph of surface water discharge followed by rapid recovery is consistent with the understanding that expanded drainage generates "flashier" hydrographs in watersheds than natural drainage [*Levavasseur et al.,* 2012; *Kaushal et al.,* 2014]. As a consequence, increased drainage results in a shorter floodwater residence time, whereas most natural floodplains with smaller drainage density often remain inundated for prolonged periods (up to a month) after a flood event [*Ludwig et al.,* 2015; *Karim et al.,* 2016].

The groundwater discharge response from an artificially drained floodplain illustrated in Figure 6 represents the simplified discharge patterns observed at the study site during flood events. We suggest that there are a number of important differences in the response of groundwater discharge to floods in floodplain catchments with different drainage density.

Firstly, there is a small increase in groundwater discharge from artificially drained floodplains which represents the "first flush" of groundwater during the rising limb of the hydrograph. This is due to the close hydraulic relation of groundwater with surface dynamics specific to this floodplain (Figure 2), although a rapid response in groundwater levels to a pulse event has been noted in a few studies on drained landscapes [*Wilcox et al.,* 2006; *Kaushal et al.,* 2014]. The mobilization of groundwater following rain events has also been reported in unmodified watersheds using the isotopic hydrograph separation technique and multiple tracers [*Sklash et al.,* 1986; *Brown et al.,* 1999; *Burns et al.,* 2001]. However, these observations are primarily reported in small, steep catchments that don't experience flooding and where stored groundwater is displaced by slope water table development [*McDonnel,* 1990]. In our simplified representation of a natural drainage floodplain we assume that this effect is dampened in flat landscapes with slower surface discharge. Secondly, the duration for which groundwater discharge is impeded by floodwater inundation is shorter for the artificially drained floodplain. This phenomenon happens in both floodplains when the surface water levels are above the surface as there is no positive vertical hydraulic head to drive groundwater discharge into surface water. This type of reduction in groundwater discharge may occur daily on a smaller scale in estuaries during flood tide or storm surges [*Wong et al.,* 2013; *Johnston et al.,* 2005]. Thirdly, groundwater discharge is greatly amplified during flood recession compared to the natural floodplain, as drainage of surface water exceeds groundwater discharge creating a steep hydraulic head. Lastly, the rapid return to negligible groundwater discharge post-flood is an important aspect on the effect artificial drainage has on the groundwater dynamics in this system, as groundwater is quickly disconnected from surface water. In more natural systems, the recession limb extends over a longer period and groundwater input into surface water is sustained, which can contribute significantly to surface flow during drier periods [*Sophocleous,* 2002; *Brannen et al.,* 2015].

The impact of drainage density on groundwater recharge has implications for management of irrigated land in areas where groundwater management is a challenge, either due to salinization associated with rising water tables [*Hatton and Nulsen,* 1999] or falling groundwater tables affecting water resource availability or the health of groundwater-dependent ecosystems [*Boulton and Hancock,* 2006].

4.5 Water balance

With continuous measurements of water balance components, our data provided insight into the interplay between water transport pathways that would not be realized in static water balance calculations. The radon modelling approach gives insight into groundwater discharge that is often a small component that is difficult to be quantified by difference in the other terms of the water balance. Consequently, total groundwater contribution to the water balance was represented over an annual cycle while providing ranges to highlight radon mass balance model uncertainties.

The total annual water balance amounted to 1,745,000 m^3 of total water inputs and 1,373,934 m^3 of total water outputs, leaving a surplus of 365,764 m^3 or around 21% unaccounted for (Figure 7A). Propagating all the errors from the loss terms resulted in an estimated uncertainty of 45,000 m^3, therefore the unaccounted for water is not represented by cumulated water balance errors.

Given the distinct shifts in surface water and groundwater hydrology driven by floods throughout the year, the total water balance was further divided into dry periods (daily rainfall <10 mm) that represented 253 days (Figure 7B) and flood events that represented 83 days (inclusive of flood and recovery to baseline conditions) of the study period (Figure 7C). The unaccounted for water surplus calculated during the flood periods clearly indicates that the total annual water surplus originated from these extreme hydrological events. Subsurface storage may have accounted for a portion of this unaccounted flux, however this is unlikely as total storage levels would recover back to baseline levels (Figure 2). The remainder of the discrepancy is likely attributed to overbank flow during surface inundation, as floodgates and levees fail to isolate the sub-catchment from adjacent creeks during large flood events.

Most of the annual discharge (75%) occurred during the flood events (Figure 7). Streamflow dominated by episodic rainfall events is a common feature of river systems in Australia, with wet seasons contributing >90% of stream flow in some catchments [*Karim et al.,* 2016]. Groundwater discharge was found to represent between 31 and 87% of the total surface water discharge, which is a substantial fraction of the surface water budget considering global estimates of 0.01 to 30% of river discharge [*Zektser and Loaiciga,* 1993]. Overall, groundwater discharge made up 9-24% of the total water balance (Figure 7), which is within the range of 2 to 30% reported from other watershed budgets [*Santos et al.,* 2008; *Saha et al.,* 2012]. Importantly, groundwater discharge dominated the water balance (i.e., >50% of the total flows) during 6-12% of the time.

The importance of quantifying groundwater discharge continuously over an annual cycle is demonstrated in the final water balance. If annual groundwater estimates were based on samples from normal surface flow conditions with minor rainfall, total groundwater discharge would be underestimated by 72 to 76%. Short term flood events had an occurrence frequency of only 1:9 days of the year yet contributed ~74% of the total groundwater discharge (Figure 7). Although many short term studies have found that flood events drive large groundwater discharge rates that are significantly higher than baseflow conditions (Table 3), this has been the first attempt made to upscale the contribution of these events to annual groundwater discharge.

5. Conclusions

In this paper, we have calculated surface water-groundwater exchanges on an annual timescale using continuous radon and groundwater level measurements. This approach is less-labour intensive and, provides additional detail compared to previous methods based on discrete samples only. Groundwater discharge rates varied over five orders of magnitude during our study, demonstrating that high temporal resolution is required to accurately estimate water balances or identify processes controlling groundwater discharge in hydrologically modified floodplains.

Drained floodplains are a common feature in agricultural landscapes and accelerate not only surface water flows, but also groundwater discharge. Groundwater discharge was driven by precipitation events that caused elevation of the water table and stimulated rapid drainage of surface waters. Peak groundwater discharge rates as high as 325 cm d^{-1} were observed, and were comparatively higher than other lowland groundwater studies during flood events. Groundwater discharge displayed large variability, fluctuating between 0-100% contributions to surface discharge following rain events. Our finding that 72 to 76% of the groundwater discharged over an annual cycle occurred after flood events and was enhanced by artificial drainage provides valuable information for improving management of groundwater in hydrologically modified floodplains.

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Tables

Table 1: Results of uncertainty analysis from model, endmember, and combined error on the annual estimate of groundwater discharge (m^3) for the minimum and maximum radon mass balance scenarios. Differences between the lower (μ - σ) and upper ($\mu + \sigma$) one standard deviation from the mean (σ) for each scenario in model and endmember were propagated to a combined error estimation.

Condition	Model error		Endmember error		Combined error	
	MIN_m	**MAX_m**	**MIN_e**	**MAX_e**	**$MIN_m MIN_e$**	**$MAX_m MAX_e$**
$\mu - \sigma$	123,006	302,743	182,299	522,026	105,627	268,049
μ	148,055	423,968	148,055	423,968	148,055	423,968
$\mu + \sigma$	176,183	552,016	124,642	356,922	184,652	568,506
Difference $\mu - (\mu - \sigma)$	17%	29%	23%	23%	29%	37%
Difference $(\mu + \sigma) - \mu$	19%	30%	16%	16%	25%	34%

Table 2: Different radon evasion scenario's for the calculation of maximum groundwater discharge (Q_{GW-MAX}) and the impact on annual Q_{GW-MAX}. Final Q_{GW} reports the range in acceptable annual groundwater discharge estimates derived from Q_{GW-MIN} and Q_{GW-MAX}.

Evasion method	Average *k* (m d^{-1})	Average evasion (dpm m^{-2} d^{-1})	Average Q_{GW} flux (m^3 d^{-1})	Annual Q_{GW} flux (m^3)	Q_{GW}/Q_{SF} (%)
Q_{GW-MAX} – w+c+d	0.83 (0.36-6.21)	26,889	4,477	1,509,176	311
Q_{GW-MAX} – c+d	0.39 (0.12-5.56)	16,626	1,258	423,968	87
Q_{GW-MAX} Diffusive,	0.12 ± 0.02	2,527	864	291,400	60
Q_{GW-MAX} No evasion	0	0	555	187,243	47
Q_{GW-MIN}	0	0	439	148,055	31
Final Q_{GW}		**-**	**439-864**	**148,055 – 423,968**	**31 - 87**

w = wind, *c* = current velocity, *d* = depth

Table 3: Aerial groundwater flux and associated drainage density (drainage length km/catchment area km^2) of other studies using a radon mass balance approach.

Environment	Drainage density (km km^{-2})	Qgw (cm day^{-1})		Study
		Baseflow	Wet conditions	
Coastal plain wetland, Australia	n/a	0.3	2	Gilfedder et al., [2015]
Coastal lake, Australia	n/a	0.67		Perkins et al., [2015]
Estuary, Australia		11		Wong et al., [2013]
River, Australia	0.04	3.6		Cook et al., [2006]
Tidal creek, Australia	0.28	35 ± 12	56 ± 13	Sadat-Noori et al., [2015]
Remediated coastal wetland, Australia	0.7-0.9	1.3 ± 0.3	51.4 ± 12.5	Jeffrey et al., [2016]
Altered coastal floodplain, Australia	1	0.58 ± 14.2		Atkins et al., [2013]
Tidal river, Australia	1.7	26.8		Makings et al., [2014]
Drained agricultural swamp, Australia	2.75	2.8-129	11.4-328	De Weys et al. [2011]
Drained tidal creeks, Florida		4-8	84-125	Peterson et al., [2010]
Altered coastal floodplain, Australia	2.1-12.4	0-35	185-325	This study

n/a means there is no drainage density as the area is mostly covered by an expanse of water

Figure captions

Figure 1: Map showing location of the sub-catchment within the Tweed Valley floodplain (modified from Smith and Melville, 2004) and schematic of drainage network within the study site.

Figure 2: Time series of hydrological parameters measured from 26 November 2014 to 26 October 2015. Shaded areas highlight flood events which caused surface inundation (flooding of soil surface above the drains, >0.15 m AHD).

Figure 3: The relationship between measured surface water radon concentrations and groundwater position. Data was fitted with a four parameter Gaussian curve equation which was applied to continuous groundwater depth measurements to model surface water radon concentrations over the annual cycle. Shaded area represents equation uncertainty.

Figure 4: Groundwater discharge modelling results, showing: (A) predicted surface water radon concentrations modelled from equation derived in Figure 3; (B) the average minimum (blue) and maximum (red) groundwater discharge rates confined by the upper and lower uncertainties (bold lines) for both extreme scenarios (note the log scale on the y-axis); (C) the average minimum and maximum groundwater discharge rates as a percentage of surface discharge; (D) the average minimum and maximum vertical flux of groundwater (note the log scale on the y-axis).

Figure 5: Summary of cumulative output water budget terms expressed as a fraction of total annual rainfall against cumulative annual rainfall (input term) over the hydrological year (2014-2015). The total surface discharge represents contributions from both surface and groundwater flows.

Figure 6: Conceptual model demonstrating the effect of hydrological connectivity on the hydrograph trends of surface water and groundwater discharge during a flood event. Blue

represents the response from the studied catchment with extensive artificial drainage and green represents a floodplain catchment retaining natural drainage. The dark shaded grey bar signifies the duration of floodplain inundation after a rain event for artificial drainage and lighter grey the inundation duration in a natural floodplain. Graph not to scale.

Figure 7: Water balance diagrams for the whole hydrological year (A) including the water budget partitioned into non-flood (B) and flood (C) events. S refers to the discrepancy in the water balance where a positive value indicates a surplus of water input and negative a deficit in water input. P = precipitation, E = evaporation from drains, ET = evapotranspiration from vegetation, D = discharge, GW = groundwater discharge, and L = leakage from surrounding creek.