Characteristics and controls of variability in soil moisture and groundwater in a headwater catchment

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Abstract

This paper presents experimental results from a headwater research catchment in New Zealand. We made distributed measurements of streamflow, soil moisture and groundwater levels, sampling across a range of aspects, hillslope positions, distances from stream and depths. Our aim was to assess the controls, types and implications of spatial and temporal variability in soil moisture and groundwater.

We found that temporal variability is strongly controlled by the climatic seasonal cycle, for both soil moisture and water table, and for both the mean and extremes of their distributions. Groundwater is a larger water storage component than soil moisture, and the difference increases with catchment wetness. The spatial standard deviation of both soil moisture and groundwater is larger in winter than in summer. It peaks during rainfall events due to partial saturation of the catchment, and also rises in spring as different locations dry out at different rates. The most important controls on spatial variability are aspect and distance from stream. South-facing and near-stream locations have higher water tables and more, larger soil moisture wetting events. Typical hydrological models do not explicitly account for aspect, but our results suggest that it is an important factor in hillslope runoff generation.

Co-measurement of soil moisture and water table level allowed us to identify interrelationships between the two. Locations where water tables peaked closest to the surface had consistently wetter soils and higher water tables. These wetter sites were the same across seasons. However, temporary patterns of strong soil moisture response to summer storms did not correspond to the wetter sites.

Total catchment spatial variability is composed of multiple variability sources, and the dominant type is sensitive to those stores that are close to a threshold such as field capacity or saturation. Therefore, we classified spatial variability as *‘summer mode’* or *‘winter mode’*. In *summer mode*, variability is controlled by shallow processes e.g. soils and vegetation. In *winter mode*, variability is controlled by deeper processes e.g. groundwater pathways and bypass flow. Double flow peaks observed during some events show the direct impact of groundwater variability on runoff generation. Our results suggest that emergent catchment behaviour depends on the combination of these multiple, time varying components of variability.

1 Introduction

Hydrological processes, including runoff generation, depend on the distribution of water in a catchment, in space and time. Understanding the distribution and its effects on dominant processes is a prerequisite for identifying organising hydrological principles (Troch et al., 2008) and building hydrological models that produce "the right answers for the right reasons" (Kirchner, 2006). However, water stores and fluxes are typically characterised by high complexity and variability at all scales (e.g. Grayson et al., 2002; Zimmer et al., 2012). The high variability of soil- and ground-water has far reaching implications for hydrological measurement, prediction and modelling. Most measurements of soil moisture or groundwater are made at the point scale, and so high variability makes it difficult and costly to estimate spatial average values. However, studies into controls on variability can give insights into the best monitoring locations and strategies to estimate spatial averages (e.g. Teuling et al., 2006 for soil moisture), and may allow us to identify sites likely to mirror the mean wetness conditions of the catchment (Grayson and Western, 1998).

Hydrological models simulate water fluxes integrated over some "model element" scale; so where variability exists below that scale, model fluxes will differ from point-scale measurements (Blöschl and Sivapalan, 1995; Western et al., 2002). This makes it difficult to compare model simulations against measured data. The same scale sensitivity affects climate models, which use land surface water content as a boundary condition (Seneviratne et al., 2010). In addition, the prevalence of high nonlinearity and thresholds in hydrological responses means that simple averaging of water content is not sufficient. For example, integrated drainage fluxes derived from soil moisture patterns with realistic variability and spatial organisation exceed those estimated from uniform soil moisture fields (Bronstert and Bardossy, 1999; Grayson and Bloschl, 2000). Model descriptions of relationships between mean soil moisture and drainage must therefore be altered to take account of soil moisture variability (e.g. Moore, 2007; Wood et al., 1992) and organisation (Lehmann et al., 2007), and may need to change seasonally as soil moisture variability changes (McMillan, 2012). Similarly, averaging of soil texture or water-holding properties should take spatial organisation into account. Threshold relationships between water content and runoff generation, which have been widely observed at the point scale, should be smoothed at the model element scale to reflect spatial variability (Kavetski et al., 2006). The critical point here is that multiple sources and characteristics of variability may exist in any catchment. To understand and model the emergent, catchment-scale processes they create, we must understand how the individual components of variability interact and change with time.

A well-established strategy to improve our understanding of hydrological variability and processes is through the development of densely instrumented research catchments (Tetzlaff et al., 2008; Sidle, 2006; Warmerdam and Stricker, 2009). Such sites expose interrelations and patterns in hydrological variables, and allow us to test hypotheses on catchment function. In recent years, improved sensor and communication technologies have increased our ability to capture space and time hydrological variability (Soulsby et al., 2008). While acknowledging the importance of breadth, as well as depth in hydrological analysis (Gupta et al., 2013), intensively-studied catchments remain a critical part of hydrological research.

In New Zealand, experiments in research catchments have uncovered the importance of vertical flow and the displacement mechanism for streamflow generation, using applied tracers (Woods et al., 2001; Mahurangi catchment) and isotope measurements (McGlynn et al., 2002; Maimai catchment). The subsequent incorporation of our revised process understanding into conceptual models of the catchments has emphasised the need to measure variability and dynamic response in groundwater as well as soil moisture (e.g. Graham and McDonnell, 2010; Fenicia et al., 2010). Groundwater dynamics and subsurface flow pathways are a key control on runoff generation and flow dynamics in a variety of different catchments (Onda et al., 2001; Soulsby et al., 2007), with strong evidence coming from hydrochemical analysis of streamwater. The hydrology of the riparian zone may be particularly sensitive to groundwater connections (Vidon and Hill, 2004). While previous NZ catchment studies have measured groundwater response in a limited number of locations (Bidwell et al., 2008) or without simultaneous surface water measurements (Gabrielli et al., 2012), a joint data set of spatio-temporal surface and groundwater measurements did not previously exist in New Zealand.

The results presented in this paper, from a research catchment in the headwaters of Waipara catchment, provide data to characterise and test hypotheses on variability and model representation of integrated surface water-groundwater physical systems. Such models are in high demand for management applications, as local governments must set allocation limits and manage supply under increasing demands for water. Although surface water and ground water systems have, historically, often been managed independently, there is now recognition that extractive use from either source impacts on the whole system (Lowry et al., 2003).

The aims of this paper are therefore to: (1) Present initial experimental data of surface and ground water responses from a research catchment in the alpine foothills of New Zealand (2) Assess the types of spatial and temporal variability in soil moisture and groundwater in this headwater catchment, the factors that control the variability, and the implications for modelling.

1.1 Soil moisture variability

New Zealand has some well-known experimental catchments, which offer information into causes and effects of hydrological variability, focused on the soil zone. In the Mahurangi catchment in Northland, Wilson et al. (2004) compared the variability of gridded soil moisture measurements in time vs in space. They found that temporal variability was approximately 5 times greater than spatial variability. Temporal variability was highly predictable, and explained by seasonality; whereas spatial variability was less easily predictable, only partly explained by terrain indices. In the same catchment, Wilson et al. (2003) compared variability of soil moisture at 0-6 cm depth vs 30 cm depth, and found differences in distribution and low correlations between the two depths. At Maimai catchment in Westland, nested arrays of tensiometers were used to estimate variability in the depth to water table. High variability was found within nests (plot scale) and between nests (hillslope scale) (McDonnell, 1990; Freer et al., 2004).

Some characteristics of the New Zealand climate and landscape may result in locally important controls on variability. Aspect is important in New Zealand hill country, due to high radiation and prevailing wind direction. Typical Penman PET is 35-50% greater on Northern than Southern facing slopes (Jackson, 1967; Bretherton et al., 2010), or more for sites exposed to the prevailing WNW wind (Lambert and Roberts, 1976). At one site, these differences translated into mean soil moisture differences of 10% (Bretherton et al., 2010). In a similar environment to the catchment described in this paper (i.e. Eastern foothills of the Southern Alps, greywacke geology), aspect-induced microclimate differences were found to promote physical and chemical soil differences, with stronger leaching and weathering on south facing slopes (Eger and Hewitt, 2008).

Controls on soil moisture are varied and may affect soil moisture mean (in either space or time), distribution (Teuling et al., 2005) and dynamics such as recession, stability or recharge rate (Kim et al., 2007). Examples from previous (international) studies are given in Table 1. Controls can also interact, such as soil type and topography (Crave and GascuelOdoux, 1997). Even though new technologies are becoming available to measure soil moisture and its variation on larger scales, including remote microwave sensing (Njoku et al., 2002) and electrical resistivity tomography (Michot et al., 2003), there is still no accurate way of predicting soil moisture patterns, with studies typically predicting less than 50% of the spatial variation.

High variability in soil moisture has many implications for hydrological process understanding and modelling. There is a large body of work investigating causes of low vs high variability, without attempting to predict exact spatial or temporal patterns, often using geostatistical methods to quantify the magnitude and the scales of variation (e.g. Western et al., 1998; Brocca et al., 2007). Causes of high variability have been found to be: dry conditions (Brocca et al., 2007), mid-wetness conditions (Ryu and Famiglietti, 2005; Rosenbaum et al., 2012), wet or dry conditions conditional on climate, soil and vegetation types (Teuling and Troch, 2005; Teuling et al., 2007), increasing scale (Famiglietti et al., 2008; Entin et al., 2000), aspects of land use and topography (Qiu et al., 2001), groundwater influence and contrasts between groundwater influenced/uninfluenced areas (Rosenbaum et al., 2012).

1.2 Groundwater variability

Studies of variability in groundwater dynamics are less common, reflecting the greater difficulty and expense in measuring groundwater levels, but a wide range of controls on groundwater levels have been identified. Detty and McGuire (2010a) considered surface topography controls, by dividing the landscape into landform units, e.g. footslopes, planar backslopes, or convex shoulders. They found statistical differences in metrics of water table hydrograph shape between different landform units. The hydrographs increased in duration and magnitude from shoulders to foot slopes, but were most sustained on backslopes. The responses also differed between the growing and dormant seasons. Anderson and Burt (1978) showed that topography can control matric potential and downslope flow: at their field site, hillslope 'hollows' had specific discharge an order of magnitude higher than hillslope spurs. Fujimoto et al. (2008) found that topography interacts with storm size to control subsurface processes. For small storms, a concave hillslope stored more water than a planar slope and produced less runoff; whereas for larger storms, transient groundwater in the concave slope caused greater expansion of the saturated area than in the planar slope, and correspondingly greater runoff. Bachmair et al. (2012) drilled 9 transects, each of 10 shallow wells (< 2 m deep) to study the effect of land use and landscape position on variability in groundwater dynamics. They found that patterns of groundwater response in winter reflected expansion of saturated areas at the base of the hillslope, whereas in summer groundwater response was controlled by transient preferential flow networks and was highly spatially variable. The wells with the strongest response also varied between events. The relationship between topography and subsurface flow dynamics has been demonstrated theoretically (Harman and Sivapalan, 2009), although bedrock topography may be more important than surface topography (Freer et al., 2002; Graham et al., 2010; Tromp-van Meerveld and McDonnell, 2006b, a).

In areas with shallow slopes, other controls dominate, such as variability in recharge. Gleeson et al. (2009) tracked snowmelt recharge to groundwater using 15 bedrock wells in a humid Canadian catchment with flat topography. In addition to widespread slow recharge, they found fast, localised recharge in areas with both thin soils and fractured bedrock. Riparian soils can form a fast conduit to groundwater, where a higher fraction of gravel leads to hydraulic conductivities an order of magnitude higher than the hillslope soils (Detty and McGuire, 2010b).

Characteristics of the groundwater aquifers are also important. Winter et al. (2008) and Tiedeman et al. (1998) monitored 31 bedrock wells and found water table gradients caused by different geological units within a catchment. Even in headwater catchments, variability in groundwater dynamics has been found due to multiple underlying aquifers (Kosugi et al., 2011; Kosugi et al., 2008). In Plynlimon catchment in Wales, Haria and Shand (2004) found that groundwaters at 1.5 m, 10 m and 30 m depth were not hydraulically connected, and were chemically stratified, with distinct pH, electrical conductivity and redox characteristics. Different groundwater pathways to the stream could therefore be identified, including discharge from fractured bedrock, and upwelling into the soil zone causing rapid lateral flow.

1.3 Soil moisture – groundwater interactions and variability

The dividing line between stored water that is considered as soil moisture or groundwater is not well defined. Soil moisture is typically measured as volumetric water content at specific depths in the unsaturated zone, although soil moisture sensors can be subsumed by perched or deeper water tables. Here, we use groundwater synonymously with water table, referring to saturated subsurface layers, which may be above or below any soil/bedrock interface. Piezometers or shallow wells to measure groundwater level can be screened along their whole length (as in our study) or at specific depths if multiple perched or confined layers are suspected. Where the geology includes fractured rock or buried lenses of gravels, groundwater levels may be highly heterogeneous.

There are many processes by which soil moisture and groundwater interact. As soil water drains downwards, layers of low hydraulic conductivity may create perched water tables. Such layers include clay pans, and the soil/bedrock interface (Tromp-van Meerveld and McDonnell, 2006a). Macropores provide a fast route for surface and soil water to recharge groundwater (Beven and Germann, 2013). They may allow water to bypass confining layers or to flow quickly along them (e.g. lateral preferential flow along the bedrock interface found by Graham et al., 2010). If groundwater rises into upper soil layers, large increases in soil matrix porosity or macropores may ‘cap’ water table levels, as additional water is quickly transported to the stream (Haught and Meerveld, 2011). Lana-Renault et al. (2014) found in a Mediterranean catchment that patterns of near-surface saturation and transient water tables were affected not only by topographic controls but also soil properties and previous agricultural land use. The riparian zone facilitates mixing between soil water and groundwater, and tracers, temperature, electrical conductivity, flow gauging and head differences may all be used to quantify the interactions (Unland et al., 2013). Using modelling and tracer data, Binley et al. (2013) found that in a 200 m river reach the upper section was connected to regional groundwater, but lower section inflows were from local lateral and down-river flow paths.

Interactions between soil moisture and groundwater provide possible explanations for relationships between the two. Results from three Nordic catchments showed a consistent negative correlation between soil moisture content and depth to water table, so that soil moisture distributions could be described as a function of depth to water table (Beldring et al., 1999). Kaplan and Munoz-Carpena (2011) studied soil moisture regime in a coastal floodplain forest in Florida, finding that groundwater and standing surface water elevations were successful predictors of soil moisture using dynamic factor analysis and regression models. Model-based studies demonstrate how capillary-rise can lead to dependencies between groundwater level and soil moisture. Kim et al. (1999) used a hillslope model to show how gravity-driven downhill groundwater flow creates downslope zones with high water tables. In those areas, capillary rise keeps soil moisture content and evaporation rates high. Similarly, the model developed by Chen and Hu (2004) showed that soil moisture in the upper 1 m of soil was 21% higher when exchange between soil moisture and groundwater was included. They inferred that groundwater variability may drive soil moisture variability.

2 Study area

The Langs Gully catchment is located in the South Island of New Zealand, in the headwaters of the Waipara River that has its source in the foothills of the Southern Alps before emptying onto alluvial plains (Figure 1). Langs Gully is typical of the Canterbury foothills landscape. This area is the source of many rivers and aquifers that provide essential irrigation water for the drier and intensively farmed plains; however the hydrology of the area is poorly understood.

The 0.7 km^2 catchment ranges from 500 - 750 m in elevation, and is drained by two tributaries. Annual precipitation ranges from 500 to 1100 mm, with a mean of 943 mm. In winter the catchment has relatively frequent frosts and occasional snow. The land cover is grazed pasture for sheep and beef cattle farming, with a partial cover of sparse Matagouri (*Discaria toumatou*) shrub. The geology is greywacke, a hard sandstone with poorly sorted angular grains set in a compact matrix. Soils are shallow gravelly silt loams derived from the underlying greywacke, and were classified as midslope, footslope or spur (Figure 2), based on expert knowledge and the S-MAP New Zealand soils map (Lilburne et al., 2004), which uses soil survey data, and topography-based interpolation (Schmidt and Hewitt, 2004). The mapping also provided estimates of fractions of stone, sand and clay for each soil type. Fractions of stone and sand decreased from spurs to footslopes, while fractions of clay increased (Table 2). Stone and sand fractions increase with depth for all soils (e.g. Footslope constituents shown in Table 3). During installation of soil moisture sensors (Section 3.2), at 6 out of 16 locations there were found to be distinct gravel-rich layers within the soil profile.

3 Materials and Methods

The aim of our experimental design was to study the temporal and spatial variability in water storage within the catchment. We installed sensors to measure rainfall, climate variables, streamflow, soil moisture and depth of shallow groundwater. Our aim was to take measurements at locations representing the variability of hydrological conditions within the catchment, and where possible to co-locate sensors in order to understand relationships between different water stores. We selected two hillslopes for detailed measurements of soil moisture and shallow groundwater, with different aspects (North and South) (Figure 1).

To support the sensor data, we took aerial photos and used GPS mapping to create a digital elevation model of the catchment (Figure 2). Aerial photos were only taken on the slope above the north-facing sites, and GPS point spacing was also closer in this area. A soils map was created using a combination of nationally available data and a field survey (Figure 2).

3.1 Climate and flow monitoring

A compact weather station was located centrally within the catchment (Figure 1). It uses a Vaisala WXT520 Weather Transmitter, which measures wind speed and direction, air temperature, barometric pressure and relative humidity. A LiCOR LI200 Pyranometer measures solar radiation. Rainfall was measured using an OTA OSK15180T 0.2mm resolution tipping bucket gauge. All weather measurements were at 5 minute intervals.

Flow was measured at three locations within the catchment (Figure 1), all at 5 minute intervals. The gauge type was chosen according to the flow magnitude: the upper two gauges are 45 cm H flumes, the downstream gauge is a v-notch weir. Periodical manual gaugings were used to confirm theoretical flow rates at all three locations.

3.2 Soil moisture and shallow groundwater monitoring

Soil moisture and water table level were monitored by 16 instrument stations. The stations are divided into 2 groups; 10 on the north-facing slope, and 6 on the south-facing slope

Our typical measurement site included Acclima TDT soil moistures sensor at 30 cm (base of the root zone) and 60 cm, these were used with factory calibration as recommended by the manufacturer (Acclima, 2014). The sites also included a well drilled to a fixed depth of 1.5 m (except where a high fraction of stones prevented the full depth being reached) equipped with a Solinst Levelogger to measure water level. The wells were sealed for the top 0.5 m to prevent ingress of surface water, with open screening below this. On each hillslope, we centred the sites around a shallow gully surface feature, with sites in the centre of the gully and on each bank. The sites were designed in two rows, at 10 m and 20 m from the stream centreline (Figure 1). In this way, we aimed to sample across multiple variables of aspect, slope position and distance from stream. All sensors recorded at 5 minute intervals, which were typically aggregated to 15 minutes before further analysis.

3.3 Telemetry

Each station aggregates sensor data and discards unneeded data. Each group is associated with a 'master' station that polls the individual stations every 5 minutes for their sensor data. The master station comprises a Unidata Satellite NRT datalogger and a proprietary short-haul radio interface. The data received by the master station is stored temporarily in the logger until it can be relayed to a central database via satellite. Data in the central database is available to end users via internet and e-mail. To conserve power in the solar-recharged batteries, the sensors and radio system are only powered up to respond to data requests.

3.4 Study period

The data used in this paper were collected between March 2012 and July 2013 (Figure 3). Climate and flow data are available for 14 months prior to this date. The largest storm event in the study period occurred in August 2012, which brought 80.6 mm of rainfall in 2 days, approximately a 1-in-2 year rainfall event when compared against the 62-year daily rainfall record from Melrose station, 2.0 km from the catchment. The 2012-13 summer was unusually dry in many parts of New Zealand; but at Melrose the summer months December/January/February recorded a rainfall total of 196 mm, only marginally below the long-term average of 210 mm.

Some data gaps occurred during the study period, with short outages due to sensor or battery failure. A long outage occurred in the aftermath of the storm event in August 2012, which caused water damage to the telemetry system on the North facing slope.

3.5 Calculation of descriptive statistics

To provide an overview of the soil moisture content and groundwater level for different time/space locations, a selection of summary measures were used. To summarise the distribution of data, we calculated the median and 5th, 25th, 75th and 95th percentiles for each data series. This allows us to compare absolute soil water content and groundwater level between sites. However, we also want to compare the extent to which each location is likely to contribute to runoff; especially as runoff generation is typically conceptualised as a threshold process (Ali et al., 2013). We therefore additionally used statistics that described the wet extremes of the data. For soil moisture, we calculated the percentage of time that the soil was saturated, as this represents the condition where the location would generate both vertical drainage and overland flow. Soil saturation points were defined individually for each sensor, using the co-located groundwater well record to determine times when the water table intersected the sensor, and taking the average soil moisture reading at those times. These values were confirmed (and in two cases adjusted) based on visual inspection of the soil moisture time series. For groundwater level, we calculated the percentage of time that the water table level was above the 75th percentile. This quantifies locations where groundwater is closer to the surface and would therefore have faster lateral velocity according to typical findings that hydraulic conductivity decreases rapidly with depth (Beven and Kirkby, 1979).

To understand how total water storage in the catchment changes through the year, we estimated the water stored in the soil moisture and groundwater components. For soil moisture, we divided the catchment by soil type, according to the classification described in Section 2. For each type, we estimated soil depth as the deepest functional soil horizon described in the S-Map database (Lilburne et al., 2004). For each time step, we derived the total soil moisture volume as:

$$\text{Total Soil Moisture}[\mathrm{m}^3] = \sum_{SoilType} \sum_{Aspect} \left[\text{Area}[\mathrm{m}^2] \,.\, \text{Soil Depth}[\mathrm{m}] \,.\, \text{Fraction Soil Moisture}\right] \quad \textit{(Eq 1)}$$

Dividing by total catchment area then gave average depth of soil water.

For groundwater, we do not know the total aquifer depth, and therefore use instead groundwater depth above minimum recorded. For each time step, we derived the total groundwater volume above minimum as:

$$\text{Total Groundwater}[\mathrm{m}^3] = \sum_{Aspect} \left[\text{Area}[\mathrm{m}^2] \,.\, \sum_{\text{Wells}} \left(\text{GW level}[\mathrm{m}] - \text{Min. GW level}[\mathrm{m}]\right) \big/ \text{Number of wells} \right] \quad \textit{(Eq 2)}$$

Dividing by total catchment area then gave average depth of groundwater above minimum.

We recognise that this calculation involves a significant and uncertain extrapolation from the 32 soil moisture time series to the remainder of the catchment. However, given that the sensor locations sampled across aspect, distance from stream, and landscape position and depth, we anticipate that the estimated storage dynamics are a reasonable guide to true behaviour.

4 Results

4.1 Temporal controls on soil moisture and groundwater

Both soil moisture and groundwater level show strong variations over event and seasonal timescales. Figure 3 shows soil moisture, and depth to groundwater for the study period; for clarity we average by location the 32 soil moisture sensors and 14 water level sensors using eight and two series respectively.

In Figure 4, we show the summary measures, split by season. The summary statistics show that both the mean and extremes of catchment water storage vary seasonally. The yearly cycle of soil moisture (Figure 3) shows an extended wet season from April/May to November, followed by a slow drying until February when the catchment reaches its summer state. The return to wet conditions occurred over a very short time period during a May storm event. Water table dynamics also display a yearly cycle (Figure 4), although the range during any season is large compared to seasonal changes. As shown in Figure 4A, soil moisture quantiles are typically lowest in summer, and water tables are lowest in summer and autumn. The driest conditions in terms of extremes (Figure 4B) occurred in late summer for both soil moisture and water table, and remains low into autumn particularly for the water table, suggesting that the lowest potential for runoff generation occurs at that time. Note that the autumn season values represent an average between the wetter conditions of the 2012 autumn and the drier conditions of the 2013 autumn, for example mean autumn (March-May) soil moisture at 0-30 cm for the upper rows of sensors was 17.9 % for 2012, 15.2 % for 2013.

Rainfall events are superimposed on the seasonal cycle. In winter, the large events cause saturation at many of the soil moisture sensors, and water tables rise in many of the wells, including some in the upper row where the water table was previously lower than the well. In early summer, rainfall can return soil moisture and water tables to winter levels, but only briefly. In summer, the catchment response to rainfall is highly subdued.

The strong seasonality of catchment conditions is due to seasonality in PET. Although rainfall depths are similar throughout the year, in summer the combination of higher temperatures, high solar radiation and frequent hot, strong winds from the north-west contributes to seasonal drying of the catchment. The effects are illustrated by storm runoff depths in winter versus summer (Figure 5A). In summer, even large rainfall events produced almost no streamflow response. To demonstrate the effect of antecedent wetness on storm runoff depths, we plotted runoff depth against the sum of antecedent soil moisture storage (ASM) and storm precipitation (Figure 5B), following Detty and McGuire (2010b; their figure 4a). Antecedent soil moisture storage was taken as the Total Soil Moisture value from Eq 1. The results show a threshold relationship between ASM + precipitation and runoff depth, although it is not linear as was found by Detty and McGuire (2010b).

4.2 Spatial controls on soil moisture and groundwater

Figure 3 shows distinct differences between the water storage dynamics of the North and South facing slopes, and between the far-stream and near-stream rows of soil moisture sensors. The near-stream sensors on the South facing slopes showing more frequent and pronounced wetting events. For example, we defined a wetting event as a period of rainfall during which soil moisture rose by at least 3%, and used this criterion to compare number and size of events by aspect and distance from stream

(Table 4). We calculated events on a per-site basis, and then averaged across sites. South facing slopes had 33% more events that were on average 22% larger than North facing slopes.

Spatial controls act differently on different water stores. These differences are illustrated in Figure 6, using the same summary statistics as in the previous section, but grouping sites by aspect and distance from stream. We did not include water table statistics for the far-stream rows as water tables only rarely rose into the wells and therefore distribution estimates would not be accurate. Figure 6A shows that when comparing North facing vs South facing slopes, soil water content at 30 cm has similar distributions, but the underlying groundwater level is on average 20 cm closer to the ground surface for the South facing slopes, and has a smaller range. Spatial controls also act differently on average vs extreme conditions; e.g. average soil moisture on the South facing slope is similar at 30 cm and 60 cm depths (Figure 6A), but the fraction of time that the soil was saturated is 11% at 60 cm against 0.5% at 30 cm (Figure 6B). Note that the statistics describing the extremes of the data are highly variable between locations (e.g. some locations are saturated much of the time; others almost never), however we show averages by location to assist interpretation of the spatial control.

4.3 Temporal changes in total water storage and variability

To quantify the relative importance of different water storage components of the catchment, we calculated the average depth of water stored as soil moisture and groundwater using the method described in Section 3.5 (Figure 7A). The groundwater component dominates, with an average depth of 0.27 m against 0.15 m for soil moisture. The difference may be further enhanced given that the part of the soil moisture volume below wilting point is not likely to be mobilised. The difference is most pronounced in the wettest conditions, with groundwater storage peaking at approximately four times that of soil moisture. During the driest summer conditions, groundwater and soil moisture components have similar depths.

To visualise the changes in variability over time for each store, we plotted the time series of spatial standard deviation in soil moisture and groundwater; separated by aspect and sensor depth (Figure 7B,C). All stores have the highest standard deviation in winter, and the lowest in summer, as the range in values tends to be compressed as the catchment dries out. Previous studies have shown that the relationship between soil moisture and soil moisture standard deviation varies by catchment (Section 1.1). Soil moisture at 60 cm maintains a high standard deviation even during summer, as both slopes have one sensor that retains high soil moisture and therefore has a strong influence on the standard deviation value.

All of the soil moisture standard deviations rise sharply during rainfall events, especially in winter, which is due to saturation of some sensors, while others remained unsaturated. Accordingly, 30 cm North facing soil moisture has smaller rises in spatial standard deviation, as none of those sensors typically saturate. Groundwater standard deviation has different behaviour by aspect: on the North facing slope, rainfall events cause the standard deviation to rise, on the South facing slope, rainfall events cause the standard deviation to fall. This finding reflects that on the South facing slope, all wells react to rainfall events, albeit at different speeds, but on the North facing slope, behaviour is more variable with one well often showing no response (i.e. water table lower than 1.5 m), and other wells split between weak or strong responses.

4.4 Controls on variability

As was apparent from the time series of flow, soil moisture and water table depth presented in Section 3.1, there is significant spatial variability between different parts of the catchment as represented by the range of sensor locations, but this variability is not constant. In this section, we investigate the specific types of variability which occur, and seek to attribute them to different catchment conditions.

We found that an overarching driver of variability is the wetness condition of the catchment. As shown in Figure 5, there is a strong seasonal differentiation in runoff coefficients. This seasonal cycle determines which of the catchment water stores are active, and where the greatest scope for variability exists. To assist our description of the seasonal changes in variability, we selected one event which is typical of, and illustrates each variability type. We selected the following events: dry-period variability: 17-27 March 2013, 15.9 mm rainfall; wet-period variability: 5-25 October 2012, 164.9 mm rainfall; winter wet-up: 15-30 April 2013, 80.0 mm rainfall; recession period: 7 September – 5 October 2012.

4.4.1 Dry-period variability caused by partial catchment response

One type of variability occurs during the driest conditions monitored: that is, some locations show a hydrological response - an increase in soil moisture or water table rise - to a rainfall event, while the others show little or no reaction. The time of onset of this type of variability varies with depth for the soil moisture probes, i.e. 60 cm probes stop reacting earlier in the summer than 30 cm probes. The fact that shallow probes are more likely to react during dry conditions suggests that the variability is caused by infiltration of precipitation that only reaches a limited depth below the surface. An example is given in Figure 8A, which shows the response of selected sensors to the March rainfall event. Figure 8B shows a spatial overview of all sensor responses for the same event. For this event, 8 of the 30 cm soil moisture probes showed a strong response, compared to 3 of the 60 cm soil moisture probes and 3 of the wells. There were two locations where the 60 cm probes responded but the 30 cm probes did not; as water tables were always below 60 cm, these cases suggest macropore flow that bypassed the upper sensor. There are 4/10 of the 30 cm soil moisture probes on the North facing slope that showed no response, compared to 1/6 on the South facing slope. This difference may be due to drier antecedent conditions on the North facing slope; North facing sensors have a mean soil moisture of 9.6% prior to the rainfall event, compared to 11.4% for the South facing sensors. Soil texture differences related to aspect may also play a role: South facing sensor locations were found to have higher clay content and higher stone content than the North facing locations.

4.4.2 Wet-period variability caused by partial saturation and groundwater response speed

In winter, the catchment is typically in a continuously wet state, and all sensors respond to rainfall events, in contrast to the summer response. Variability between sensors is introduced because some locations experience saturation (either transiently or for prolonged periods), while others do not. Saturation is characterised by high peaks or plateaux in the soil moisture signal. For both the North and South facing slopes, saturation occurs earlier and more extensively for probes at 60 cm than at 30 cm, and is limited to the sites at 10 m from the stream, indicating a rise in the catchment water table to these probes, rather than transient or perched saturated layers in the soil column. Cross-checking against measured groundwater levels also shows that the peaks in the water tables reach the depths of the soil moisture sensors showing saturation, although they do not typically reach the land surface. Wells in the upper locations may also react at this time. The rise of the near-stream water table into the soil is consistent with our knowledge of the soil and bedrock structures, as there are no evident confining layers, rather an increase in cobbles and rock fragments with depth.

Figure 9 gives an example of the response of soil moisture and groundwater level to a series of storm events in October (3 distinct peaks over 15 days) occurring in the already-wet catchment. Saturation only occurs in 30 cm or 60 cm probes when lower probes also show saturation. 3 out of 4 locations where saturation occurred at the 60 cm probes in this event were locations that showed a water table response during the summer event previously described. All locations that had a water table response in the summer event also had a water table response during this event. The consistency of locations suggests that relative groundwater levels are maintained across seasons, with the same locations always the most likely to display a groundwater response. These locations were not related to the gully/ridge features in the catchment, in conflict with our prior hypothesis, but instead may indicate preferential groundwater flow paths which channel water from the upper slopes. Such preferential paths were previously reported at Maimai catchment where there is a clearly defined bedrock interface (Graham et al., 2010; Woods and Rowe, 1996); our results suggest a similar outcome in the Langs Gully catchment despite the gradual transition from soil to broken bedrock. The cross-slope gradients needed to generate the preferential paths could be caused by deeper bedrock structures, or by local areas with high permeability such as the gravel-rich soil layers observed during installation of the soil moisture sensors. At Maimai, suggested causes were temporary hydraulic gradients in the soil, and variations in vertical drainage due to patterns of soil moisture deficit (Woods and Rowe, 1996).

Figure 9A (third panel) shows distinct differences in the speed of the groundwater response between locations. In some locations, there is a fast groundwater peak followed by a fast decline. In other locations, the groundwater is slower to rise, reaching a peak approximately 24 hours later than the fast-response site, and much slower to decline. The characterisation of each site as either a fast or slow responder is consistent through the three consecutive events in Figure 9. During some storm events, these two response types cause a double peak, or prolonged flat peak, in the storm hydrograph (lower panel). The differing responses are mapped in Figure 9C. There is some spatial correlation with the saturation response shown in Figure 9B, whereby locations with a flashy groundwater response correspond to locations where saturation rose to the 60 cm soil moisture sensor. Locations where the water table was detected in the upper row of sensors were classified as slow groundwater responses (i.e. a later and prolonged peak), but they peak slightly before the downslope slow-response sites, which could indicate a delayed groundwater flow path from upslope.

Our results suggest that relative groundwater levels, and the classification of sites as fast or slow groundwater responses, are consistent between events. Previous work reviewed in the introduction (Section 1.3) showed that groundwater level can influence soil moisture distribution. We therefore hypothesise that groundwater behaviour might help to define distinct spatial zones of the catchment. To test this, we firstly classified sites by maximum groundwater level, separating sites where the water table rose as high as the 30 cm soil moisture probe at any point during the study period ('Saturating sites'), against those where it did not ('Non-Saturating sites'). We only used near-stream sites to remove the influence of distance to stream. Secondly we classified sites by speed of groundwater response, as described in the previous paragraph. Other sites where groundwater rarely responds were not included as only the peaks of groundwater responses are measured, and therefore these sites could not be easily classified. We calculated the distributions of the soil moisture and water table level for each classification (Figure 10). The results show that the Saturating vs. Non-Saturating classification clearly delineates two zones with consistent differences in soil moisture content at 30 cm and 60 cm, and water table level. The fast vs. slow groundwater response classification is much less distinct, with the two zones having similar soil moisture distributions. The slow groundwater response zone has slightly deeper water tables, although this is partly because it includes two far-stream sites.

4.4.3 Variability in seasonal dynamics: winter wet-up

The wetting up of the catchment at the start of winter is a major event (Figure 3). In 2013 this occurred in late April, quickly transitioning the catchment from its dry summer state, to the wet state that it maintained throughout the winter. The typical pattern for soil moisture is a sharp rise over less than 24 hours (e.g. Figure 11A, red lines), however some locations have a more gradual response (Figure 11A, blue lines). On the South facing slope, this sharp rise is reflected in a sharp water table rise in some locations, and a more gradual rise in others. On the North facing slope, the water table rises only gradually in all locations (Figure 11B,C). The two locations with gradual soil moisture response had a soil layer containing larger rocks (5-10 cm diameter) at 45-60 cm depth. This feature may promote fast drainage and therefore slow the soil wetting process.

The winter wet-up is a critical event in terms of flow prediction, as was previously shown in Figure 5 which illustrates the stark differences in run-off coefficients in winter vs summer. However, the spatial variation shown here in the speed and magnitude of the wet-up illustrates that it is a complex phenomenon which occurs differently for hillslopes with a different aspect.

4.4.4 Variability in event dynamics: recession characteristics

During a dry period, catchment soils, water table and flows undergo a recession. It is common to collate flow recessions, to specify a master recession shape which can then be used directly to calculate model parameters relating to baseflow generation. Recessions are typically expected to show a convex shape; initial drying occurs quickly from loosely-bound water, but drying slows as only more tightly-bound water remains. In the Langs Gully catchment, we were surprised to find strong variations in recession shapes. This is illustrated in Figure 12, which shows the recession shapes of soil moisture at 30 cm on the North facing slope after a September rainfall event, including both convex and concave shapes. We found that at different times of the year, the same soil moisture sensor at the same soil moisture content could display either convex or concave behaviour, suggesting that this finding is not an artefact of the soil moisture sensor calibration or the particular soil tension characteristics. We also found that the shape (i.e. convex or concave) of the corresponding 60 cm soil moisture response was typically the same as the 30 cm sensor (not shown). It can also occur across the range of soil moisture contents. Instead, the difference in recession shapes could be due to either transient downslope flow towards the sensor, similar to the theoretical case described by Henderson and Wooding (1964), or seasonally varying vegetation characteristics. For example, the unusual concave responses could be due to plants exhausting near-surface soil water stores and therefore starting to extract water from the slightly deeper location of the soil moisture sensor.

5 Summary and implications of variability

Our results have shown multiple modes of spatial and temporal variability in the Langs Gully catchment. Here we summarise the temporal variability in soil moisture and groundwater, followed by spatial variability in soil moisture and groundwater. We then consider connections between them, i.e. temporal changes in spatial variability. Lastly we consider implications of variability for catchment runoff response and prediction.

Temporal variability is characterised by a strong seasonal cycle in catchment wetness; the mean and extremes of the soil moisture and water table distributions are higher in winter than summer. The cycle is driven by PET rather than rainfall depth, and causes significantly higher runoff coefficients in winter. The seasonal cycle in soil moisture shows a long, high winter plateau; compared to water table levels that respond mainly to individual events. The catchment wets up quickly in autumn, but takes longer to dry out in spring, and spring rainfall can briefly return soil moisture and water table levels to their winter state. The volume of stored water in the catchment also has a seasonal cycle, mostly due to increased groundwater in winter, especially during the largest storms.

Spatial variability is controlled most strongly by aspect and distance from stream. South facing slopes have similar mean soil moisture to North facing slopes, but have more soil moisture wetting events, and experience soil saturation more often. Water table levels are higher in South facing slopes and more consistent between locations within the South-facing slope. Near stream locations have higher soil moisture for both mean and extremes, and experience more wetting events. Near-stream locations frequently record saturation in winter, whereas far-stream locations have water tables below the soil moisture sensors and wells for almost the whole study period. We found a strong interaction between groundwater level and soil moisture distribution. Sites where water tables peaked above the 30 cm sensor had a significantly wetter soil moisture distribution compared to sites where water table remained below 30 cm for the whole study period. The finding that soil moisture distribution is dependent on water table depth agrees with measurements in Nordic catchments by Beldring et al. (1999).

Our conclusion that aspect is an important control on soil moisture echoes the results of previous studies in NZ hill country (e.g. Bretherton et al., 2010; Lambert and Roberts, 1976). The mechanisms linking aspect with soil moisture are varied. For example, Lambert and Roberts (1976) found complex interactions between air temperature, soil temperature and ET, driven by wind direction and aspect-induced radiation differences. They note that the specific heat capacity of soil drops as it dries, leading to a positive feedback cycle. In the Langs Gully catchment, the South facing slopes are also steeper than the North facing slopes. This is not obviously due to geological bedding – the main trend of syncline-anticline pairs in the wider Waipara catchment is Northwest-Southeast (transverse to catchment slopes), and in the immediate area of Langs Gully, known dip directions are highly variable. However, feedbacks are likely to exist between slope angle, vegetation (denser shrub cover on South-facing slopes), soil depth (thinner on South-facing slopes) and downslope sediment transport. Shading by denser vegetation and increased lateral flow are possible causes of the increased number of wetting events on the South-facing slope. Typical hydrological models do not account for aspect, but our results suggest that this is an important factor to consider in hillslope runoff generation.

Temporal changes in spatial variability. We suggest that spatial variability can be classified as being in '*summer mode'* or *'winter mode'*. These modes are illustrated as a schematic diagram in Figure 13. In *summer mode*, variability is controlled by shallow processes e.g. soils and vegetation. Water does not typically penetrate to deeper soil moisture or groundwater. Summer variability is therefore disconnected from the channel, and will not directly affect the flow response. However, summer variability affects land surface processes such as evapotranspiration, and may have a lagged effect on the autumn/winter wetting-up process. An example of the disconnect is that the 30 cm soil moisture sites that reacted most strongly to the selected summer rainfall event did not correspond to the

'Saturating' sites identified in Section 4.4.2 as having consistently wetter soil moisture and shallow water tables.

In *winter mode*, variability is controlled by deeper processes e.g. groundwater pathways and bypass flow. The change from shallow vertical flow in dry conditions to vertical bypass flow and lateral flows from upslope in wet conditions is very similar to that found by Detty and McGuire (2010a). However, the summer and winter modes in Langs Gully differ from those found by Bachmair et al. (2012). In their catchment, intense summer storms onto dry soil caused preferential flow and fast, strong, spatially variable water table responses throughout the hillslope. In contrast, their winter storms led to slower water table responses that were strongest at near-stream locations.

In the shoulder seasons, there is a spatially variable shift between the summer and winter modes. Sensors in near-stream locations, particularly those with responsive water tables, spend longer in winter mode. As locations switch between summer and winter modes at varying speeds, spatial variability is increased. This effect is particularly evident on the North facing slope, where soil moisture standard deviation at 30 and 60 cm has a sustained rise during the spring drying period. Rosenbaum et al. (2012) similarly found that seasonal differences between groundwater-influenced and groundwater-distant locations had a strong effect on soil moisture standard deviation. This effect provides one explanation for why high spatial and temporal variability tend to co-occur, as has been found in previous work in New Zealand (McMillan et al., 2014).

Implications for prediction of runoff generation. It is common for some parts of the Langs Gully catchment to wet-up or become saturated, and hence potentially contribute to a runoff response, while other parts of the catchment remain dry. Near-stream and South-facing locations have higher water tables and experience more wetting events. We were able to classify the near-stream sensors into 'Saturating zones' and 'Non-saturating zones'. The saturating zones had higher water table distributions and wetter soil moisture distributions. These zones remained distinct throughout the year.

The Saturating zones are likely to be dominant areas for runoff generation, as wetter soils facilitate vertical drainage and high water tables increase lateral transmissivity. For example, Jencso et al. (2010) found that connectivity between hillslopes and riparian zones led to fast turnover times of riparian groundwater. However, the Saturating/Non-saturating zones did not correspond with the pattern of sensors wetted by infiltration during a summer storm event. The different patterns imply that shallow soil moisture storm responses may not provide a good guide to winter run-off generation pathways, as also found by Tromp-van Meerveld and McDonnell (2005). Rainfall-runoff model structures that delineate catchment landscape components according to dominant processes (e.g. Gharari et al., 2011) may need to use different spatial disaggregations for shallow soil water and ground water.

Understanding catchment variability has further implications for predictions of catchment behaviour. Variability controls which parts of the catchment are generating runoff and controlling water partitioning: it therefore controls uncertainty in flow predictions, depending on our knowledge or lack of knowledge about those water stores or fluxes. Similarly, variability controls how quickly water flows through a catchment, as the different response modes direct water into flow paths with different transit times (Heidbuechel et al., 2013). Variability also provides clues into unmeasured fluxes which are important for catchment response; for example areas with more rapid water table movement suggest locations of preferential flow paths, either vertical or horizontal. Signatures of the catchment variability are seen in the flow response, such as a double or prolonged peak caused by slower groundwater pathways (also found by Bachmair et al. (2012)) , and seasonally variable changes in contributions between different hillslopes. These features suggest that understanding catchment variability is essential to predicting the hydrograph.

6 Conclusion

We made distributed measurements of flow, soil moisture and depth to groundwater in a New Zealand headwater catchment, to characterise controls on variability. The data showed that temporal variability was dominated by a strong climatic seasonal cycle, with event dynamics superimposed. The volume of stored water in the catchment had a corresponding seasonal cycle, mostly due to increased groundwater in winter. Spatial variability is controlled most strongly by aspect and distance from stream: South-facing and near-stream sites are typically wetter, and in particular have more and larger wetting events. The relative wetness of different locations was stable: high water table locations were consistent across seasons, and sites where water tables peaked above 30 cm depth had consistently wetter soils. Temporal dynamics vary spatially, including timing of winter wet up (faster on South-facing slopes), different speeds of groundwater response (slow at far-stream sites) and different recession shapes (no clear spatial pattern).

We examined soil moisture and groundwater responses to rainfall, for dry vs. wet antecedent conditions, and found significant differences in the patterns of response. This led us to classify catchment variability as being in *'summer mode'* or *'winter mode'*. In *summer mode*, variability is controlled by shallow processes e.g. soils and vegetation, and sites where soil moisture reacts strongly to a rainfall event may not correspond with the usual wetter locations. In *winter mode*, variability is controlled by deeper processes e.g. groundwater pathways and bypass flow. In both cases, variability is strongest for stores where typical water content is close to a threshold such as saturation. Because spatial variability changes with season, we suggest that methods to predict emergent catchment behaviour arising from small-scale variability may also need to change with season.

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Reference	Control	Relationship
Brocca et al. (2007)	Upslope area	Positive spatial correlation between soil moisture and ln(upslope area) and soil moisture, found at all 14 sampling times.
Qiu et al. (2001)	Land use and topography descriptors including slope, aspect and elevation	Statistically significant spatial correlation between mean soil moisture and classifications of land use (higher soil moisture for crops than forest), aspect (higher soil moisture for North aspect) and slope position (higher soil moisture for downslope locations).
Kim et al. (2007)	Topographic position	Topographic zones (upper, buffer and flow path zones) defined by contributing area and distance to flow path. Qualitative differences in soil moisture dynamics found between zones.
Penna et al. (2009)	Slope, topographic index	At 5 sites and 3 depths, Pearson's correlation typically positive between soil moisture and topographic wetness index, always negative between soil moisture and slope.
Nyberg (1996)	Topographic index	Significant positive Spearman correlation between soil moisture and topographic wetness index.
Crave and GascuelOdoux (1997)	Height above the nearest drainage	Fitted negative exponential relationship between soil moisture and height above the nearest drainage.

Table 1: Examples of controls on soil moisture distribution found in international studies.

	Stones	Sand	Clay
Spurs	30-80%	10-50%	10-25%
Footslopes	5-20%	5-40%	20-35%

Table 2: Fractions of stones, sand, clay for typical spur and footslope soils at 0-30 cm depth. Sand and clay values are excluding the coarse fraction.

	Stones	**Sand**
0 - 30 cm	5-20%	5-40%
30 - 60 cm	35-80%	10-40%

Table 3: Fractions of stones and sand for typical footslope soils at 0-30 cm and 30-60 cm depth.

		Number of wetting events	Mean soil moisture increase in the 10 largest events
South facing	**Near-stream**	16	16%
	Far-stream	12	6%
North facing	**Near-stream**	12	12%
	Far-stream	9	6%

Table 4: Number and size of soil moisture wetting events by aspect and distance from stream.

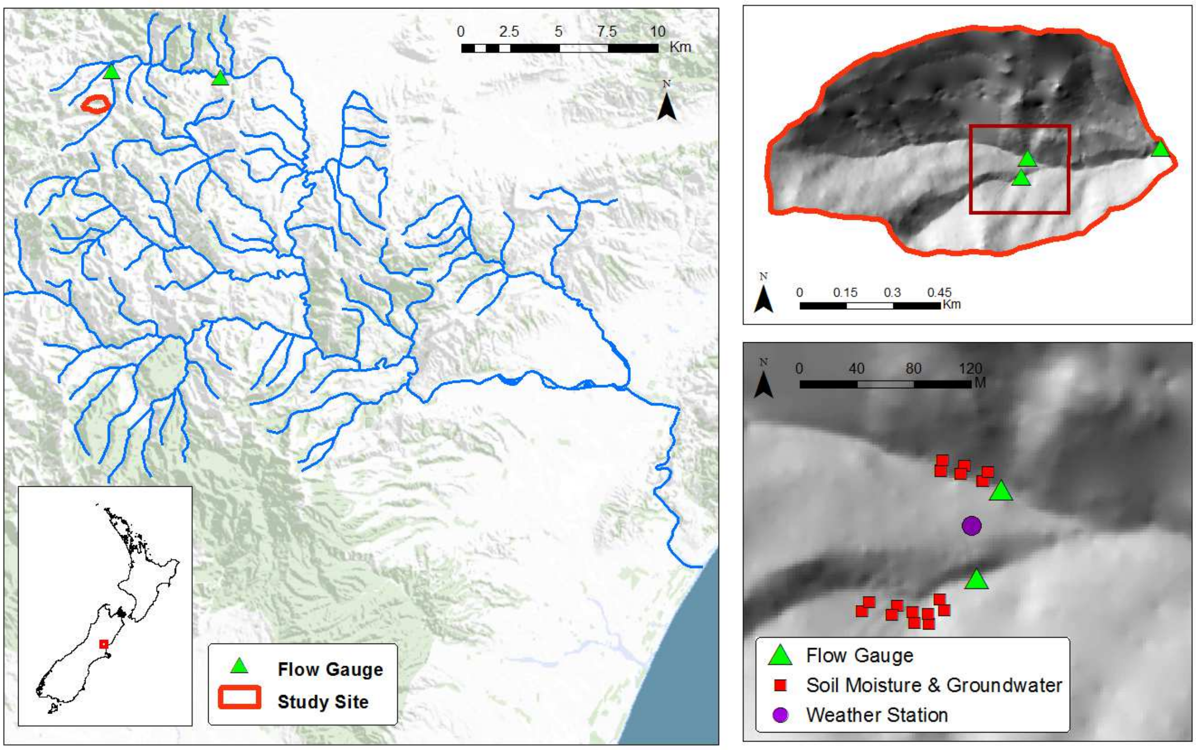

Figure 1: Catchment location and Instrumentation

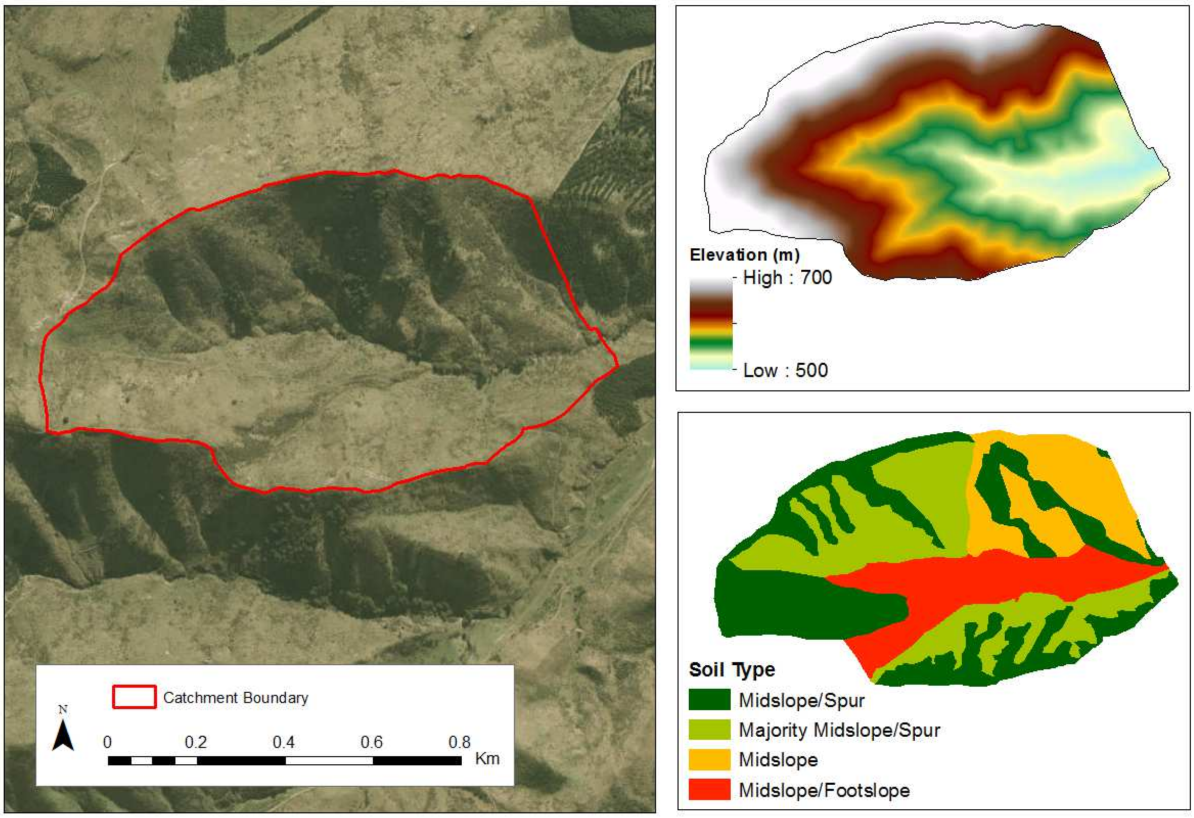

Figure 2: Catchment aerial photo, topography and soils

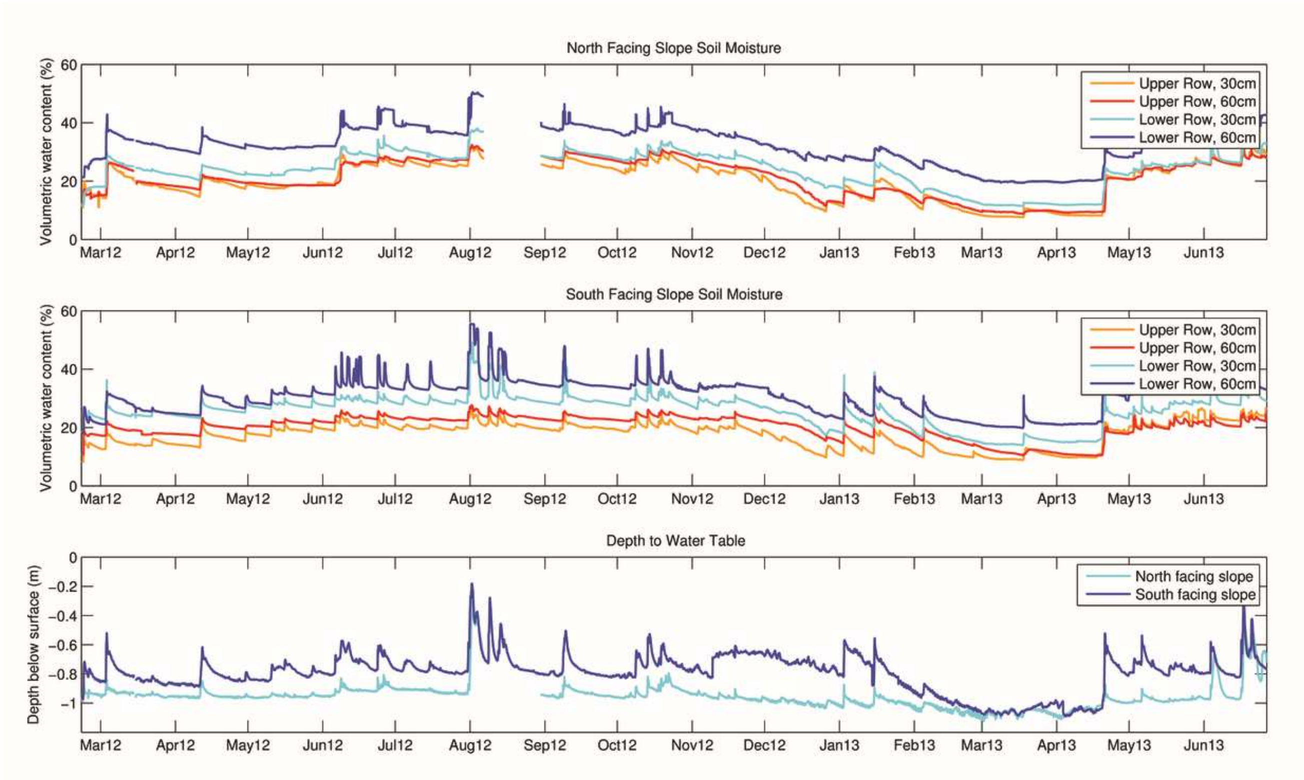

Figure 3: Time series of average soil moisture and groundwater level for the complete study period.

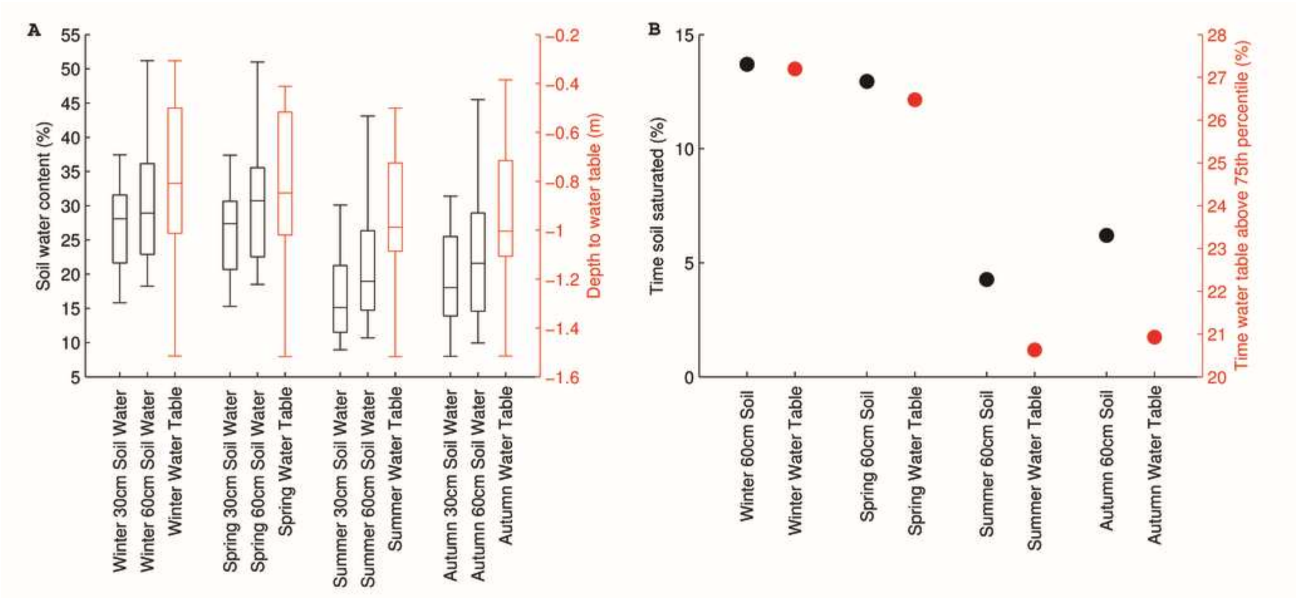

Figure 4: Summary statistics of soil moisture and depth to water table by season. (A) Distributions of measured values. (B) Summary of wet extremes.

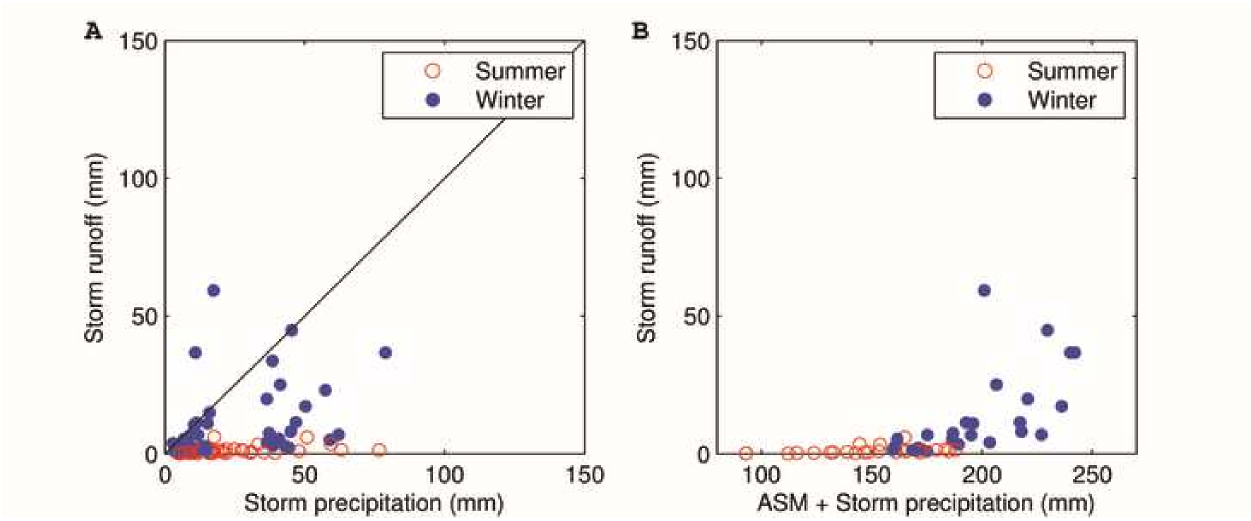

Figure 5: (A) Storm Runoff against Storm Precipitation, split by season. This figure was created after pre-processing of the data to define storm and inter-storm periods, based on the method of (McMillan et al. (2014)) using thresholds for precipitation depth and inter-storm duration, and without baseflow separation. (B) Storm Runoff against the sum of Storm Precipitation and Antecedent Soil Moisture storage (ASM), split by season. ASM was taken as the Total Soil Moisture value from Eq 1.

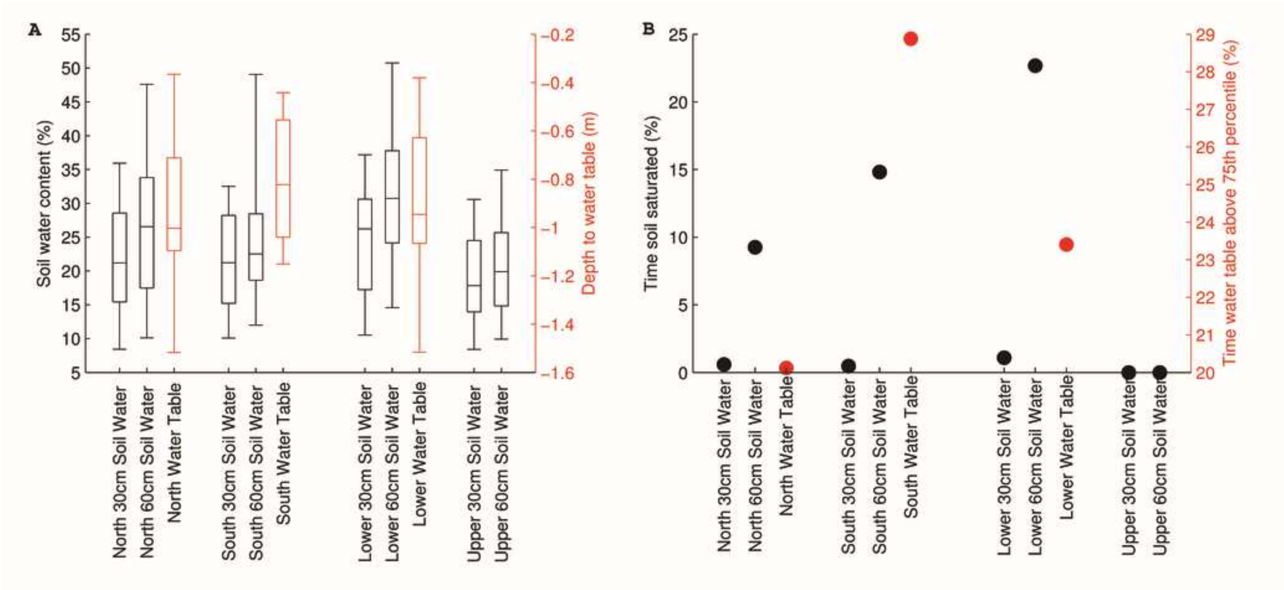

Figure 6: Summary statistics of soil moisture and depth to water table by location. (A) Distributions of measured values. (B) Summary of wet extremes.

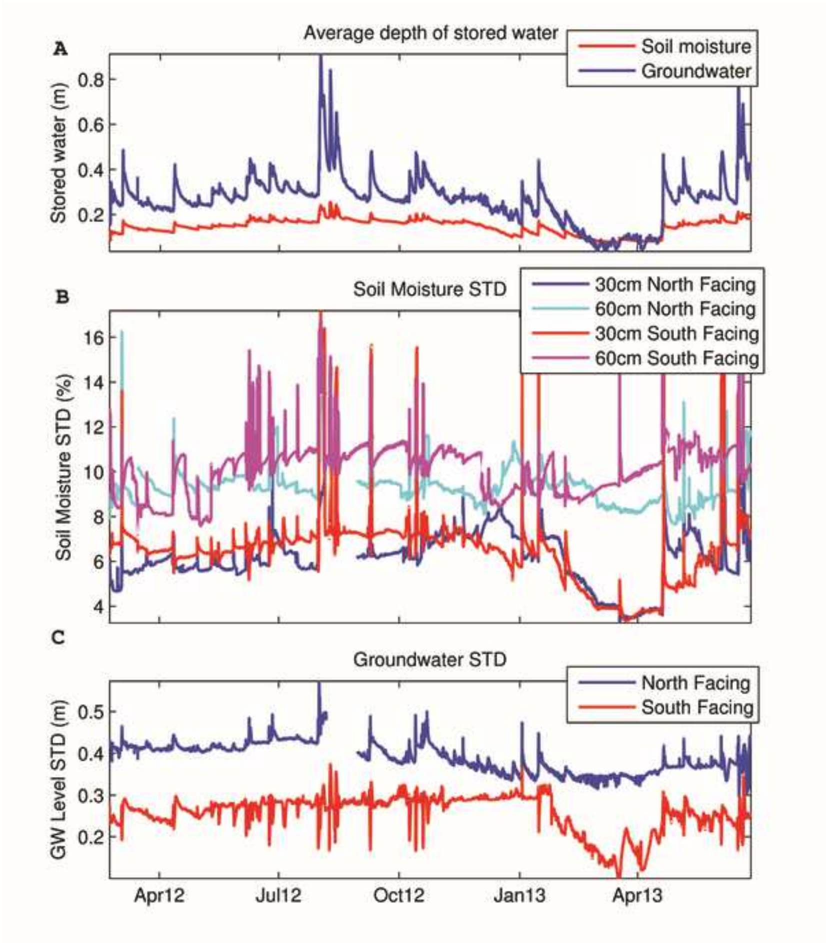

Figure 7: (A) Average depth of water stored in the catchment as soil moisture and groundwater (B) Spatial standard deviation of soil moisture values, by aspect and depth (C) Spatial standard deviation of groundwater levels, by aspect.

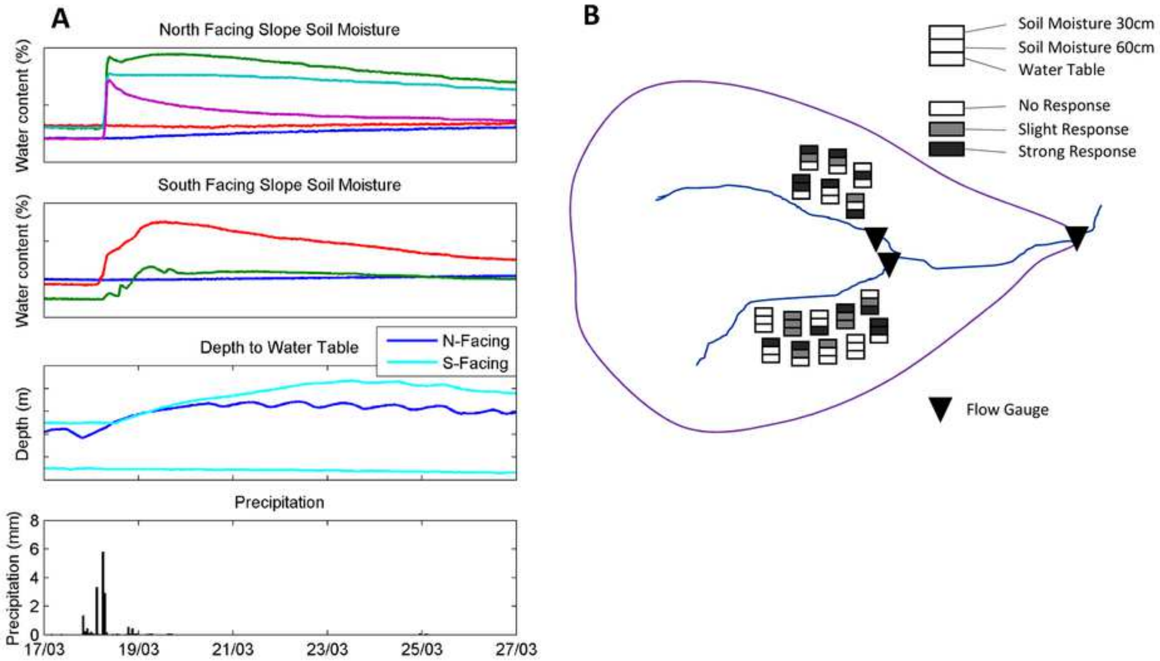

Figure 8: (A) Response of selected sensors to a March rainfall event. First and second panels: Soil moisture responses in North- and South-facing slopes respectively. Colours are used only for visual clarity. Third panel: Depth to water table. Fourth panel: Storm precipitation (B) Spatial overview of strength of soil moisture and water table sensor responses to the March rainfall event.

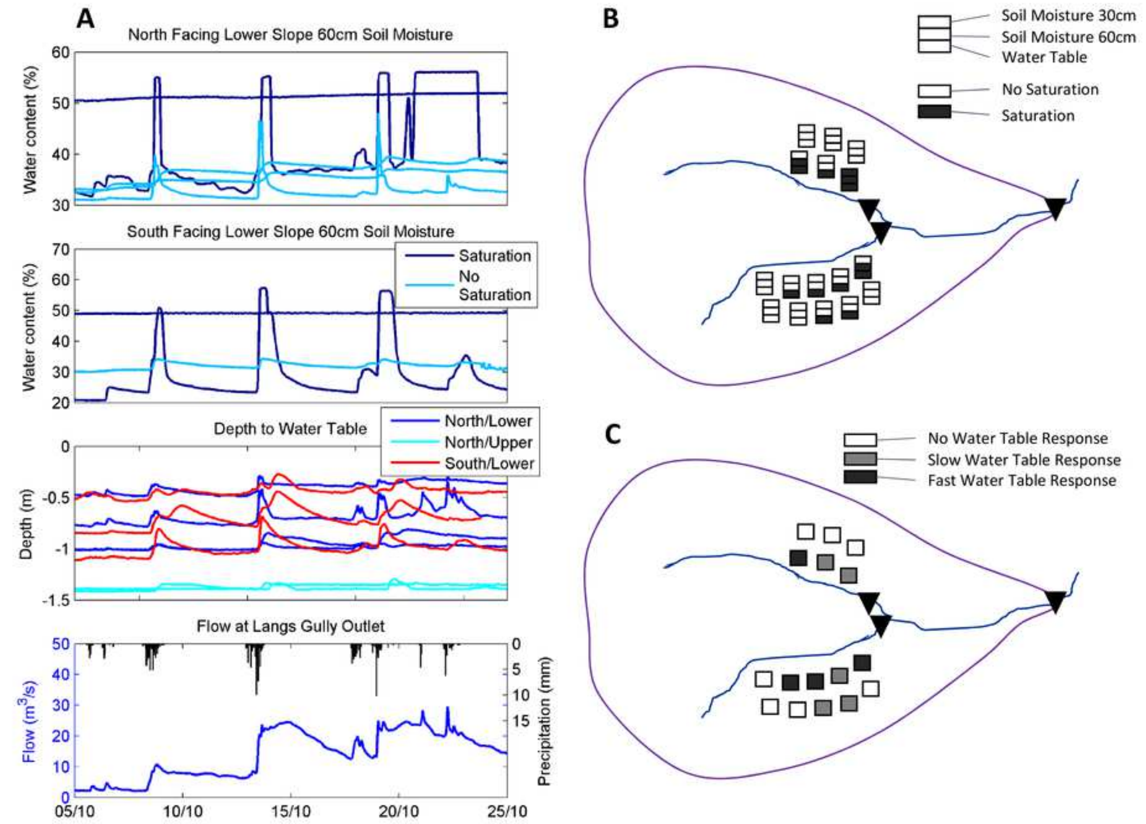

Figure 9: (A) Response of selected sensors to a Winter rainfall event. First and second panels: Soil moisture responses in North- and South-facing slopes respectively. Dark lines show sensors where saturation occurred. Third panel: Depth to water table by well location. Fourth panel: Storm precipitation and flow measured at the catchment outlet gauge. (B) Overview of saturation response to the Winter rainfall event (C) Overview of speed of water table response to the Winter rainfall event

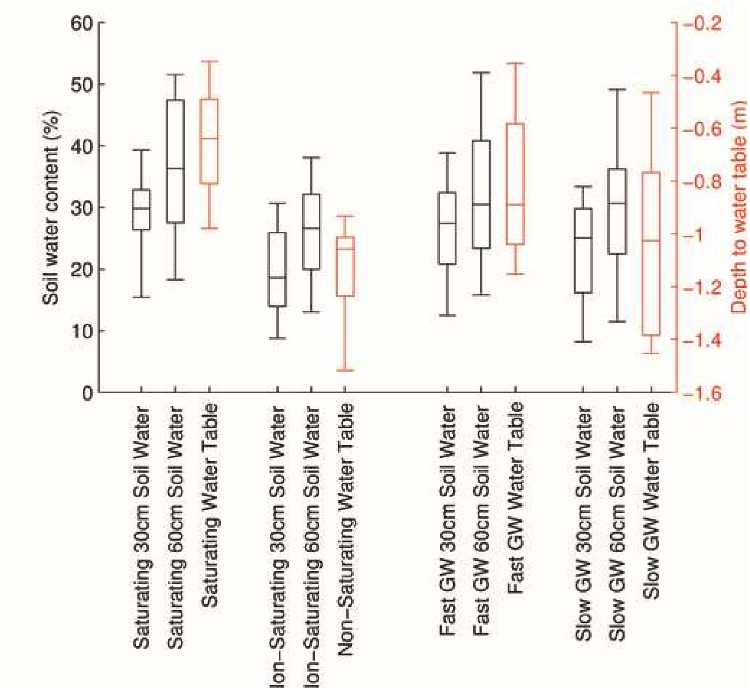

Figure 10: Distributions of soil moisture and depth to water table, classified as Saturating/Non-saturating sites, and Fast/Slow groundwater response sites. Saturating sites were defined as those where the water table rose as high as the 30 cm soil moisture probe at any point during the study period. Fast/Slow sites were classified according to the speed of groundwater response as described in Section 4.4.2 and Figure 9C.

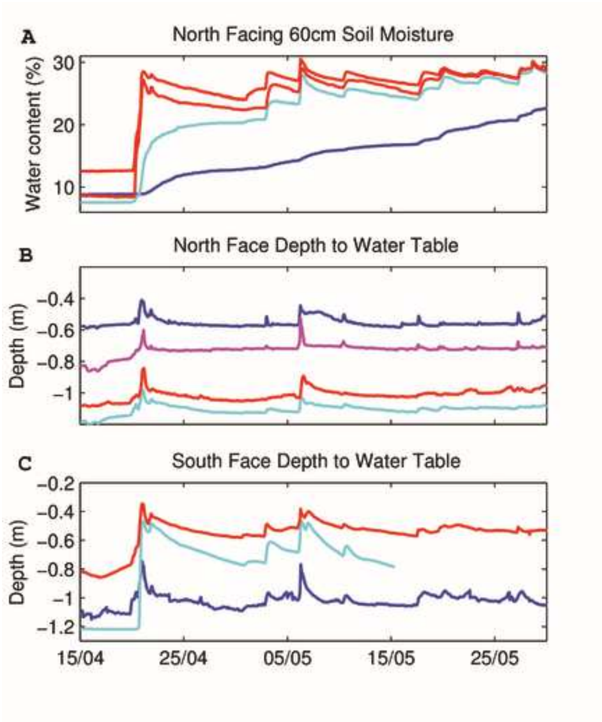

Figure 11: Winter wet-up response of selected soil moisture and water table sensors. (A): Soil moisture on the North-facing slope. Red lines show locations with a fast wet-up; Blue line show locations with a gradual wet-up. (B)/(C): Depth to water table at North- and South-facing slopes. Colours are used only for visual clarity.

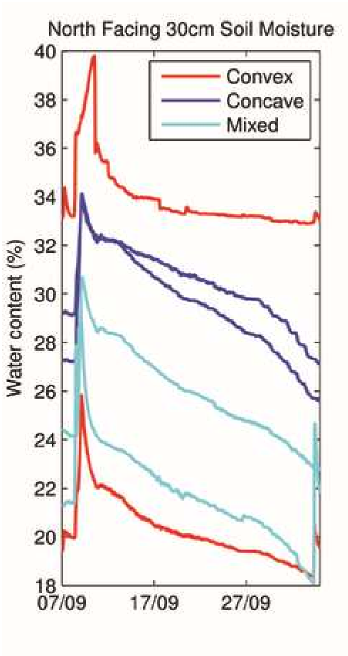

Figure 12: Selected North-facing 30 cm soil moisture sensor responses during a recession, having convex, concave or mixed response shapes.

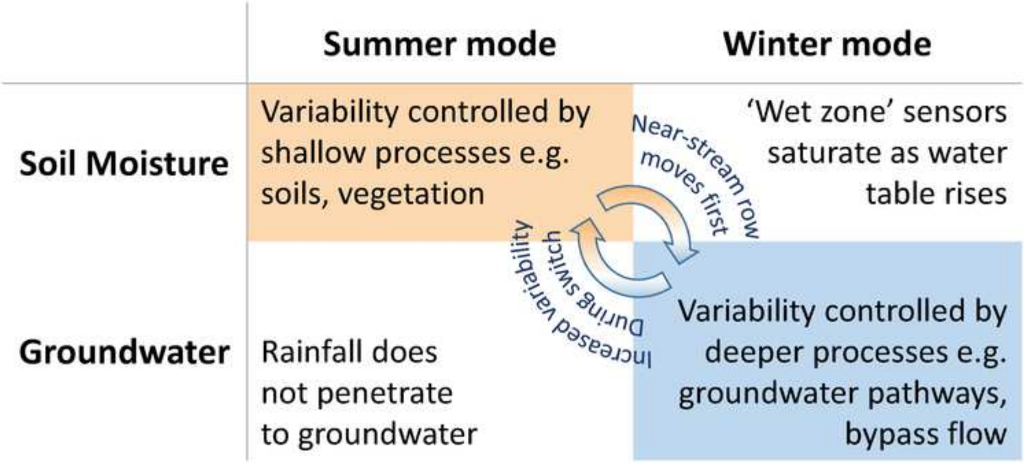

Figure 13: Schematic diagram of the seasonal cycle of catchment variability between 'Summer mode' and 'Winter mode'.