

A preliminary assessment of water partitioning and ecohydrological coupling in northern headwaters using stable isotopes and conceptual runoff models

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Abstract:

We combined a conceptual rainfall-runoff model and input–output relationships of stable isotopes to understand ecohydrological influences on hydrological partitioning in snow-influenced northern catchments. Six sites in Sweden (Krycklan), Canada (Wolf Creek; Baker Creek; Dorset), Scotland (Gimock) and the USA (Dry Creek) span moisture and energy gradients found at high latitudes. A meta-analysis was carried out using the Hydrologiska Byråns Vattenbalansavdelning (HBV) model to estimate the main storage changes characterizing annual water balances. Annual snowpack storage importance was ranked as Wolf Creek > Krycklan > Dorset > Baker Creek > Dry Creek > Gimock. The subsequent rate and longevity of melt were reflected in calibrated parameters that determine partitioning of waters between more rapid and slower flowpaths and associated variations in soil and groundwater storage. Variability of stream water isotopic composition depends on the following: (i) rate and duration of spring snowmelt; (ii) significance of summer/autumn rainfall; and (iii) relative importance of near-surface and deeper flowpaths in routing water to the stream. Flowpath partitioning also regulates influences of summer evaporation on drainage waters. Deviations of isotope data from the Global Meteoric Water Line showed subtle effects of internal catchment processes on isotopic fractionation most likely through evaporation. Such effects are highly variable among sites and with seasonal differences at some sites. After accounting for climate, evaporative fractionation is strongest at sites where lakes and near-surface runoff processes in wet riparian soils can mobilize isotopically enriched water during summer and autumn. Given close soil–vegetation coupling, this may result in spatial variability in soil water isotope pools available for plant uptake. We argue that stable isotope studies are crucial in addressing the many open questions on hydrological functioning of northern environments. © 2015 The Authors. *Hydrological Processes* published by John Wiley & Sons Ltd.

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INTRODUCTION

The biomes of the upper latitudes of the Northern hemisphere extend from the edge of northern mixed temperate forests, through boreal forests into tundra (Figure 1). These energy-limited areas, where snow is an important component of the annual water balance, cover a significant proportion of the Earth's terrestrial surfaces with boreal forest covering 9% and tundra covering 10% of the land mass. Critically, much of this region is currently subject to rapid rates of climate change; warming trends are

apparent in many areas with implications for the influence of snow on the water balance and the annual distribution of precipitation (Kundzewicz *et al.*, 2007). These changes have fundamental implications as they will affect pedogenic processes, produce vegetation changes and alter flowpaths and river flow regimes (Dye, 2003; McClelland *et al.*, 2006; Tetzlaff *et al.*, 2013). The hydrological function of this globally important region is not well understood. The sheer scale, diversity and remoteness of the North dictate that empirical studies are logistically constrained and are less dense than in other, more accessible environments (Pomeroy *et al.*, 2013; Tetzlaff *et al.*, 2015). This means that there is a limited scientific evidence base for predicting and managing environmental change (Barnett *et al.*, 2005). Addressing this need requires

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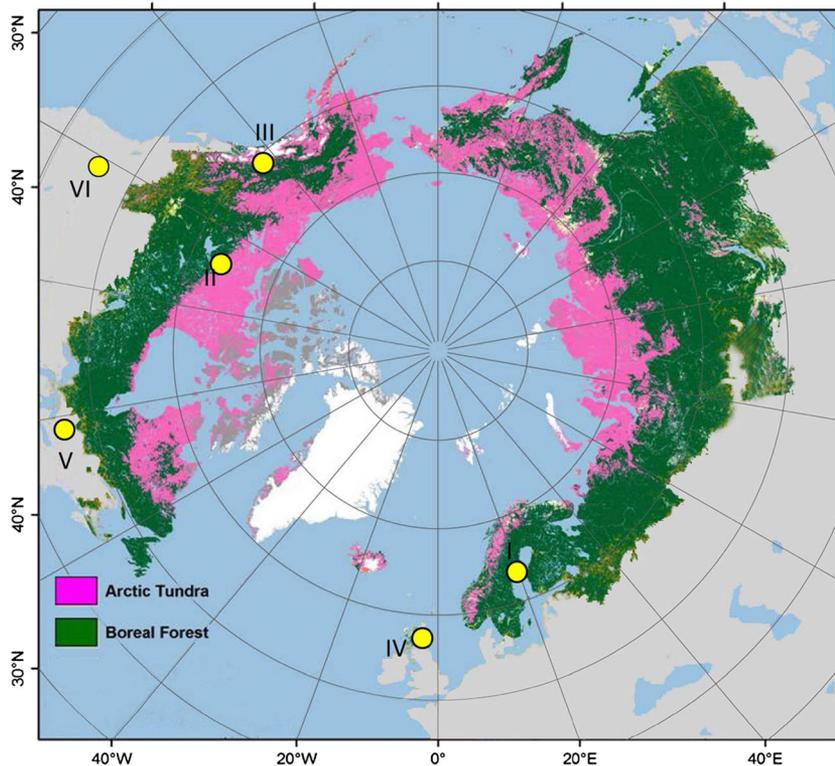


Figure 1. Location of the VeWa experimental study sites: I Krycklan (Sweden); II Baker Creek (Canada); III Wolf Creek (Canada); IV Girnock (Scotland); V Dorset (Canada); VI Dry Creek (USA)

the use of integrative hydrological tools that can rapidly improve understanding and test hypotheses about hydrological function to provide a basis for future projections across a range of scales. Applications of stable isotopes (e.g. Stadnyk *et al.*, 2013), as well as runoff modelling (Dornes *et al.*, 2008; Rasouli *et al.*, 2014), have been used successfully in this way to integrate across scales and are likely to be invaluable in developing the understanding of northern regions hydrology that is urgently needed to inform policy.

Much of the complexity of northern landscapes is a consequence of the moisture and temperature gradients of their diverse hydroclimatic regimes, which in turn affect the timing and rates of snowpack accumulation and melt (DeWalle and Rango, 2011; Anderson *et al.*, 2014). Seasonal snow melt, and even ground thaw, strongly influences or even dominates the annual hydrological regime (e.g. Carey and Woo, 2001a; Pomeroy *et al.*, 2007). Also, there is usually a strong legacy of glaciation and contrasting glacial drift that consequently has an important influence on pedogenesis with highly heterogeneous soils and associated vegetation communities affecting hydrological partitioning, flowpaths and storage (Spence and Woo, 2003; Devito *et al.*, 2005; Verry *et al.*, 2011; Tetzlaff *et al.*, 2014). Organic peat soils (Histosols) often mantle extensive low-lying areas of riparian zones, where runoff generation is dominated by saturation-excess overland

flow or rapid flow in transmissive near-surface horizons (Quinton and Marsh, 1999; Carey and Woo, 2001b; Soulsby *et al.*, 2006). Further upslope, these areas transition to more freely draining soils that serve as recharge zones (Spence and Woo, 2003; Grabs *et al.*, 2009). Subsurface permeability and hillslope connectedness to riparian areas and the stream network are also important determinants of hydrological catchment response (Tetzlaff *et al.*, 2014).

These different landscape units may have contrasting land covers and vegetation communities. Both ecosystem components are expected to respond most rapidly to climatic change (Hinzman *et al.*, 2005; Donohue *et al.*, 2007). Forest cover is prevalent on more freely draining soils, with Sphagnum moss, grasses and shrubs dominant in wetlands that can be extensive in northern regions (Tetzlaff *et al.*, 2013). Lakes are also common in northern glaciated regions (Gibson and Reid, 2010; Phillips *et al.*, 2011). These landscape structures exert a strong influence on the hydrology of these regions. Of course, relationships between landscape organization and catchment hydrological response vary in different climatic and geomorphic zones (Tetzlaff *et al.*, 2009; Carey *et al.*, 2013a). However, the role of vegetation on the hydrological response of northern regions is still a major research gap that needs to be urgently addressed, particularly given the uncertain hydrological implications

of the marked vegetation change underway in many areas (Newman *et al.*, 2006; Danby and Hik, 2007; Naito and Cairns, 2011, 2015; Tetzlaff *et al.*, 2014). Generically, these implications include alteration to precipitation phase, patterns and processes of snow redistribution and melt, interception, transpiration and soil structure; all of which can exert a profound effect on water partitioning in the vegetation canopy or soil. Given a backdrop of projected marked climate change, there is greater uncertainty regarding the future hydrology in these areas and the associated implications for related processes such as carbon cycling (Guo *et al.*, 2007; Frey and McClelland, 2009; Laudon *et al.*, 2012).

These complex changes and subtle interactions highlight potential challenges in using stable isotopes in northern catchments. Stable isotope records from precipitation, stream flow and various geographical runoff sources have proved useful in giving insights into catchment-scale runoff responses and the timing of dominant flowpaths (e.g. Soulsby *et al.*, 2000; Laudon *et al.*, 2004; Carey *et al.*, 2013b; Peralta-Tapia *et al.*, 2015a). However, recent work has highlighted the importance of ecohydrological coupling in mediating isotope transformations in the landscapes (Brooks *et al.*, 2010; Dawson and Simonin, 2011). While the effect of snowpack accumulation and melt on isotope fractionation is reasonably well-understood (Cooper, 1998; Rodhe, 1998; Taylor *et al.*, 2002), the influence of vegetation and soils on water isotope pools in northern systems is much less clear (McDonnell, 2014). Moreover, the lack of comprehensive tracer data sets still hinders the development of a generalized understanding of the internal functioning of northern catchments, which can be conceptualized into models. This would be an important step forward as such tracer-aided models have significant potential in testing conceptualizations of physical processes (Birkel and Soulsby, 2015).

As a step towards integrating insights from tracers and modelling studies in northern landscapes, we provide a preliminary analysis of six well-known experimental catchments (Figure 1). These are part of the VeWa (*Vegetation effects on water flow and mixing in high-latitude ecosystems*) project, which aims to understand the nature of water partitioning along cross-regional hydroclimatic gradients of moisture and temperature and to assess the role of vegetation, soils and other landscape features. The six VeWa catchments are located in geomorphic provinces ranging from the Precambrian (Dorset and Baker Creek in Canada) and Fennoscandian (Krycklan in Sweden) Shields to mountainous areas such as the Cairngorms (Girnock in Scotland) and the North American Cordilleran (Wolf Creek in Yukon Territory, Canada). Vegetation cover varies from mixed hardwood-conifer forest (Dorset), through boreal (Krycklan) and

subarctic (Baker Creek) forest, to shrub/tundra (Girnock, Wolf Creek). The more southerly US site at Dry Creek was selected to include a snow-influenced area with higher energy inputs. The elevation range and the complex topography in Dry Creek produce environmental conditions varying from high desert environments characteristic of low latitudes to cold, snow-dominated environments characteristic of high latitudes in a single catchment. Extensive process-based research provides a strong background of hydrological understanding for all catchments. In several cases, this knowledge has been integrated into modelling studies, and all catchments have at least several years of input–output isotope data.

A catchment inter-comparison was conducted in order to: (1) characterize the water balance and hydrological function of the sites using a common, simple modelling framework with a particular focus on hydrological partitioning at each site; (2) use input–output stable isotope data to assess the influence of landscape characteristics on water partitioning and runoff processes; and (3) examine whether signs of fractionation during water flow through the catchment observed in the stream water can be used to infer the nature of ecohydrological separation of waters in the catchments. Our goal in carrying out this comparative analysis was to identify open research questions on the ecohydrological partitioning and coupling of waters in northern catchments and assess research priorities.

STUDY SITES

Hydroclimatic characteristics of the VeWa catchments are summarized in Table I. The most northerly site is Krycklan (S7 subcatchment, 0.5 km²), in the Swedish Boreal Forest biome (Laudon *et al.*, 2013). Half of the runoff occurs during the snow-free period (June to November), and a third during 3–4 weeks of spring flood in April/May when maximum flows usually occur (Figure 2). Low flows occur throughout winter, with an average duration of 170 days. Most of the land cover is forest (82%), mainly Scots pine (*Pinus sylvestris*) in upper drier locations and Norwegian spruce (*Picea abies*) in lower wetter areas. A minerogenic mire, dominated by *Sphagnum* moss, covers 18% of the catchment. Podzol soils have developed on compact basal till (overlying metasediments) with hydraulic conductivity that decreases with depth. Near the stream channel, riparian peat soils have a profound impact on stream hydrology and hydrochemistry (Grabs *et al.*, 2012). Subsurface pathways dominate runoff delivery in the forested landscape elements, while overland flow occurs on the mire during both snow melt and rain events (Peralta-Tapia *et al.*, 2015b).

Table I. Catchment characteristics

| Catchment (Site) | Start date | End date | Latitude | Longitude | Size (km ²) | Mean altitude (masl) | MAT (°C) | MAP (mm) | MAQ (mm) | Mean runoff ratio(-) | BFI (-) | Land cover (%) | Dominant soils | Dominant geology |
|---|------------|------------|----------|-----------|-------------------------|----------------------|----------|----------|----------|----------------------|---------|---|---|---|
| Krycklan Sweden (S7) | 01/10/1982 | 31/12/2007 | 64.2 | 19.87 | 0.5 | 250 | 1.8 | 622 | 310 | 0.49 | 0.42 | 82/18 Coniferous forest/wetland | Podzols | Meta-sediments |
| Baker Creek, Canada (Moss) | 01/01/2009 | 31/12/2012 | 62.6 | -114.43 | 9.3 | 230 | -3.1 | 311 | 88 | 0.27 | 0.65 | 39/25/21/15 Bedrock/black spruce/ lakes/wetland | Podzols and organic cryosols | Precambrian volcanic and sedimentary bedrock intruded by Archean batholiths and plutons |
| Wolf Creek, Canada (Granger) | 1/04/1999 | 04/10/2008 | 60.71 | -135.7 | 7.6 | 1650 | -2 | 471 | 358 | 0.75 | 0.64 | 42/13/46 Tall shrub (willow), short shrub (birch), tundra | Organic soils overlying till; regosols at high elevation | Primarily sedimentary, till mantle with thickness from cm-10 m |
| Gimcock Scotland (Littlemill) | 01/01/2000 | 31/12/2011 | 57.05 | -3.12 | 31 | 350 | 6.6 | 1001 | 602 | 0.59 | 0.41 | 20/70/10 Wetland/heather moorland/conifer | Peaty gleys; peaty podzols | Low permeability igneous + metamorphic rocks; In valley bottom areas: fine textured drifts |
| Dorset, Canada (Harp-4) | 01/11/1977 | 27/10/2002 | 45.38 | -79.14 | 1.2 | 370 | 4.8 | 1020 | 536 | 0.52 | 0.41 | 8/88/1/3 bedrock/hardwood forest/ ponds/wetlands | Uplands: brunisols and podzols Valley bottoms: histosols | Granitized biotite and hornblende gneiss; surficial geology: bedrock outcrops to sandy till > 10 m thick, with peat in wetlands |
| Dry Creek, US (Q: lower gauge; P,T: treeline) | 01/01/2000 | 31/12/2012 | 43.7 | -116.17 | 28 | 1470 | 9 | 653 | 122 | 0.19 | 0.80 | 49, 47, 4 sagebrush steppe, mixed conifer, riparian phreatophytes | Argixerolls, Haploxerolls, and Haplocambids | Biotite granodiorite |

MAT, mean annual temperature; MAP, mean annual precipitation; MAQ, mean annual discharge; BFI, baseflow index

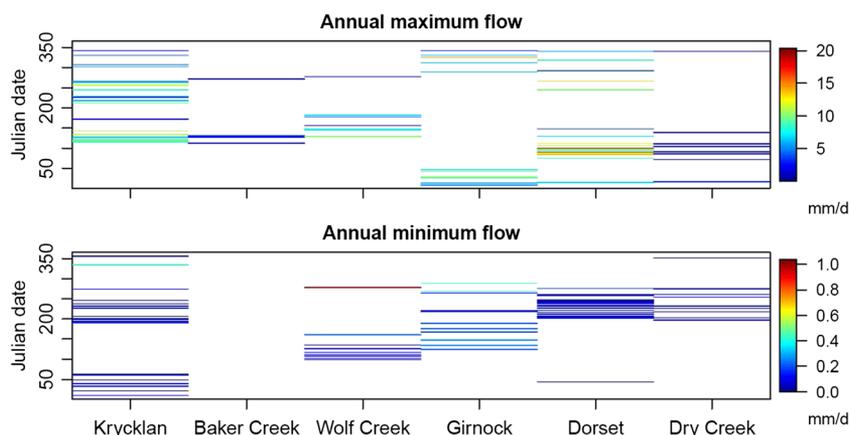


Figure 2. Timing and magnitude of annual maximum and minimum flow for the six VeWa catchments for the period given in Table 1 (note: for Baker Creek, there were no discharge data available during winter months; therefore, annual minimum flows are not shown). The magnitude of the maximum flow is based on the 1-day maximum flow for Baker Creek for the other catchments on the mean of the 7-day maximum flow

Moss Creek (9.3 km²) is a tributary of the Baker Creek Research Basin in Canada's Northwest Territories. Forty-two per cent of annual precipitation falls as snow. Maximum flows are generally in April/May melts (Figure 2). The landscape is dominated by lakes connected by short channels that cover 21% of the basin area. The predominant vegetation is open black spruce forest that occupies 25% of the basin. Bogs, fens and peat plateaus cover 15% of the basin. The basin is in the zone of discontinuous permafrost, and upland soils are either podzols or organic cryosols formed over Precambrian bedrock, which outcrops over 39% of the basin.

Wolf Creek (7.6 km², Granger sub-basin) is in the Yukon Territory. Geology is primarily sedimentary overlain by glacial till ranging from a thin veneer to several metres in thickness. Fine textured alluvium covers the valley floor, whereas upper elevations have shallow colluvial deposits with frequent bedrock outcrops (Mougout and Smith, 1994). Permafrost underlies approximately 70% of the basin, particularly north-facing slopes and higher elevation areas, whereas seasonal frost predominates on southerly exposures (Lewkowicz and Ednie, 2004). In permafrost zones and the riparian zones, soils are capped by an organic layer up to 0.4 m thick consisting of peat, lichens, mosses, sedges and grasses. Only a few scattered white spruce (*Picea glauca*) occur within the basin, which is considered above treeline (ca. 1200 m). Vegetation consists predominantly of willow (*Salix*) and birch (*Betula*) shrubs. A small perennial snowpack exists on the western edge of the catchment.

The Scottish catchment is the Girnock Burn (31 km²). Peak flows can occur through the year, but are usually between November and February, often with a snowmelt component. Underlying geology is mainly granite with associated metamorphic rocks. Below 400 m, the bedrock is covered by drift deposits (mainly poorly sorted till),

which can be up to 40 m deep in the wide valley bottoms. Riparian areas are characterized by *Sphagnum spp* and *Molina caerulea*-dominated blanket bog, while drier steeper slopes are heather (*Calluna vulgaris*) dominated. Forest cover (mainly Scots Pine (*Pinus sylvestris*)) is restricted to small areas on the steeper hillslopes. Organic-rich soils dominate, with deep (>1 m) peats (Histosols) in valley bottoms and shallow (<0.5 m) peat on lower hillslopes covering 22% of the catchment. Steeper slopes have podzols with a 0.1–0.2 m deep O horizon overlying free-draining mineral subsoil. Isotope data were used from the Bruntland Burn, a 3-km² tributary of the Girnock, which has a similar distribution of soils and landscape properties as the Girnock.

The most southerly Canadian site is the Harp-4 catchment, a 1.2-km² headwater near Dorset, Ontario in the southern Boreal ecozone. Approximately one-third of the precipitation falls as snow. Fifty per cent or more of the runoff occurs in the March-to-early-May snowmelt period (Figure 2). Mean catchment slope is 5% on impermeable Precambrian Shield bedrock (Jeffries and Snyder, 1983). Surficial geology is approximately 52% deeper (1–10 m deep) till, 36% thin (<1 m deep) till with rock ridges and 8% peat coverage. There is a ~3.2 ha pond in the central area of the catchment through which the western area drains. Upland soils are podzols. Combined with the impermeable bedrock, these thin soils result in the development of only local aquifers (Devito, 1995). Upland forests are mixed hardwoods (dominated by *Acer saccharum* and *A. rubrum*), while poorly drained areas have conifer (*Picea mariana*, *Thuja occidentalis*) cover and *Sphagnum* and peaty humisol soils.

The most southerly VeWa site is Dry Creek, in the granite hills above Boise, Idaho, USA. Instrumented sub-basins drain areas ranging from 0.012 km²–27 km². The catchment is dissected into V-shaped fluvial valleys with 20–40° slope angles. Soils are thin (<1 m), highly

permeable sands underlain by fractured granodiorite (Smith *et al.*, 2011). Vegetation at higher elevations (>1500 m) is primarily Douglas fir (*Pseudotsuga menziesii*) and Ponderosa Pine (*Pinus ponderosa*), while lower elevations are dominated by grass and sage. Riparian areas are lush, with dense stands of cottonwoods (*Populus spp*) and other hardwoods. In general, large woody plants are confined to the riparian areas and seeps at low elevations.

DATA AND METHODS

Hydrometric data and runoff modelling

We applied the HBV model (the 'HBV light' version; Seibert and McDonnell, 2010; Soulsby *et al.*, 2011; Seibert and Vis, 2012) for initial inter-site comparisons. HBV (Lindström *et al.*, 1997) is a well-known conceptual runoff model developed for snow-influenced catchments that simulates runoff on a daily time step. The model is driven by time series of precipitation and air temperatures and estimates average monthly potential evaporation. Snow-melt is calculated using the degree-day method; groundwater recharge and actual evaporation are computed as functions of estimated soil water storage; runoff generation is conceptualized by three linear reservoir equations; and channel routing is estimated by a triangular weighting function (see Table II for model parameters). The HBV model parameters each have a physical meaning, but cannot be measured directly as they represent effective catchment scale values. To keep the model simple, all routines were used in a lumped representation of each catchment with one vegetation/elevation zone. Full details of the model are available online (<http://www.geo.uzh.ch/en/units/h2k/services/hbv-model>).

Daily time series of discharge, precipitation and air temperature were used, although the length of available time series differed among catchments (Table I). Catchment average precipitation and temperature were used; at some sites, particularly Dry Creek, altitudinal variation influences precipitation amounts and phase, but in this meta-analysis, model structures were kept simple to produce catchment-average storage estimates. Monthly average potential evapotranspiration values were calculated using the Thornthwaite method (Thornthwaite, 1948) based on observed temperature data. The model was calibrated for the full time periods to estimate parameter values. Based on the Nash–Sutcliffe efficiency (Nash and Sutcliffe, 1970), the 20 best parameter sets were chosen from 1 million Monte Carlo runs and used to determine changes in storage (Table II). To further assess model performance, the mean difference between observed and simulated discharge was determined as well over the whole simulation period (expressed as

volumetric error). The annual change in active storage (e.g. McNamara *et al.*, 2011) for soil and groundwater was calculated as the mean of the difference between maximum and minimum values for each year of each model run, which was then averaged for all 20 model runs. The average annual mean snowpack (expressed as snow water equivalent, *SWE*) was calculated by first averaging the snowpack for each separate year and then averaging over the whole period giving information about the differences in the amount of snow between the catchments. The average annual maximum storage in the snowpack (also expressed as *SWE*) was determined from the mean annual maximum *SWE* per model run averaged for all 20 model runs.

Stable isotope data

Precipitation and stream water have been sampled for isotope analysis at all VeWa sites. Where fresh snow samples were collected, these were allowed to melt prior to analysis. Sampling took place daily, weekly to bi-weekly or as logistics and access to the streams allowed. Samples were collected in scintillation vials and capped tightly to avoid evaporation. If taken with autosamplers, the samples were preserved by paraffin. At Wolf Creek and Dorset, snowmelt lysimeters were used to collect meltwater, i.e. snow melts into a simple constructed steel pan or Teflon-lined wood construction that funnels water into a container. When water reports, it is sampled; therefore, sample frequency was daily to every few days. Samples were analysed with a Los Gatos DLT-100 laser isotope analyser (for Girnock, Dorset) and a Picarro cavity ringdown laser spectrometer (L1102-i) and the vaporizer module (V1102-i) (for Krycklan) following standard measurement protocols (precision $\pm 0.1\text{‰}$ for $\delta^{18}\text{O}$; $\pm 0.4\text{‰}$ for $\delta^2\text{H}$). Samples from Baker (Moss) Creek were analysed at the National Hydrology Research Centre in Saskatoon, Canada using an isotope ratio mass spectrometer following Lis *et al.* (2008). Wolf Creek samples were analysed with a Thermo Fischer Delta Plus Advantage mass spectrometer connected to a GFL 1086 equilibration device (for 2001–2004; precision $\pm 0.1\text{‰}$ for $\delta^{18}\text{O}$; $\pm 1\text{‰}$ for $\delta^2\text{H}$) and a Finnigan MAT Delta plus XP + GasBench isotope ratio mass spectrometer (for 2005–2009; precision $\pm 0.1\text{‰}$ for $\delta^{18}\text{O}$; $\pm 0.5\text{‰}$ for $\delta^2\text{H}$). Isotope signatures are reported in the δ -notation (‰) after calibration using Vienna Standard Mean Ocean Water standards. Stable isotope data for Dry Creek were obtained from Tappa (2013).

The precipitation and streamwater samples for each site were plotted against the Global Meteoric Water Line (GMWL). This shows the relative change of $\delta^{18}\text{O}$ to $\delta^2\text{H}$ in precipitation assuming equilibrium fractionation in the atmosphere, which produces a slope of 8 (see Kendall and Caldwell, 1998 for full details). In terrestrial hydrological systems, kinetic fractionation of isotopes occurs during phase changes of water (e.g. between vapour and liquid,

Table IIa. Description of parameters of the HBV model

| | Units | Description |
|--------|---------|--|
| TT | °C | Threshold temperature |
| CFMAX | mm/d °C | Degree-day factor |
| SFCF | — | Snowfall correction factor |
| CFR | — | Refreezing coefficient |
| CWH | — | Water holding capacity |
| FC | mm | Maximum soil moisture storage |
| LP | — | Soil moisture value above which ET_{act} reaches ET_{pot} |
| BETA | — | Determines relative contribution to runoff from rain or snowmelt |
| PERC | mm/d | Maximum percolation rate from upper to lower groundwater store |
| UZL | mm | Threshold parameter |
| K0 | l/d | Recession coefficient |
| K1 | l/d | Recession coefficient |
| K2 | l/d | Recession coefficient |
| MAXBAS | d | Length of triangular weighting function |

Table IIb. Initial and final parameter ranges for the HBV model for the different VeWa sites

| | Initial values | | Krycklan | | Baker Creek* | | Wolf Creek | | Gimrock | | Dorset | | Dry Creek | | |
|--------|----------------|-------|----------|-------|--------------|--------|------------|------|---------|-------|--------|-------|-----------|--------|-------|
| | Lower | Upper | Lower | Upper | Mean | Lower* | Upper* | Mean | Lower | Upper | Mean | Lower | Upper | Mean | |
| TT | -2 | 2.5 | -1.5 | 1.5 | -0.03 | 0.39 | -1.5 | 1.5 | 1.01 | -1.5 | 1.5 | -1.5 | 1.5 | -0.83 | 1.5 |
| CFMAX | 1 | 10 | 1 | 10 | 1.58 | 3.9 | 1 | 6 | 1.47 | 1 | 10 | 1 | 10 | 1.78 | 8 |
| SFCF | 0.5 | 1.2 | 0.5 | 0.9 | 0.64 | 0.85 | 0.5 | 0.9 | 0.86 | 0.5 | 0.9 | 0.5 | 0.9 | 0.54 | 0.9 |
| CFR** | 0 | 0.1 | 0.05 | 0.05 | | | 0.05 | | 0.05 | 0.05 | | 0.05 | | 0.05 | 0.05 |
| CWH** | 0 | 0.2 | 0.05 | 0.05 | | | 0.1 | | 0.1 | 0.1 | | 0.1 | | 0.1 | 0.1 |
| FC | 50 | 600 | 200 | 600 | 270.91 | 280.79 | 200 | 600 | 436.51 | 50 | 500 | 50 | 500 | 217.05 | 200 |
| LP | 0.3 | 1 | 0.3 | 1 | 0.86 | 0.94 | 0.3 | 1 | 0.92 | 0.3 | 1 | 0.3 | 1 | 0.67 | 0.3 |
| BETA | 1 | 6 | 1 | 5 | 1.56 | 2.56 | 1 | 5 | 1.19 | 1 | 5 | 1 | 5 | 3.32 | 1 |
| PERC | 0 | 4 | 0 | 4 | 1.81 | 0.06 | 0 | 4 | 2.35 | 0 | 4 | 0 | 4 | 2.04 | 0 |
| UZL** | 0 | 70 | 30 | | | 60 | | | | 20 | | 20 | | | 20 |
| K0 | 0.1 | 0.5 | 0.1 | 0.5 | 0.28 | 0.32 | 0.1 | 0.5 | 0.34 | 0.1 | 0.5 | 0.1 | 0.5 | 0.2 | 0.1 |
| K1 | 0.05 | 0.2 | 0.05 | 0.3 | 0.23 | 0.04 | 0.01 | 0.2 | 0.16 | 0.05 | 0.3 | 0.05 | 0.3 | 0.19 | 0.05 |
| K2 | 0.001 | 0.1 | 0.001 | 0.1 | 0.07 | 0.04 | 0.001 | 0.1 | 0.07 | 0.001 | 0.1 | 0.04 | 0.1 | 0.07 | 0.001 |
| MAXBAS | 1 | 7 | 1 | 7 | 1.52 | 4.59 | 1 | 7 | 1.6 | 1 | 5 | 1 | 5 | 1.9 | 1 |

*Same parameter range used for Wolf Creek and Baker Creek;

**CFR, CWH and UZL were fixed parameters

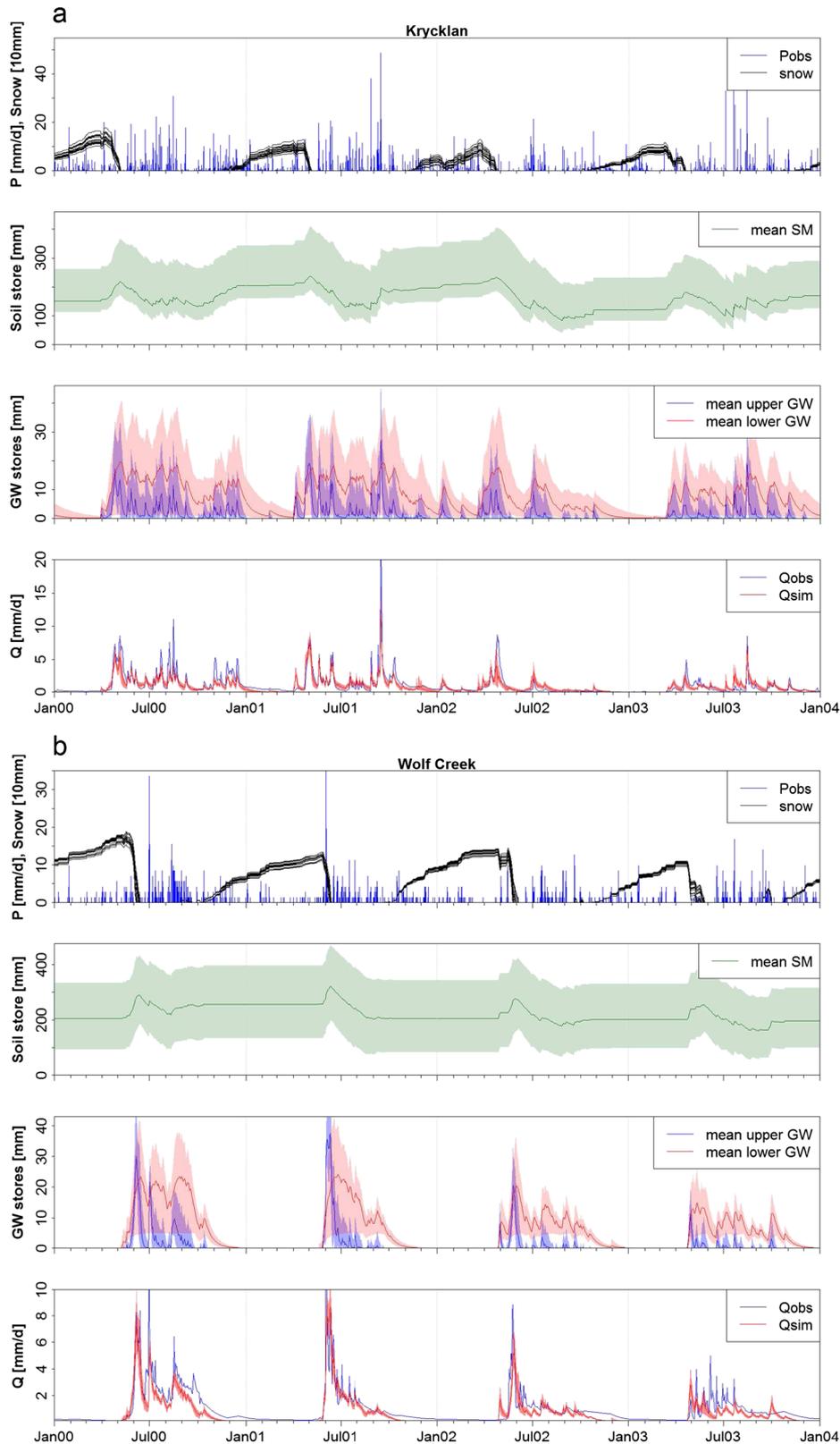


Figure 3. Examples of 4-year time series of model results for (a) Krycklan, (b) Wolf Creek, (c) Girnock and (d) Dry Creek. Top panel: observed precipitation and modelled snow (The unit for snow is mm, it is given as the water equivalent of the snowpack. For some catchments this is plotted in cm (10 mm), for other catchments, it is mm, this depends on the snow amount, results from 20 model runs are shown); second and third panels: average soil moisture and groundwater over the 20 model runs with bands indicating the range for 20 model runs; bottom panel: observed and simulated discharge (average and range over 20 model runs)

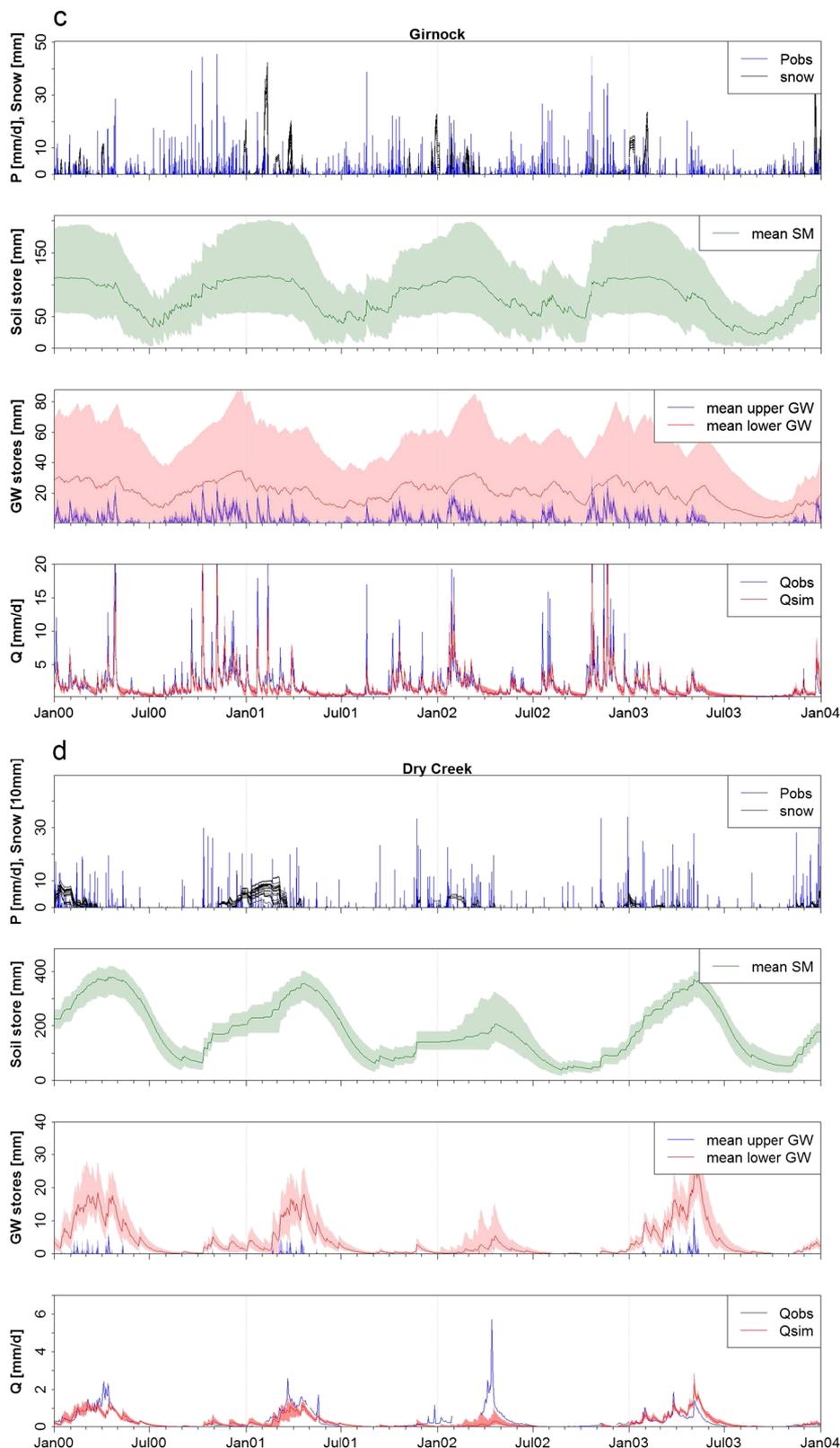


Figure 3. Examples of 4-year time series of model results for (a) Krycklan, (b) Wolf Creek, (c) Girnock and (d) Dry Creek. Top panel: observed precipitation and modelled snow (The unit for snow is mm, it is given as the water equivalent of the snowpack. For some catchments this is plotted in cm (10 mm), for other catchments, it is mm, this depends on the snow amount, results from 20 model runs are shown); second and third panels: average soil moisture and groundwater over the 20 model runs with bands indicating the range for 20 model runs; bottom panel: observed and simulated discharge (average and range over 20 model runs)

liquid and solid and solid and vapour) as ‘light’ isotopes diffuse more rapidly. Thus, evaporation (or sublimation) produces vapour that is lighter than the source water and enriched in $\delta^2\text{H}$ relative to $\delta^{18}\text{O}$. This leaves the source water relatively depleted in $\delta^2\text{H}$ compared with $\delta^{18}\text{O}$ and can be used to infer evaporative effects by a change in gradient of the regression equation relative to the GMWL. Similarly, snowmelt is usually initially isotopically depleted compared with the bulk snowpack because of differences in the melting temperatures of waters with different isotope ratios (Cooper, 1998). This consequently enriches both the remaining snowpack and later melt water. This changing composition and fractionation of stream waters were explored relative to the GMWL (i.e. changing slopes of the $\delta^{18}\text{O} - \delta^2\text{H}$ regression) to tentatively explore for evidence of ecohydrological separation.

Time series analysis of isotopic tracers in precipitation and stream water has been widely applied to explore transit times and transit time distributions (McGuire and McDonnell, 2006). Here, we employed a simpler approach of using the standard deviation of the isotopic composition in stream water as an index of the variability in streamwater isotopic signatures. Previous studies have shown the standard deviation of stream water alone gives a reasonable first approximation of tracer damping that correlates closely with travel times; lower standard deviations infer longer travel times and vice versa (McGuire *et al.*, 2005; Soulsby and Tetzlaff, 2008). It can be used as a simple travel time proxy for qualitative inter-catchment comparison. Unfortunately, differences in the sample frequency and snow influence precluded comparisons to the standard deviation or range of precipitation (Tetzlaff *et al.*, 2009).

RESULTS

Differences in hydrological partitioning

The HBV model provided reasonable simulations of the runoff response of all the catchments (Figure 3 shows four of the sites as examples). Initial and calibrated parameter ranges are shown for the 14 model parameters in Table II. Model performance was reasonable, given the very simple model structure with mean Nash–Sutcliffe efficiencies of >0.6 for Krycklan, Girnock and Dorset (see examples in Figure 3a and c), but only >0.5 for Baker Creek, Wolf Creek and Dry Creek. However, the volumetric errors were generally within 30 mm at all sites except Wolf Creek (ca. 100 mm). The lower efficiencies for Wolf Creek (Figure 3b) and Baker Creek are mainly due to difficulties measuring winter low flows when the stream surface is frozen, and the very specific timing of the melt peak, which is difficult to capture on the correct day using such a simple model. Degree day methods are

Table III. Storage estimates based on the HBV model simulations

| Site | Average annual maximum snowpack SWE (mm) | Average annual mean snowpack SWE (mm) | Average annual storage, soil store (mm) | Average annual storage change, soil store (mm) | Average annual storage change, soil store (mm) | Average annual storage change, two GW stores (mm) | Average annual storage change, lower GW store (mm) | Average annual storage change, upper GW store (mm) |
|-------------|--|---------------------------------------|---|--|--|---|--|--|
| Krycklan | 125 | 63 | 162 | 120 | 34 | 19 | 19 | 19 |
| Baker Creek | 99 | 56 | 160 | 108 | 40 | 7 | 40 | 40 |
| Wolf Creek | 160 | 93 | 219 | 103 | 46 | 24 | 27 | 27 |
| Girnock | 21 | 5 | 82 | 74 | 37 | 22 | 19 | 19 |
| Dorset | 117 | 60 | 171 | 131 | 51 | 23 | 34 | 34 |
| Dry Creek | 66 | 29 | 203 | 289 | 28 | 24 | 9 | 9 |

GW, groundwater

less effective in more northerly environments where radiation (rather than temperature) is the main driver of melt. For Dry Creek (Figure 3d), the main problem is that the simple model structure used for this initial comparison does not take the large elevation differences into account; thus, higher altitude snowpack accumulation and melt are sometimes not captured.

The modelling was used as a basis for a preliminary inter-site comparison of dynamic storage changes in the snow, soil and groundwater compartments simulated by the best 20 models (Table III and Figure 3). General snowpack dynamics simulated by the model were broadly similar to those measured at the sites with the largest, and longest, storage from snowpack accumulation at

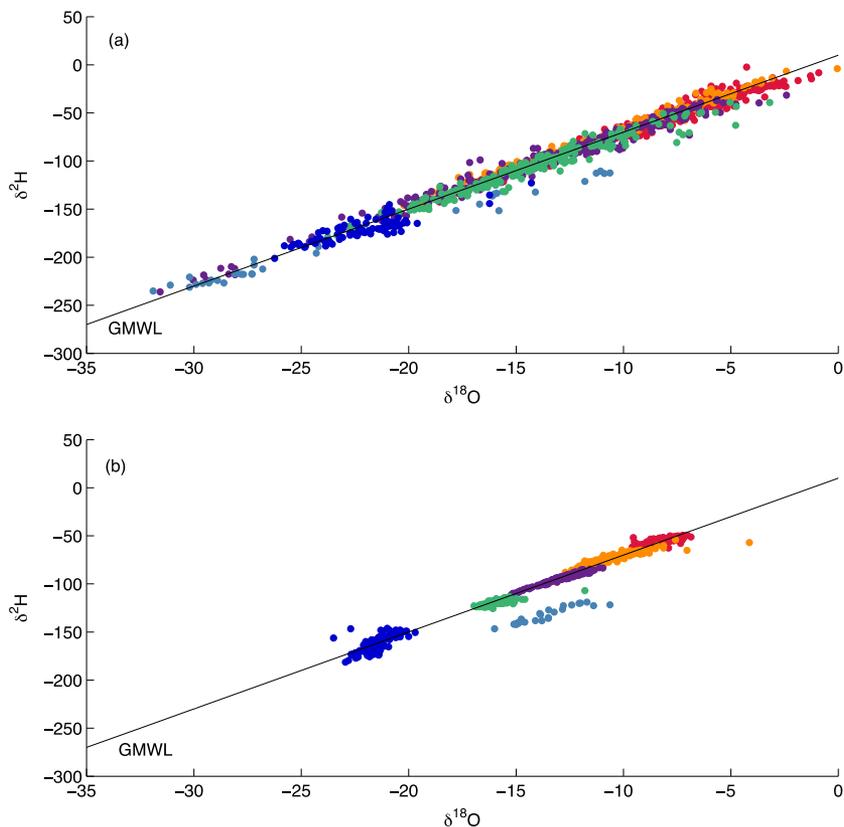


Figure 4. Stable isotopes at all sites plotted along the Global Meteoric Water Line: (a) precipitation and (b) streamwater. Colour of dots: Krycklan–purple; Baker Creek (data from Moss Creek, which is a tributary of Baker Creek)–light blue; Wolf Creek–dark blue; Girnock (data from Bruntland, which is a tributary of the Girnock)–red; Dorset–orange; Dry Creek–green

Table IV. Summary statistics of $\delta^2\text{H}$ values (in brackets are site names of tributaries the isotope data are from)

| Site | Min | Max | Median | Standard deviation |
|--------------------------|--------|--------|--------|--------------------|
| Precipitation | | | | |
| Krycklan | −236.0 | −31.5 | −94.6 | 41.2 |
| Baker Creek (Moss Creek) | −235.1 | −110.6 | −208.2 | 43.4 |
| Wolf Creek (Granger) | −201.2 | −122.8 | −169.9 | 12.3 |
| Girnock (Bruntland Burn) | −147.3 | −2.5 | −50.6 | 23.0 |
| Dorset | −148.0 | −3.9 | −79.6 | 36.2 |
| Dry Creek | −179.2 | −39.1 | −105.0 | 28.8 |
| Streamwater | | | | |
| Krycklan | −111.6 | −83.2 | −94.5 | 5.3 |
| Baker Creek (Moss Creek) | −146.7 | −118.9 | −135.5 | 8.5 |
| Wolf Creek (Granger) | −181.5 | −146.1 | −164.8 | 7.9 |
| Girnock (Bruntland Burn) | −73.1 | −49.5 | −56.2 | 3.5 |
| Dorset | −87.9 | −54.2 | −73.2 | 5.9 |
| Dry Creek | −125.7 | −107.0 | −119.3 | 2.5 |

Krycklan, Wolf Creek and Dorset (cf. Carey and Quinton, 2004; Laudon *et al.*, 2004; Fu *et al.*, 2014). Smaller snowpack accumulations were still a dominant influence on the water balance at Baker Creek and Dry Creek. In contrast, snow influence tends to be very transient at Girnock, with a relatively small snowpack developing for short periods.

The relative influence of snowmelt on the annual runoff regime differs markedly among catchments (Figure 2). At Baker Creek and Wolf Creek, snowmelt clearly drives the annual hydrological response, building to a maximum discharge in spring (May–June and April/May, respectively). This dominant influence of spring snowmelt on

annual peak discharge is also the case at Krycklan, Dorset and Dry Creek, although autumn rainfalls can produce annual maximum runoff in some years at these sites. The influence of snow on flows in the Girnock is much less pronounced, as rainfall events usually produce the greatest runoff responses; nevertheless, winter rain-on-snow events can produce some of the largest runoff responses.

Modelled changes in soil water storage exceeded those for groundwater at all sites. At most sites, snowmelt exerted a strong influence on replenishment of soil moisture in the spring. The mean calibrated maximum soil water storage parameter (FC) was highest in Dry Creek (ca. 570 mm) and lowest for the Girnock (ca. 100 mm)

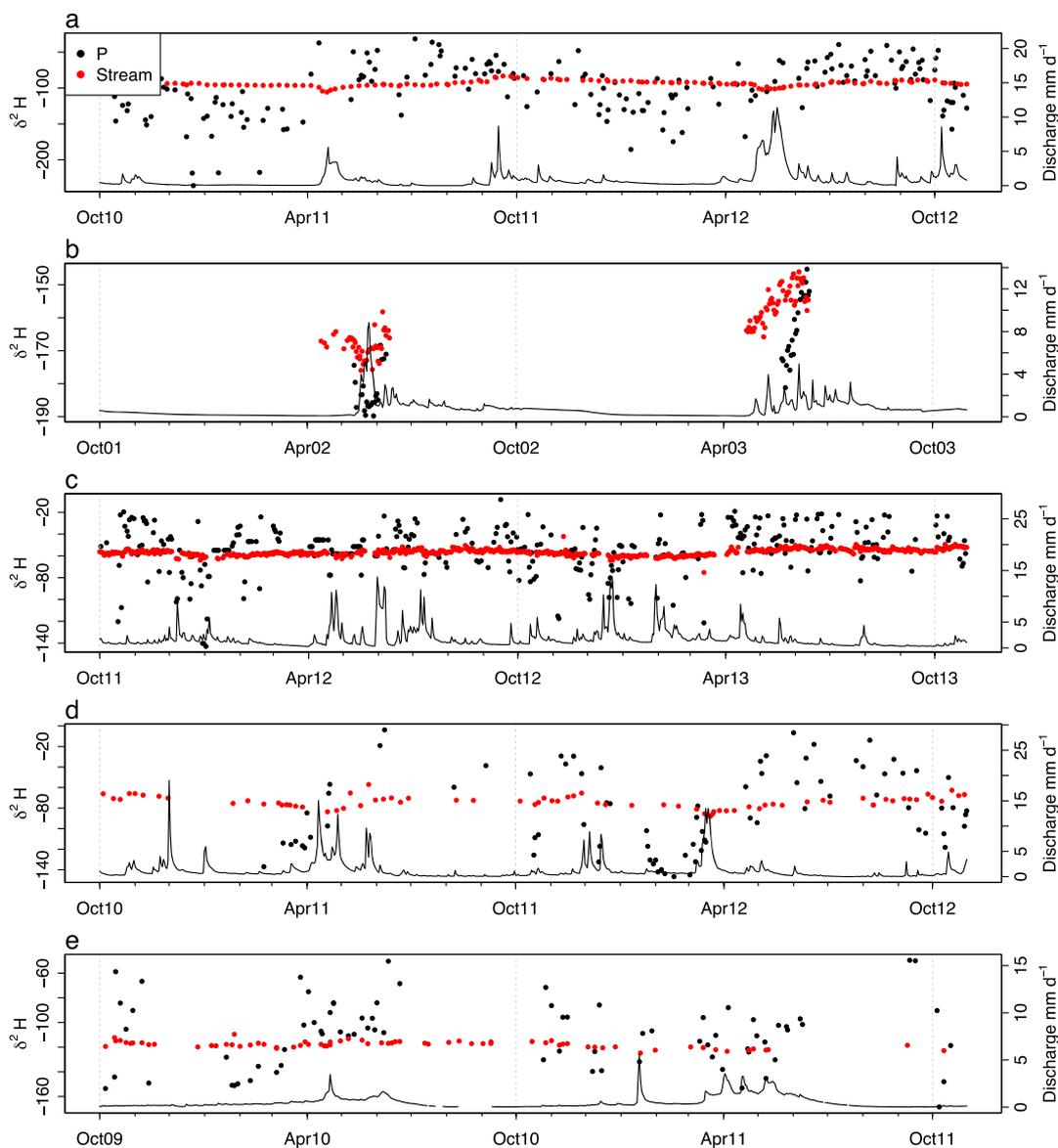


Figure 5. Time series of continuous discharge and $\delta^2\text{H}$ for Krycklan, Wolf Creek, Girnock (isotope data from the Bruntland Burn tributary), Dorset and Dry Creek (Baker Creek is not shown here, given gaps in the isotope and flow data)

(Table II). The remaining catchments were in the 200–400 mm range. These rankings were similar for the annual soil water storage change, which varied between 289 mm for Dry Creek and 76 mm for the Girnock, with the other catchments exhibiting changes of ~ 100 mm (Table III). At most sites, the largest storage change reflected snowmelt inputs in spring and subsequent summer evaporation losses, although soil moisture could be replenished by summer rainfall at Krycklan, Dorset and particularly Girnock.

Storage changes in the two groundwater boxes in HBV are the main drivers of the modelled storm runoff response, and simulated changes are smaller than that for the soil store. Recharge into the deeper groundwater store is a calibrated parameter, and again, Dry Creek and Girnock showed the highest and lowest recharge rates, respectively (defined as the Maximum percolation (PERC) parameter in mm/d), with most of the other catchments all being similar. Of these, Baker Creek had the lowest values reflecting more limited deep

groundwater influence. In terms of runoff partitioning between the rapid (upper) and slower (lower) responding groundwater reservoirs boxes, the Girnock had the highest calibrated K1 parameter and the greatest contribution of runoff from the upper box. This is consistent with its rapid hydrological response, rainfall-driven flow regime and rapid rates of runoff. In contrast, Dry Creek has the most attenuated catchment response, with a small upper groundwater recession coefficient and most streamflow contributed from the slow groundwater store. The other catchments were less different; Wolf Creek had the second lowest K1 parameter, with Krycklan the second highest. The partitioning of flows between the fast and slow flow reservoirs was similar.

Isotope dynamics

Figure 4 and Table IV summarize the isotopic composition of precipitation and stream flow for at least 2 years at each site. The space occupied by each site along the meteoric water line broadly reflects the combined

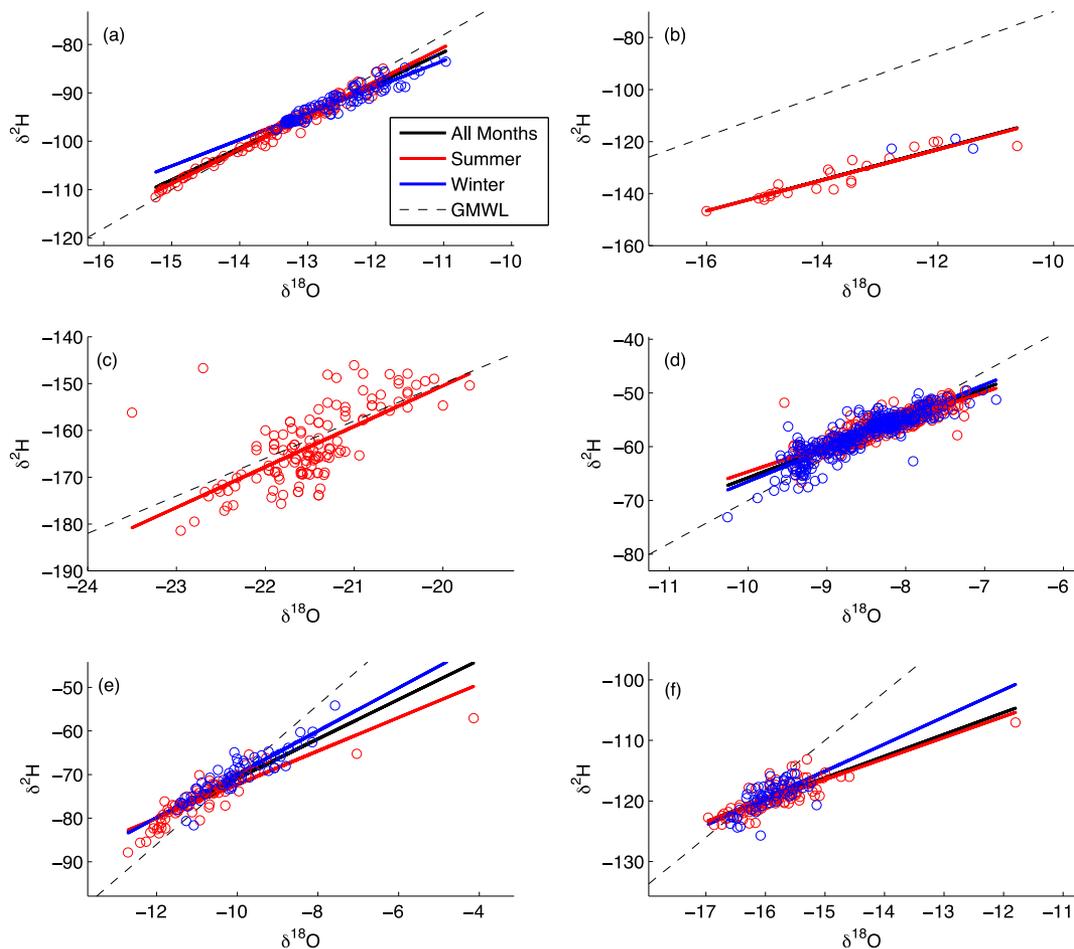


Figure 6. Deviation of streamwater $\delta^2\text{H}$ from the Global Meteoric Water Line (GMWL) in summer (red line) and winter (blue line) months. (slope of GMWL is $m=8$). Slopes of the deviations from GMWL are given in Table V. (a) Krycklan; (b) Baker Creek (isotope data from Moss Creek tributary); (c) Wolf Creek; (d) Girnock (isotope data from Bruntland Bum tributary); (e) Dorset; and (f) Dry Creek

Table V. Regression equations for annual, summer and winter isotope relationships showing deviation from the Global Meteoric Water Line streamwater isotope data (in brackets are site names of tributaries the isotope data are from)

| Site | Summer | Winter | All year |
|--------------------------|-------------------------------------|-------------------------------------|-------------------------------------|
| Krycklan | $y = 7.06x - 2.91$ $R^2 = 0.97$ | $y = 5.44x - 23.53$ $R^2 = 0.89$ | $y = 6.60x - 8.97$ $R^2 = 0.95$ |
| Baker Creek (Moss Creek) | $y = 5.90x - 52.30$ $R^2 = 0.87$ | No data | |
| Wolf Creek (Granger) | $y = 8.66x + 22.74$ $R^2 = 0.46$ | No data | |
| Girnock (Bruntland Burn) | $y = 4.93x - 15.32$ $R^2 = 0.76$ | $y = 6.01x - 6.42$ $R^2 = 0.78$ | $y = 5.53x - 10.42$ $R^2 = 0.78$ |
| Dorset | $y = 3.85x - 33.83$ $R^2 = 0.77$ | $y = 4.97x - 20.29$ $R^2 = 0.83$ | $y = 4.53x - 25.61$ $R^2 = 0.81$ |
| Dry Creek | $y = 3.47x - 64.46$ $R^2 = 0.69$ | $y = 4.48x - 47.95$ $R^2 = 0.48$ | $y = 3.60x - 62.19$ $R^2 = 0.61$ |

effects of latitude and continentality. Thus, precipitation at Baker Creek and Wolf Creek is the most depleted, followed by Dry Creek, Krycklan and Dorset. The Girnock site (Bruntland Burn tributary) is least depleted, reflecting its southerly, more maritime location. There is a marked seasonality in precipitation, which at all sites is more depleted in winter and more enriched in summer, although intra-seasonal variation can also be significant (Figure 5).

The temporal variability of the isotopic composition of stream waters is damped as compared with that of precipitation. The position on the meteoric water line reflects the same influences of latitude and continentality as for precipitation. In terms of median values, $\delta^2\text{H}$ -values in stream water are most depleted in the order Wolf Creek < Baker Creek < Dry Creek < Krycklan < Dorset < Girnock (Table IV). Baker Creek strikingly plots below the GMWL, presumably because of the fractionation from open water evaporation from the lakes covering 21% of the catchment (see the following). Variability in stream isotopic composition is indexed by the range and standard

deviation, which are greatest at Wolf Creek and Baker Creek, followed by Dorset and Krycklan. Dry Creek has the lowest variability, with the Girnock second lowest.

Much of the intra-site variability in stream water $\delta^2\text{H}$ -values reflects the annual snowmelt influence, which delivers depleted water to the stream channel in the initial phases of melt via rapid flowpaths (Figure 5). Thus, the timing of the measured greatest stream water depletion coincides with peak flows and modelled spring melt at most sites (cf. Figure 3). Variability in snowmelt contribution to streamflow is evident in the degree and longevity of depletion. This contribution from depleted meltwater to streamflow is most clearly evident in the Krycklan and Dorset data (Figure 5), where inter-annual variability in longevity and depletion also occurs. However, the importance of late summer/early autumn rainfall (see also Figure 2) is apparent in routing more enriched summer precipitation and possibly displacing soil water that has been subjected to evaporation to the stream channel along rapid flowpaths.

Table VI. Pearson correlation coefficient for correlations between standard deviation of streamwater signatures and slopes of the $\delta^{18}\text{O}$ - $\delta^2\text{H}$ plots against catchment characteristics (in bold r coefficients above 0.7)

| | Stdev streamwater | Slope summer | Slope winter | Slope all year |
|--|-------------------|---------------|----------------|----------------|
| Average annual storage change of the sum of two GW stores (mm) | 0.64 | 0.22 | 0.13 | 0.35 |
| Average annual maximum snowpack SWE (mm) | 0.70* | 0.65 | -0.37 | 0.62 |
| Average annual mean snowpack SWE (mm) | 0.77* | 0.70* | -0.36 | 0.67 |
| Average annual storage change, soil store (mm) | -0.55 | -0.54 | -0.90** | -0.63 |
| Average annual storage, soil store (mm) | 0.30 | 0.26 | -0.95** | 0.20 |
| Average annual storage change, lower GW store (mm) | -0.57 | -0.13 | -0.56 | -0.07 |
| Average annual storage change, upper GW store (mm) | 0.86** | 0.22 | 0.18 | 0.28 |
| Mean annual T (°C) | -0.97** | -0.75* | -0.34 | -0.74* |
| Mean annual P (mm) | -0.64 | -0.51 | 0.60 | -0.40 |
| Mean runoff ratio | 0.26 | 0.59 | 0.85** | 0.71* |
| Baseflow index | 0.02 | -0.04 | -0.76** | -0.10 |

SWE, snow water equivalent

** $p < 0.05$;

* $p < 0.1$

Data for Wolf Creek are limited (Figure 5) to the spring/summer field season, but depression of the streamflow isotopic signature during rapid spring melt in 2002 is evident. This is followed by marked variability in the summer that is probably influenced by the effects of altitude and aspect on spatially variable melt as well as thawing of ground frost and inputs of enriched summer precipitation (Carey *et al.*, 2013a).

The Girnock shows more frequent and gradual isotope changes in response to rainfall and snowmelt throughout the year (Figure 5). Depending on season, these can produce depletion during winter or enrichment in the summer, giving a sinusoidal streamwater signal over the course of the year. As the Bruntland Burn in the Girnock has the highest resolution data, a sinusoidal streamwater signal is most clear for this site, although it is also evident at other sites such as Baker Creek (not shown, but see Gibson and Reid, 2010). For Dry Creek, the data are less comprehensive; nevertheless, the lack of variability in stream water $\delta^2\text{H}$ -values is apparent (Figure 5e), such that translation of depleted meltwater to the stream during snowmelt is barely discernible.

In addition to the hydrologically driven variation in stream water $\delta^2\text{H}$ -values at higher flows, deviations from the GMWL can show more subtle effects of internal catchment processes on isotopic fractionation most likely through evaporation. Such effects are highly variable among sites, and there are some suggestions of seasonal differences between winter and summer at some sites as evidenced by the different slopes of the lines, although these differences were not statistically significant (Figure 6 and Table V). Regression lines show the slopes for the summer increase from Dry Creek < Dorset < Girnock < Baker Creek < Krycklan < Wolf Creek, which broadly reflects the declining energy gradient and effects on evaporative fractionation (Figure 6). This summer effect of fractionation remains evident for the whole year data sets where the same sequence is maintained. At each individual site, the timing of the more fractionated water varies, and it is difficult to separate the effects of enriched summer precipitation from the effects of evaporation. At Krycklan and Dorset, the most enriched and fractionated samples generally occurred in late summer/autumn. Some spring samples for Dorset were also enriched and fractionated, which may indicate enriched rain-on-snow inputs. In the Girnock, the most fractionated waters were evident in midsummer. For Dry Creek, the most fractionated sample occurred in mid-summer, which was also in response to enriched summer precipitation. For Baker Creek, the fractionation was more clearly marked than at any other site in terms of plotting below the GMWL, reflecting the strong lake influence, which may be exacerbated by slow movement of water through seasonally connected wetlands and surface waters.

Influence of catchment characteristics on isotope dynamics

Table VI shows the Pearson correlation coefficients for metrics derived from the time series of stream isotope $\delta^2\text{H}$ -values and various catchment characteristics, including selected storage features derived from the HBV modelling. Given the small number of sites, few relationships are statistically significant; however, those that are provide important insights. The average annual maximum snowpack and annual mean snowpack storage were positively correlated with the standard deviation of stream water $\delta^2\text{H}$ -values used as a tentative proxy of both catchment damping and travel times. This implies the dominant role of snowmelt on stream water isotopic composition at these sites. The standard deviation of stream water $\delta^2\text{H}$ -values is also strongly positively correlated with the average annual storage change in the upper groundwater store, which would be consistent with the importance of responsive hydrological flowpaths in routing melt water isotopic signatures to the stream. The mean runoff ratio was also positively correlated with the slope of the regression equation for stream waters in winter and for the whole year, implying less fractionation in the most responsive catchments. This was supported by the negative correlation between the baseflow index and the slope for winter isotopes.

DISCUSSION

Hydrological partitioning in the study catchments

Despite differences in both hydroclimate and catchment characteristics, the hydrological response and isotope dynamics of the VeWa sites exhibit many commonalities reflecting the distinct hydrological characteristics of northern catchments (Tetzlaff *et al.*, 2015). The dominance of snow at all sites, with the exception of the more rainfall-driven Girnock, affects the annual flow regime, the water partitioning inferred by the HBV modelling and the response of $\delta^2\text{H}$ in streamwater. Annual snowmelt usually determines the annual flow maxima at most catchments as well as controlling the timing of the most depleted stream water isotope signals, as suggested by correlations between indices of isotope variation and annual snowpack accumulation and model-based metrics of rapidly responding hydrological pathways (Table VI).

Detailed field experimentation in the study catchments has characterized the nature of these responsive flowpaths. At Krycklan, spatial differences in runoff processes driven by annual snowmelt have been reported by Laudon *et al.* (2007). Here, tracer responses in rapid runoff from forested areas during melt are relatively damped as a result of mixing with soil waters as transmissivity feedback occurs in unfrozen soils. In contrast, the wetland areas freeze and generate contributions of more marked and less mixed

meltwater signals to the stream channel network from the overland flow (Peralta-Tapia *et al.*, 2015b). However, summer and autumn rainfall is also important in the overall water balance and can displace water from both forest and wetland areas. In contrast, winter baseflows are mainly generated from deeper groundwater when water sources nearer the surface ‘switch off’ in the period of snow accumulation (Peralta-Tapia *et al.*, 2015b).

High spatial and temporal complexity has been reported during snowmelt for the Moss Creek tributary of the Baker Creek catchment (Spence, 2007). Marked temporal variations in the connectivity of surface water sources in the basin’s complex and lake-influenced channel network generally restrict the area influencing runoff as the summer progresses (Phillips *et al.*, 2011; Spence *et al.*, 2014). However, the fractionation effects of the lakes transform the isotope composition of waters as they transit through the system (Gibson and Reid, 2010).

At Wolf Creek, the evolution of rapid, shallow flowpaths in organic-mantled soils coupled with the gradual descent of the frost table and spatial patterns of snowmelt has been shown (Carey and Quinton, 2004; Quinton *et al.*, 2009; Carey *et al.*, 2014). The runoff generation from initial snowmelt on south-facing slopes progresses to runoff from north-facing, permafrost-underlain slopes with expected altitudinal delays (McCartney *et al.*, 2006). Ambiguities in tracer-based hydrograph separation studies have hinted at the complex spatial and temporal sequencing of runoff generation in the catchment, which underpins the variability around the stream water isotope response to melt (Carey and Quinton, 2004; Carey and Quinton, 2005; Boucher and Carey, 2010; Carey *et al.*, 2013b).

In the Girnock catchment, detailed process studies in the Bruntland Burn tributary provide a rich set of isotope data for precipitation, soil water, groundwater and streams (Geris *et al.*, 2015). Overland flow from riparian peat soils dominates storm runoff generation, which has a non-linear relationship with catchment wetness, as freely draining podzols on steeper hillslopes can connect with the wetlands (via groundwater exfiltration) in the wettest conditions (Tetzlaff *et al.*, 2014). While deeper groundwater sources sustain low flows, there is usually some component of surface seepage from the wetlands (Birkel *et al.*, 2014). This results in rapid routing of precipitation $\delta^2\text{H}$ -values, albeit highly damped, into the stream, which explains the summer dominance of enriched isotopes and winter depletion.

There have been less detailed hydrological process studies in the Harp-4 catchment at Dorset, although McDonnell and Taylor (1987) showed the importance of snowmelt in triggering surface and subsurface flowpaths in routing water to streams. More recent work has shown the importance of riparian wetlands in generating runoff (Fu *et al.*, 2014), although previous work has documented

how such wetlands can dry out in summer (Devito *et al.*, 1999) and serve as sinks for hillslope runoff contributions rather than source areas. Nevertheless, unconfined aquifers in deeper till cover in portions of the catchment (Hinton *et al.*, 1993) can sustain baseflow.

Both the modelling and the stable isotope data show that Dry Creek is the most groundwater-dominated of the VeWa catchments (Kelleners *et al.*, 2010), even though snowmelt can contribute directly to stream flow generation (Eiriksson *et al.*, 2013). Winter and spring melt mainly replenishes soil moisture and groundwater (Miller *et al.*, 2008; Kormos *et al.*, 2014; Aishlin and McNamara, 2011), although rates and timing are spatially variable in relation to altitude and aspect (Stratton *et al.*, 2009).

Isotope damping and fractionation

Although isotopic variability in stream waters was damped, our analysis suggests that there was a strong snow melt influence, which helped explain inter-site differences (Table VI). Hydrological partitioning largely determines how rapidly the seasonally varying inputs in snowmelt and rainfall are routed to streams. However, analysis of departures from the meteoric water line, as well as the process understanding from individual catchments, highlights the limitations of input–output relationships alone. Evidence of fractionation in stream waters strongly hints at the importance of internal catchment processes in influencing the isotopic composition of outputs, and these appear to have different controls among catchments. Regardless, input–output data provide limited insight into these processes, a problem compounded in a meta-analysis such as this where data have different sampling frequency and are not strictly comparable.

At all sites, there is the potential for fractionation in the snowpack and preferential release in melt waters, which is well known from previous work (see summary by Rodhe, 1998; Taylor *et al.*, 2002). However, large energy inputs in summer generally have the strongest influence on potential evaporative fractionation. While energy inputs (and hence, latitude and specific hydroclimate) are first-order controls on fractionation, catchment characteristics (particularly landcover and soils) are also important, particularly at sites where lakes or wetlands with surface pools persist through the summer.

Consequently, Dry Creek, the most southerly site with the highest summer energy inputs, shows the most marked effects of evaporative fractionation in stream water. This may be caused by evaporation from surface soils during and after snowmelt, which influences the composition of soil moisture and groundwater. Both Dorset and the Girnock also show evidence of summer fractionation, presumably as a result of evaporation from wetlands soils that are connected to the channel network under

moderate/dry flow conditions (Devito *et al.*, 1999; Birkel *et al.*, 2011). Dorset also shows heavy fractionated water in the autumn, which does not only partly relates to the effect of enriched precipitation but may also reflect the displacement of soil water with enriched isotopic composition that has been subjected to evaporation from the drier forest soils during autumn rain events.

In Wolf Creek, the response is complex, as snowmelt and seasonal permafrost thaw are dependent on aspect and altitude, and thus exert a highly spatially variable influence on stream flow composition (Lessels *et al.*,). The influence of extended surface water stores is particularly marked at Baker Creek, which shows the effect of lakes in terms of dramatic impacts on evaporative losses in summer (Spence, 2007). Lakes comprise 21% of the catchment area, and their role in evaporative fractionation of water contributions to streamflow may be augmented by the extensive peatlands, which will also have surface water pools that can be fractionated (Gibson and Reid, 2010). However, the effect of wetlands in headwaters on stream water isotope signatures at the catchment outlet may be reduced by declining connectivity in the active stream network in the post-melt period (Phillips *et al.*, 2011).

At Krycklan, the high slope of the summer waters in the isotope plot (Figure 6) reflects the continuing effect of snowmelt through the early part of the summer. Autumn streamflows show evidence of fractionated waters that are more enriched than summer precipitation, indicating fluxes of soil water, with enriched isotopic composition that has been subjected to evaporation, into streams (Peralta-Tapia *et al.*, 2015b). It is unclear whether this contribution is from wetland or forested areas, or both.

The role of land cover and soils on water partitioning: open questions

Despite these implied controls on stream water isotope composition, disentangling the relative influence of hydroclimate, land cover and soil characteristics is not possible with simple input–output data alone. Although such an integrated framework as used in this study of combining stable isotopes and modelling provides an insightful basis for inter-site comparison, it is clear that new, more detailed data at a range of spatial and temporal scales for process understanding and the use of models for hypothesis testing will aid future projections. The general relationship between hydrology and catchment characteristics is underpinned by more complex interactions, and it seems that the roles of vegetation and soils in the so-called ‘critical zone’ is of particular significance in terms of the water cycle and isotope systematics (McDonnell, 2014). This ‘critical zone’ is the uppermost land surface layer where energy and water fluxes drive coupled physical, chemical and biological processes that support life (Brantley *et al.*, 2007). Whereas individual studies have

examined detailed processes in specific catchments, the paucity of studies in such a geographically extensive area as the biomes of the upper latitudes of the Northern hemisphere suggests we are a long way from any synthesis. This is of concern at a time when climate change is resulting in vegetation changes over extensive areas (Wookey *et al.*, 2009; Epstein *et al.*, 2013; Tetzlaff *et al.*, 2013). The following paragraphs seek to highlight some of the fundamental questions that need to be addressed to understand water partitioning and ecohydrological coupling in high-latitude biomes. It becomes clear that the unique characteristics of northern colder regions mean that dominant paradigms and theories derived from temperate zone hydrology do not necessarily transfer to northern landscapes (Quinton and Carey, 2008).

The role of vegetation in affecting patterns of snow accumulation and subsequent melt is of critical importance, and while the physical effects are increasingly well understood, the implications for isotope dynamics are not. Vegetation canopies can intercept snowfall resulting in transient storage, as well as providing roughness elements for trapping wind-blown snow (Essery and Pomeroy, 2004). How this affects the volumes of subsequent sublimation, local microclimate, melt, evaporation and redistribution as effective precipitation to the soil surface is likely to be highly variable both within and among catchments (Ellis *et al.*, 2010). It will also be affected by land cover change (Reid *et al.*, 2014). Implications for the isotopic composition of water entering the soil are likely to be similarly variable in space and time, needing better characterization (Cooper, 1998).

The resulting pulses of melt water through soils will also be heterogeneous. Through detailed hydrometric studies, we have insight into the spatial and temporal variability of different flowpaths and connectivity to channel networks (e.g. for Dorset McDonnell and Taylor, 1987; Devito *et al.*, 1996; for the Girmock e.g. Birkel *et al.*, 2015). However, the effects of mixing processes on the isotopic composition of soil water are still unclear. Kirchner *et al.* (2001) proposed a general model of advection–dispersion along hillslope flowpaths, which can explain the integration of younger and older waters in stream flow. This provides a testable hypothesis of how tracer mixing occurs vertically in soil profiles and laterally along hillslopes as shown by Tetzlaff *et al.* (2014). This also has implications for the recharge of groundwater, which is thought to have a relatively consistent isotope composition as a result of damping of near-surface heterogeneities in the isotopic signature of inputs and mixing with older waters (Peralta-Tapia *et al.*, 2015b). Further testing is needed to examine whether general theories developed in less extreme hydroclimatic zones transfer to colder environments where deeper penetration of snowmelt signatures may occur in partially frozen soils.

Investigations of isotopic changes along soil flowpaths have recently focused on the rooting zone as an important area of transformation, both in terms of water partitioning (i.e. interception, transpiration, soil evaporation and groundwater recharge) and associated isotope systematics and fractionation (Dawson and Simonin, 2011). Plant–water relations play a key role here, and recent attention has focused on ecohydrological partitioning of water into plant-available water and runoff water that drains to streams (McDonnell, 2014). It has been suggested, at least for strongly seasonal climates with a warm, dry season, that water in the early re-wetting phase is rapidly adsorbed into smaller soil pores for plant use in the following summer (Brooks *et al.*, 2010). This has been established from analysis of soil waters and xylem from trees. However, alternative possible explanations have been put forward in terms of plant physiological controls and root uptake from the finer soil pores as well as factors such as hydraulic lift and possible re-distribution of soil water via root exudates (Dawson and Simonin, 2011). It also may be that aeration of the soil zone may lead to re-equilibration of water isotopic signatures with the soil atmosphere, which may affect the isotope composition of plant-available water as soils dry (Barnes and Turner, 1998). The degree to which any of these processes is active in more energy-limited northern regions is highly uncertain as a result of a paucity of studies. However, preliminary data from Geris *et al.* (2015) at the Girnock catchment provided little evidence to support isotopic separation of soil waters between water that may be uptaken by plants or discharged to streams. Rather, the data imply that fractionated water in plant xylem may reflect summer uptake from upper soil horizons where soil water, which has been subjected to evaporation, is fractionated in the main rooting zone. The degree to which such interactions between water moving via macropores and soil matrix affect stream water isotope values is currently an open question. While detailed studies indicate a high level of spatial variability and heterogeneity at the soil profile and hillslope scale, the input–output relationships are often rather predictable and consistent with relatively well-mixed sources of water (Tetzlaff *et al.*, 2014). It is unclear whether this may reflect emergent properties of the catchment response or limitations in data that result in overly simplistic interpretations.

While there have been extensive advances in empirical and modelling studies of the hydrology of northern regions, particularly in Europe and North America, isotope studies have lagged behind (Tetzlaff *et al.*, 2015). The increased availability of inexpensive, reliable spectrometers offers the opportunity for advances in this regard, and the potential to generate the kinds of data needed to answer the questions presented here. Given the

likely trajectories of climate and associated vegetation change in these areas (e.g. Tetzlaff *et al.*, 2013), such developments are timely. In terms of climate trajectories, changes in magnitude, timing and phase of precipitation, along with changing temperatures, will affect hydrological processes, water balance and short-term hydrological dynamics via melt rates, as well as flowpath partitioning throughout the broader North (Kundzewicz *et al.*, 2007). In terms of changes in their composition and distribution but possibly also species physiology, vegetation communities will respond to rising temperatures and an increasing proportion of precipitation falling as rain (Donohue *et al.*, 2007; Wookey *et al.*, 2009). The potential to integrate isotope data into tracer-aided models that can be used to test hypotheses in a learning framework will facilitate rapid advances to expand the meta-analysis provided here.

CONCLUSIONS

This meta-analysis of six northern catchments has shown that snowpack accumulation and melt typically dominate the catchments' runoff and isotope dynamics. While simple input–output relationships can link stream flow generation to hydrological flowpaths that route water to streams, intra-annual variability and evidence of evaporative fractionation at each site are indicative of more complex internal processes that can be influenced by energy inputs and catchment characteristics. We understand relatively little about how such small-scale processes affect stream water dynamics, and whether and how they should be conceptualized in models, particularly for northern catchments. Major challenges still exist to understand the role and interplay of vegetation and soils on hydrological partitioning in these regions. These questions provide an opportunity for hypothesis-driven isotope studies in northern environments.

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