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Spatio-temporal Variations of d^{18}O Isotope Signatures of Hydrological Components within a Glacierised Mountainous Basin

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ABSTRACT

Understanding the spatio-temporal variations of isotopic tracers in complex mountainous environments can assist with seasonal and inter-annual hydrological process studies. In addition, if components of a hydrological balance can be isotopically characterised through time at a basin end-point, then they can potentially be used to evaluate the hydrological balance predicted in a model run. Over one hundred samples were collected of bulk snowpack, glacial ice, rainfall and various forms of groundwater/baseflow from the Bow Valley in the Canadian Rockies in an attempt to characterise the spatio-temporal variations of stable oxygen isotope compositions for each of these end members. A discussion of the observed changes in each of the flow component signatures with regard to the hydrological processes that may be controlling these dynamic changes in signature is provided. The paper ends with a generalisation (or conceptual model) of the oxygen isotope signatures for each of the major hydrological balance components at a monthly time step for the years 1996 – 1999.

INTRODUCTION

High mountain hydrology is complex and basin hydrographs display the amalgamation of events and baseflow from widely varying spatial, temporal, elevational and lithological origins. In addition, the hydrological origin of runoff in such environments begins its journey in many forms: rain, snow, ice or old groundwater. In large mountainous basins it is virtually impossible to accurately separate the seasonal or interannual hydrograph into all of its component parts and as a result hydrological models used in such areas grossly simplify the distribution of hydrological processes in space and time. One way to test the accuracy of model simplifications is to evaluate hydrological model output using geochemical tracers (see companion paper – Hopkinson *et al*, this issue). However, before models can be assessed using tracers over seasonal and longer time periods, the end-member signature of each tracer must be characterised in space and time.

The purpose of this paper is to characterise spatio-temporal stable oxygen isotope signatures for hydrological balance end members during hydrologically diverse years from 1996 to 1999 to assist with: 1) the interpretation of hydrological processes and 2) hydrological model evaluation within a mountainous environment. This second objective is discussed in the companion paper (Hopkinson *et al,* this issue). The hydrological input components of interest are snowmelt, rainfall and glacial icemelt. The changing characteristics of long-term storage components that make up inter-annual baseflow in the river system are considered a separate but more dynamic end-member. For further discussion of the patterns and uses of oxygen isotope tracers in hydrology, the reader is referred to Dansgarrd (1964), Gaspar (1987) and Clark and Fritz (1997).

STUDY AREA

The Bow Valley above Banff (Figure 1) has an area of 2230 km^2, ranges in elevation from 1350–3500 m.a.s.l and drains the eastern side of the Canadian Rockies between latitudes 51 and 52^o (N). It is largely underlain by carbonate rocks and deep tills, has about 50% forest coverage, is 3% glacierised and has a dendritic river network containing many small lakes. A characteristic of the Bow Valley is the marked difference in glacier cover between the two sides of the basin. The west side is significantly glacierised, whereas on the east side there is generally very little glacier cover. Upstream of Banff, there are several major tributaries, the most significant of these being the Pipestone River, which drains the northeastern slopes (2 % glacierised) and joins the Bow River, which drains the northwestern slopes (12 % glacierised) at Lake Louise.

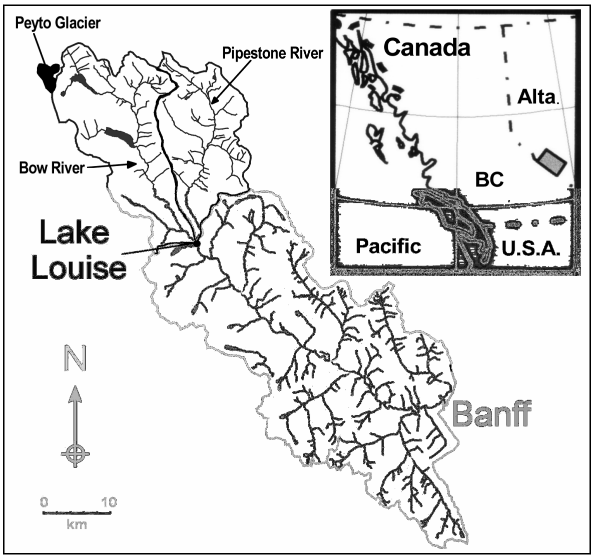

Figure 1. The Bow Valley above Banff showing river network.

The hydrological regime in the basin has been investigated in recent studies. Zawadzki (1997) used the UBC watershed model to forecast changes to the river's flow regime in a scenario of future climatic change. The hydrological balance predicted by the model was brought into question and the need for in depth studies of basin wide hydrological processes over interannual time scales was highlighted (Zawadzki, 1997). To this end, some work was carried out by Hopkinson & Young (1998) to discern proportional contributions of interannual glacier wastage to river flow. During the late summer months, the small glacier cover within the basin can be a significant contributor to flow, particularly during warm and dry years. Hopkinson and Young (1998) estimated that for the low flow year of 1970, glacial wastage contributed around 12% to annual basin yield rising to over 50% for the month of August. However, the hydrological process that dominates timing and volume of annual runoff at Banff is spring snowmelt with peak runoff in June. Therefore, basin runoff varies markedly from year to year, largely as a result of winter snowpack accumulation. The years investigated in this study, 1996 to 1999, displayed widely varying hydrometeorological characteristics with 1996, 1997 and 1999 receiving deep snowpack accumulations while 1998 had a very shallow snowpack due to *El Niño*.

METHODS

From 1996 to 1999, approximately 120 hydrological end-member samples were collected from the Bow Valley upstream of Banff. The end-members sampled were bulk snowpack, rainfall, glacial ice and basin baseflow. All samples were stored in 22 ml gas capped scintillation vials after collection and sent to the University of Waterloo Stable Isotope Laboratory for oxygen isotope analysis. Samples were analysed on a Varian Mass Spectrophotometer. For every twenty samples processed on the spectrophotometer, three samples of standard mean ocean water were included to standardise the data set. Analytical error was in the range of 0.1 to 0.2 $^o/_{oo}$.

During the four years studied a total of 20 snowpack samples were collected at maximum snowpack accumulation. Each sample was a composite of snow from all layers of the snowpack depth profile and was collected in a zip-lock bag and allowed to melt naturally before being transferred to a scintillation vial. Samples were collected between 1400 to 2900 m.a.s.l (virtually the entire basin range) and the average sample elevation was approximately 2290 m.a.s.l. (the modal basin height or elevation of maximum ground cover). The samples were generally collected from different locations throughout the basin but three snowpits were dug in the area of Peyto Glacier in mid May 1999 at 2650, 2150 and 1900 m.a.s.l. to investigate localised elevation effects.

From 1996 to 1999, 23 glacial ice samples were collected during the summer when ice was exposed. All of the samples were collected from below the ice weathering crust on glacier surfaces from various locations on the Wapta Icefield and hanging glaciers on the east side of the Bow Valley. Samples were collected between 2100 to 2700 m.a.s.l (the entire range of ice exposure) and the average sample elevation was approximately 2310 m.a.s.l. (a little above the modal basin height but slightly below the median ice exposure elevation). It is reasonable to assume that the average elevation of ice melt contribution is not far above the average sample height, given the greater potential for ice melt at lower elevations. Localised elevational patterns in oxygen isotope signatures over glacial ice surfaces were investigated by taking samples over the surface of Peyto Glacier above and below the long-term equilibrium line altitude (ELA) (~ 2650 m.a.s.l.).

There was some difficulty sampling rainfall within a basin as large as the Bow Valley over a four year time period and so a composite sampling strategy was adopted where rainwater from several events and from various elevations was mixed together. From 1996 to 1999, 16 composite rain samples were collected during the summer months of May to September. Samples were taken from a range of elevations during single storm events or were taken from one location but several events over a few days. The range of sample elevations was 1400 to 2250 m.a.s.l. with an average height of 1900 m.a.s.l.

Nine samples were collected of interannual baseflow at Banff (defined as any discharge below 20 m^3s^{-1}) directly from the river each winter. To study spatial and temporal variations in baseflow, basin wide samples were taken from the river during baseflow conditions (at least three days following any major events in the basin) several times throughout the study period. To this end, late winter and late summer longitudinal profiles of the Bow River and its main tributaries were sampled. To further investigate the seasonal variability in baseflow conditions, groundwater from three headwater wells located in the valley bottom upstream of Lake Louise (1800 to 2000 m.a.s.l.) of approximately 10 to 20 m in depth and a hard rock spring near Peyto Glacier were sampled every two to five weeks during the summer months.

RESULTS AND DISCUSSION

For comparative purposes, the δ^{18}O results of the three hydrological balance components (snow, ice and rain) have been plotted with elevation in Figure 2. It is apparent that although there was some overlap in the δ^{18}O signatures for snow and ice, there was a noticeable difference between each of the end-members. The following sub-sections discuss the spatio-temporal variabilities of each of these end-members. The results and discussion of the baseflow end-member study will be presented at the end of this section.

Snowpack d^{18}O Signature

δ^{18}O values ranged between –19 $^o/_{oo}$ to –24.1 $^o/_{oo}$ with an average of –22.2 $^o/_{oo}$ and a standard deviation of 1.4 $^o/_{oo}$ (Figure 2). When all were plotted together there was no evidence of an elevational dependence. However, the two upper snowpits on Peyto Glacier did indicate a slight depletion of average snowpack ^{18}O with height (less than 0.1 $^o/_{oo}$ per 100 m rise in elevation). This isotopic depletion of ^{18}O is less than the normal range between 0.15 and 0.35 $^o/_{oo}$ per 100 m for precipitation, quoted in the literature (e.g. Schotterer, *et al.* 1996). Also, the lower pit, by the side of a lake in a forested area was isotopically lighter than both the higher pits. Although the data suggest that there was no elevational control on the isotopic composition of snowpack at the Basin scale, this did not preclude the possibility of localised elevation effects.

Although no elevational patterns in snowpack isotopic composition were established, it was found that the signature varies from year to year. For the two high snowpack years of 1996 and 1999 average δ^{18}O compositions of –22.3 $^o/_{oo}$ (n = 6) and –22.6 $^o/_{oo}$ (n = 6) were respectively returned. For the low snowpack year of 1998 an isotopic composition of –21.6 $^o/_{oo}$ (n = 7) was found. An average could not be calculated for 1997, however, similar snowpack conditions to 1996 were experienced and it would seem fair to assume that the isotopic signatures would also be similar. The data suggest that during years of greater snow pack accumulation the isotopic composition may be more depleted than during years of little snowpack accumulation.

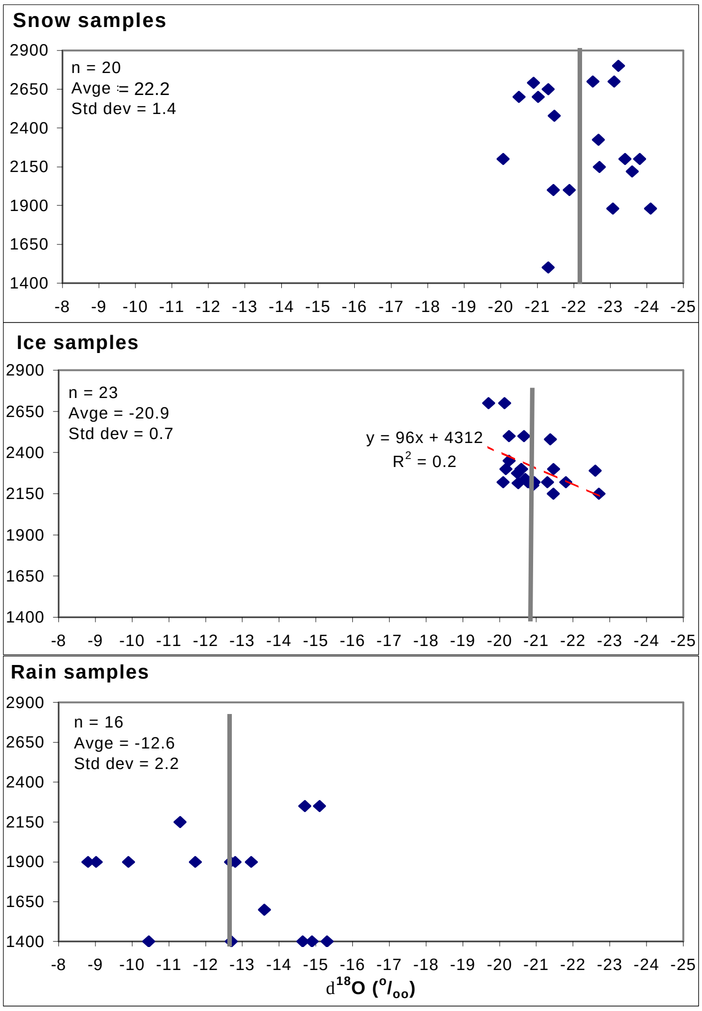

Figure 2. Hydrological balance sample δ^{18}O values with elevation (m.a.s.l.). Vertical line = mean oxygen isotopic concentration.

It would be convenient to assume that the snowpack end-member signature would be constant throughout each year, however, there is one physical process that may somewhat invalidate this assumption. At the onset of melt, the lighter ^{16}O isotope snow grains are prone to melt first (Maule and Stein, 1990). Thus as the deep snowpacks of the Rockies lose water, they should become isotopically enriched with ^{18}O. There are no data available from the present study but from other studies in deep snowpack environments, enrichment up to 1 $^{o}/_{oo}$ between initial and final snowpack melt conditions (e.g. Raben and Theakstone, 1998) has been observed. It should be noted that this enrichment does not alter the overall snowmelt signature, only its temporal distribution. For example, the initial melt signature should be more depleted than average snowpack and the final melt signature enriched. The magnitude of this change depends on the speed of snowmelt and the time step of interest. For a seasonal study of large basin hydrology, a monthly time step is probably sufficient. In valley locations much of the snowpack melts out in less time than this following melt onset in late April to early May and therefore a constant signature causes no problem here. High on the icefields, however, much snow remains year round and any melt from these areas will occur late in the summer (June to August) and likely be more depleted than the average signature. In contrast to this however, snowmelt from lower alpine areas (i.e. just above treeline) at this time will probably have already lost their early isotopically depleted snowmelt and be more enriched. Considering: 1) the complexities surrounding the temporal distribution of basin wide snowmelt signature at a monthly time step, 2) the potentially low overall temporal variation in range of values and 3) the balancing factors introduced due to the changing isotopic composition of snow as the transient snowline (TSL) progresses upward, it was not feasible to characterise the potential seasonal variation of snowmelt at the basin scale and a constant annual $\delta^{18}O$ signature was considered a reasonable approximation.

Glacial Ice d^{18}O Signature

$\delta^{18}O$ values ranged between –19.7 $^o/_{oo}$ to –22.6 $^o/_{oo}$ with an average of –20.9 $^o/_{oo}$ and a standard deviation of 0.7 $^o/_{oo}$ (Figure 2). The average value was further corroborated by data collected by West (1972) in which a study of 86 ice samples taken from Peyto Glacier tongue revealed an average isotopic signature of –20.7 $^o/_{oo}$. The ice isotope signature was, therefore, slightly enriched compared to snow and this was as to be expected for the following reasons:

1. During firnification of the snowpack in the accumulation zone, melting and/or sublimation/ evaporation lead to a preferential loss of lighter isotopes (e.g. Sommerfield *et al.* 1991).
2. Any summer time rain events at high elevation mix rain with the snowpack and again lead to a more enriched overall signature.

The lower overall range and standard deviation in ice $\delta^{18}O$ composition was probably partly due to homogenisation of the snowpack during firnification (e.g. Raben and Theakstone, 1998). In addition, it is possible that elevation and landcover influences may be in part responsible for the greater variability in snowpack signatures. Basin wide snowpack covers two times the elevational range than glacial ice and encounters many more different types of landcover. Therefore, there are likely to be more physical processes controlling the snowpack isotope signature than for ice. For example, heterogeneous tree cover at low elevations could lead to widely variable changes in the snowpack isotope signature through processes of interception and sublimation.

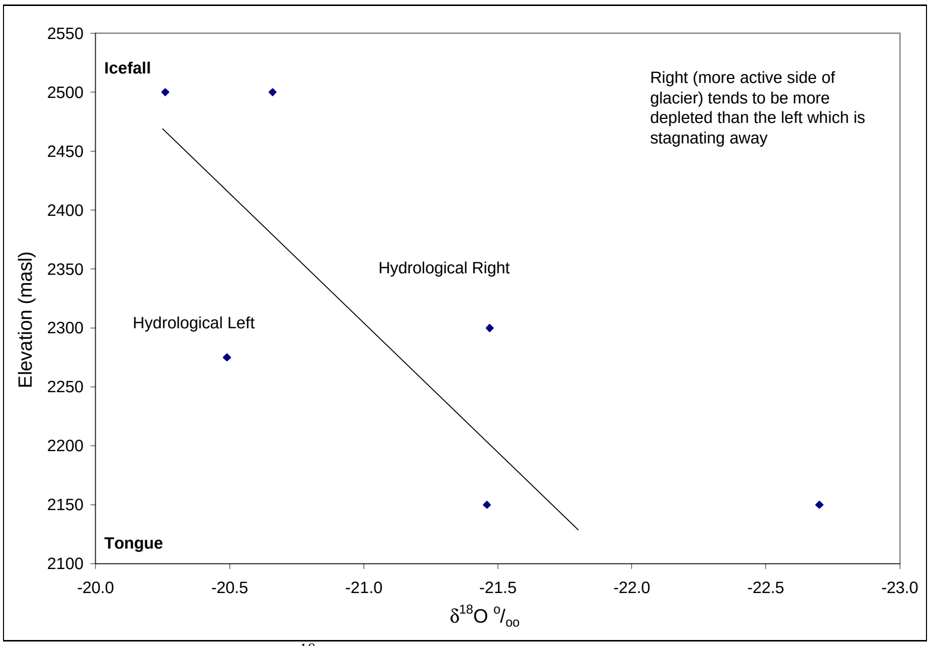

Figure 3. Elevational $\delta^{18}O$ isotope pattern (per mil), Peyto Glacier, summer, 1999

When all ice samples were plotted together (Figure 2) there was little evidence of a change in isotope signature with elevation. However, six samples taken from Peyto Glacier in the summer of 1999 did illustrate a localised pattern (see Figure 3). It can be seen that from the snout at 2150 m.a.s.l. up to 2450 m.a.s.l., there was an enrichment of the isotope signature with elevation. The same pattern and approximately the same range of values were reported at these elevations on

Peyto Glacier by West (1972). The reason given in West (1972) was a reverse elevation effect resulting from ice emerging at the snout originating high in the accumulation area with ice just below the ELA originating just above the ELA. Reid (1896) proposed this pattern of ice movement through a glacier and a reverse isotopic gradient in ice below the ELA is considered evidence for this. (The elevational isotope gradient in ice is probably due to high elevation snowpack undergoing little melt thus preserving much of its isotopically depleted signature whereas snowpack just above the ELA undergoes enrichment from increased melt and mixture with rainfall.) In both West's (1972) and the current study the range of isotopic compositions from Peyto's snout to ELA was approximately –22 $^{o}/_{oo}$ to –20 $^{o}/_{oo.}$ Slightly above the long-term ELA on Peyto, two samples were obtained near the basin divide at 2750 m.a.s.l. in late summer 1998. These samples were both around –20 $^{o}/_{oo,}$ similar to the samples collected just below the ELA, and therefore fit the dynamic ice movement model proposed by Reid (1896).

It should be borne in mind that patterns observed at Peyto may be localised and not representative of other glacial covers. For example, from Figure 3 it can be seen that although ^{18}O enriched with elevation on both sides of the glacier, the right side was consistently more depleted. The left side of the Glacier is currently undergoing stagnation and this is likely the reason for the disparity in isotope composition. Supporting evidence that the ^{18}O difference between both sides of the glacier is linked to recent stagnation is that in West's study (1972), undertaken when both sides were active, the same pattern was not apparent. Within the entire Bow Basin, ablating glaciers are likely to be a mixture of active and stagnating ice and therefore, applying a characteristic signature to any one location at any point in time is difficult.

It was possible, however, to generate a simple δ^{18}O model to the whole basin that accounted for the observations in the data, the rising TSL and increasing ice exposure throughout the season. The generalised ice δ^{18}O signature started at –21.5 $^{o}/_{oo}$ in May then gradually enriched to –20.5 $^{o}/_{oo}$ in September. During August, the time of maximum icemelt, the modeled signature would be –20.8 $^{o}/_{oo,}$ approximately the average basin ice value.

Rainfall d^{18}**O Signature**

The average rain signature was found to be -12.6 $^{o}/_{oo}$ with a standard deviation of 2.2$^{o}/_{oo}$. Rainfall is generally enriched compared to snow or ice for the following reasons:

1. Rainfall occurs mainly during the summer when temperatures are warmer (Dansgaard, 1964);
2. The dominant weather patterns between winter and summer are markedly different. In winter, much of the snowfall is precipitated from westerly airmasses whereas during summer, much of the rainfall comes from Idaho highs, which originate further to the south and have a more enriched signature (unpublished data by Yonge in Grasby, 1997).

It should be noted that the average height of rainfall samples was lower than snow or ice due to the logistical difficulties of sample collection. However, rainfall occurs at all elevations and the low sample height could have biased the isotope signatures slightly if there was an elevational gradient. It was appropriate, therefore, to estimate the effect elevation might have on the rainfall δ^{18}O signature and quantify the level of potential bias.

δ^{18}O isotope gradients in precipitation have been reported to deplete by 0.15 and 0.35 $^{o}/_{oo}$ per 100 m rise in elevation (Yurtsever & Araguas, 1993). The range of elevation in the Bow Basin is approximately 1600 m and thus adopting the upper limit of 0.35 $^{o}/_{oo}$ per 100 m it was found that the range in δ^{18}O signature due to elevation alone could be up to 5.6 $^{o}/_{oo.}$ Estimating the δ^{18}O composition for the modal basin height of 2200 m.a.s.l. (to compare with snow and ice) the rainfall signature became –13.7 $^{o}/_{oo}$ instead of –12.6 $^{o}/_{oo}$. This new value was still significantly different to that of snow and ice. This test assumed that an elevational depletion of isotope signature was likely in the Bow Valley. However, the samples collected displayed no evidence of an elevational dependence within the 850 m range sampled.

The observations suggest that the relatively small range of elevations experienced in the Rockies was insufficient to lead to a noticeable elevation effect apart, perhaps, from that associated with a change of phase from snow to rain (e.g Shanley *et al*, 1995). Air mass turbulence and precipitation redistribution probably masked altitude effects and as such no particular elevation was more or less representative of the basin as a whole. For this reason, a δ^{18}O value of –13 $^{o}/_{oo}$ was considered a reasonable approximation of the summertime rainfall signature.

Baseflow d^{18}O Signature

River Baseflow

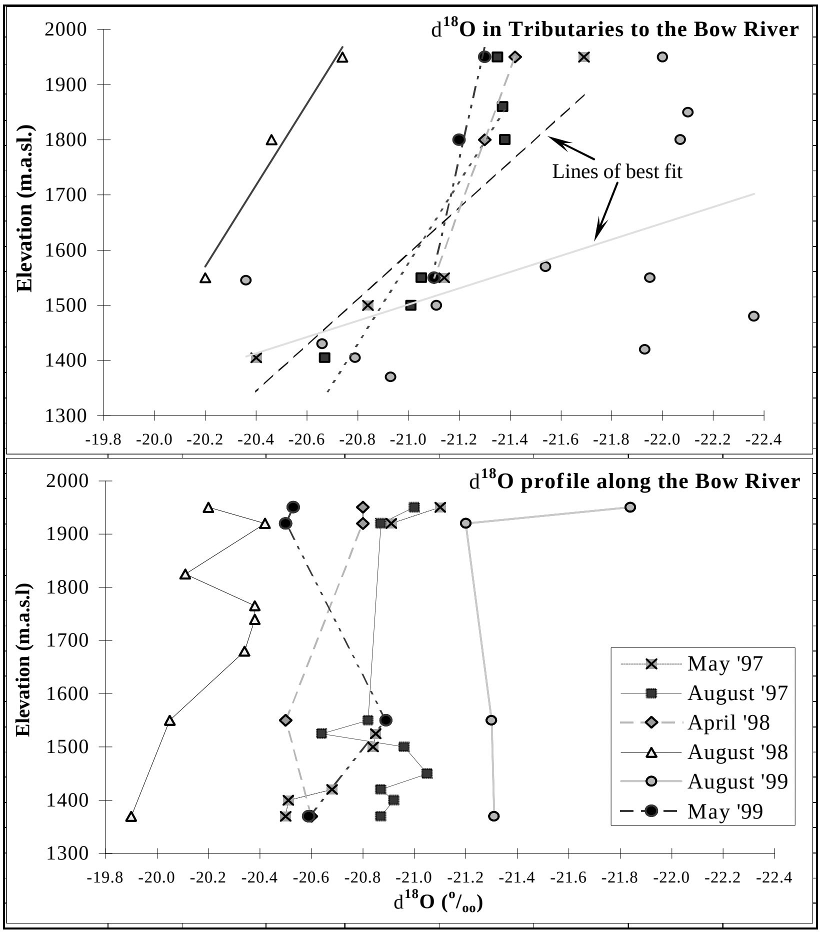

Figure 4. Oxygen isotope profiles with elevation along Bow River and tributaries from Banff up to the headwaters. Samples taken during baseflow conditions.

Baseflow samples collected at Banff each winter from 1996 to 1999 indicated that this signature was stable and had a consistent $\delta^{18}O$ signature of –20.5 $^{o}/_{oo}$ (n = 9, s = 0.05 $^{o}/_{oo}$). Along the Bow River profile there was no significant depletion of the $\delta^{18}O$ signature with elevation (see Figure 4). However, samples collected from the major tributaries indicated an average depletion of approximately 0.15 $^{o}/_{oo}$ per 100 m. There are various possible reasons for the discrepancy:

1. Lakes are an important landcover in the upper half of the basin and could homogenise the elevational isotope pattern in the river through mixing and evaporation processes.
2. The main river could exchange water with regional groundwater aquifers and thus the influence of tributary runoff could be minimised.
3. The pattern observed in tributary runoff may not be representative of the true inputs to the river. The values shown illustrate nothing of the volumetric importance of individual samples.
4. Efforts were made to sample most major tributaries but there were difficulties obtaining samples on the western slopes of the Bow Valley headwaters. These tributaries have a higher proportion of glacier cover than more accessible tributaries. If glacial melt inputs were more enriched than others during baseflow conditions in the main valley, this could explain the reduced elevational $\delta^{18}O$ gradient in the Bow River.

Depletion of the $\delta^{18}O$ signature with elevation was expected. However, the standard reasons given are the elevational gradient in precipitation and the reduction in average air temperature with height. Such reasons are often given for isotopic depletion of groundwater at higher elevations (e.g. Yonge *et al.* 1989). This may often be the case but in this study, there was little evidence for an isotopic gradient in snow or rainfall and it is thought that within the range of elevations investigated, precipitation was susceptible to atmospheric mixing and redistribution, thus homogenising the $\delta^{18}O$ signature. If this was the case, then the reason for depleted baseflow with elevation is more likely the higher proportion of snowfall over rainfall at higher elevations. From meteorological records (provided by the Meteorological Service of Canada), it was found that from 1961 to 1991 the proportion of precipitation falling as snow increased from Banff (1350 m) up to Lake Louise (1650 m) from 46% to 55%, respectively. Assuming that both rainfall and snow melt contribute to groundwater recharge, more depleted baseflow would be expected higher up.

Another observation from Figure 4 is that in both charts the data are closely grouped except for the summers of 1998 and 1999. To explain this, it was necessary to consider the hydrometeorological conditions in the Bow Basin during these time periods. Deep snowpacks and above average basin yields (Water Survey of Canada discharge data for Bow River at Banff) were encountered during 1996, 1997 and 1999. In 1998, however, there was an extremely shallow winter snowpack, which melted out rapidly and basin yield was well below average. The enrichment of 1998 summer baseflow could be explained by the relatively enriched snowpack of 1998 and a reduced snowmelt to rainfall ratio. The isotopically depleted baseflow signature in 1999 could be partially linked to the light snowpack for this year. However, a complimentary explanation could be that the reduced snowpack of 1998, led to lower regional groundwater levels prior to the 1999 spring freshet and more snowmelt infiltrated to groundwater than normal. The high proportion of snowmelt in groundwater at the beginning of the 1999 melt season may have led to depleted baseflow signatures later in the summer.

Groundwater Wells

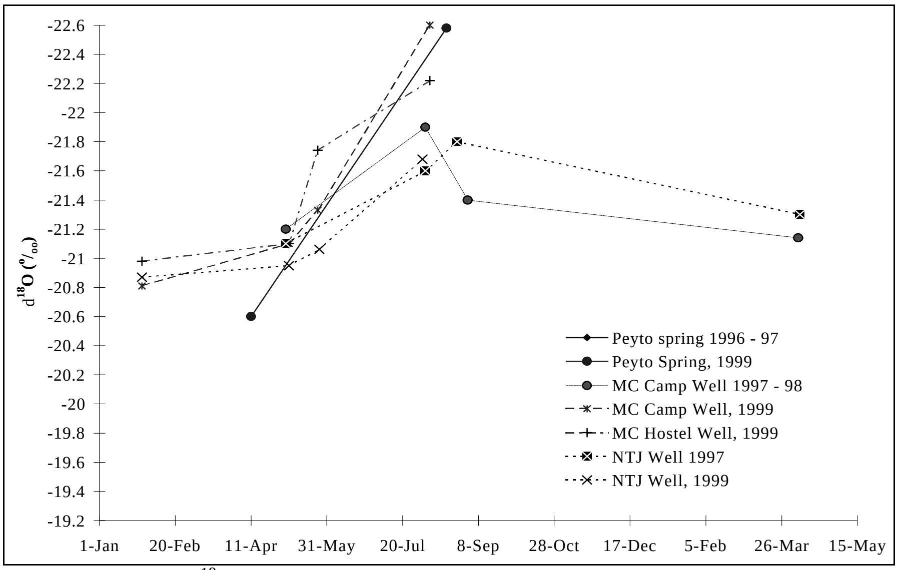

Figure 5. Seasonal $\delta^{18}O$ signatures in groundwater from various locations in the upper Bow Basin during years of deep snowpack (1996, 1997 and 1999).

Figure 5 clearly demonstrates that during years of deep snowpack, both deep groundwater (10 to 20 m) and year round spring water had a seasonality which led to the most depleted signatures in mid to late summer. This was thought to reflect the infiltration of snowmelt a few months earlier. From these data, it would appear that summer time baseflow maintained by groundwater during deep snowpack years was isotopically most depleted in ^{18}O during the month of August. In addition, the marked depletion noticed in summer time baseflow for 1999 was also noticed in the well and spring data presented. This observation provides supporting evidence that patterns observed in groundwater wells and springs can be used to help characterise river baseflow.

Further evidence that the groundwater wells displayed a snowmelt signature, which was lagged by a few months, is that the groundwater $\delta^{18}O$ signatures for 1999 were the most depleted of all years sampled (as was snowpack) with August displaying signatures that were very close to average annual snowpack (Figure 5). In addition, 1998, the year of most enriched snowpack, displayed the most enriched baseflow (Figure 4) and groundwater (Figure 6).

The observation that groundwater and baseflow isotope signatures in late summer may be close to average annual snowpack has important implications for the point raised earlier regarding the lack of an elevational gradient in the Bow River isotope profile. The higher elevation tributary samples were from sub-basins with little glacial cover, however, there were several other sub-basins containing significant glacial cover that could not be sampled. During the time of most depleted groundwater baseflow in August, the glacial melt signature was estimated to be relatively enriched (around –20.8 $^{o}/_{oo}$). This was around the mean value experienced in the Bow River at its headwaters during late summer despite depleted groundwater (Figure 5) and tributary (Figure 4) samples. Volumetrically, glacial melt was probably the main flow contributor in the headwaters at this time and thus it was likely that the glacial melt signature masked the more depleted groundwater baseflow signature present in other less glacierised tributaries.

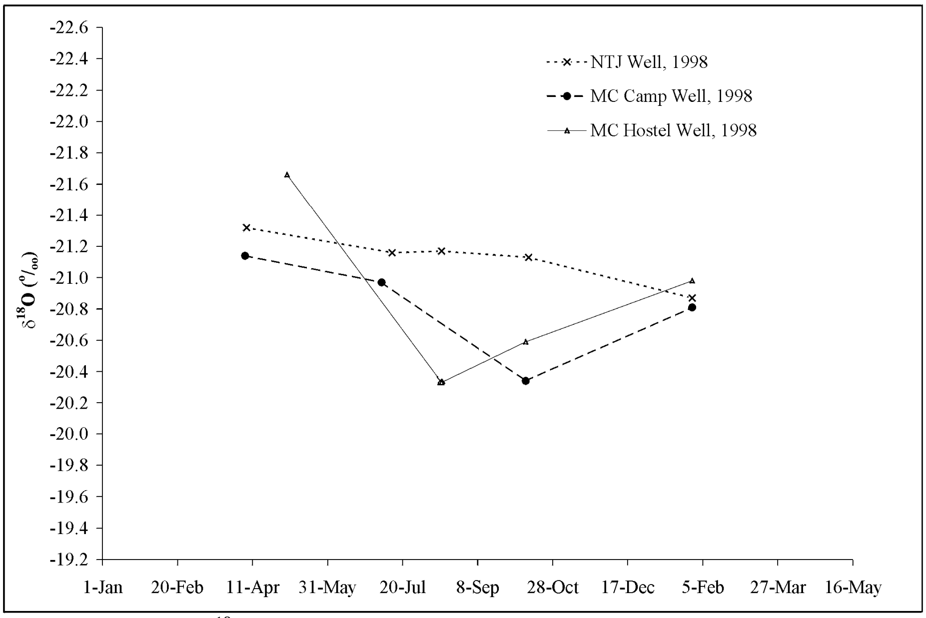

Figure 6. Seasonal δ^{18}O signatures in groundwater wells from various locations in the upper Bow Basin during the shallow snowpack year, 1998

In Figure 6, data for the same groundwater wells are displayed as in Figure 5 (no hard rock spring data were collected in 1998). It can be seen that the distinct seasonality observed during years of deep snowpack was not evident during 1998. If anything, the seasonality was slightly reversed at two of the wells, with enriched ^{18}O levels during early summer depleting as winter approached. This might have been due to lower water tables at the end of the year and groundwater subsequently being replenished rapidly from higher up in the basin. However, one thing common to all groundwater samples in 1998 was that each location was more enriched at the end of the year than the beginning. This may have been due to a greater proportion of rainfall infiltrating to ground during 1998 or it could suggest that a higher proportion of older and more enriched water was in the ground than normal

From the data presented, it was possible to generalise the seasonal behaviour of basin wide baseflow δ^{18}O signature. At Banff a value of –20.5 $^{o}/_{oo}$ was observed during each winter prior to spring melt from 1996 to 1999. For the deep snowpack year of 1997 the groundwater signature from well data became isotopically depleted by about 0.6 $^{o}/_{oo}$ from winter to summer, with most depletion occurring in August (no groundwater well data were available for 1996 so patterns were assumed to be similar to 1997 due to the similarity in hydrometeorological conditions). For 1999, the level of summertime groundwater well depletion was around 1.2 $^{o}/_{oo}$. For both 1997 and 1999, however, the in river baseflow samples at Banff displayed a subdued isotopic depletion of 0.4 $^{o}/_{oo}$ and 0.8 $^{o}/_{oo,}$ respectively. During 1998, groundwater was slightly enriched later in the year compared to the beginning. However, this enrichment of groundwater well data for 1998 was not evident in Bow River baseflow at Banff. The apparent discrepancies between river baseflow and groundwater well observations may be due to a higher proportion of baseflow at Banff being maintained by regional groundwater responding more slowly to annual conditions than the seasonal signature observed in headwater groundwater. The objective here was to characterise basin baseflow conditions and therefore the river observations were considered a more reliable indicator of absolute variability in isotopic composition than headwater groundwater wells.

Following is a generalisation of the basin baseflow δ^{18}O signature during 1996 to 1999:

1. The oxygen isotope signature for each year was most enriched during winter when baseflow was relatively stable i.e. December to April.
2. The most isotopically depleted month of each year was August as the snowmelt “wave” transcended groundwater aquifers and was released to baseflow.
3. The change in basin baseflow δ^{18}O signature from relatively enriched winter to relatively depleted summer baseflow was approximately linear.
4. For each year, the winter months from December to April had a constant δ^{18}O signature of –20.5 $^{o}/_{oo}$. Baseflow for the months of August 1996, 1997, 1998 and 1999 had, respectively, the following δ^{18}O signatures: - 21.0, -21.0, -20.6 and –21.4.

CONCLUDING REMARKS

A summary of the average and temporally variable δ^{18}O signatures during 1996 to 1999 for the hydrological balance and storage components studied is provided in Table 1. The dynamic, or temporally variable, isotope signatures presented are best estimates based upon the data collected. It is highly likely that any component sampled at any location will not display these signatures. However, for the basin as a whole, the signatures suggested should provide an overall approximation of the patterns for each of the years in question and have been used as the end-member δ^{18}O signatures for a hydrological model evaluation in an accompanying paper (Hopkinson *et al*, this issue).

Table 1. Generalisation of hydrological end-member d^{18}O signatures within the Bow Valley above Banff, based upon observations from 1996 – 1999. (Values in $^{o}/_{oo}$)

Static isotope signatures		**Dynamic isotope signatures**
Snow	Avg = -22.2 s = 1.4 n = 20	Hydrological years: 1996 = -22.3 (s = 1.5, n = 6) 1997 = -22.3 (estimated from 1996) 1998 = -21.6 (s = 1.6, n = 7) 1999 = -22.6 (s = 1.0, n = 6)
Ice	Avg = -20.9 s = 0.7 n = 23	All years: May = -21.5 June = -21.3 July = -21.0 August = -20.8 September = -20.5
Rain	Avg = -12.6 s = 2.2 n = 16	Constant signature: -13 (note change in decimal places)
Baseflow	Avg = -20.5 s = 0.05 n = 9 (winter samples only)	December to April of each year = -20.5 Baseflow depletes to a minimum during August each year: 1996 = -21.0 1997 = -21.0 1998 = -20.6 1999 = -21.4

Snowpack tended to have a reasonably consistent isotopic composition but the variation from year to year likely reflected the changes in atmospheric patterns during the winter months. 1998, for example, was an *El Niño* year with very little snowfall as the jet stream was pushed to the north bringing in drier and warmer weather from the south. These air masses would have been isotopically more enriched than normal. In contrast 1999 was a *La Nina* year with opposing conditions to the prior year, which probably brought in more precipitation from northerly regions with a relatively depleted δ^{18}O signature.

Random sampling throughout the basin at a variety of elevations suggested that glacial ice did not have an elevational δ^{18}O gradient. However, from localised studies on Peyto Glacier and from the literature (West, 1972) it was found, that localised changes in δ^{18}O with height did occur. The dynamic δ^{18}O signature associated with icemelt began relatively depleted in the spring becoming gradually more enriched as the TSL approached the ELA. This pattern reflected the reverse isotope elevation gradient in the $\delta^{18}O$ signature of glacial ice. Ice emerging at the snout with a depleted signature was theoretically formed high in the accumulation zone and ice below the ELA was originally formed just above it.

Atmospheric mixing appeared to reduce the effect of altitude on rainfall $\delta^{18}O$ signature in this basin. Rain samples collected in the spring were just as enriched as those collected later in the year. Suggesting that the seasonal variation in precipitation may be largely due to the difference in $\delta^{18}O$ signatures between rain and snow and or the changing dominance of weather systems from westerlies in winter to more southerly weather in summer. The normal explanation for the difference between summer and winter precipitation (and altitude variations) is changing temperature. However, the data collected in this study did not support this hypothesis.

Winter baseflow samples, river and tributary profiles and groundwater well samples were all collected and analysed to characterise the inter-annual seasonal baseflow $\delta^{18}O$ signature. In general, annual river baseflow had a constant signature during winter months gradually depleting during summer in response to spring snowmelt infiltration to groundwater. It was found that 1998 was different from other years, in that the low snowpack did not replenish the groundwater store as much as would normally be expected and therefore the summer time baseflow signature was only slightly enriched compared to winter. During 1999, however, the relatively enriched snowpack and low antecedent groundwater levels probably led to a high proportion of snowpack recharging groundwater aquifers and thus a depleted summer time baseflow signature.

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