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Bedrock infiltration and mountain block recharge accounting using chloride mass balance

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Abstract:

Mountain front catchment net groundwater recharge (NR) represents the upper end of mountain block recharge (MBR) groundwater flow paths. Using environmental chloride in precipitation, streamflow and groundwater, we apply chloride mass balance (CMB) to estimate NR at multiple catchment scales within the 27 km^2 Dry Creek Experimental Watershed (DCEW) on the Boise Front, southwestern Idaho. The estimate for average annual precipitation partitioning to NR is approximately 14% for DCEW. In contrast, as much as 44% of annual precipitation routes to NR in ephemeral headwater catchments. NR in headwater catchments is likely routed to downgradient springs, baseflow, and MBR, while downgradient streamflow losses contribute further to MBR. A key assumption in the CMB approach is that the change in stored chloride during the study period is zero. We found that this assumption is violated in some individual years, but that a 5-year integration period is sufficient. Copyright © 2011 John Wiley & Sons, Ltd.

KEY WORDS water balance; chloride mass balance; mountain block recharge; groundwater recharge; bedrock infiltration; catchment hydrology

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INTRODUCTION

Mountain front recharge (MFR) is an important source of water to valley aquifers in arid and semi-arid regions (Wilson *et al*., 2004; Figure 1). The subsurface component of MFR, called mountain block recharge (MBR), hydraulically connects upland catchments through bedrock flow paths to valley aquifers. Clearly, upland catchments must lose water to underlying bedrock for MBR to occur. Yet, the role of bedrock in catchment water balances has until recently received little attention (Tromp-van Meerveld *et al*., 2006; Katsuyama et al, 2005). Historically, catchment hydrologists conveniently assumed underlying bedrock to be impermeable, conceptually eliminating the source of MBR, which conflicts with knowledge that upland catchments contribute water to valley aquifers (Hutchings *et al*., 2001; Russell and Minor, 2002; Manning and Solomon, 2003). This paradox calls for a greater understanding of the hydraulic connections between mountains and adjacent aquifers. In particular, identifying upland sources of MBR and the spatially variable rates of infiltration into bedrock systems from small to large catchments is essential. This is particularly true for improved water resource management where land use change and/or climate change impacts mountain hydrology.

Quantifying MBR at the downgradient end of mountain block flow paths, i.e. the valley aquifer, is common. In the context of a mountain system and/or aquifer water balance, MBR has been estimated using Darcy flow equations applied to the bedrock/basin-fill contact (Clark *et al*., 1985), derived as a residual in the mountain block water balance given estimates of precipitation, evapotranspiration, and streamflow (Cederberg *et al*., 2009) and modeled (Flint *et al*., 2004; Cederberg *et al*., 2009; Magruder *et al*., 2009). Outflow methods equate groundwater discharge to recharge relative to source area or use source area recharge efficiencies to estimate MBR (Maxey and Eakin, 1949). Similarly, aquifer response methods estimate MBR through inverse modeling of aquifer measurements or environmental tracer enrichment. Recently, environmental tracers, including noble gases, tritium, C^{14}, and major ions, have been used to assess proportions of valley groundwater that originated as higher elevation recharge (Hutchings *et al*., 2001; Manning and Solomon, 2003; Petrich, 2004).

In contrast, quantifying MBR at the upgradient beginning of mountain front flow paths is not common. Rather, bedrock hydrology is often treated as a complication to upland catchment water balance and runoff generation studies. The paucity of information about infiltration into bedrock is perhaps because of the difficulty in applying hydrometric methods to measure and estimate bedrock infiltration, particularly where bedrock permeability is dominated by fractures beneath the soil mantle. Water that infiltrates bedrock may be routed to adjacent streams or move into deep bedrock groundwater systems as a 'loss' relative to the catchment water balance to potentially become MBR (Anderson *et al*., 1997; Nyberg *et al*., 1999; Katsuyama *et al*., 2005; Tromp-van Meerveld *et al*., 2006). This interaction of catchment water with bedrock has received a variety of labels including deep seepage, deep percolation, and bedrock

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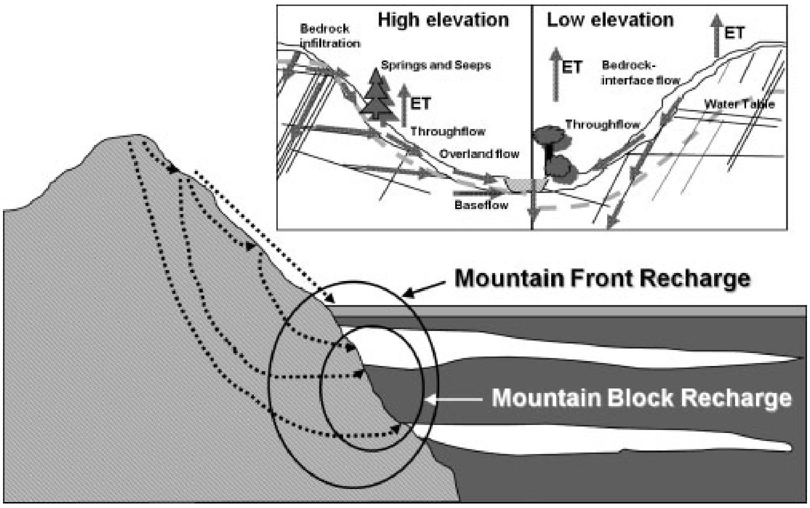

Figure 1. Conceptual diagram for MFR and MBR. Insert diagram depicts upland subcatchment routing of precipitation that may include discharge of groundwater to streamflow and subsequent streamflow loss to groundwater recharge via channel seepage

infiltration. In this article, we use the term net groundwater recharge (NR) as water that infiltrates bedrock and does not re-emerge within the catchment boundary (Figure 1). A tightly constrained catchment water balance in which all stores and fluxes are adequately measured may be used to estimate NR. For example, using detailed uncertainty and error analysis, Graham *et al.* (2010) attribute as much as 44% of a hillslope irrigation water balance in a forested catchment to deep seepage. Recharge estimates based on catchment water balance residuals inherit errors from estimation of other water balance components, particularly evapotranspiration and storage. Given the increased potential for error inherent in upscaling these components and subsequent need for error analysis, it is not feasible to estimate loss as a water balance residual in moderately sized catchments or mountain fronts.

Chloride mass balance (CMB) provides an attractive alternative to water balance methods, because evapotranspiration does not transport chloride and is, therefore, removed from the balance equation. Consequently, the mass of chloride input to a catchment is accounted for by the mass that leaves as streamflow and the mass that enters the groundwater system (Eriksson, 1960; Phillips, 1994). Environmental chloride has been used to estimate point scale and mountain block groundwater recharge at numerous locations worldwide since its suggestion by Eriksson in 1960. Eriksson and Khunakasem (1969) utilized CMB to estimate groundwater recharge in Israel according to the premise that evaporation is responsible for increased chloride concentration in the system. Dettinger (1989) applied reconnaissance level CMB to 16 intermountain basins in Nevada, including the Las Vegas basin, with results comparable to existing Maxey–Eakin and water balance estimates, averaging 5, 6 and 7% of annual precipitation, respectively. An elevation-dependent CMB application by Russell and Minor (2002) to basins in southern Nevada produced estimates of MBR between 3 and 5% of annual precipitation, that were, on average, greater than Maxey–Eakin estimates. Zhu *et al.* (2003) reported 3% of present day annual precipitation routed as MBR to the 14 000 km^2 regional aquifer in Black Mesa Basin, NE Arizona, and $4\% \pm 2\%$ over the past 5000 years. In a less dry environment on the western flank of the South Bridger Range, Montana, CMB estimation of aquifer recharge via MFR (Figure 1) is estimated as 20%, in contrast to 34% determined from physical water balance (Hay, 1997). Using measurement of local flux to groundwater in concert with CMB estimation of stream discharge over a 3-year application, Claassen *et al.* (1986) determined recharge rates as 9–15% of annual precipitation, with the 3-year average at 13% of 360 mm, for a 28-km^2 mountain catchment in the San Juan Mountains, southern Colorado.

The CMB method requires many assumptions (Wood, 1999). First, net storage of chloride in the unsaturated zone must be zero during the period of integration. Second, the chloride mass flux over the time period must be fully accounted for. Third, chloride must behave conservatively in the system. Fourth, there must be no external surface water or groundwater input. Fifth, all runoff from the system must be measured. The latter three assumptions are relatively simple to meet. The first two, however, present challenges. The first assumption is easily violated if chloride balances are performed on less than an annual time scale. In arid and semi-arid climates, nearly all rainfall during the summer months evaporates, leaving chloride to accumulate in the vadose zone. When fall rain and spring snowmelt travels through the soil, infiltrating water presumably dissolves and transports the stranded chloride. Therefore, use of CMB to determine NR for a catchment requires a minimum of 1 year data collection to assure zero net chloride storage in the unsaturated zone. In arid and semi-arid environments with highly variable inter-annual precipitation regimes and potential for annual positive net chloride storage in the unsaturated zone, a multi-annual period of integration may be necessary, as well as advantageous in representing average annual groundwater recharge. The second assumption can be violated by weathering of geologic formations high

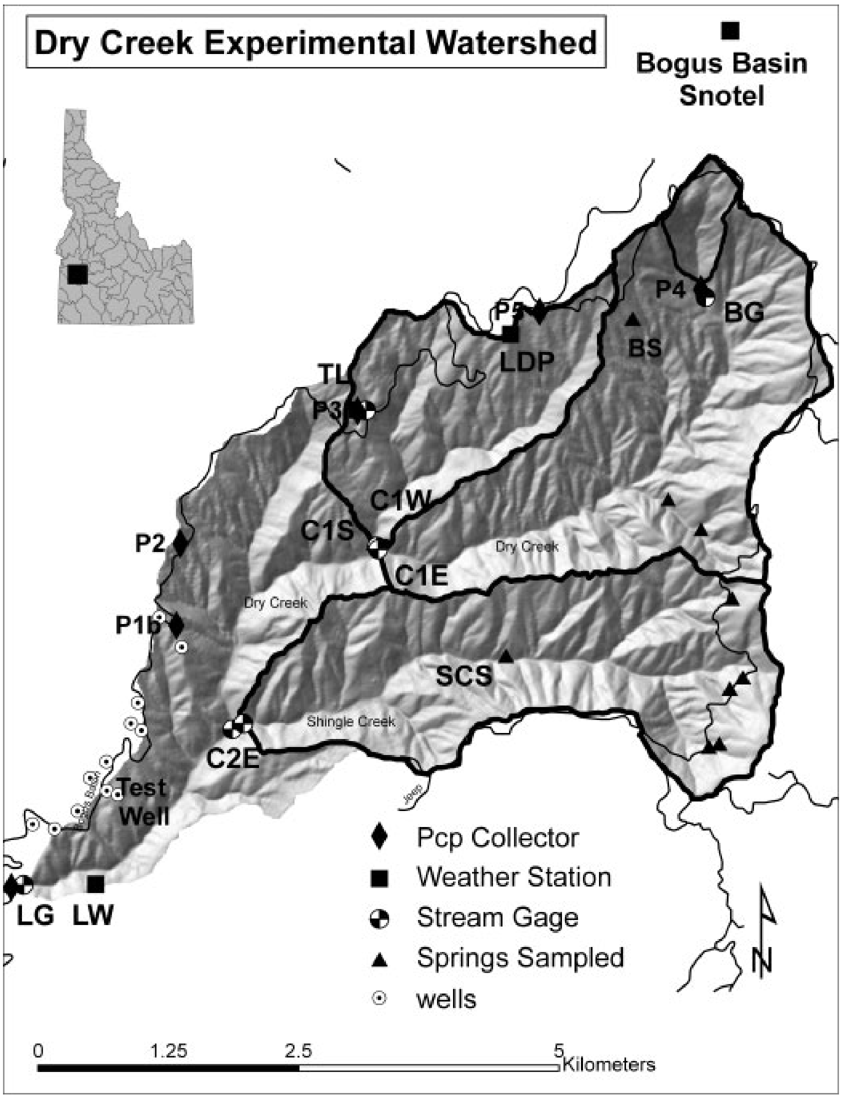

Figure 2. DCEW. Measurement site instrumentation is shown with study catchments outlined as defined by outlet measurement points LG, C2M, C2E, C1E, C1W, TL, and BG. Springs C1S and SCS are also identified

in chloride, anthropogenic activities such as road salting or by integrating over a too short a time period relative groundwater age.

The western Snake River Plain aquifer system in semi-arid southwestern Idaho abuts the granitic Boise Front Range (Figure 2). Noble gas thermometry suggests that high elevation zones are important sources of recharge to Boise valley aquifers (Thoma, 2008). In addition, Wood and Burnham (1987) suggest that a significant amount of water in the deep geothermal aquifer system underlying the Boise Valley is derived from recharge into fractured granite in the adjacent mountain block. These studies imply that upland catchment must lose water to underlying bedrock. Indeed, by calibrating a physically based hydrology model with bedrock permeability as a parameter, Kelleners *et al*. (2009; 2010) concluded that up to 35% of annual precipitation must recharge deep groundwater in small upland catchment in the Boise Front. Because of the small scale of that study, it was unknown if NR in the small catchment re-emerged within the mountain block or migrated through deep flow paths to valley aquifers. A comprehensive understanding of the hydraulic connections between mountain blocks and valley aquifers requires NR estimates at larger spatial scales. In this study, we evaluate NR at multiple catchment scales within a 27-km^2 catchment in the Boise Front Range using CMB, paying particular attention to CMB model assumptions. We seek to understand the role of bedrock infiltration in the water balance of mountain catchments and to provide a starting point for constructing a comprehensive view of MBR from upland catchments to valley aquifers.

STUDY SITE

The Boise Front Range in semi-arid southwestern Idaho provides the edge of the Snake River Plain at Boise, Idaho, and includes the 27-km^2 Dry Creek Experimental Watershed (DCEW). Elevations in DCEW range from 1000 to 2200 m. Bedrock is fractured Cretaceous granodiorite (Mitchell and Bennett, 1979). The lower portion of the Boise Front is surfaced by fluvial, lacustrine, and minor volcanic strata of the Tertiary Idaho Group. Terrain on the Boise Front is steep with thin sandy soils. Within DCEW, soil depths range from 0 to 370 cm with an average of approximately 82 cm (Tesfa *et al.*, 2009). Vegetation is primarily grass and brush groundcover in the lower elevations, contrasted by Douglas Fir and Ponderosa Pine forest in the upper elevations with lush, though narrow zones of riparian vegetation along the length of Dry Creek. Dry Creek and Shingle Creek flow perennially. Headwater springs from fractures (Gates *et al.*, 1994) provide perennial flow, while lower elevation springs provide less consistent flow later in the dry season. In some years late dry season flow ceases within the lowest reach of Dry Creek. Groundwater in the high elevation fractured bedrock aquifer is from 9 to 23 years old (Gates *et al.*, 1994).

DCEW contains six stream instrumentation sites, three weather stations, and multiple soil moisture sensing sites (Figure 2). Road maintenance crews apply sand with some salt to Bogus Basin road along the western catchment boundary during the winter months. Consequently, subcatchments not bordered by the road, Treeline (TL), South Bogus (BG), Confluence 1 East (C1E), and Confluence 2 East (C2E) are unaffected by road salt, while Confluence 1 West (C1W) and Lower Gauge (LG), the outlet point for DCEW, are potentially affected by road salt chloride. We conduct our analyses on subcatchments not affected by road salt, but attempt to account for road salt in the larger catchment LG.

METHODS

Chloride mass balance to estimate catchment NR

NR is a loss of water volume, relative to the catchment boundary, in the steady-state catchment water balance

$$P - (\mathrm{ET} + Q + \mathrm{NR}) = \Delta S \tag{1}$$

where P, ET, Q, and NR are precipitation, evapotranspiration, stream discharge, and NR volumes, respectively, and ΔS is the change in catchment water storage. Assuming that chloride transport by evapotranspiration is negligible, the steady-state catchment CMB solved for NR is

$$\mathrm{NR} = \frac{(P)(\mathrm{Cl}_P) - (Q)(\mathrm{Cl}_q)}{\mathrm{Cl}_r} \tag{2}$$

where $\mathrm{Cl_p}$ is catchment chloride concentration including bulk wet and dry deposition, $\mathrm{Cl_r}$ is catchment groundwater chloride concentration and $\mathrm{Cl_q}$ is temporal-volume-weighted average chloride concentration in stream discharge at the catchment outlet. To assess spatial variation in groundwater recharge, we apply the 5-year integration CMB to subcatchments with outlet points designated as follows: Treeline (TL), Bogus Gauge South (BG), Confluence 1 East (C1E), Confluence 2 East (C2E), and Lower Gage (LG) (Figure 2).

Data collection

Cumulative precipitation was measured at four elevations, including two DCEW weather stations equipped with paired shielded and unshielded weighing-bucket precipitation gauges, a Bureau of Reclamation unshielded weighing-bucket gauge at the base of the Boise Front, and a shielded weighing-bucket gauge paired with a snow pillow for snow water equivalent (SWE) measurement at the Bogus Basin Snotel site (Figure 2). Raw cumulative precipitation data from the DCEW sites for the period of record, 2005–2009, were corrected for noise and wind effects using the automated precipitation correction program (Nayak *et al.*, 2008) and corrected for noise 2000–2004. Monthly sums of cumulative precipitation and/or SWE were acquired for the Bureau and Snotel sites. Hypsometric methods were used to spatially distribute the four station monthly precipitation depths. From the resultant monthly regressions of precipitation *versus* elevation, at 100 m elevation intervals, annual precipitation volumes (P) were calculated for each study catchment. Stream discharge (Q) for DCEW was determined by applying empirical stage-discharge rating curves to continuous stage data at LG for the period of record 2000–2009 and, similarly, for each subcatchment 2005–2009. Data gaps were addressed using temporal interpolation of discharge at a given site and/or linear regression with adjacent station data.

We sampled for bulk wet and dry chloride deposition at five elevations in DCEW during water years 2005 and 2006 (Figure 2). Collectors were sampled and emptied promptly following precipitation events. The collectors remained in place between events to provide sampling of combined wet (precipitation delivered) and dry fall (eolian deposition) chloride concentration. Collectors were 20-cm diameter screened funnels attached to graduated collectors wrapped in reflective tape to minimize evaporation. Fresh-fallen snow was grab-sampled and melted under refrigerated conditions. Also, snowmelt collectors, for each water year 2005 and 2006, were emplaced near precipitation collector P5 (Figure 2) as winter snowfall began. These buckets were left undisturbed beneath snow pack and sampled at completion of site snowmelt. Spring and stream locations were sampled for chloride approximately bi-weekly July 2004 through June 2009 at each catchment outlet and spring shown in Figure 2, except for the Shingle Creek and north-facing C1E springs which were sampled late summer. Well water was sampled during an extensive pump test in October 2006 at a ridge top well located between C1E and LG (Figure 2). Additional sampling of surface water, including a range of tributaries to Dry Creek, was conducted on three separate dates in February, April, and May 2006 to facilitate assessment of suspected road salt-derived chloride transport into the study area. Before sampling, bottles, 30 or 60 ml high-density propylene, were thrice rinsed in the water to be sampled. Samples were filtered on site and refrigerated until delivered to laboratory for analysis. Analysis was conducted by ion chromatography in the Boise State University BioTrace Laboratory.

The 2-year precipitation chloride data set was analysed for temporal and spatial variability from which an elevation-bounded, spatially-weighted average value for Cl_p was determined for each study catchment. Cl_p values determined for each catchment are applied to all annual and multi-annual CMB integration periods. Annual and 5-year integration stream chloride concentrations, Cl_q, for 2005–2009, were determined as volume-weighted averages of monthly streamflow chloride concentrations for each collection site. Catchments TL, BG, C1E, and C2E are isolated from unmeasured anthropogenic chloride sources. The western boundary of DCEW, however, is exposed to road salt at the intersection of Bogus Basin road, particularly within catchment C1W and portions of Bogus Basin road below C1W. To account for this addition, we determined adjusted annual values of Cl_q at LG, as Cl_{qn}. We define Cl_{qn} as the chloride concentration exclusively attributable to routed wet and dry deposition chloride. As an initial step, we quantified total annual chloride mass discharged at LG as a volume-weighted mass determined from monthly average chloride concentrations at the outlet. From this total mass, chloride mass attributable to road salt, determined according to three scenarios, was subtracted to arrive at a mass relevant to derivation of Cl_{qn}. Methods applied to address road salt additions are described further in the section below on unsaturated zone stored chloride and subsequent section on Cl_{qn}.

Groundwater chloride concentrations were determined by several methods. For headwater catchments with perennial streams, BG, C1E and C2E, multi-annual groundwater chloride concentrations, Cl_r, were determined from average late dry season (August and September) outlet streamflow chloride concentrations and, alternately, from late dry season springflow. The TL subcatchment does not contain a spring, nor does it provide late dry season streamflow. Consequently, we used the average of the late dry season spring samples from C1E and C2E headwater catchments to represent groundwater beneath the TL catchment. Groundwater chloride concentration for DCEW at LG was determined as an area-weighted average of headwater catchment, C1E and C2E, groundwater chloride concentrations, rounded up to one significant digit, and lower elevation groundwater chloride concentration. Lower elevation groundwater samples were provided by the low elevation ridge test well (Figure 2). Values utilized in area–volume weighting represent high-end values of Cl_r for their respective catchment areas. The use of Cl_r for C1E and C2E in the area-weighting assumes no downgradient discharge of NR within DCEW, as evidenced by dry season cessation of springflow at C1S and SCS (Figure 2), as well as measured losses along stream reach C1E to LG.

Assessing unsaturated zone stored chloride

The assumption of zero net unsaturated zone stored chloride is assessed for DCEW by quantifying annual slope-derived chloride delivered to stream reach C1E to LG (Figure 2) relative to estimates of wet and dry fall chloride deposition on contributing reach slopes, assuming no additional chloride slope additions. This lower elevation reach receives notably less precipitation, occurring predominantly as rainfall, in contrast to the upper elevations, and is, therefore, considered most likely to incur inter-annual unsaturated zone stored chloride. Chloride mass gained along the stream may occur due to mobilization of unsaturated zone stored chloride along reach slopes or from groundwater additions. Net gain/loss along the reach is determined at monthly time steps by subtracting the monthly inputs at C1W, C1E, C2E, and C1S from the output at LG, for both discharge and chloride mass, as shown in Equations 3 and 4.

$$Q_{\mathrm{lg}} - (Q_{\mathrm{clw}} + Q_{\mathrm{cls}} + Q_{\mathrm{cle}} + Q_{\mathrm{c2e}}) = \Delta Q \quad (3)$$

$$M_{\mathrm{lg}} - (M_{\mathrm{c1w}} + M_{\mathrm{c1s}} + M_{\mathrm{c1e}} + M_{\mathrm{c2e}}) = \Delta M \quad (4)$$

where Q_{c1w}, Q_{c1s}, Q_{c1e}, Q_{c2e}, and Q_{lg} are stream discharge volumes for a given monthly time step at designated catchment outlets and ΔQ is the water volume gain/loss along the reach for the given time step. Similarly, M_{clw}, M_{c1s}, M_{c1e}, M_{c2e}, and M_{lg} represent chloride mass discharged during the same time step at designated catchment outlets, and ΔM is the chloride mass gain/loss along the reach for the given time step. Chloride mass, M_i, at each outlet, i, is calculated as the product of Q_i and Cl_i. Chloride concentrations, Cl_i, are average sample chloride concentrations, determined for each monthly time step from samples collected and processed as described above. A positive result represents a net gain, while a negative result represents a net loss.

A pseudo chloride concentration, Cl_x, for streamflow gain is calculated to estimate the chloride concentration of combined gain sources

$$Cl_x = \Delta M / \Delta Q \quad (5)$$

It is assumed that any concentration of chloride along the stream reach due to evapotranspiration will be negligible under flowing stream conditions. It is also assumed that any channel seepage losses along the reach occur at ambient stream chloride concentrations, thus, would not affect the value of Cl_x. When net gain in chloride occurs concurrent with net gain in streamflow, the source is groundwater and/or adjacent slopes. Under conditions of net gain of both chloride mass and streamflow, a value of Cl_x (Equation 5) greater than Cl_r indicates contribution from unsaturated zone stored chloride along reach slopes, occurring via surface runoff, throughflow, and/or bedrock interface flow. Values of Cl_x less than Cl_r indicate low chloride input from adjacent slopes. Low chloride input may occur as overland flow, rain on snow runoff, lateral transport of snowmelt within snowpack, macropore flow, or throughflow/bedrock interface flow in slopes that have minimal unsaturated zone stored chloride. In either case, coincident groundwater contribution to streamflow is possible. Only when chloride mass net gain occurs concurrent with streamflow net loss, in the absence of evapotranspiration, can it be stated that groundwater contribution is not a consistently occurring process along the stream reach. Using these qualifying conditions, $Cl_x > Cl_r$ at LG, using Cl_r determined as an area-weighted average for DCEW at LG as described above, and/or streamflow loss in the absence of evapotranspiration, qualifying monthly gains were selected to quantify delivery of slope unsaturated zone stored chloride to the stream reach.

Monthly chloride mass gains along the reach that meet the qualifying conditions were summed to annual values and compared to estimates of annual chloride delivered to reach slopes to assess transport of annual unsaturated zone stored chloride. Chloride delivered to reach slopes was calculated using estimates of chloride mass delivered by precipitation to slopes along the reach, combined with estimates of road salt chloride addition from the west side of this catchment. Chloride mass delivered by precipitation is determined using the mean 2-year Cl_p value for DCEW elevation below 1500 m and hypsometrically determined annual precipitation volumes for the slope area contributing runoff to reach C1E to LG, namely DCEW exclusive of C1W, C1E, and C2E (Figure 2). The annual mass of road salt-derived chloride input to reach C1E to LG is estimated as half that added to the C1W catchment. Although the catchment inclusive road lengths are similar for C1W and the western boundary of DCEW below C1W, road sand/salt application occurs primarily above the elevation marked by precipitation collector P2 (Figure 2), making this a conservative, high estimate. Road salt-derived chloride mass for the C1W catchment is estimated from annual streamflow chloride concentrations at C1W in excess of expected natural streamflow chloride concentrations. Natural chloride concentration is assumed equal to C1E chloride concentrations, with determination made at monthly time steps. As stated above, to provide stream reach chloride mass gain values defendable as unsaturated zone stored chloride, only chloride mass gained under conditions of streamflow loss in absence of evapotranspiration or associated values of $Cl_x > Cl_r$ are utilized. Accordingly, the resultant estimate of mobilized and transported unsaturated zone stored chloride is considered a minimum value.

Accounting for chloride derived from road salt

In addition to accounting for road salt to assess transport of unsaturated zone stored chloride, we must account for road salt attributing to continuous streamflow chloride concentrations at an outlet for which CMB will be performed. All study catchments are considered free of non-wet and dry fall deposited chloride, except for the large DCEW catchment defined at LG. Road salt additions occur in catchment C1W, tributary to LG below C1E and possibly along the west side of DCEW below C1W, as noted in the section above. We define chloride in streamflow that was derived from natural wet and dry deposition as Cl_{qn}. To determine Cl_{qn} for LG we apply three estimation scenarios for road salt-derived chloride. Each scenario includes measurement-based estimation for road salt-derived chloride contributed by catchment C1W, each year, 2005–2009, as described above. Additionally included is estimation of road salt-derived chloride delivered annually to the reach by C1S. Road salt-derived chloride mass for C1S is similarly estimated from annual streamflow chloride concentrations at C1S in excess of expected natural streamflow chloride concentrations. Estimation for road salt-derived chloride originating from road sources below C1W and C1S varied in assumption from no road salt-derived chloride, road salt-derived chloride contributing only year, 2006, to, thirdly, road salt-derived chloride occurring every year. The amount of road salt added in an assumed low elevation road salt application year is, as stated in the above section, estimated as half that incurred in catchment C1W. The 2006 road salt scenario is based upon road sand/salt occurrence, field observation, and inference from anomalous high surface water chloride concentrations during 2006. As snowfall rarely persists below 5000 ft elevation on the Boise Front, road sanding/salting is rare below C1W. However, snowfall for winter 2005–2006 was extensive and road sanding/salting was applied below 5000 ft. Most remarkable was the extensive surface runoff in spring 2006. Widespread low elevation tributary flow occurred which had not been observed during prior or post years within the study period. Catchment-wide sampling of surface runoff conducted during snowmelt episodes, February, April, and May 2006, revealed anomalously high chloride concentrations in tributaries draining from Bogus Basin road relative to non-road-draining tributaries.

Error and uncertainty

The combined uncertainty and error range on CMB results for catchment NR (Equation 2) is determined using ±20% projected maximum error in precipitation, ±20% error in stream discharge, and ±5% error in lab-determined chloride concentrations, combined with uncertainty in specific Cl_q and Cl_r values as outlined below. This provides the full range of reasonable NR estimates, with the high end NR estimates representing gross undercatch of precipitation, calculated as noise and wind-corrected measured plus 20%, concurrent with gross over calculation of stream discharge, measured less 20%, appropriate ±5% lab error, and low end Cl_r values. Low end NR estimates are produced using the opposite end of each parameter error and uncertainty range. Our precipitation maximum error is determined from wind-corrected precipitation values utilized for 2005–2009 which were 9–20% higher than uncorrected cumulative shielded gauge data. The 20% maximum difference between shielded gauge data and wind-corrected data provides our maximum projected error likely to occur as undercatch after wind correction or resulting from overcorrection. This large error range is assumed to also address hypsometric regression error for which the monthly average regression root mean square error is

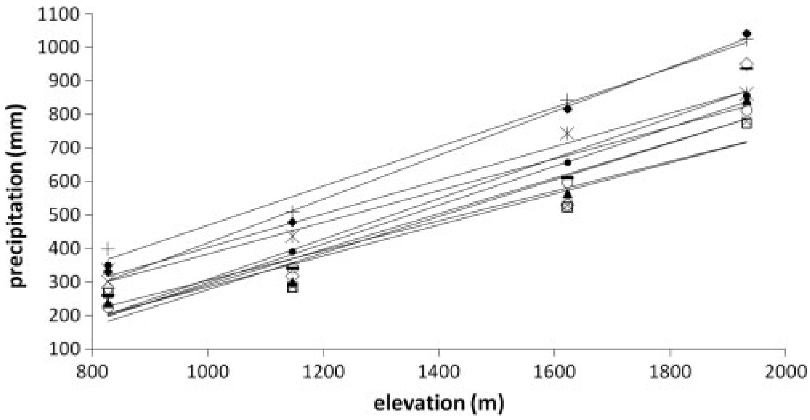

Figure 3. Annual precipitation *versus* elevation along the Boise Front, July–June water years 2000–2009. Data points are from Boise Agrimet (827 m), Lower weather station (1146 m), TL weather station (1622 m), and Bogus Snotel station (1932 m). Years with the highest precipitation are 2006 and 2009

0·94, minimum 0·81, for primary precipitation months, October through June. Stream discharge values are based upon frequent stream discharge measurement applied to quality-controlled continuous stage data. The ±20% error stated for stream discharged is an estimate based upon our replication of United States Geological Survey discharge measurement using standard pygmy current meters and acoustic Doppler meter in small turbulent streams and annual rating curve regression for which the maximum error is projected to be 20%. Analysis of our rating curve regression error, combined with the projected maximum 20% error in discharge measurement and stage recording precision, places combined error as no more than 20%.

Uncertainty may be considered for Cl_p relative to unknown variance in Cl_p for years before or following sampled years 2005–2006. Regional records of chloride deposition from National Atmospheric Deposition Program (NADP) sites and calculated long-term deposition rates, as explained in results below, indicate long-term stability in average annual Cl_p with a decreasing trend occurring during the past two decades. As a result, Cl_p applied in this study may contribute to underestimation of NR where groundwater sampled represents recharge occurring before the study period. Uncertainty in Cl_q values, as Cl_{qn}, are determined for LG as a function of estimates for lower reach road salt-derived chloride, in addition to correction for road salt chloride at C1W and C1S. Uncertainty exists in catchment Cl_r values because of limited spring and well groundwater measurement points, as well as the lack of bedrock flow path delineation. To provide a plausible range in Cl_r values for BG, C1E, and C2E, values are derived from average late dry season streamflow at respective catchment outlets and as average late dry season springflow. We consider late dry season streamflow to represent the integrated chloride concentration of baseflow for the given catchment, while late dry season springflow from high-elevation headwater springs represents the integrated chloride concentration of limited areas upgradient of a given spring. For C1E, this includes south-facing ridgeline seep area BS (Figure 2) and two unnamed southwest-facing springs (Figure 2). Given the lack of springflow in the TL catchment, Cl_r values for TL are determined as an average of catchment-wide late dry season headwater spring values, with uncertainty placed at the range of all spring chloride values. The uncertainty range in Cl_r applied for LG is the catchment-weighted groundwater value described previously, with additional values assigned from average late dry season streamflow at LG and the test well samples. In our analysis, we found the CMB equation to be least sensitive to groundwater and streamflow chloride concentration with similar sensitivity among the remaining parameters.

RESULTS

Precipitation and streamflow

Average annual precipitation across the Boise Front increases with elevation from a low 200 mm at the base of the Boise Front to nearly 1000 mm near its peak, as measured over the past 10 years (Figure 3). Most of this precipitation occurred November through May, primarily as snowfall above 1500 m December through April (Figure 4). Annual average precipitation for DCEW during the period of record, 2000–2009, is 635 mm, in contrast to 691 mm for the 2005–2009 integration period. The 30-year average precipitation for DCEW is inferred to be 656 mm, based on comparison of the DCEW 10-year precipitation record to regional SNOTEL 30-year records. In this comparison, 2004 is identified as an average precipitation year, and years 2000–2003 are below average.

Annual streamflow at LG ranged from 92 mm in 2005 to 246 mm in 2006. In general, annual streamflow is minimal July through August followed by gradual discharge increase through autumn and winter months, punctuated with March–April snowmelt-induced peak discharge and concluding with gradual discharge decrease into June. Annual peak discharge ranged from a minimum peak 0·41 m^3/s in 2007 to a maximum 4·15 m^3/s in 2002 with a close second occurring in 2006. Discharge to precipitation ratios at LG ranged between 0·14 in 2005 and 0·31 in 2006, with a 10-year average 0·25. Spatial variation in Q/P is notable between catchments at 0·22,

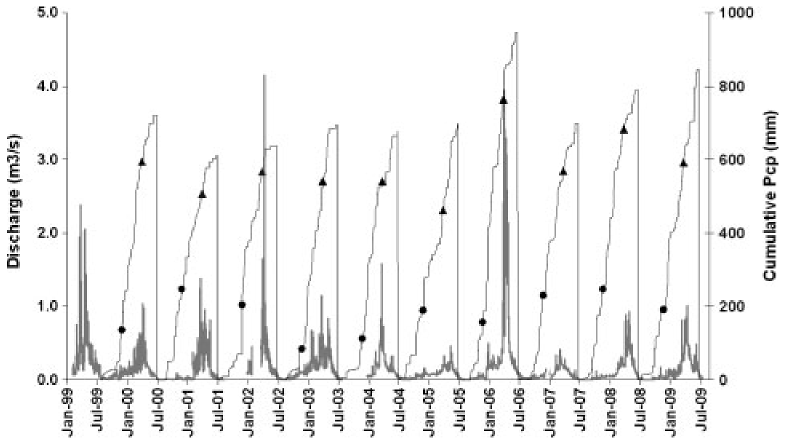

Figure 4. Cumulative precipitation measured at Bogus Basin Snotel site (1932 m) and stream discharge measured at LG for July–June water years 2000–2009. Closed circles denote cumulative precipitation as of 1 December and triangles denote cumulative precipitation as of 1 April, defining the usual snowfall accumulation period for DCEW above 1500 m

0·44, 0·34, 0·21, and 0·20 for TL, BG, C1E, C2E, and LG, respectively, as 5-year averages.

Chloride concentrations

Because of differences between the bulk wet and dry deposition sample population below 1500 m *versus* the population sample above 1500 m (Figure 5), we determined mean bulk wet and dry fall chloride concentrations for these two elevation zones. Mean chloride concentration below 1500 m was 0·442 ppm, range 1·81 ppm, and 0·378 ppm, range 0·962 ppm, above 1500 m. This elevation zone difference in chloride concentration is likely because of snow-dominated precipitation at the higher elevations. The bulk wet and dry deposition chloride concentration sample values for each elevation-bound sample population were sampled using Monte Carlo Bootstrap resampling to determine spatially weighted average precipitation bulk chloride concentrations for each study catchment. For DCEW at LG, the area-weighted result is 0·407 ppm (95% confidence interval 0·378–0·430 ppm). The area-weighted results were 0·378, 0·378, 0·389, and 0·408 ppm for catchments TL, BG, C1E, and C2E, respectively (Table I). This value range for average annual bulk wet and dry deposition chloride is reasonable relative to global (Hem, 1985; Oberg, 2003) and regional chloride deposition records, as detailed below.

The average annual precipitation volume-weighted wet deposition concentration for nearby Reynolds Creek is 0·10 ppm [standard deviation (SD) 0·04 ppm], while the Smiths Ferry value is 0·07 ppm (SD 0·02 ppm) based upon 22 years data (NADP). Dry deposition may be conservatively estimated as twice the wet deposition for a given location (Oberg, 2003), which translates these regional wet deposition values to a conservative estimate of regional mean bulk chloride deposition as 0·25 ppm. The higher value for DCEW bulk chloride deposition may be representative of anthropogenic input and eolian transport of lower elevation hillslope, agricultural field, and lakebed sediment. This is particularly relevant in comparison to Reynolds Creek and Smiths Ferry which are remote, high elevation, and snow-dominated. In contrast, wet deposition chloride concentration values are higher at the Great Basin NADP site in northern Nevada, 0·15 ppm (SD 0·06 ppm) for which the conservative bulk wet and dry concentration would be 0·45 ppm. The Reynolds and Smiths Ferry sites show a decreasing trend in chloride concentration over the period of record, 25% lower during the study period relative to the 20-year record. This indicates that our precipitation sampling may provide values which are low relative to chloride concentrations contributing to groundwater recharge before our study period. Assuming the 20-year average Cl_p to be 25% higher than that measured for the study period, the resultant 20-year average Cl_p value for DCEW would be 0·509 ppm. Additional sampling conducted at 74 sites across Nevada presents a mean wet deposition value of 0·4 ppm (66 sites) and 0·6 ppm for bulk wet and dry deposition (8 sites) (Russell and Minor, 2002). In addition to sampling local precipitation and utilizing limited regional measurements of chloride concentration in precipitation, recent studies (Fabryka-Martin *et al.*, 2000; Russell and Minor, 2002; Zhu *et al.*, 2003) have utilized $^{36}Cl/Cl$ ratios in recharging water and long-term ^{36}Cl deposition rates. Using these methods and ^{36}Cl deposition rates established for 44° north latitude, Russell and Minor (2002) provide results for mean bulk wet and dry chloride concentration for study sites in Southern Nevada as 0·431 ppm (SD 0·083 ppm, range 0·257–0·884 ppm). For this same region, Zhu *et al.* (2003) show average annual Cl_p to be relatively constant over the past 5000 years. Considering these comparisons, our values for Cl_p may be slightly low relative to long-term chloride deposition and, thus, would contribute to underestimation of net recharge where groundwater sampled represents recharge occurring before the study period.

The average chloride concentration in late dry season (August and September) headwater springs includes 1·067 ppm (SD 0·402 ppm, ten samples; Table I) at south-facing BS, in contrast to 0·612 and 0·859 ppm for two eastern, north-facing springs in catchment C1E

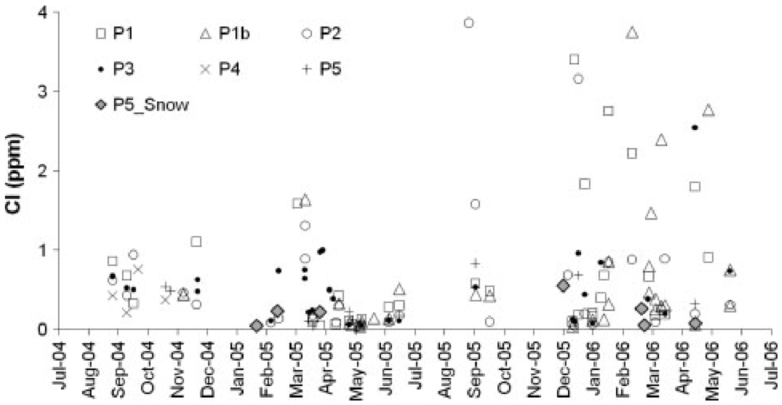

Figure 5. Bulk wet and dry fall chloride concentrations in precipitation. Collector locations are shown in Figure 2

Table I. CMB results for the 2005–2009 5-year integration period. For BG, integration period is 2008–2009. Ranges in NR/P values are based on $\pm20\%$ P, $\pm20\%$ Q, $\pm5\%$ lab error and uncertainty in Cl_q and Cl_r

	Site				
	TL	BG	C1E	C2E	LG
Gage elevation (m)	1607	1698	1335	1158	1036
Drainage area (km^2)	0·02	0·52	8·58	7·5	26·93
Cl_p (ppm)	0·378	0·378	0·389	0·408	0·407
Cl_q (ppm)	0·5	0·56	0·64	0·92	0·95 (0·82,1·06)
Clr (ppm)	0·78 (0·6–1·5)	0·57 (1·07)	0·62 (1·07)	0·90 (0·8)	1·5 (1,2·8)
P annual average (mm)	705	899	754	736	691
Q annual average (mm)	158	399	259	154	140
NR annual average (mm)	245	201	207	175	99
NR/P	0·34 (0·17–0·44)	0·22 (0·03–0·42)	0·27 (0·08–0·43)	0·24 (0·14–0·39)	0·14 (0·04–0·31)
Q/P	0·22	0·44	0·34	0·21	0·20
ET/P	0·44 (0·34–0·6)	0·34 (0·14–0·53)	0·39 (0·23–0·58)	0·55 (0·40–0·65)	0·65 (0·35–0·83)

and 0·795 ppm (SD 0·223 ppm, 20 samples) at C2E headwater springs. As C1E is predominantly south facing, the 1·067 ppm value is utilized. Results for late dry season streamflow chloride at headwater catchment outlets are BG 0·572 ppm (SD 0·328 ppm, 6 samples), C1E 0·624 ppm (SD 0·255 ppm, 11 samples), and C2E 0·903 ppm (SD 0·191 ppm, 13 samples). These values are derived from samples taken each year of the study period 2005–2009. The low elevation groundwater well sample was analysed as 2·8 ppm ($\pm$0·144 ppm as 5% lab error). In general, these results follow the trend of increasing chloride concentration in groundwater with decreasing elevation found by Russell and Minor (2002) in Nevada Mountain Front investigations. An exception to this trend is apparent relative to aspect wherein lower chloride concentrations occur in low elevation north-facing aspects of C1E and northwest-facing C2E springs relative to high elevation south-facing spring BS.

At a given location, groundwater chloride concentration represents water recharged upgradient of the sample location (Russell and Minor, 2002). In consideration of this statement and field observations in DCEW, we state the following assumptions for groundwater chloride concentrations applied to catchment CMB: (1) late dry season streamflow for headwater outlets BG, C1E, and C2E is derived solely from upgradient groundwater and (2) some recharge occurring in these headwater catchments may not be represented in the sampled late-season streamflow. Similarly, it must be understood that the low elevation well, drilled to 226 m and sampled from a screened 100 m section of unconfined fractured granitic bedrock aquifer, does not represent groundwater recharge that may occur via channel seepage along stream reach C1E to LG, nor water recharged on the opposite north-facing aspect of the stream reach. We applied 2·8 ppm as representative of groundwater recharge occurring on slopes downgradient of headwater catchments C1E and C2E, to an area-weighted chloride concentration for DCEW groundwater. On the basis of our observation of varied groundwater concentration relative to slope aspect, this value is considered conservatively high. Using 1 ppm for groundwater recharged in C1E and C2E catchments as the average conservatively high value, the area-weighted result for the larger catchment Cl_r at LG is 1·5 ppm. The average late-season headwater spring chloride concentration used as Cl_r in the TL subcatchment is 0·775 ppm (range 0·491–1·867, using all DCEW spring samples).

Streamflow chloride concentrations increase from late fall through early spring and decrease to minimum concentrations near the end of spring runoff. This trend occurred consistently for each outlet with peaks in chloride concentration occurring earlier and with less distinction at higher elevations (Figure 6). Annual

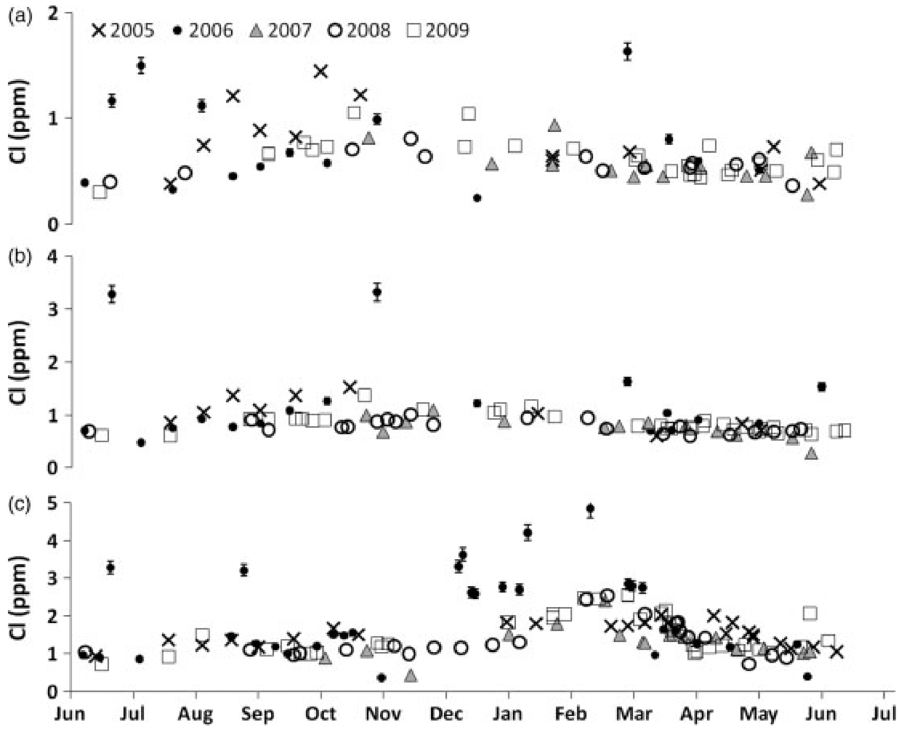

Figure 6. Annual time series for streamflow chloride concentrations: (a) C1E, (b) C2E, and (c) LG, water years June–July; 5% lab error shown for highest values, 2006

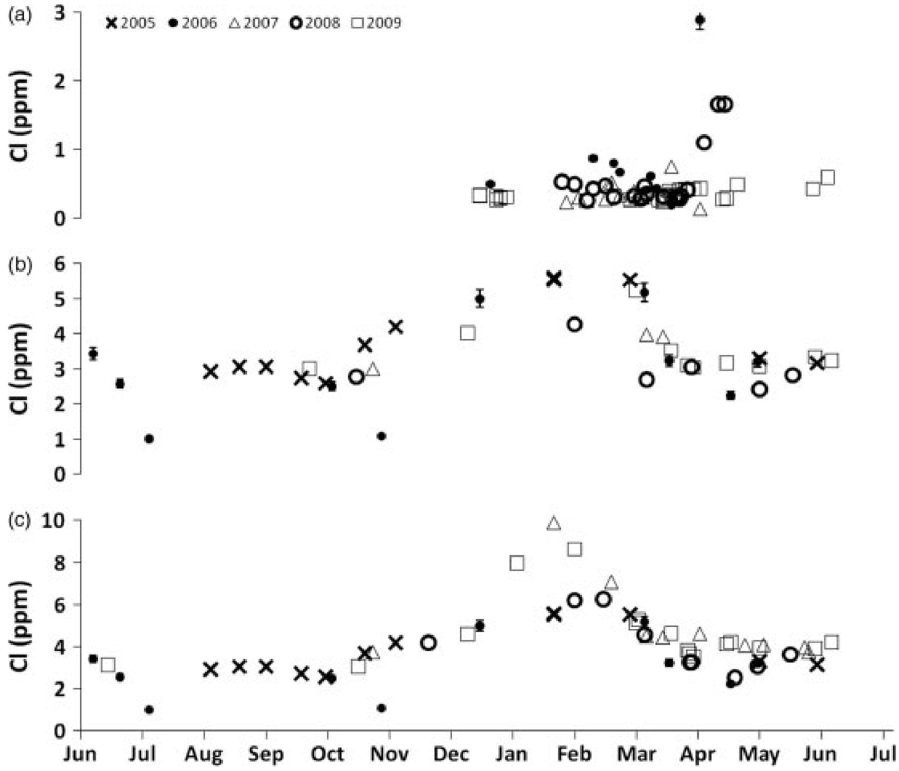

Figure 7. Annual time series for streamflow chloride concentrations: (a) TL, (b) C1S, and (c) C1W, water years June–July; 5% lab error shown for highest values, 2006

volume-weighted average streamflow chloride concentrations were consistently greater downgradient, with exception of C1S and C1W, wherein chloride concentrations are anomalously high relative to upgradient TL and adjacent subcatchments (Figure 7). The 5-year average streamflow chloride concentrations, Cl_q, determined from monthly volume-weighted average chloride concentrations, are 0·5, 0·56, 0·64, 0·92, and 1·57 ppm for TL, BG, C1E, C2E, and LG, respectively (Table I). The 5-year volume-weighted average LG Cl_{qn} values for the three

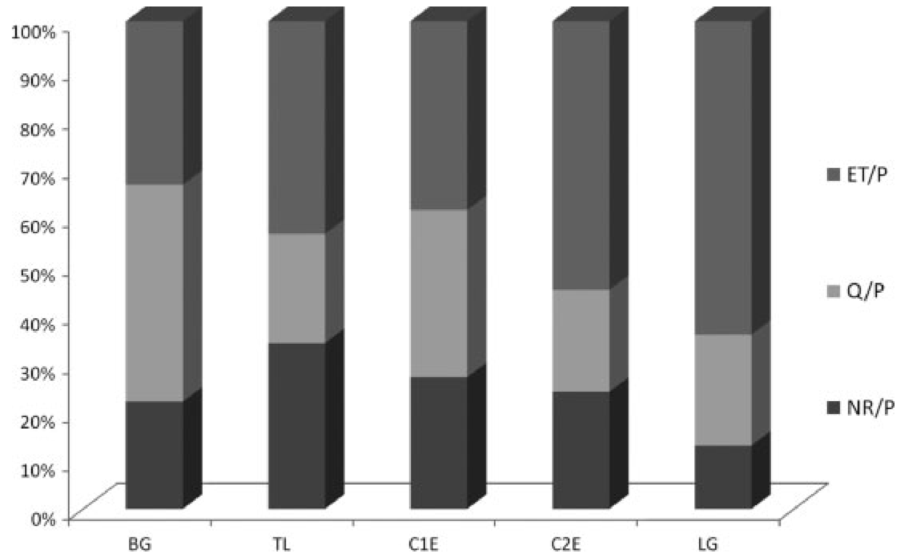

Figure 8. Results for water balance using measured stream discharge, catchment NR calculated by CMB, and evapotranspiration as the residual

Table II. Annual and 5-year values for chloride input to stream reach C1E–LG, compared to partial year gains in chloride mass along the reach. Months were selected based on gains in both streamflow and chloride mass with $Cl_x > Cl_r$ at LG (1·5 ppm), or conditions of gain in chloride mass concurrent with streamflow loss as channel seepage

Water year, July–June	Annual *P* Cl (kg)	Annual road salt Cl (kg)	Total annual Cl (kg)	Annual Cl gain along reach (kg)	Annual gain/ Annual input	Stream reach gain minus annual input (kg)	Months for Cl mass gain along reach
2005	1455	612	2067	1192	0·54	−875	January–June 2005
2006	1763	2309	4072	5772	1·42	1700	September 2005, January–May 2006
2007	1352	518	1870	350	0·19	−1520	March 2007
2008	1492	275	1768	1645	0·93	−123	November 2007, March–April 2008
2009	1734	1349	3082	2056	0·67	−1027	March–April 2009
5 year	7796	5064	12 860	11 015	0·86	−1845	

Note ‘chloride flushing’ apparent in 2006.

estimation scenarios are 1·06, 0·948, and 0·819 ppm, assuming, respectively, no road salt below C1W, road salt below C1W in 2006 only, and road salt below C1W every year.

NR and catchment water balance

NR in headwater catchments TL, BG, C1E, and C2E, determined for the 5-year integration period 2005–2009, ranged between 175 and 245 mm/year, or 0·22–0·34 NR/*P* (Table I; Figure 8). For DCEW at LG, NR was 99 mm/year, or 0·14 NR/*P*. For NR/*P* at LG, the range is 0·04–0·31 for the 5-year integration, compared to a range 0·14–0·42 for TL. Calculation of ET as water balance residual reflects the error and uncertainty range of NR/*P* values. In the small TL headwater catchment, NR/*P* was greatest at 0·34 (0·17–0·44). This high rate agrees with independently modeled estimates by Kelleners *et al.* (2009; 2010). In contrast, NR/*P* in headwater catchment BG was 0·22 (0·03–0·42). Downstream of BG, at C1E, NR/*P* is 0·27 (0·08–0·43), in contrast to 0·24 (0·14–0·39) in the similarly sized C2E catchment. Annual ET/P using Equation 1 ranged from 0·34 (0·14–0·53) at the BG catchment to 0·65 (0·35–0·83) for DCEW overall (Table I; Figure 8).

Unsaturated zone stored chloride

Estimates of bulk atmospheric chloride and road salt-derived chloride input to reach slopes exceeded slope-derived chloride gains along reach C1E to LG attributable to unsaturated zone stored chloride (Table II). The 5-year result is 86% gain in chloride mass relative to estimated input. Assessed annually, the results were 58, 142, 19, 93, and 67% for each of the five water years 2005–2009. The number of months for which gains were characterized as attributable to unsaturated zone stored chloride varied as 6, 6, 1, 3, and 2 months, respectively, with March and April spring runoff months providing the most consistent slope-attributable chloride gains under the conditions of combined gain in streamflow and chloride mass. Values for $Cl_x > 1·5$ ppm ranged from 1·64 ppm to 25·19 ppm, with a median value 6·36 ppm, for the included months experiencing gains in both streamflow and chloride mass. It is understood that during conditions of combined streamflow and chloride mass gains, some of the streamflow and chloride contribution may be attributable to baseflow. Such potential baseflow contributions are, however, assumed minimal relative to slope contributions during the relevant peak spring runoff months, particularly in light of verified consistent streamflow losses during fall and winter months along the reach. Rare conditions of gain in chloride combined with loss in streamflow occurred in January and February and/or September, November. Relative to field observations and prior investigations (McNamara *et al.*, 2005), we interpret these conditions to result from the initial lateral soil water flux that occurs once soil moisture is sufficient and slope hydraulic connection is accomplished following fall and early winter wet-up along stream reach slopes.

DISCUSSION

NR and catchment water balance

Water loss by infiltration into underlying bedrock is an important component of the catchment water balance at all scales in this study (Figure 8). The relative importance of NR within each catchment depends on many interacting factors such as the competing demands of ET with NR, the occurrence of infiltration pathways in fractured bedrock and groundwater discharge to springs and baseflow. Water discharging to springs and gaining streams likely originates as NR in upgradient catchments. For example, consider the ephemeral TL catchment, which had the highest rate of NR. Water that infiltrated into the bedrock within the TL catchment boundary becomes NR for that catchment only and likely emerges within the larger C1W catchment through a rapid-response fractured aquifer system, as evidenced by seasonal fluctuation of discharge and chloride concentration at downgradient spring C1S (Figures 2 and 7). C1S is located several meters upslope and NW of C1W, also downgradient of the TL catchment, and flows longer during the dry season than the C1W tributary. NR is lower in the high elevation perennial spring-fed BG catchment than in the TL catchment. BG and other spring-fed headwater catchments receive groundwater discharge all year, yet still lose water annually to NR. This demonstrates the complexities fractured bedrock can present, incurring both lateral and deeper groundwater flow paths. It is unknown if NR at the largest scale in our study ultimately recharges the aquifers in the valley adjacent to the mountain block and/or emerges within the creek downstream on the mountain front.

Perennial streamflow discharge from the headwater catchments, combined with downgradient streamflow loss to groundwater, indicates that downgradient streamflow loss to groundwater may be an important component of NR at the large catchment scale and, consequently, an important component of MBR. The location of C1S (C1W catchment) and SCS (C2E catchment), of similar elevation combined with their unique qualifications as the lowest elevation observed springs and tendency to incur cessation of flow in the dry season, indicates their elevation as that below which groundwater flux to streamflow is minimal. On the basis of these observations, combined with measured loosing conditions along stream reach C1E to LG, we suspect predominantly loosing stream conditions across the mountain front below these springs (Figure 2). Conceptually, these streamflow losses are routed downgradient as deep groundwater flowpaths in the mountain block. The trend of increasing ET with decreasing elevation, combined with less total precipitation occurring predominantly as rainfall at lower elevations, suggests lower rates of NR at lower elevation slopes across the mountain front. This further emphasizes the importance of mountain front streamflow loss to NR relative to NR incurred at hillslopes.

Evapotranspiration can be calculated as the residual of the catchment water balance with these estimates of NR (Table I). The CMB-based estimate of average annual ET, 452 mm at LG, is considerably lower than previous efforts using the Soil Water Assessment Tool (SWAT) model (Stratton *et al.*, 2009). If we neglect NR in the annual water balance, as is commonly done, and calculate ET as $P - Q$, annual average ET is 551 mm. This closely matches the estimate in Stratton *et al.* (2009). It is possible that the SWAT model is not adequately accounting for losses to groundwater recharge, leaving too much water available for ET. This suggests that the SWAT model may be calculating ET correctly, but arriving at an incorrect value based upon incomplete water balance constraints. For an overall evaluation of ET in our mountain front water balance, we consider trends presented in our results. Both groundwater chloride concentrations and CMB water balance residual calculations (Table I) indicate increasing evapotranspiration rates with decreasing elevation. As shown by ecohydrologic process modeling (Magruder *et al.*, 2009), and indicated by soil moisture studies in DCEW (McNamara *et al.*, 2005; Williams, 2005; Smith, 2010), rainfall occurring in semi-arid mountainous regions rarely moistens soil at levels adequate to incur infiltration to bedrock. In contrast, snow accumulation and subsequent snowmelt occur at higher elevation on the mountain front and provide the primary flux for hillslope bedrock infiltration (Magruder *et al.*, 2009). It is likely that the gradual progression of precipitation regime from snowfall to rainfall with decreasing elevation on the mountain front may primarily facilitate increased precipitation loss to ET with decreasing elevation. This possibility warrants further research where climate change forces rising snowline elevations.

The assumption of zero net storage of chloride

The fundamental assumption of zero net change in unsaturated zone stored chloride during the integration period has been assessed both qualitatively and quantitatively in this study. Water year 2006 experienced a notably high Q/P ratio associated with above normal peak spring runoff (Figure 4), the occurrence of rarely observed low elevation tributary flow and anomalously high streamflow chloride concentrations at LG (Figure 6), indicating mobilization and transport of unsaturated zone stored chloride. The anomalously high streamflow chloride concentrations at LG in 2006, occurring predominantly early spring, are interpreted as evidence of inter-annual 'chloride flushing', i.e. transport of unsaturated zone chloride which had been stored during prior low Q/P water years. The overall chaotic pattern of chloride concentration in streamflow at LG in 2006 may be attributed to a number of hydrologic conditions, including cessation of lower reach flow late August 2005, followed by abundant November rainfall and rain on snow and periodic warming-induced rapid snowmelt events. Though inconsistent year to year, mobilization of unsaturated zone stored chloride is indicated by rising streamflow chloride concentrations with fall, winter, and/or spring runoff during each of the study years 2005–2009. At 86%, chloride mass gains along reach C1E to LG relative to estimated chloride input to slopes does not clearly verify zero net unsaturated zone stored chloride for the 5-year period. However, given the conservatively high estimates for road salt chloride delivery to reach slopes and minimal use of monthly chloride gains along the stream reach, 86% does not preclude zero net unsaturated zone chloride for the integration period. Road maintenance records and observation indicate occasional (<20% of the C1W catchment application rate) road salt application during the study period for C1E to LG reach slopes, occurring primarily in 2006, in contrast to our annual 50% estimate. If we calculate accordingly for road salt application occurring exclusively in 2006, the resultant gain/input ratio is 109%. The occurrence of chloride gains via baseflow during the months applied to the analysis and/or slope chloride additions in excess of our estimates would counter this assessment. If the 86% gain/input value is accurate, we have overestimated NR by 1% (13% NR/P *versus* 14% NR/P. Although precipitation in this semi-arid environment may be inadequate within a given year to accomplish full soil profile wet-up necessary to transport unsaturated zone stored chloride, rapid and complete transport of unsaturated zone stored chloride is likely given the thin sandy nature of the soil and occasional wet year.

The assumption of well-constrained chloride concentrations

The assumption that chloride concentrations are well constrained is challenging in three ways because of the difficulties in assessing the spatial variability of chloride concentrations in recharge water, conducting an experiment long enough to reconcile with groundwater ages, and accounting for anthropogenic inputs of chloride. The concentration of chloride in recharge water is perhaps one of the most significant sources of error in the CMB method because deep groundwater beneath sub-catchments may integrate larger areas. The uncertainty is clear in a catchment, such as the TL catchment for which a groundwater sample is unavailable. As a rough approximation of what groundwater concentrations should be in TL, we may consider the fate of chloride within a year. Chloride that is stranded in the soil by summer evaporation is remobilized and transported to the water table when precipitation resumes. Fall rains likely do not transport chloride to NR because soil profiles are not fully wet (McNamara *et al.*, 2005). Winter and spring snowmelt likely transports its own chloride plus the chloride that is stored in the soil. Some of this chloride is routed to streamflow, with the remainder routed to NR. Thus, chloride concentration in NR water should be equal to the mass of chloride delivered by total annual dryfall and precipitation ($P \times \mathrm{Cl_p}$), less the chloride mass discharged by annual streamflow ($Q \times \mathrm{Cl_q}$), divided by the volume of snowmelt less the seasonal streamflow.

$$\mathrm{Cl_r} = (P \times \mathrm{Cl_p} - Q \times \mathrm{Cl_q})/(S - Q) \tag{6}$$

where $\mathrm{Cl_r}$ is the chloride concentration in groundwater, P the volume of rainfall received in the catchment since the last snowmelt period, $\mathrm{Cl_p}$ the average bulk chloride concentration of the dryfall and precipitation, $\mathrm{Cl_q}$ the volume-weighted average chloride concentration of stream discharge, Q the stream discharge volume for the applied period, and S the volume of annual snowmelt for the catchment. If we assume minimal evaporative losses from the snowpack, the estimated chloride concentration in recharge water below the TL catchment is 0·73 ppm, which is very close to the average late-season streamflow value that we used for $\mathrm{CL_r}$ in the TL catchment. If we assume 25% evaporative loss from the snowpack (Fassnacht, 2004), and additionally assume all November snowfall to be sublimated, the 5-year average $\mathrm{Cl_r}$ value for TL is 1·5 ppm; this value of $\mathrm{Cl_r}$ approximates the highest springflow chloride concentration and is used to calculate the low estimate of NR/*P* for TL, 0·17.

The CMB integration period should constrain chloride mass flux over the time period represented by the recharged water. Groundwater sampled at the headwater catchments conjoining the study area is 9–23 years old (Gates *et al.*, 1994). Our 5-year integration period does not satisfy this requirement. However, regional long-term records, for both precipitation and atmospheric chloride input (Reynolds Creek, South Mountain and Smiths Ferry, Idaho, NRCS and NADP data), indicate stationarity in precipitation over the recent 30-year period and a decreasing trend in wet chloride deposition. If this trend is true for the study area, then the chloride concentration in precipitation applied presents an underestimate of groundwater recharge, while the precipitation volume for the study period may contribute to a slight overestimation. Assuming stream discharge to be a function of meteorologic conditions, it may be considered an essentially stationary parameter, particularly over multi-year integrations, where meteorologic conditions have been stationary. The final consideration is groundwater chloride concentration.

A potentiometric surface constructed from well water level data (Aishlin, 2006), inclusive of the test well (Figure 2), indicates that the source area for the sampled test well water is limited to the western, low elevation DCEW ridgeline. The aquifer test conducted at this ridgeline test well estimates maximum hydraulic conductivity at 0·24 cm/day, in contrast to 5·18 cm/day at a well immediately adjacent to the Bogus Basin SNOTEL site (Figure 2), indicating the likelihood that low elevation groundwater age is >9–23 years stated for the Bogus Basin test well (Aishlin, 2006). The estimated (86%) mobilization and transport of C1E to LG slope-stored unsaturated zone chloride for the 5-year CMB application leaves in question historic delivery of chloride to groundwater as sampled at the test well. As an alternate approach to acquiring a groundwater chloride concentration appropriate to the study integration period for C1E to LG reach slopes, we can consider potential bedrock infiltration conditions for the reach slopes during the study period. Assuming that soil profile wetting resulting in tributary flow generation concurrently facilitates vertical infiltration on ridgetops and slopes into bedrock where matrix or fracture permeability is adequate (McNamara *et al.*, 2005), it may be stated that vertical infiltration and percolation to groundwater will occur beneath slopes during months when precipitation conditions are adequate for tributary flow generation. Such conditions were observed January through June along reach C1E to LG in 2006 with January–March measured Cl_q at an average 2·2 ppm. This chloride concentration is high relative to average measured streamflow chloride concentration along the reach and likely occurred as a result of the observed rare tributary flow. The average tributary chloride concentration for this reach may be inferred from the January through June 2006 volume-weighted average concentration of C1E to LG gains, Cl_x, 3·55 ppm chloride. If 3·55 ppm chloride is assumed to represent the recent 5-year average concentration of vertically infiltrating and percolating water along the lower elevation ridge and slopes, the resulting estimated Cl_r value at LG would be a spatially weighted average of 3·55 and 1 ppm *versus* 2·8 and 1 ppm, resulting in Cl_r 1·7 ppm *versus* the previously stated 1·5 ppm. This results in NR/P 0·13 rather than 0·14.

Road salt is a considerable challenge in this study where it affects the lower portion of DCEW. It should be noted that under-accounting for road salt would lead to erroneously high values of Cl_{qn} and under estimation of NR for the LG catchment. Although we attempted to account for road salt additions, this may explain the relatively low values for the LG catchment. All other catchments in Table II have considerably high NR magnitudes and are presumably not affected by road salt because of their locations.

CONCLUSIONS

Catchment-scale NR into underlying bedrock is an important component of the mountain front water balance, as well as the source of MBR to valley aquifers. In this study, spatial variation in NR indicates the relative importance of specific mountain front sub-environments to total MBR, including headwater catchments with ephemeral streams, north-facing slopes, and springflow catchments contributing streamflow for downgradient channel seepage to groundwater. Approximately 14% of average annual precipitation routes to NR in DCEW. In contrast, as much as 44% of annual precipitation routes to NR in ephemeral headwater catchments. NR in headwater catchments likely supplies water to lower elevation springs, baseflow, and MBR.

We find application of the CMB method viable to solution of catchment NR in arid or semi-arid mountain front environments wherein thin sandy soils combined with steep slopes facilitate mobilization and transport of unsaturated zone stored chloride. This is particularly true of high elevation catchments with precipitation rates adequate for annual flushing of unsaturated zone stored chloride. Although the CMB approach may be invalid for some individual years in larger catchments, requiring multi-annual integration, the utility of assessing variation in annual NR relative to variation in climatic conditions and hydrologic response motivates the effort of varied integrations. This is especially true when considering the question of climate or land use change effects on runoff, net catchment groundwater recharge, and MBR. The degree of fall wet-up, air temperature fluctuations, the volume of snowmelt, and timing of snowmelt may be particularly relevant to annual NR.

ACKNOWLEDGEMENTS

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