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Regional subsurface mountain-block recharge (MBR) is viewed as a key component of basin aquifer systems found in semi-arid environments. Yet water resource managers do not have a commonly available and reasonably invoked quantitative method to constrain possible MBR rates. Recent advances in landscape-scale ecohydrologic process modeling offer the possibility that weather, climate, and land surface physical and vegetative conditions can be used to estimate MBR. We present an approach that uses remotely sensed physiographic data to model a mountain water balance including the component of MBR. In this approach, we evaluate the ecosystem process model Biome-BGC (Running and Hunt, 1993; Thornton et al., 2002), used in tandem with the mountain climate simulation program MT-CLIM (Running et al., 1987; Kimball et al., 1997; Thornton and Running, 1999), to calculate the annual MBR within a 24,600 ha watershed. The modeling tool is also used to investigate how climatic and vegetative controls influence recharge dynamics along the basin-mountain physiographic gradient. Our work estimated mean annual MBR flux in this crystalline bedrock terrain to be 99,000 m3/d or approximately 19% of annual precipitation. Data analyses indicate that vegetative control on soil moisture flux is significant only at lower elevations and snowmelt is the only significant annual recharge source occurring on a macroscale in this environment. Results also demonstrate that evapotranspiration (ET) is radiation limited in wet years and moisture limited in dry years, and consequently potential recharge to groundwater is significantly higher during wet climate cycles. The application of ecohydrologic modeling to estimate MBR shows promise for modeling MBR at the mountain-scale. However, future efforts will need to incorporate a more advanced understanding of mountain recharge processes and refined ability to simulate those processes at varying and appropriate scales.

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INTRODUCTION

Many of the world's people and sensitive riparian ecosystems found in semi-arid regions are dependent on groundwater derived from adjacent mountain ranges. Often in developed areas of the world, surface water sources alone are no longer capable of meeting societal needs. Increasingly, growing urban populations, and industrial and agricultural interests are relying on mountain margin alluvial aquifers for water supply. However, the development of basin aquifers often proceeds without a clear groundwater budget, mainly because mountain-block recharge is difficult to quantify. In addition, in regions where these aquifers are being exploited, impacts to valley rivers and riparian areas are often poorly understood. In this setting, water supplies and resolution of environmental issues will remain tenuous without the development of more comprehensive methods to define basin hydrologic budgets.

It has been shown that recharge from the mountain block can contribute a significant proportion of the water to a basin aquifer (Wilson and Guan, 2004; Maurer and Berger, 1997; Gannett et al., 2001; Manning and Solomon, 2004 and 2005). Characterization of the processes that control mountain-block recharge above the soil-bedrock interface is confounded by the heterogeneity of mountain meteorology, topography, the spatially complex distribution of vegetation communities, soil/bedrock types, and the lack of site instrumentation and monitoring data. As described by Wilson and Guan (2004) in their overview of mountain-block hydrology, mountain front recharge can be categorized into the following components: 1) focused subsurface recharge that follows flowpaths within faults and fractures; 2) diffuse subsurface recharge through primary permeability in the bedrock matrix; 3) focused near-surface recharge of shallow groundwater transmitted in the sediments of streams which drain the mountain-mass and recharge from streambed infiltration; 4) diffuse near-surface recharge which is the infiltration and deep soil drainage that occurs during episodic runoff events in ephemeral drainages at the mountain front. Quantifying recharge to the basin aquifer from stream loss is less problematic than quantifying other components of mountain-front recharge as standard stream gaging techniques can be used and combined with shallow monitoring well networks (Goodrich et al., 2004). In contrast, the other components of recharge that feed lateral groundwater flux at the mountain-front, which we will refer to as mountain-block recharge (MBR), is particularly difficult to estimate and is the focus in this paper.

In the semiarid to arid basins of the Southwestern United States attempts to quantify mountain front recharge were first developed using empirical precipitation-mountain front recharge regression analyses (Maxey and Eakin, 1949). Other authors have presented analyses of mountain front recharge based on water balance formulation, wherein ET was estimated from empirical data or mathematical models (Feth et al., 1966; Huntley, 1979). The increase in computational power provided by modern computers over the last four decades has allowed researchers to develop more process based methods including the simulation of deep soil percolation and groundwater recharge (Bauer and Vaccaro, 1987; Hevesi et al., 2002; Khazaei et al., 2003). Gogolev (2002) investigated deriving groundwater recharge by coupling a water balance model capable of simulating flux at the soil-atmosphere boundary with an unsaturated flow model based on the Richards equation. In an alternate recharge approach, Dettinger (1989) used basin water chemistry to quantify mountain front recharge using the chloride balance technique. Other studies compare estimates of mountain front recharge based on chloride balances with estimates derived from precipitation-runoff regression and Darcy's Law (Anderholm, 2000; Maurer and Berger, 1997). Estimates of mountain front recharge have also been obtained by using basin centered numerical modeling approaches (e.g.: Tiedeman et al., 1998; Sanford et al., 2000). Dickinson et al. (2004) modeled the water level response in a synthetic basin to develop an analytical model relating water level fluctuations to basin recharge. Flint et al. (2002) provide a comprehensive comparison of techniques for quantifying spatially distributed recharge at Yucca Mountain, Nevada, USA. Manning (2002) and Manning and Solomon (2004) applied an environmental tracer approach by combining age dating of basin groundwater to constrain recharge flux with noble gas concentration to isolate the fraction of mountain front recharge attributable to high elevation MBR. This environmental tracer approach was further refined by integrating chemical data with a numerical model of heat and fluid flow that is calibrated to groundwater temperature and age (Manning and Solomon, 2005).

At the same time that these advances in mountain recharge science were occurring, scientists studying landscape scale ecohydrologic relationships were gaining an advanced understanding of the influence of plant physiological processes on runoff and soil moisture movement. Researchers recognized the relationship between the measurable vegetative parameter leaf area index (LAI) and groundwater recharge (Finch, 1998; Hatton et al., 1993; Zhang et al., 1999a). The control that plant stomatal resistance exerts on ET is a fundamental driver of plant-soil water dynamics and therefore one of the principal factors controlling deep soil water percolation that becomes groundwater recharge (Phillips et al., 2004). It has been exemplified that plants provide significant control on soil water movement especially as aridity increases (Seyfried et al., 2005). The dynamics of plant growth, maturity, and senescence provide important feedbacks with soil water and the energy budget and hence ET of a particular biome. Realizing the importance of plant processes on the water cycle, Rodriquez-Iturbe (2000) suggests that much of past hydrologic research has failed to adequately consider ecosystem-hydrologic process linkage. More recently, researchers have begun to incorporate soil-vegetation-atmosphere (SVAT) models into water balance-groundwater recharge approaches. Much of the research applying SVAT models to water-balance recharge estimates has been used to determine the hydrological implications of land use change on excess recharge and soil salization in Southeast Australia (Hatton et al., 1993; Pierce et al., 1993; Zhang et al., 1999b). Both 1-dimensional SVAT models (Zhang et al., 1996; Zhang et al., 1999a; Gannet et al., 2001) and quasi 3-dimensional SVAT models (Dawes et al., 1997; Zhang et al., 1999c; Arnold et al., 2000; Gogolev, 2002; Walker et al., 2002) have shown utility for modeling recharge processes at variable scales.

In contrast to the basin and range province of the Southwestern U.S. where the majority of mountain recharge studies have been focused, climate in the Northern Rocky Mountains is relatively temperate, a large area of the landscape is mountainous, and annual precipitation on mountain crests may exceed 250 cm (Western Regional Climate Center data). The quantification of mountain recharge under climate conditions typical of more humid mountain regions has seen considerably less research than arid regions. Additionally, the role that mountain ecosystems play in determining the fate of precipitation and recharge to basin aquifers is poorly constrained as these regions are remote and difficult to instrument. To our knowledge, published studies using ecosystem process or SVAT models to investigate mountain recharge are not available.

In this paper we investigate the application of an ecosystem process model, Biome-BGC (Running and Hunt, 1993; Thornton et al., 2002) to provide estimates of the annual MBR to an adjacent alluvial basin groundwater system located in the Rocky Mountains in Montana, U.S.A. Our application of Biome-BGC (biogeochemical cycle) tests the effectiveness of a 1-dimensional ecosystem process model to calculate ET and soil water storage in the heterogeneous environment typical of an alpine mountain range. Ecosystem process models have provided ecologists insight into the functioning of ecosystems from the tree-stand to global scale and have also benefited the hydrological sciences in revealing linkage between atmospheric water, vegetation, and soil moisture. Importantly, these models appear to provide an opportunity for mountain recharge researchers to better understand the partitioning of precipitation into runoff, ET, soil moisture, and groundwater across varied climate and physiographic gradients. The appeal in applying an ecosystem process model to a mountain water balance problem lies in incorporating a more complete plant life cycle into the model, allowing a more detailed consideration of the feedbacks between plant physiological response, climate, and groundwater recharge. An additional appeal of the use of an ecosystem process model, or similarly, of a SVAT model in mountain recharge modeling is that the model can be used to reveal how climate and vegetation patterns influence recharge in specific settings within a mountain environment.

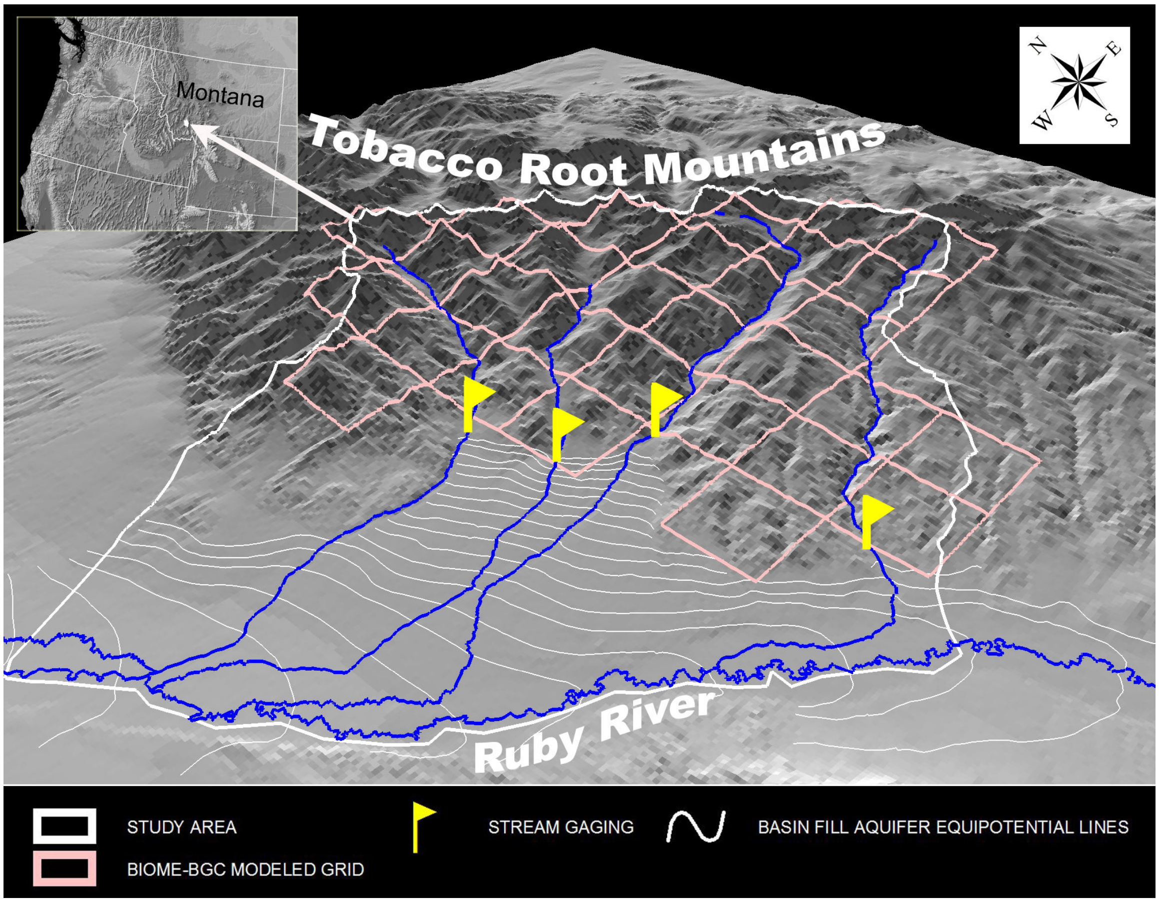
The purpose of this work is to assess the advantages and limitations of using an ecohydrologic model to generate representations of MBR in northern Rocky Mountain landscapes. The first objective of this research is to use Biome-BGC to analyze how climate and vegetation gradients in a mountain range influence the recharge processes. This is accomplished by investigating the relationship between modeled soil water outflow (including groundwater recharge and stream discharge), climate, net primary production, and ET in a mountain range over the course of a 13 year period of climate record. Climatic and vegetative controls on recharge are also revealed by testing the sensitivity of soil water outflow to soil and vegetation parameters across the climatic and physiographic gradient present in the study area. The second objective is to assess if a climate and landscape driven ecosystem process model can be used to generate reasonable estimates of annual MBR to basin aquifers at the mountain-scale. This is accomplished by developing a mountain-scale water balance and comparing the resulting recharge estimate with the results of other MBR studies. We further attempt to constrain the range of possible MBR rates by applying the resulting MBR to a numerical model of the basin alluvial aquifer that is calibrated to measured groundwater head and stream-groundwater exchange locations and rates. The third objective is to present an integrated evaluation of our use of Biome-BGC in estimating MBR. We review other studies relevant to modeling recharge processes to provide insight into how process models can be enhanced to more accurately assess mountain recharge rates.

EXPERIMENT AND DATA

The study area encompasses the southwestern portion of the Tobacco Root Mountains and adjacent Ruby Valley basin in Montana, U.S.A (Figure 1). The study area is coincident with four mountain watersheds and we assume that bedrock groundwater flow divides are coincident with topographic divides. Bedrock within the study area includes Archean quartzofeldspathic gneiss and amphibolite with an underlying Cretaceous granite pluton (Ruppel et al., 1993). Mineral exploration drilling into this bedrock has encountered high artesian pressure at several hundred meters depth providing anecdotal evidence of regional bedrock groundwater flow. Basin fill geology is characterized by a sequence of fine grained Tertiary silts and clays with intermittent sand and gravel conglomerate up to 1.3 km thick (KirK Environmental, 2004b). Relatively coarse grained Quaternary glaciofluvial and alluvial deposits up to 50 m thick overlie the Tertiary basin fill and host the principle unconfined aquifer.

Our ecohydrologic modeling includes the 24,600 ha bedrock portion of the study area. Elevation within the modeling domain ranges from 1600-3200 m. Mean annual precipitation in the bedrock of the mountain range varies from 28 cm/yr at the piedmont zone to 107 cm/yr near the crest of the mountain range (Oregon Climate Service, 1998). The dominant land use within the adjacent Ruby Valley basin is irrigated hay. Peak irrigation demand is approximately 18 m^3/s. Water loss from irrigation water conveyance and field application drives the water table hydrograph and causes seasonal increases in groundwater discharge to streams in the valley bottom (KirK Environmental, 2004a).

Figure 1: Study area location.



Visual Biome-BGC Version 0.69b is a process-based model that calculates the flux and storage of energy, water, carbon, and nitrogen between the atmosphere, plant, and soil components of an ecosystem. Biome-BGC has undergone over a decade of model validation and improvement. The hydrologic output of Biome-BGC and it's predecessor Forest-BGC have proven to accurately predict the timing of snowmelt and surface water discharge when averaged spatially over a watershed (Coughlan and Running, 1997; Running, 1994; White et al., 1998; Kremer and Running, 1996). Additionally, Biome-BGC is programmed to work loosely coupled with the mountain climate simulation program MT-CLIM (Running et al., 1987; Kimball et al., 1997; Thornton and Running, 1999). These attributes make Biome-BGC well suited to applications involving mountain recharge where the interest is in the long-term average of MBR over a mountainous watershed. Biome-BGC requires daily maximum and minimum air temperature, humidity, incident solar radiation, and precipitation as climate inputs. MT-CLIM Version 4.3 provides the necessary climate input by interpolating daily near-surface meteorological parameters across elevation gradients and requires only the temperature and precipitation data that is typically recorded at automated weather stations.

Biome-BGC uses a bucket model for soil moisture storage and drainage. It does not simulate infiltration rates, preferential flow, or lateral moisture flux. Biome-BGC routes precipitation minus canopy interception into soil water or snowpack as a function of daily temperature. Precipitation throughfall and snowmelt become available in the soil compartment for root uptake. ET is calculated by the Penman-Monteith equation using extrapolated site micrometeorology. Actual plant transpiration is modulated by considering the soil water content, vapor pressure deficit, and temperature. By modeling ET sensitivity to plant water stress, Biome-BGC provides a realistic mechanism for modeling soil moisture depletion and actual ET in regions that experience an annual dry period during the growing season. Parameterization of Biome-BGC was accomplished by using the remote sensing and meteorological datasets shown in Table 1. To develop primary modeling units, we partitioned the study area into a grid with 2.9 km^2 cells using standard GIS techniques to determine average soil moisture properties, precipitation, elevation, and slope and dominant aspect and vegetation. Model parameterization and execution was performed manually for each cell necessitating the use of this large-scale grid. Biome-BGC uses ecophysiologic constants files (epc) for parameterization and the current model version includes default epc files for generalized biome types of C3 and C4 grasses, deciduous and evergreen broadleaf and needleleaf forests, and evergreen shrubs. We use the default epc files for evergreen needleleaf forest, shrub, and C3 Grass for sites based on LANDSAT Thematic Mapper classified cover types of conifer forest, dry shrub, and upland grassland respectively.

Table 1: Biome-BGC / MTCLIM Input Parameter Data Sources.

Daily temperature max/min, precipitation	*USDA Snotel*
Elevation, slope, aspect	*USGS DEM*
Biome type	*LANDSAT Thematic Mapper*
Soil texture	*USDA STATSGO*
Annual precipitation	*University of Montana NTSG Daymet*
Annual Nitrogen deposition	*NADP*
Shortwave albedo	*Matthews (1984)*

A mountain weather station (U.S. Department of Agriculture Natural Resource Conservation Service SNOTEL) located 2.5 km outside the study area at 2400 m. elevation provided meteorological data for the period 1991-2003 as input to MT-CLIM. Periodic runoff gaging was performed during the period May 2002-October 2003 using standard U.S. Geological Survey flow gaging techniques (Rantz et al., 1982).

The ecohydrologic water balance approach calculates the annual MBR for an October 1 to September 31 water year by the following equation:

$$MBR_{annual} = \Sigma_{(oct\ 1\ to\ sept\ 31)} [P - ET - \Delta S] - Q_{sw\ annual}$$

Where:

P = modeled daily precipitation

ET = modeled daily evapotranspiration

ΔS = modeled daily change in soil moisture storage

Q_{sw} = measured surface water runoff for the water year

MT-CLIM handles the daily precipitation budget. Biome-BGC provides daily water budgeting of ET as well as soil water storage and outflow. In this paper, the term soil water outflow refers to both percolating water that becomes mountain bedrock groundwater and surface water runoff.

A steady state MODFLOW groundwater model of the basin fill alluvial aquifer was developed to provide constraints on possible MBR rates. The MODLOW domain corresponds to the basin fill alluvium of the northern three watersheds where there are head, hydraulic conductivity, and flux data (Figure 1) and for purposes of MODFLOW modeling, a portion of the MBR was assigned to the mountain-front boundary. The model consists of a uniform 100x100 grid of 136x121 m. cells and seven layers. The active model domain is an 11,900 hectare portion of the basin-fill alluvium from the mountain front to the basin river. Active cells in layers 2-7 follow bedrock topography determined from the gravity survey presented in KirK Environmental (2004b). Conceptually, the approximately 25 m. saturated thickness of the top layer represents the unconfined Quaternary alluvium while layers 2-7 simulate the finer grained confined Tertiary alluvium. Layer 2 is 75 m. thick and represents the relatively transmissive units within the upper Tertiary aquifer system as determined from well logs and aquifer testing (KirK Environmental, 2004b). Layers 3-7 represent the uncharacterized deep Tertiary sediments and properties of these units were adjusted during model calibration. Layers 3-6 are 100 m. thick and layer 7 is approximately 800 m. thick. $K_{x,y}$ in layer 1 ranges from 0.3-76 m/d. $K_{x,y}$ in layers 2-7 ranges 3×10^{-3}-1.5 m/d and $K_{x,y}$ in layers 2-7 decreases with depth. K_z was adjusted from $^1/_{10}$ to $^1/_5$ of $K_{x,y}$ during calibration. Model boundaries include flux boundaries at the mountain front, lateral no-flow boundaries coincident with bounding flowpaths, a river boundary, and a specified head boundary coincident with an equipotential line to simulate basin groundwater underflow. We partitioned the total MBR flux into two components, diffuse bedrock recharge and recharge representing alluvial valley underflow of mountain streams that enter the basin. Diffuse bedrock flux was divided proportionally among layers 1-6 according to layer saturated thickness. Alluvial underflow was calculated based on the measured hydraulic gradient in wells at the mountain front and representative hydraulic conductivities for the geologic formations described in alluvial well logs. Diffuse bedrock MBR and aerial recharge was added using the recharge array in layer 1, while injection wells were used for applying MBR to layers 2-6. Conceptually, diffuse bedrock flux is applied evenly to the upper 500 m. of basin aquifer to approximate diffuse bedrock flux within a decompressed zone as demonstrated by Marechal and Etcheverry (2003). The relatively concentrated recharge flux of stream underflow was applied to the Quaternary alluvium at the mouths of mountain stream valleys. The MODFLOW model was calibrated to surface to groundwater flow exchange in basin streams, measured using synoptic stream gaging. We then attempt to provide constraints on the MBR flux by evaluating the range of values for the mountain front diffuse bedrock recharge boundary which resulted in stream to groundwater exchange falling within our calibration targets. The MODFLOW model was run with mountain stream underflow parameterized as calculated above and one additional simulation with 100% of the calculated MBR as stream underflow to evaluate the instance where bedrock is impermeable.

RESULTS AND DISCUSSION

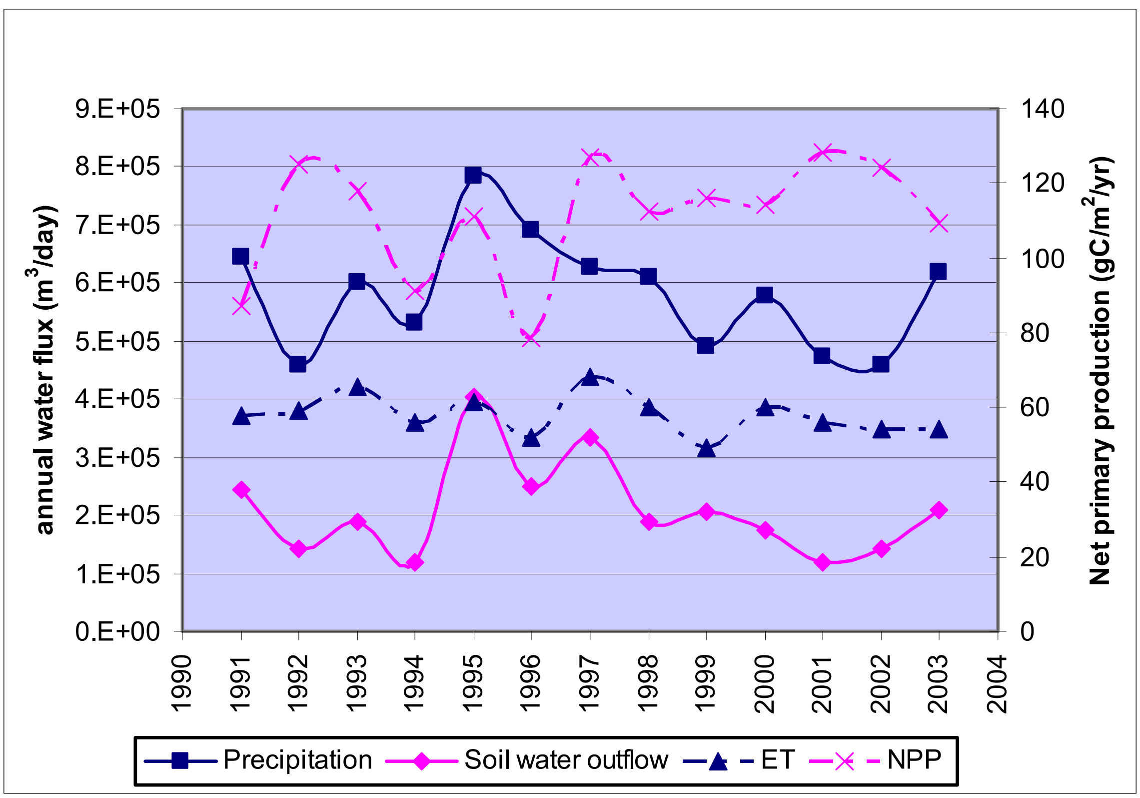
The following sections present the results of the Biome-BGC ecohydrologic modeling, the mountain-scale water balance and calculated MBR, and analysis of the sensitivity of the ecohydrologic model to soil and vegetative land surface parameters. We then provide an evaluation of the MBR estimate which compares our results to other published studies and describes our efforts at constraining MBR to the basin-fill aquifer using MODFLOW. We then present considerations for future efforts using water balance recharge models to represent mountain environments.

Biome-BGC Modeling

Biome-BGC modeled annual precipitation, ET, soil water outflow and net primary production (NPP) was summed for the entire study area (Figure 2). Soil water outflow and precipitation show the only strong correlation ($R^2 = 0.70$) present in these data. No correlation is evident between soil water outflow and ET ($R^2 = 0.14$) or between ET and precipitation ($R^2 = 0.10$). Additionally, mean annual temperature data analyzed but not depicted in Figure 2 demonstrates a weak, but inverse relationship between ET and temperature ($R^2 = 0.26$). The apparent correlation between precipitation and soil water outflow (which includes runoff and MBR components) and the lack of correlation between soil water outflow and ET is evidence that the dominant control on outflow is by precipitation exceeding soil moisture capacity and suggests that the annual evaporative energy budget has a comparatively less significant influence on outflow. The lack of response in ET to temperature and precipitation can be explained in part by characteristics of the model and in part by the climate of the study area. In MT-CLIM, daily incident solar radiation is determined using the algorithm developed by Bristow and

Campbell (1984) which defines an inverse relationship between modeled atmospheric transmittance (cloud cover) and the daily minimum to maximum temperature range. In theory, sky-cover during wetter periods reduces nighttime radiational cooling and lessens daily temperature fluctuation. In practice, the model responds by reducing atmospheric transmittance, reducing the daily radiation load. This theoretical relationship is partially supported by local climate data which shows a weak inverse correlation between precipitation and temperature (R^2 = 0.33). Global climate cycles including El Nino Southern Oscillation (ENSO) favor a more northerly orientation of the North American jet stream during extended periods of wet climate (La Nina) and may also be a cause of the inverse relationship between precipitation and temperature seen in these data.

Figure 2: Biome-BGC modeled annual water flux and net primary production.

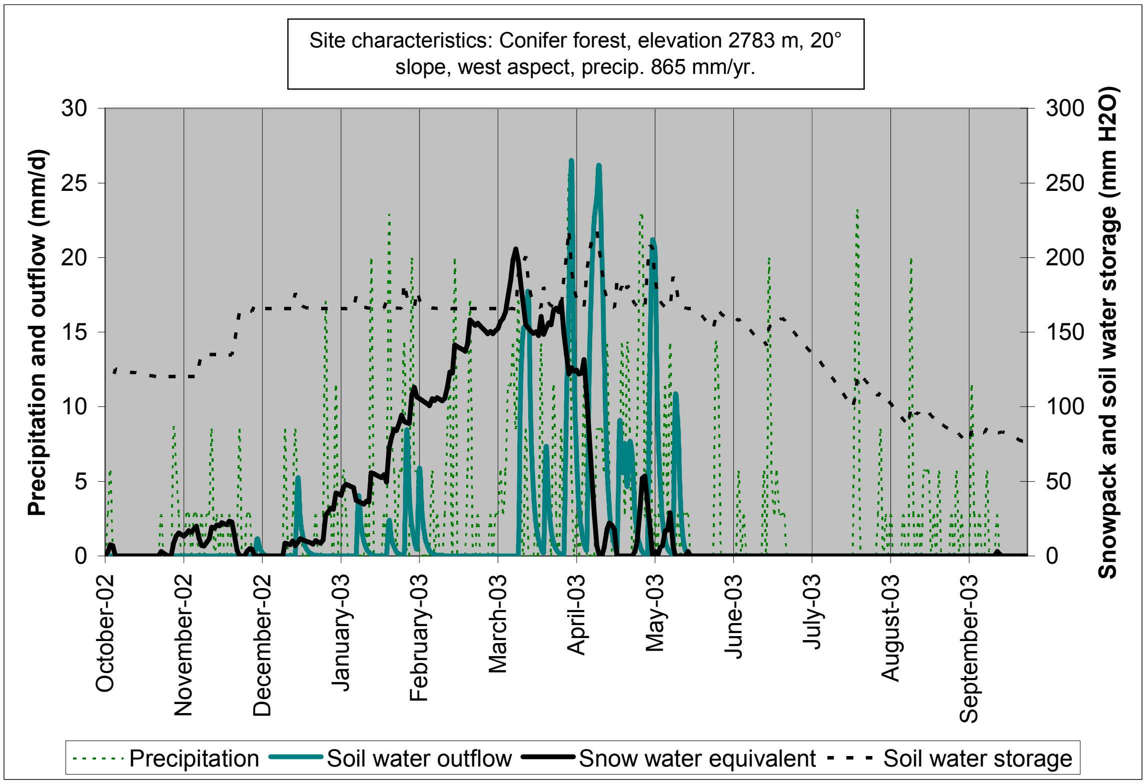


Analysis of NPP and ET trends also demonstrates how ecological controls on the water balance respond to climate. The relationship between annual NPP and mean annual temperature is insignificant ($R^2 = 0.13$) and a weak, but inverse correlation is apparent between NPP and annual precipitation ($R^2 = 0.23$). This inverse relationship between NPP and precipitation combined with the lack of ET response to precipitation suggests ecosystem productivity is radiation limited during years of above average precipitation and moisture limited during years of below average precipitation. The outcome of this climate regime on plant water use is that inter-annual variability in ET is approximately 1/3rd of the variability in annual precipitation. The occurrence of significant variability in precipitation between years and the relatively constant nature of the ET signal appears to be an important factor controlling the interannual variability in soil water outflow which in turn controls the amount of subsoil water available for MBR.

To demonstrate the hydrologic response of a typical temperate semiarid mountain biome, daily water flux and storage states for a high elevation coniferous forest site were modeled (Figure 3). From October through May, soil water storage recovers quickly from the dry-season water deficit and is maintained near field capacity. The October to May period is also coincident with snowpack accumulation and during this period soil moisture flux from snowmelt and rain contribute to soil water outflow. Figure 3 indicates that outside of the snow season not a single precipitation event, including the larger summer storms of magnitude 2 cm/d, raises macro-scale soil moisture above storage capacity and ET quickly depletes additional soil moisture from these storms. This suggests that snowmelt and rain occurring during snowmelt, drives the only significant recharge process occurring on a macroscale in this temperate mountain environment.

The suggestion that MBR is derived predominantly from snowmelt occurring simultaneously with a period of minimal ET presents important implications as to how warmer global temperatures that either raise the elevation of seasonal snow accumulation or cause snowmelt to occur sporadically throughout winter months would effect MBR dynamics. In the climate of the study area, the 13-years of temperature data reasonably correlate with (R^2 = 0.58) a positive linear trend in mean annual temperature with a slope of 0.2° C/yr, indicating a significant warming trend in recent years. Annual NPP does not show as conclusive of a response to this climate trend. As described previously, NPP shows very little correlation with mean annual temperature and an inverse correlation with mean annual precipitation. Although NPP appears to remain elevated throughout the second half of the time period (Figure 2) the increase in annual NPP coincides with a period of drought, suggesting that the higher rates of NPP may be related to increased solar radiation loads. Additional research is needed to fully characterize ecosystem response to climate in this environment and to predict the behavior of ecosystem productivity and associated water budget components under prospective climate change scenarios.

Figure 3: Example daily hydrologic response of a temperate semiarid mountain biome.



Mountain-Block Water Balance and MBR Estimate

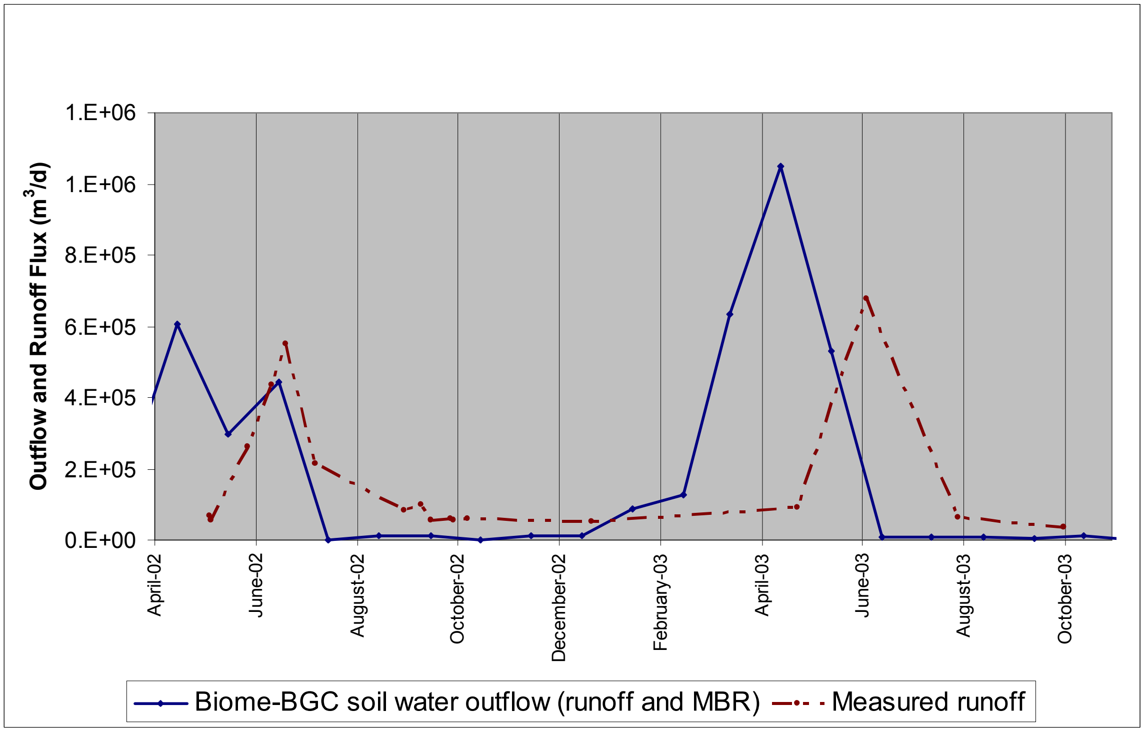
Table 2 shows mountain-block water balance calculations for the 2003 October to September water year. When the annual stream flow (108,000 m^3/d) is subtracted from the soil water outflow then the calculated MBR flux is 99,000 m^3/d or approximately 19% of annual precipitation. We found it necessary to use the October to September water year, rather than the calendar year, to avoid carry-over of snowpack accumulating during the northern autumn into the annual water budget of the subsequent calendar year. The Biome-BGC modeled 2003 water year total annual soil water outflow of 7.58 x 10^7 m^3, which includes runoff and MBR, is approximately equal to the 14 -year mean annual outflow of 7.63 x 10^7 m^3. Considering this, we make the assumption that MBR calculated during the 2003 water year is representative of an average rate for current climatic conditions.

Table 2: Mountain-block water balance 2003 water year.

Water Balance Component *(source in parentheses)*	**Annual Average Flux (m^3/d)**	**% of total water balance**
Precipitation ***(MT-CLIM)***	**532,000**	**100%**
Soil Water Outflow ***(Biome-BGC)***	**207,000**	**39%**
Runoff ***(gaged)***	**108,000**	**20%**
MBR ***(Calculated)***	**99,000**	**19%**

Figure 4 compares monthly Biome-BGC modeled outflow with gaged mountain runoff for the period of runoff record May 2002-September 2003. The modeled snowmelt induced peak in the outflow hydrograph occurs approximately 1-2 months prior to actual peak runoff. The premature timing of the simulated soil water outflow may be an artifact of the coarse resolution of the model grid introducing a bias in the site aspect towards the southwest. The study area is located in the southwest portion of the mountain range and the characteristic accumulation of deeper snowpack and slower melting of snow on north through east facing slopes is presumably lost at the 2.9 km grid size. The shift in the modeled hydrograph may also be partially explained by the lack of model treatment for shallow soil and bedrock flux which presumably delays the yield of a portion of the snowmelt water flux.

Figure 4: Comparison of modeled monthly soil water outflow and measured runoff.



Modeled Soil Water Outflow - Sensitivity to Vegetative and Soil Properties

Soil and vegetation parameters were varied for modeled sites across the orographical climate gradient present in the study area to investigate the sensitivity of modeled outflow to changes in these parameters as a function of landscape position. The sensitivity experiment varied biome type and soil depth (soil depth can be regarded as analogous to available water holding capacity in Biome-BGC's 1-layer soil model) for three sites, one at the mountain front near the piedmont zone, a second site at mid-elevation, and a third site at the mountain crest. Changing the modeled biome type at the low elevation site to either an evergreen shrubland or evergreen forest invokes a 21% reduction in modeled soil water outflow, indicating a significant degree of vegetation control of soil moisture outflow (Table 3). In contrast to this, at the two higher elevation sites outflow is not substantially influenced by the modeled biome type.

Table 3: Soil water outflow sensitivity to modeled biome type.

Modeled biome type	Site physiographic setting (gradient from lower mountain front to mountain crest): Grassland (C3 phenology), elev. 1714 m, 12° slope, west aspect, precip. 406 mm/yr.		Conifer forest, elev. 2370 m, 21° slope, south aspect, precip 690 mm/yr.		Conifer forest, elev. 2783 m, 20° slope, west aspect, precip. 865 mm/yr.	
	1991-2003 mean annual soil water outflow (mm/yr)	% change in outflow	1991-2003 mean annual soil water outflow (mm/yr)	% change in outflow	1991-2003 mean annual soil water outflow (mm/yr)	% change in outflow
Grassland (C3 phenology)	47	*standard*	312	1%	504	-4%
Shrubland	37	-21%	305	-2%	521	-1%
Conifer forest	37	-21%	310	*standard*	525	*standard*

Soil depth sensitivity demonstrates a significant degree of correlation between the aridity of a site and the relative soil water outflow sensitivity to the modeled soil depth (Table 4). Comparison of the range in soil water outflow for each site demonstrates that when sensitivity is expressed as a percentage of the site's total outflow, it is highest at the low elevation, most arid site. In contrast, when sensitivity is expressed as the range in flux magnitude, the mid elevation site is most sensitive to modeled soil depth. At the high elevation site, increasing the modeled soil depth from 16 to 260 cm invokes only a 15% reduction in modeled outflow and a relatively small variability in outflow magnitude compared to the lower sites.

Considering the results of both sensitivity experiments it is apparent that vegetation plays a more critical role in controlling the hydrologic response of the arid, low elevation site. The lack of model sensitivity to the simulated biome type at both the mid and high elevation sites supports the suggestion that snowmelt driven moisture flux is the dominant control on MBR in this environment for elevations above mid-mountain.

Despite the relative influence of vegetation at the low elevation site, soil depth invokes a much greater influence over soil water outflow than does the modeled biome type at all sites. Considering the relative sensitivity to soil depth with elevation, we are able to resolve the water balance with a lower percent uncertainty in those areas producing the highest soil moisture outflow, limiting total uncertainty in the water balance over the entire model domain.

Table 4: Soil water outflow sensitivity to modeled soil depth.

	Site physiographic setting *(gradient from lower mountain front to mountain crest)*					
	Grassland (C3 phenology), elev. 1714 m, 12° slope, west aspect, precip. 406 mm/yr.		Conifer forest, elev. 2370 m, 21° slope, south aspect, precip 690 mm/yr.		Conifer forest, elev. 2783 m, 20° slope, west aspect, precip. 865 mm/yr.	
Modeled soil depth (cm)	1991-2003 mean annual soil water outflow (mm/yr)	% change in outflow	1991-2003 mean annual soil water outflow (mm/yr)	% change in outflow	1991-2003 mean annual soil water outflow (mm/yr)	% change in outflow
16	*174*	*270%*	*397*	*28%*	*572*	*9%*
33	*124*	*163%*	*365*	*18%*	*555*	*6%*
65	*47*	*standard*	*310*	*standard*	*525*	*standard*
130	*12*	*-75%*	*203*	*-35%*	*484*	*-8%*
260	*0*	*-100%*	*51*	*-84%*	*486*	*-7%*

Evaluation of the MBR Estimate

To provide a first-level approach to evaluating our ecohydrologic model water balance approach and resulting estimate of MBR, we compared our estimate of MBR with values reported by other authors (Table 5). These studies represent MBR estimates from mountain ranges that have a fairly similar climate and physiographic setting and are examples of MBR estimated at the mountain-scale. Gannett et al. (2001) is the exception to this and is included here to represent an upper limit of MBR rates (annual precipitation up to 5000 mm/yr and porous volcanic). Gannett et al. (2001) use the Deep Percolation Model (Bauer and Vaccaro, 1987) which calculates the water budget and diffuse recharge for individual cells in an 1829 m raster grid. Our MBR estimate for the southwest portion of the Tobacco Root Mountain Range compares reasonably well with these studies from semiarid settings in the Rocky Mountains. Feth et al. (1966), Huntley (1979), Gannett et al. (2001) as well as our approach all employ a water balance method and as such they do provide a comparison to an entirely unique approach for estimating MBR. Manning et al. (2005) does provide a unique comparison in their integrated modeling approach which uses a combined heat and fluid flow model calibrated to groundwater age and temperature. Their integrated modeling approach provides a well constrained example of MBR in a semiarid mountain setting.

To provide a second level of evaluation of our MBR estimate, the MODFLOW model was used to constrain the possible range of MBR flux. Modeled stream to groundwater exchange is relatively insensitive to MBR because decreases in diffuse bedrock recharge is in part compensated by increases in loss from streams that cross the mountain front alluvial fan and likewise, increases in diffuse bedrock recharge correspond to decreases in stream loss (Table 6).

Based on the MODFLOW water budget, our MBR estimate accounts for 36% of the annual recharge to the basin aquifer while surface water loss from alluvial fan streams represents 40% and aerial recharge from irrigation and precipitation infiltration accounts for 24% of total recharge. Table 6 demonstrates that varying the modeled diffuse bedrock flux by 50% from the base simulation invokes a corresponding change of only 8% in modeled stream loss and 9% in stream gain, both of which are within the error of our synoptic stream gaging measurements. The relative insensitivity of the model to MBR does not allow us to provide useful constraints on MBR and attests to the benefits of using a groundwater age-date calibrated numerical model to provide constraints.

Table 5: Calculated MBR: comparison to published studies.

Study	MBR (% of mean annual precipitation)	Mean annual precipitation (mm/yr)	Method	Location and dominant bedrock geology
this study	19%	887	Ecohydrologic water balance modeling.	Tobacco Root Mountains, Northern Rocky Mountains in Montana, U.S.A. Gneiss/granite.
Feth et al. (1966)	22 %	926	Water balance, incremental precipitation and empirical ET with elevation.	Wasatch Range, Central Rocky Mountains in Utah, U.S.A. Gneiss/schist/minor carbonate.
Huntley (1979)	14 %	not reported	Water balance with ET estimated by analytical equation.	Sangre de Cristo Range, Southern Rocky Mountains in Colorado, U.S.A. Schist/gneiss/granite.
Manning et al. (2005)	7-16 %	1107[1]	Integrated environmental tracer combined with modeling of age calibrated fluid flux and calibrated heat flux.	Wasatch Range, Central Rocky Mountains in Utah, U.S.A. Granite/quartzite-shale/minor carbonate.
Gannett et al. (2001)	up to 70 %	up to 5000	Modeled water balance for individual 1829 m cells using Deep Percolation Model (Bauer and Vaccaro, 1987).	Upper Deschutes Basin, Cascade Range in Oregon, U.S.A. Basaltic/andesitic lava.

1- Precipitation derived from values in Hely et al. (1971)

In one additional MODFLOW simulation, our MBR estimate was applied to the model as focused stream underflow to represent the scenario in which bedrock is impermeable and all mountain front recharge occurs through stream alluvium. In this model run stream loss was 12% greater and stream gain within 1% of the base simulation. Again, the relative insensitivity of the modeled stream flux does allow us to provide useful constraints on the configuration of MBR flowpaths. However, modeled heads were uniformly 2 to 13 m. low across the mountain front alluvial fan except immediately adjacent to the mountain stream valleys where head residuals were 4 to 8 m. high suggesting that considerable recharge occurs through bedrock flowpaths.

Table 6: MODFLOW basin-fill model, sensitivity to MBR. (flux values in m^3/d)

MODFLOW Boundary Description	***MBR diffuse bedrock boundary flux (% change from base simulation)*** -100%	-90%	-50%	0%	50%	100%
MBR (diffuse bedrock flux)	*0*	*5,000*	*25,000*	*51,000*	*76,000*	*101,000*
MBR (focused stream underflow)	*24,000*					
Irrigation loss and basin precipitation (aerial recharge flux)	*51,000*					
Model Response						
Surface water to groundwater flux	*99,000*	*97,000*	*91,000*	*84,000*	*77,000*	*71,000*
Surface water to groundwater, % change	*18%*	*15%*	*8%*	*0%*	*-8%*	*-16%*
Groundwater to surface water flux	*152,000*	*154,000*	*166,000*	*182,000*	*198,000*	*214,000*
Groundwater to surface water, % change	*-17%*	*-15%*	*-9%*	*0%*	*9%*	*18%*

Considerations for Refining Water Balance Mountain-Block Recharge Modeling

Our experience using an ecohydrologic model to estimate MBR suggests that this approach holds promise as a tool for water managers to estimate the component of MBR to basin aquifers. However, the science of using process models to simulate large-scale groundwater recharge has not fully matured. In the following sections we integrate our experience using Biome-BGC with a review of other studies that have incorporated ecohydrologic processes to estimate recharge. We attempt to isolate several of the major obstacles that need to be addressed in improving recharge process models. We discuss the need for independent methods to provide reliable constraints on MBR so that estimates produced by ecohydrologic-modeling approaches can be quantitatively evaluated. We also discuss the need for future research that needs to determine which recharge processes are critical to model at the mountain-scale and to understand how to incorporate small scale processes into a macro-scale model.

Need for Constraints on Mountain-Block Recharge Estimates

Future refinements in modeling of mountain recharge processes are necessary to reduce uncertainty in water balance approaches. However, the practical development of water balance techniques will require reliable estimates of actual MBR as well as techniques to quantify the uncertainty in a water balance.

There are several approaches for assessing MBR discussed in the literature that are independent of the mountain water balance and show promise for providing reliable constraints on MBR rates. Dettinger (1989) developed the chloride-balance technique wherein the percentage of precipitation that becomes groundwater recharge is calculated from the mass-balance of chloride ion concentrations measured in both precipitation and groundwater. To our knowledge, published reports using the chloride-balance technique to estimate recharge have been limited to areas with an arid climate (e.g.: Dettinger, 1989; Maurer and Berger, 1997; Anderholm, 2000; Flint et al., 2002). One potential drawback to chloride-balance interpretation is that the method is susceptible to error caused by chloride concentration representing paleoclimatic recharge rates. The climate in the Northern Rockies has varied considerably in the Holocene and there is mounting evidence indicating that the present climate is in a stage of relatively rapid change suggesting that the chloride-balance technique may not be widely applicable to this environment. Chloride-balance techniques are also susceptible to error from sources chloride other than precipitation.

Sanford (2002) reviews the use of groundwater-age calibrated numerical modeling for estimating recharge rates. Age calibrated numerical models can provide a more precise method to estimate MBR in areas where groundwater flux is predominantly horizontal and where there is minimal mixing with groundwater from recharge sources below the mountain front. Manning and Solomon (2005) present an integrated environmental tracer approach in which recharge elevation is determined using basin groundwater noble gas concentration. They then use an integrated numerical groundwater flow and heat flux model which is calibrated to groundwater age and temperature profiles to compute MBR. In the valley-fill sediments of the Salt Lake Valley, Utah groundwater originating from a mountain block source was found to be less than 20 years old. This work suggests that in this relatively high precipitation mountain environment, mountain to valley flux times can be on the order of several decades. These relatively short mountain recharge residence times indicate that climate variability may not limit the use of environmental tracer and chloride-balance techniques in more humid environments. Mountain-block water balance modeling would benefit from research comparing water balance derived estimates of MBR for a study site with constraints derived from integrated geochemical and physical flow modeling techniques.

Identifying Key Recharge Processes

Soil bypass/macropore flux and lateral routing of surface and subsurface flow have not been incorporated into water balance models for topographically complex terrain, despite the influence of these processes on spatial patterns of soil moisture, ET, and recharge in steep topography. Mountain terrain often contains large areas of shallow soil or bare rock that are likely important areas for localized recharge. Additionally, steep and irregular terrain provides an ideal mechanism wherein overland flow may be captured by bare rock fractures, or collect and infiltrate in localized depressions and flat areas. Several methods for representing soil bypass flow in non-mountainous watersheds are presented in the literature. Finch (1998) incorporates bypass flow into a spatially distributed recharge model by incorporating the algorithm of Rushton and Ward (1979) which defines a threshold magnitude of daily precipitation at which bypass flow occurs. Alternatively, Zhang et al. (1999b) suggest that bypass flow can be accounted for by adjusting the saturated hydraulic conductivity of modeled soils when calibrating a model to soil moisture or runoff data. As an example of this approach, Tague and Band (2001) propose that in their calibrated forest runoff model that soil transmissivity is essentially a tuning parameter which accounts for the actual spatial heterogeneity in soil matrix and macropore flux. However, they demonstrate that using soil transmissivity as a tuning parameter for bypass flow can cause modeled baseflow recession to occur unrealistically fast because soil transmissivity values have to be uniformly high to account for bypass flow. Further research is needed to determine the contribution of localized recharge processes in complex topography and how to best represent bypass flow in a mountain-block recharge model.

Spatially distributed hydrologic modeling in mountainous terrain presents special problems in accounting for the affects of the lateral redistribution of water and the resulting patterns of soil moisture, ET, and recharge. Lateral soil moisture flux, overland flow, and local groundwater systems operate on a small scale that is difficult to characterize at the scale of the mountain range, making an accurate representation of these processes a challenge. Despite these challenges, representative depiction of the spatial differences in antecedent soil water is critical to accurate modeling of diffuse recharge. Our 1-dimensional application of Biome-BGC does not consider either ecosystem or hydrological processes that are connected in horizontal space.

Lateral redistribution of water is addressed in several published watershed-scale recharge studies using SVAT models; however, none of the models used are specifically suited to mountain terrain. Arnold et al. (2000) evaluate recharge at the large scale of the Upper Mississippi River basin (492,000 km^2) using the Soil and Water Assessment Tool (SWAT), a model designed to quantify impacts from land management practices. As implemented by Arnold et al. (2000), SWAT includes sub-models to simulate processes operating in horizontal space including surface runoff, lateral soil moisture flux, and stream baseflow. SWAT handles shallow groundwater as a storage compartment wherein recharge volume is added to shallow groundwater storage and baseflow is routed to the stream network based on a modeler-defined baseflow recession constant.

Another model which appears to handle lateral redistribution well is the quasi 3-dimensional SVAT model TOPOG (Dawes and Hatton, 1993). TOPOG is intended for use in watersheds smaller than 1000 ha and invokes common flow accumulation techniques for overland flow routing and simulates saturated subsurface soil water flux using Darcy's Law. Hatton et al. (1995) use TOPOG to demonstrate that it is critical to use a 3-dimensional model when modeling an environment where there is sufficient precipitation, slope, and soil hydraulic conductivity to allow significant lateral soil moisture and groundwater flux. In contrast to this, Zhang et al. (1999c) and Dawes et al. (1997) use TOPOG to demonstrate surprisingly little lateral soil water flux in a watershed with only 60 m of topographic relief in New South Wales, Australia and these authors suggest that a 1-dimensional model could accurately predict recharge in that environment. Comparison of these studies demonstrates that there is a threshold combination of soil conductivity, terrain steepness, and climate at which 3-dimensional modeling is necessary to capture both lateral moisture flux and an accurate spatial representation of soil moisture and recharge. Specifically, Hatton et al. (1995) supports the need for 3-dimensional modeling to correctly simulate the spatial distribution of soil moisture in environments where steep elevation gradients cause hill slopes to drain towards valley bottoms. Dawes et al. (1997) and Zhang et al. (1999c) demonstrate that this is possible using TOPOG in an intra-annual simulation of a hilly watershed which successfully represented measured temporal and spatial patterns of ET and soil moisture. The handling of lateral soil flux and groundwater storage and release in TOPOG and SWAT shows promising simplicity and accuracy in light of the complex nature of 3-dimensional hillslope hydrologic processes. To date, published studies using both the

SWAT and TOPOG models have been performed in relatively low relief, agricultural watersheds and application of these or similar 3-dimensional models to high relief topography is needed to test how well these 3-dimensional models can approximate mountain hillslope hydrology. In light of the unique environments present in the world's mountain ranges, it seems probable that the special conditions of the mountain environment will demand the development of new techniques to model 3-dimensional recharge processes.

Scaling Recharge Processes to the Model Unit

Improved understandings of process-scale relationships as well as an improved understanding of how to simulate recharge processes at varying spatial scales are precursors to the development of reliable water balance modeling techniques. Testing different modeling approaches at variable scales will in part allow hydrologic scientists to understand how to scale recharge processes to the landscape size needed for management of water resources. Evaluation of modeling techniques for scaling modeled processes in this manner will require comparing model outputs to measurable physical parameters such as soil moisture, runoff, and net primary production.

Wilson and Guan (2004) suggest a two level hierarchy with the hillslope and mountain block as the two essential spatial scales relevant to mountain block hydrology. Examples of hydrologic processes showing heterogeneity at the hillslope-scale include snowpack accumulation and melt, radiation loading and ET, and soil moisture storage (Band et al., 1991). However, owing to lateral redistribution of surface and subsurface water, hillslope position can have a profound affect on soil moisture content (Band et al., 1991; Hatton et al., 1995). Further work is needed to answer the question of how to best model recharge processes operating on sub-hillslope spatial scales. Wilson and Guan (2004) also propose that the mountain-block scale be used to research how recharge is differentiated between surface runoff, local bedrock flowpaths and regional MBR flux. The results of our research demonstrate the effectiveness of estimating MBR to basin aquifers at the mountain-scale.

Another issue needing resolution is how to best incorporate the heterogeneity of the physical attributes of a mountain environment into a model. The relationship of a specific recharge processes' response to physical parameters determines the manner in which spatially heterogeneous parameters must be treated in a model. In general, the linearity of a relationship affects the models ability to produce accurate predictions when simple averaging procedures are used when scaling up processes that are strongly influenced by spatial heterogeneity in the driving parameters. Studies relevant to characterizing parameter-process response relationships include Finch (1998) who uses a water balance model to investigate the sensitivity of normalized recharge response to model input parameters. Finch (1998) demonstrates a predominantly linear relationship between modeled recharge and proportion bypass flow, bypass flow threshold, and vegetation root distribution among soil sub-layers parameters. Additionally, Finch (1998) shows that varying fractional available water content results in a fairly linear relationship except at values less than about 0.15 where recharge increases more rapidly. The linear form of the recharge response to parameter variance suggests that the heterogeneity of these spatial parameters can be represented by mean values in process modeling. In contrast, Finch (1998) demonstrates a relatively high degree of nonlinearity in recharge-LAI and also in recharge-leaf stomatal resistance relationships. The nonlinearity present in the response to these parameters indicates that these parameters may require a statistical model representation.

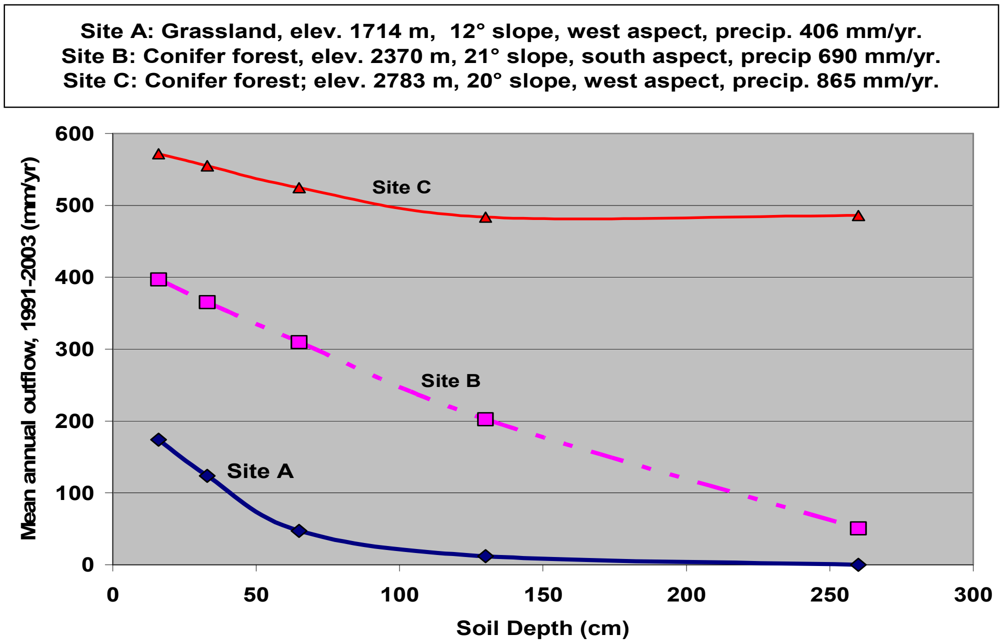
The results of our soil sensitivity experiment demonstrate the possibility for significant non-linearity in the relationship between soil depth (water holding capacity) and soil water outflow. The degree of linearity in the soil depth-outflow relationship is a function of the physiographic position and climatic influences at a particular site (Figure 5). The soil water outflow response demonstrated by the low elevation and comparatively arid Site A suggests a moisture limited environment in which nonlinearity is controlled by a rapid decline in soil water outflow in soils with higher water holding capacity and that are capable of storing the entire winter's soil moisture recharge for summertime evapotranspiration. Opposite of this, the high elevation Site C demonstrates a radiation limited environment in which the annual evapotranspiration in not able consume all of the soil moisture stored in soils 150 cm deep or more. The specific climate at these sites influences the shape of the soil moisture outflow relationship and importantly, climate affects the range of water holding capacity under which the relationship is fairly linear. In between these two extremes, Site B demonstrates an environment in which relative moisture and radiation availability leads to a fairly linear soil depth-outflow relationship and a wide range in outflow magnitude with varying soil depth (Table 4).

Considering the inherent limitations in parameter datasets for mountainous areas, the affects of soil parameter-outflow/recharge nonlinearity has important implications for modeling water flux in mountain soils. High resolution soil maps for mountainous areas are not widely available and the coarse resolution maps that are available, such as the

STATSGO data used in this study, often represent mountain soils as large areas with relatively homogeneous soil properties and the representation of actual soil heterogeneity in these maps is very suspect. Using the spatially averaged soil properties depicted on low resolution soil maps may introduce significant error in spatially distributed recharge modeling where process response to soil parameters is highly non-linear. In light of this, the affect of soil depth on the soil depth-outflow relationship suggests that the response of shallow soils, which are typical of many mountain settings, has a relatively linear form (Figure 5). However, recharge models that use average soil properties in environments with a high standard deviation in soil water holding capacity and affected by climate influence as seen in our results are susceptible to error. Further research is needed to characterize nonlinearity in parameter-response relationships for the physical environments of other mountain ranges to determine whether the parameter-response relationships discussed here have broad application or whether they are specific to the environment studied.

Wood et al. (1988) explains the concept of a representative elementary area (REA) potentially answering some of the questions regarding process scaling. The REA is the process and environment specific threshold scale at which a process can be accurately portrayed with an aggregated representation of the dominant parameter heterogeneity. At scales larger than the REA, sufficient sampling of parameter heterogeneity occurs such that the response of an aggregated-input model is nearly identical to the response of a model scaled to the size of the actual heterogeneity in the input parameters. Using parameter data for the Kings Creek watershed in Kansas, U.S.A, Wood (1995) presents the affects of using variable scales of catchment partitioning on the magnitude of ET and runoff flux derived from a fine-scale spatially distributed water balance model versus a macro-scale model with averaged input parameters. The comparison demonstrates that at a sample size of approximately 1 - 2 km^2 the response of the various levels of catchment partitioning approach common ET and runoff values, suggesting 1 - 2 km^2 as the REA-scale for the dominant parameters controlling both ET and runoff response in that environment. Wood (1995) proposes that soil and topographic heterogeneities occurring at scales of 10^2-10^3 m are dominant in controlling runoff and ET and that the results suggests that the REA-scale for a particular process will be on the order of 1½-3 times the scale of the dominant parameter heterogeneity. For a modeled process scaled larger than the homogeneity of dominant input parameters, defining the REA may hold the key to determining the minimum scale at which we can represent heterogeneous parameters in aggregate form. Although certain landscape components can potentially be treated as spatially homogeneous with a minimum of model bias, a large homogeneous forest stand for example, processes such as bypass flow and localized soil and groundwater flowpaths respond to heterogeneity at relatively small scales. Consideration of the REA-scale for these processes has practical implications for their treatment in a water balance recharge model. It is interesting to note that the dominant heterogeneities evaluated in Wood (1995) occur at the hillslope scale, which supports Wilson and Guan's (2004) proposal of using this as the base scale for understanding mountain recharge processes. Additional research into the REA-scale for specific combinations of parameter heterogeneity and processes response will provide insight into whether the hillslope is the appropriate base scale at which to incorporate processes into models of mountain recharge processes.

Figure 5: Soil depth – soil water outflow linearity.



CONCLUSIONS

Our application of the 1-dimensional Biome-BGC to determine a mountain-scale water balance was not intended to address all of the factors needing consideration in mountain recharge modeling. Instead, we apply Biome-BGC as a first step in evaluating the suitability of ecosystem process modeling for computing a mountain-scale water balance. The MBR estimate from our study does contain a relatively high degree of uncertainty due to nonlinearities in model processes and our use of mean parameter values in the relatively coarse resolution of our model grid. A complete evaluation of the error introduced into our water balance by simplistic 1-dimensional modeling and lack of statistical treatment for spatial heterogeneity in the input parameters would require testing of 3-dimensional models of mountain recharge at variable spatial resolution. Modeled and measured runoff results showed that mountain snowmelt does not scale linearly from the hillslope scale at which snowmelt occurs to the 2.9 km grid size. Advanced techniques for landscape partitioning as well as automated distributed modeling using Biome-BGC algorithms are presented in Band et al. (1991) and White et al. (1998) and we believe that the use of similar partitioning methodology will benefit future research using Biome-BGC for assessing MBR. While 1-dimensional models can give a reasonable approximation of total precipitation interception and ET at the watershed scale, Hatton et al. (1995) shows the inherent limitations of a 1-dimensional representation in their comparison of measured and modeled soil moisture distribution and catchment yield. We expect similar behavior and uncertainty in the modeled soil moisture and outflow in a 1-dimensional application of Biome-BGC.

Despite these shortcomings, our application of Biome-BGC does capture the snowmelt-recharge process that is apparently dominant in the temperate and semiarid climate of many mid-latitude continental mountain ranges. Recharge in this environment occurs predominantly during spring snowmelt when water is abundant, plants are not water stressed, and soil moisture is fairly uniformly saturated. We anticipate that systematic error, as described in Walker et al. (2002), owing to simplifying assumptions made in our parameterization is minimized to a degree because nonlinearity in ET and soil moisture flux are minimized when soils are saturated. Uncertainty in modeled soil moisture is relatively small compared to the quantity of water transmitted from high mountain snowpack suggesting that our model provides a reasonable representation of the dominant snowmelt driven recharge process.

Results of our study show the utility of calculating MBR at the mountain-scale as opposed to the cell-based water balance approach that is used in the other SVAT recharge models reviewed. By calculating the water balance at the mountain-scale we incorporate many of the smaller scale processes that transfer water laterally to streams into the runoff measurement, thereby reducing the total error of the water balance. To elucidate the usefulness of calculating MBR at the mountain-scale, consider the challenges present in simulating hillslope-scale bedrock groundwater flux into a mountain recharge model. Parry et al. (2000) demonstrates that stratigraphy as well as bedrock structure controls the lateral movement and discharge of bedrock groundwater to mountain springs and streams suggesting that in a cell-based water balance approach the fate of small scale bedrock flowpaths must be accurately accounted for. Considering this, the question remains as to how to adequately characterize bedrock hydrogeology over the extent of a mountain range without conducting detailed measurements on each hillslope. By incorporating bedrock groundwater discharge into the runoff component of the water balance we have allowed a simplified water balance, and in our opinion one realistic approach to addressing these issues.

In face of the challenges in describing and modeling mountain hydrology, refining water balance techniques will ultimately depend on our ability to develop methods that simplify complex hydrologic processes while retaining an accurate response in the modeled outcome. Future research is needed in several areas including understanding those processes which control regional MBR at the mountain-scale, parameter-process response nonlinearity, and process scaling such that we can incorporate and accurately scale processes into future mountain water balance models. The large size and relative difficulty in accessing mountain terrain makes calibration by traditional field measurement difficult and expensive to undertake. For purposes of calibrating and validating recharge models, researchers can look to remote sensing techniques capable of describing patterns of soil moisture and ET flux at hillslope resolution. Ecosystem process modeling gives the added benefit that NPP can be calibrated to satellite derived indices of vegetation greenness (Normalized Difference Vegetation Index (NDVI)/Enhanced Vegetation Index (EVI)). Although remote sensing techniques have been touted as a possible panacea for characterizing spatially distributed groundwater recharge, remotely sensed datasets remain essentially a snapshot in time and current methods to remotely measure the water stresses of an ecosystem do not work well for cloudy periods when many recharge processes are most active. This implies that a combination of remote sensing and process modeling will provide the temporal resolution necessary to capture recharge processes in mountainous areas and elsewhere.

The successes of the studies reviewed in this paper demonstrate the potential of process model based water balance approaches. The ecohydrologic approach we describe is a functional method for determining the water balance and regional MBR for mountainous areas with snowmelt dominated recharge where stream flow records are available. It has allowed us to interpret the relative influence of plant versus soil parameters across the climate and physiographic gradient present in our study. Improvements in the understanding of mountain recharge processes and an ability to translate that understanding into models will allow researchers to better quantify and reduce the uncertainty in water balance approaches. Perhaps the development of automated modeling programs that allow the partitioning of landscapes into a desired model unit and model parameterization will provide researchers the efficiency and flexibility needed to evaluate available process modeling approaches and tailor modeling techniques to the specific processes governing mountain hydrology.

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Appendix A: Hydrogeologic and hydrologic data

The hydrogeologic and hydrologic data used in this thesis is a subset of data from the Lower Ruby Valley Groundwater Management Plan and Water Resource Data Report (KirK Environmental, 2004a,b). The data collected for the Lower Ruby Valley Groundwater Management Plan characterize the physical and chemical baseline conditions of groundwater resources in the lower Ruby Valley. The study area for this thesis is a portion of the Lower Ruby Valley encompassing the drainages of Wisconsin, Indian, Mill, Sand, and Ramshorn Creeks from their confluence with the Ruby River to their respective watershed divides in the southwest side of the Tobacco Root Mountains. The hydrogeologic and hydrologic data used in this thesis include: 1) measured water levels, 2) synoptic flows on Wisconsin, Indian, Mill, and Sand Creeks and Leonard Slough on the Ruby River floodplain as well synoptic flows on the Ruby River, 3) mountain runoff measurements and stage readings at staff gages situated on Wisconsin, Indian, Mill, and Ramshorn Creeks near where each creek leaves the bedrock mountain mass and enters the alluvial mountain-front fan, 4) irrigation loss estimates based on irrigation type from a field inventory of irrigation practices, and 5) modeled gravity data. These 5 subsets of data are described below.

1) Measured water levels:

Water levels used in this study were taken from existing domestic, stock, and commercial wells completed in the alluvial aquifer of the Lower Ruby Valley basin fill sediments and adjacent bedrock of the mountain-front. A single complete set of water levels from the period May 22, 2002 – June 20, 2002 and associated equipotential maps was used;

measured water levels are tabulated in the Microsoft Excel spreadsheet *Magruder_2006_waterlevels.xls*. Depth to water was measured with an electric sounder to the hundredth of a foot. Well locations were mapped with a GPS. Ground surface elevations were determined from USGS 30 m. digital elevation model (DEM) or 1:24,000 quadrangle maps. Head was determined by adding the ground surface elevation and water level measuring point height and subtracting the depth of the water level measurement.

2) Synoptic flows:

Synoptic flows in Wisconsin, Indian, Mill, and Sand Creeks and Leonard Slough were measured using the velocity-area method with velocity measured at 0.6 depth (Rantz, 1982). Synoptic flow measurements are included in the Microsoft Excel spreadsheet *Magruder_2006_synoptic_table.xls*. Flows were measured using a standard wading rod and USGS pygmy meter and an Aquacalc 5000 handheld flow meter. With the exception of two synoptic flow sequences completed on Mill Creek, synoptic flows were obtained during the non-irrigation season. Generally, there are too many irrigation diversions and return flows from the streams to make synoptic flow measurements practical or accurate during the late May through September irrigation season. Additional synoptic flows were taken on Mill Creek to describe the transient nature of stream-groundwater flux during relatively higher spring flows on 4/25/02 and during irrigation season on 9/14/02. Because of dangerous stream depths and velocities, no peak runoff data is available. Wisconsin, Indian, and Mill Creeks are losing streams from where they exit the mountain-front to lower on the alluvial fan below the town of Sheridan. In general, stream losses appear to increase with stage and flow on the Sheridan Fan. Sand Creek did not flow at all on the alluvial fan during the May 2002 to October 2003 data collection period (KirK Environmental, 2004a,b). Both Indian Creek and Wisconsin Creek combine to form Leonard Slough on the floodplain of the Ruby River. Wisconsin, Mill, Sand, and Ramshorn Creeks and Leonard Slough are all gaining streams on the floodplain of the Ruby River. Stream gains on the Ruby River Floodplain increase during the irrigation season. Synoptic exchange in unmeasured reaches of Leonard Slough were estimated by assuming that the average flux per mile in the measured reaches of Mill Creek and Leonard Slough was the same as that in the unmeasured reaches of Leonard Slough (see *Magruder_2006_synoptic_table.xls)*. The stream-groundwater exchange in Wisconsin and Indian Creeks and Leonard Slough during higher spring flows and during irrigation season was estimated by assuming that the change in flux from those measured during December 2002 and April 2003 is proportional to the change in flux measured in Mill Creek between the respective dates measured.

Synoptic flows on the Ruby River were accomplished by two methods, 1) reading staff gages on each bridge crossing the river and using stage rating curves developed by the Montana Department of Natural Resources and Conservation (DNRC), and 2) by floating the River and using a standard wading rod and USGS pygmy meter and an Aquacalc 5000 handheld flow meter. For staff gage synoptic readings, where access was available creek inflows between synoptic sites on the Ruby River were measured at locations near their confluence with the Ruby River. Access was not gained to lower Wisconsin Creek. In the synoptic table the flow is estimated where Wisconsin Creek crosses Middle Road by comparing flows taken at the mountain-front on 12/20/02 and 4/25/03 and at Middle Rd on 12/20/03 and assuming that the flow change at Middle Rd in April was proportional to the flow change at the mountain-front gage.

To address problems associated with synoptic calculations based on staff gage readings without control of inflows/diversions between gages, the Ruby River was floated from Harrington Bridge to Seyler Lane by canoe on 9/23 and 9/24/06. This synoptic flow run produced quite different synoptic flow change results that those calculated using staff gage readings. Flows tabulated on the synoptic table spreadsheet indicate that one section of the Ruby River between Harrington and Wheatley Bridge which was losing approximately 5 cfs in April 2003 was gaining 23 cfs in September 2006. Additionally, while the staff gage readings show the Ruby River generally losing over its lower reaches, the measured flows show the river generally gaining water. Reasons for the discrepancy may include the transient nature of groundwater heads and groundwater-surface water exchange. However, the lack of control on synoptic calculations based on staff gage readings, due to unmeasured diversions or inflows to the river, may also affect the April 2003 synoptic calculations.

3) Mountain runoff measurements:

Staff gages were constructed on Wisconsin, Indian, Mill, and Ramshorn Creeks which are perennial at the mountain-front. Sand Creek did not flow at the mountain-front or on the alluvial fan during the May 2002 to October 2003 data collection period of KirK Environmental (2004a,b). Runoff measurements were made according to the same techniques described in the synoptic flow section above. Staff gage readings were taken during field visits when flow measurements were not taken. Stage – discharge relationships were determined for each staff gage by using Microsoft Excel to determine a trend line for graphs of the stage – discharge data. A power function trend line (power = 2.1013) was found to best fit the stage – discharge data from Indian Creek. The remaining creeks were best fit by a second power polynomial equation. It is notable that Indian Creek has a noticeably steeper gradient at the gaging station than do the other three creeks. The formula for each stage – discharge trend line was then used to calculate flows for field visits where a stage reading, but no flow measurement was taken. Peak spring runoff flows had to be estimated because the creeks were not safe to wade due to the high flow conditions. Flow on Indian Creek on 6/6/03 was visually estimated by tossing a piece of wood into the flow and estimating the velocity of the wood and water surface and multiplying this by estimated cross-sectional flow area. Peak flows on Mill and Wisconsin Creeks were estimated by extrapolating the stage-discharge curves beyond measured flows. All measured flows, stage readings, stage – discharge graphs and formulas, calculated flows, and stream hydrographs are included in the Microsoft Excel spreadsheet *Magruder_2006_creek_flows.xls*. Annual runoff volumes for the 2003 water year were calculated by plotting the individual creek hydrographs on graph paper and summing the area under the hydrograph curves.

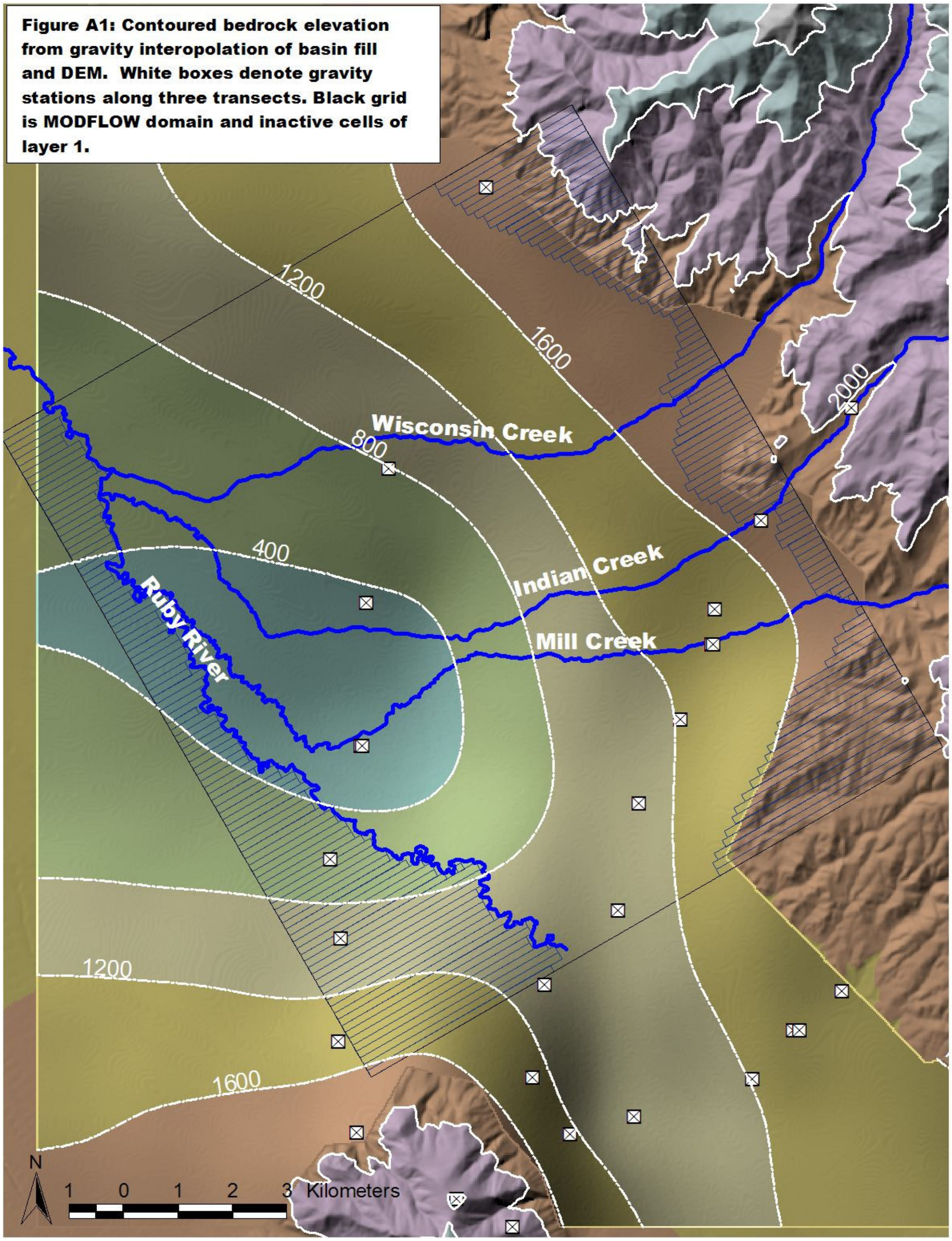
4) Irrigation loss estimates:

Irrigation practices were mapped throughout the Lower Ruby Valley in spring of 2003. Irrigation practices (central pivot, hand/wheel line, flood) were identified in the field and drawn on 1:24.000 scale aerial photo maps. Aerial photo maps were later digitized and attributed using a digitization table and ESRI GIS software. Irrigation efficiency was calculated using the NRCS Farm Irrigation Rating Index computer software. The Farm Irrigation Rating Index computer software was attributed with local soils, Amesha Loam, Crago Gravelly Loam, Kalsted Sandy Loam as suggested by the NRCS. The irrigation efficiency determined for each irrigation type is an average for these three soil types. Annual crop water use was calculated assuming two irrigation applications for a grass/alfalfa mix (50%/50%). Flood irrigation was assumed to have a 2000 ft unlined contour ditch delivery system. Total irrigated acreage, irrigation efficiency, and calculated water loss are tabulated in the Microsoft Excel spreadsheet *Magruder_2006_irrigation_efficiency.xls.* The valley-wide irrigation map was clipped to the area of the active MODFLOW domain (Figure C1) for use in parameterizing the recharge boundary in layer one of the model and is included in the ESRI shapefile *Magruder_2006_irrigation_map.shp*.

5) Gravity Data

KirK Environmental (2004b) presents gravity Data from the Defense Mapping Agency processed to develop the total Bouguer anomaly field for the Lower Ruby Valley. The residual gravity field was calculated for 4 transects of the Lower Ruby Valley assuming linear regional gravity trends. Residual basin depth profiles were modeled using GravCadW (Sheriff, 1997). Three of these GravCadW basin depth profiles within the MODFLOW modeling domain were interpolated to a 30 m. grid using ESRI ArcView Inverse Distance Weighted (IDW) interpolator with a fixed radius of 8 km and a power parameter of 2. The grid was interpolated only over the area of the basin fill alluvium with bedrock contacts attributed as a basin depth of 0. The resulting grid was imported into Golden Software Surfer 8 for smoothing. In Surfer, spline smooth function was used to coarsen the grid to 990 m and a user defined moving average filter with a 3X3 cell window was applied to the 990 m grid. Spline smooth was then used to return the grid to a 30m cell size to match the resolution of the USGS DEM of land surface elevation. The USGS DEM was processed in Surfer in the same fashion to produce a smooth elevation grid. The smoothed basin depth grid was then subtracted from the smoothed DEM resulting in a grid of approximate bedrock elevation for the Lower Ruby River basin (Figure A1).

Figure A1: Contoured bedrock elevation from gravity interpolation.



Appendix B: Geographic datasets, MT-CLIM, and Biome-BGC files

This appendix covers the data sources and processing methods used to attribute the mountain climate simulation program MT-CLIM and ecosystem process model Biome-BGC. The geographic datasets described herein are used to develop the model units and attributes for Biome-BGC. This appendix is divided into three sections which describe 1) geographic datasets and processing, 2) MT-CLIM data and model execution, 3) Biome-BGC attributes and model execution.

1) Georeferenced datasets and processing

The following attributes were needed for ecosystem modeling: latitude, elevation, slope, aspect, biome type, soil volumetric water holding capacity, soil texture, and average annual precipitation. An 8x8 grid with 2868.75 m cells size was determined to best fit the aerial dimensions of the modeled bedrock area. Geographic data sources ranged in resolution from 30 meters to 1 kilometer. Processing of these datasets to arrive at the final 2868.75 m grid size for Biome-BGC modeling was accomplished using scripts written in Arc Avenue used in ESRI Arcview 3.3 software in the following manner:

a. Elevation, slope, and aspect were derived from a 30 m. U.S.G.S DEM. Aspect was reclassified from degrees to cardinal direction.
b. Biome type was derived from 30 m. land cover classification from Thematic Mapper LANDSAT imagery from a satellite flight on 7/22/91.
c. Weighted average soil available water holding capacity and percent clay were derived from 1:250,000 scale U.S.D.A. State Soil Geographic Database

(STATSGO) by using the method described under the section *STATSGO Map Development* on pages 7-13 of USDA (1994). Vector soil maps were gridded at a 30m cell size.

d. Annual average precipitation DAYMET data at 1 km resolution from the University of Montana Numerical Terradynamic Simulation Group (NTSG) was resampled at 30m cell size to facilitate clipping to the study area.

e. The respective datasets were all clipped to the bedrock study area giving cells outside of the study area null values.

f. Either a "blockstat_mean" or "blockstat_majority" command with a neighborhood size of 2868.75 meters was performed by the Arc Avenue script to determine the average or dominant parameter within the area of each cell. Mean values were used for elevation, slope, soil available water content, soil percent clay, and annual average precipitation. Majority values were used for aspect and land cover. In this resampling, cells in which more than half of the sample consists of null values become null values. All data layers were subsequently resampled at a 2868.75 m cell size.

g. The latitude of the centroid of each cell was computed using GIS.

Resampling to the 2868.75 m cell size reduced the number of cells with non-null values to 30. The attributes from these 30 cells were used in climate and ecosystem modeling. Processed 2868.75 m ESRI ArcInfo grids are included on the CD in the folder /Appendix B Data/ESRI ArcInfo Grids. The Excel spreadsheet *attributes_for_BGC_input.xls* included on the CD in the folder /Appendix B Data/Biome-BGC/Inputs contains tabulated attributes from the individual grid cells.

2) MT-CLIM data and model execution

The MT-CLIM version 4.3 mountain climate simulation program requires daily maximum and minimum temperature and precipitation for a base weather station. MT-CLIM uses the average annual precipitation for the base station and for the modeled site to calculate precipitation lapse rates. MT-CLIM also requires the latitude, elevation, slope, and aspect of the modeled site. MT-CLIM was used to derive daily climate for the period 1/1/91 to 12/31/03 for each of the 30 modeled cells. MT-CLIM data was assembled in the following manner:

a. Temperature and precipitation data for the period of record from the U.S.D.A. Natural Resource Conservation Service National Water and Climate Network Lower Twin SNOTEL site was downloaded from http://www.wcc.nrcs.usda.gov/snotel/snotel.pl?sitenum=603&state=mt. This remote automated SNOTEL weather station is located at 2409 m elevation in the Tobacco Root Mountain adjacent to the study area. These data were formatted for MT-CLIM use.
b. The base station annual precipitation isohyet was determined from the 1 km DAYMET data. Site annual precipitation isohyets were determined from the 2868.75 m resampled DAYMET data.
c. Elevation, slope, and aspect were determined for each site from the 2868.75 m model grid. Latitude was determined from the centroid of each modeled cell.

d. Site east and west horizon inclination was not practical to model at the large cell size and was not parameterized.

e. Default MT-CLIM temperature lapse rates were used.

The MT-CLIM initialization .ini files and the output .mtc43 files for all 30 modeled sites and the input .mtcin climate data file for the base station are included on the CD in the folder /Appendix B Data/MT-CLIM.

3) Biome-BGC attributes and model execution

Parameterization of soil properties in Biome-BGC requires soil percent sand, silt, and clay and soil depth. The available STATSGO volumetric water holding capacity data was used to derive soil texture for Biome-BGC. The section of the Biome-BGC code which includes functions that relate soil water potential as a function of volumetric water content to soil texture was provide by Dr. Matt Jolly, a Biome-BGC programmer at NTSG. The following parameters are defined in the Biome-BGC code, and referenced to Cosby et al. (1984) and Saxton et al. (1986), where clay, silt, and sand are given in percent:

i. Slope of log (ψ) versus log (soil relative water content) = soil_b = -(3.10 + 0.157 * clay – 0.003 * sand)

ii. Volumetric water content at saturation = vwc_sat = (50.5 – 0.142 * sand – 0.037 * clay) / 100

iii. Soil matric potential at saturation = psi_sat = - (e $^{((1.54 - 0.0095 * sand + 0.0063 * silt) * log(10))}$ * 9.8 * 10^{-5})

iv. Volumetric water content at field capacity = vwc_fc = vwc_sat * ((-0.015 / psi_sat)$^{(1/soil\_b)}$

These equations were used to back calculate values of percent sand and silt, using STATSGO defined values of percent clay. The soil property calculations made use of the following assumptions:

i. All soils have an assumed depth of 0.65 m.

ii. Volumetric water content was related to volumetric available water holding capacity by the following equation: awhc (m) = (vwc_fc - wilting coefficient) * soil depth (m).

iii. A volumetric water content of 0.06, representative of the wilting coefficient of a loamy sand (ASCE, 1990), was assumed for all modeled soils.

Visual Biome-BGC version 0.69b was used. The 13 year period of climate record was used in a spinup run for each site to create restart files for the actual simulations. These spinup runs used identical parameterization to the actual model runs. Each model run was set to read the site-specific 13 year .mtc43 climate file. The Excel spreadsheet *attributes_for_BGC_input.xls* described under section 1 above was used for site elevation and latitude parameters in Biome-BGC. Biome-BGC requires an ecophysiological constant (.epc) file. Biome-BGC default .epc files for evergreen needleleaf forest, shrub, and C3 Grass were used for sites with LANDSAT Thematic Mapper classified land cover types of conifer forest, dry shrub, and upland grassland respectively. Modeled cell 1 which is in the alpine terrain of the study area has a dominant land cover type of exposed rock. For this exposed rock terrain, the default .epc file for evergreen needleleaf forest was also used, but soil depth was set to $1x10^{-7}$ m and soil texture was set to 100% sand to account for the lack of soil cover. The annual wet and dry nitrogen deposition rate given in the National Atmospheric Deposition Program (NADP) National Trends Network 2002 Annual & Seasonal Data Summary for Site MT07 in Clancy, Montana was used. Annual averages of the seasonal albedo values given in Matthews (1984) were used as described below:

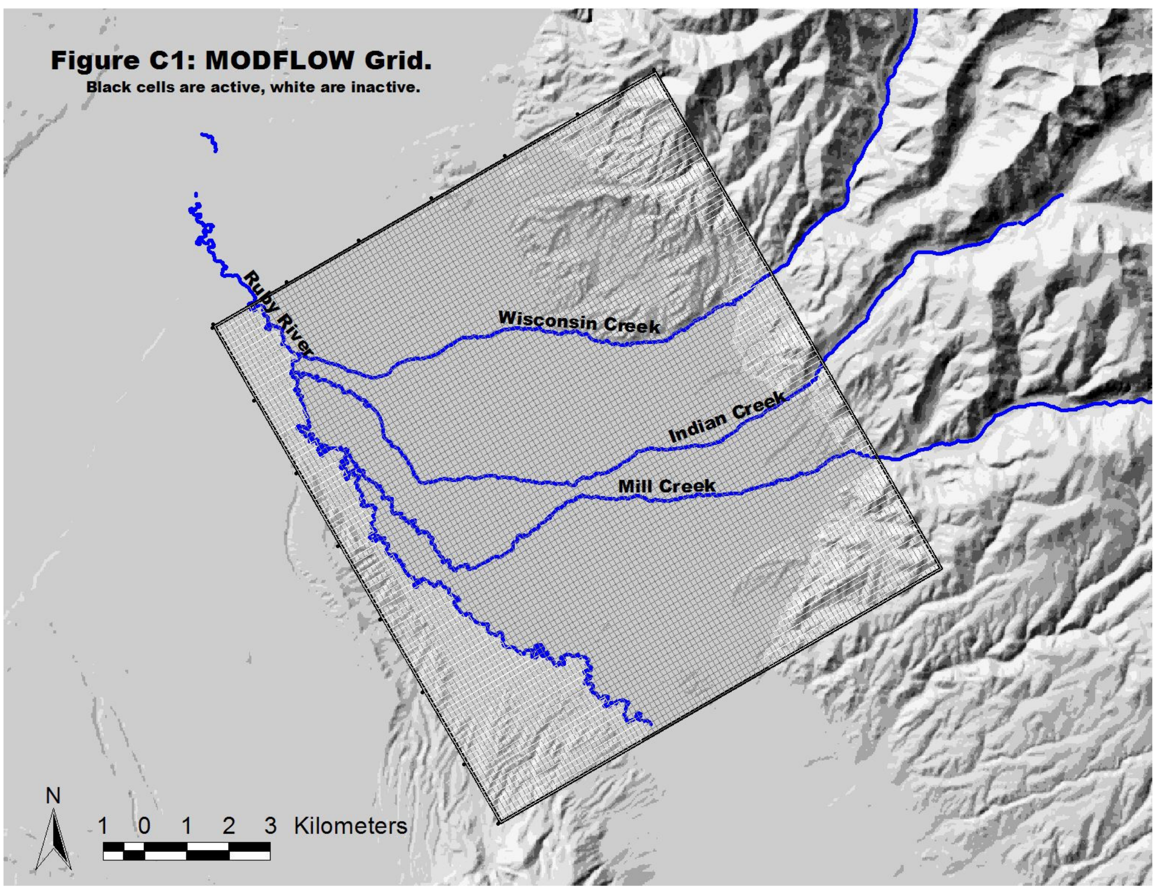
i. Evergreen needleleaved woodland, average value 14.5, was used for all conifer forest biome types.
ii. Tall/medium/short grassland with shrub cover, average value 19.75, was used for dry shrub biome types.
iii. Meadow, short grassland, no woody cover, average value 18.5, was used for upland grassland biome types
iv. Desert, average value 30, was used for exposed rock land cover types.

Default values were used for all other parameters in Biome-BGC. Biome-BGC runs used to test the sensitivity to soil depth and biome type were parameterized as stated previously except for the following changes. In the soil sensitivity test the soil depth in both spinup and model runs was simulated as 16.25 cm, 32.5 cm, 130 cm, and 260 cm deep. In the biome sensitivity test, both spinup and model runs were also simulated with each of the two other dominant land cover types present in the study area, evergreen needleleaf forest, shrub, or C3 Grass, that was not used in the standard model run. All Biome-BGC input files and output files are available on the CD in the folder /Appendix B Data/Biome-BGC/Inputs.

Appendix C: MODFLOW numerical groundwater modeling

This appendix further describes the methods and results of the MODFLOW model that was used to evaluate the mountain-block recharge rate determined in the water balance and associated ecosystem modeling described in the main paper. The goals of the groundwater modeling exercise were 1) to evaluate whether the MBR rate estimated from the mountain water balance is reasonable given constraints on other physical parameters of the basin aquifer system; and 2) to investigate if the model could provide useful constraints on MBR.

Figure C1: MODFLOW Grid.



The aerial configuration of the MODFLOW model grid is shown in Figure C1. Additional details of the model grid are included in the Experiment and Data chapter of the accompanying thesis. Active model cells represent the basin fill alluvium. Model cells overlying bedrock contacts were assigned to be inactive in layer 1. In layers 2-7, inactive cells were assigned corresponding to the interpolated bedrock surface described in Appendix A. The model was run in steady state to simulate the average MBR flux into the basin aquifer. Use of a steady state model is justified in this investigation because the mountain water balance is used to determine average MBR and does not describe potential transient changes in regional bedrock flux. Additionally, a steady state model is sufficient to gauge whether the calculated MBR is a reasonable flux given the available data on the basin aquifer. Lastly, the data needs to describe seasonally transient stream-groundwater exchange conditions in the Lower Ruby Valley would be exceptionally difficult to collect as water is managed for agriculture and stream and ditch flows may vary considerable hour to hour.

The three streams and the Ruby River were simulated using MODFLOW's River Package with stage specified equal to elevation of the USGS 30 m. DEM. The riverbed bottom was assumed to be 1 m. below the elevation of the DEM. Riverbed conductance was adjusted during model calibration. Alluvial underflow in the Ruby River floodplain was simulated by constant head boundaries along equipotential lines interpolated from water level measurements taken during spring 2002 and presented in KirK Environmental (2004b).

Recharge boundaries assigned to layer 1 of the model include the irrigation loss estimates described in Appendix A, aerial recharge from precipitation, and calculated diffuse bedrock MBR along the mountain-front. During model calibration, recharge from irrigation loss was adjusted to half of the calculated values presented in Appendix A to account for the transient nature of irrigation loss in the steady state model (calibration heads are from spring months immediately prior to during initial seasonal irrigation). Gannett et al. (2001) use of the Deep Percolation Model to estimate 5% of annual precipitation becomes aerial recharge to groundwater in the Deschutes Basin. Five percent of mean annual precipitation in Twin Bridges, Montana located in the valley bottom adjacent to the MODFLOW modeled area is 12 mm/yr. Biome-BGC modeling results from this study indicate that soil water outflow for the bedrock areas adjacent to the Sheridan Fan average 47 mm/yr. Using these aerial recharge estimates as a possible range for the Lower Ruby Valley, aerial recharge from precipitation was assumed to be 25 mm/yr for all cells of the MODFLOW model.

MBR was applied at a mountain-front boundary in model layers 1-6. The complete mountain water balance for water year 2003 was parsed to include only those drainages within the MODFLOW model domain (Table C1). The total MBR was divided into two components, the alluvial underflow of mountain streams that enter the basin and diffuse bedrock flux at the mountain-front boundary of the basin model. Alluvial underflow was estimated given available data for the study area (Table C2). Alluvial underflow was simulated using injection wells in cells adjacent to the mountain-front.

Table C1: Water Balance 2003 water year for MODFLOW.

Water Balance Component *(source in parentheses)*	Annual Average Flux (m^3/d)	% of total water balance
Precipitation *(MT-CLIM)*	**423,000**	**100%**
Soil Water Outflow *(Biome-BGC)*	**172,000**	**41%**
Runoff *(gaged)*	**98,000**	**23%**
MBR *(Calculated)*	**75,000**	**18%**
Alluvial Underflow (Estimated)	**24,000**	**6%**
Mountain-front Diffuse Bedrock Flux	**51,000**	**12%**

Table C2: Mountain stream alluvial underflow calculations.

Tributary Name	*Alluvium width (ft* [1]	*Gradient*[2]	*Alluvium depth (ft)*[3]	*Area (ft^2)*[4]	*K (ft/d)*[5]	*Q (m^3/d)*
Mill Crk	400	0.063	45	9000	1200	19379
Indian Crk	333	0.165	30	4995	1	23
Wisconsin Crk	225	0.079	30	3375	600	4505
					Total Alluvial Underflow	23907

1- Airphoto used for measurement.
2- Assumed groundwater gradient is equal to USGS digital elevation model valley slope.
3- Estimated from well logs proximal to the stream valleys.
4- Assuming area = $^1/_2$ width x depth.
5- Estimated from Driscoll (1986) Figure 5.14.

The numerical model was calibrated to the measured water levels and measured groundwater to surface water exchange data described in appendix A. The final calibrated steady state model Visual MODFLOW files are located on the CD in the folder

/Appendix C data/Visual_MODFLOW/ Sheridan_3d_new_K_zones. Measured heads used for calibration are from one set of complete water level measurements obtained between 5/22 and 6/20/02 and are compiled in the Excel spreadsheet *calibration_heads.xls*.

Flux calibration targets are based on the synoptic flow calculations presented in Appendix A. The flux of water between streams and groundwater in the Lower Ruby Valley is highly transient as described in Appendix A and therefore the flux calibration targets include the range of measured and estimated stream-groundwater exchange values (Table C3). The simulated average stream-groundwater flux of the final calibrated model is presented in Table C4.

Table C3: Flux Targets.

	cfs		***m^3/d***	
	Low	High	Low	High
Stream Loss	9	33	22000	81000
Stream Gain	45	109	110000	267000
River Exchange	-19	25	-46000	61000

Table C4: Calibrated Model Flux **(m3/d)**

Stream Loss	81000
Stream Gain	140000
River Loss	3000
River Gain	42000

The resulting head residuals, flow field (Figures C2 and C3) and simulated stream-groundwater exchange of the calibrated model indicate that the MBR estimate is reasonable given available data for the basin aquifer system. To investigate constraints on MBR, the diffuse bedrock flux into the basin model was adjusted by a factor of 0, 0.1, 0.5, 1.5, and 2.0 of the rate calculated from the mountain water balance. The resulting head residual calibrations graphs and head equipotential maps are presented in the CD in the folder /Appendix C data/Calibration Figures. The diffuse bedrock flux was adjusted evenly across the model to maintain a uniform lateral flux rate into the upper 500 m. of the model. As shown in Table 6 of the main body of this thesis, varying the MBR rate by +/-100% corresponds to an 18% change in surface-groundwater flux. Because of the transient nature of stream-groundwater exchange and inaccuracy owing to lack of control of surface water diversions and inflows as well as instrumental error, it was not possible to achieve a high enough level of accuracy in measured flux to constrain the MBR estimate within +/- 100%. The results of the sensitivity experiment demonstrate that it is not possible to provide useful constraints on MBR given the large range in the flux calibration targets and attest to the utility of using groundwater age dating to obtain average long term groundwater flux when attempting to constrain MBR.

Figure C2: Calibrated model head residual graph. (Note: Head residuals shown in graph are interpolated between cell nodes.)

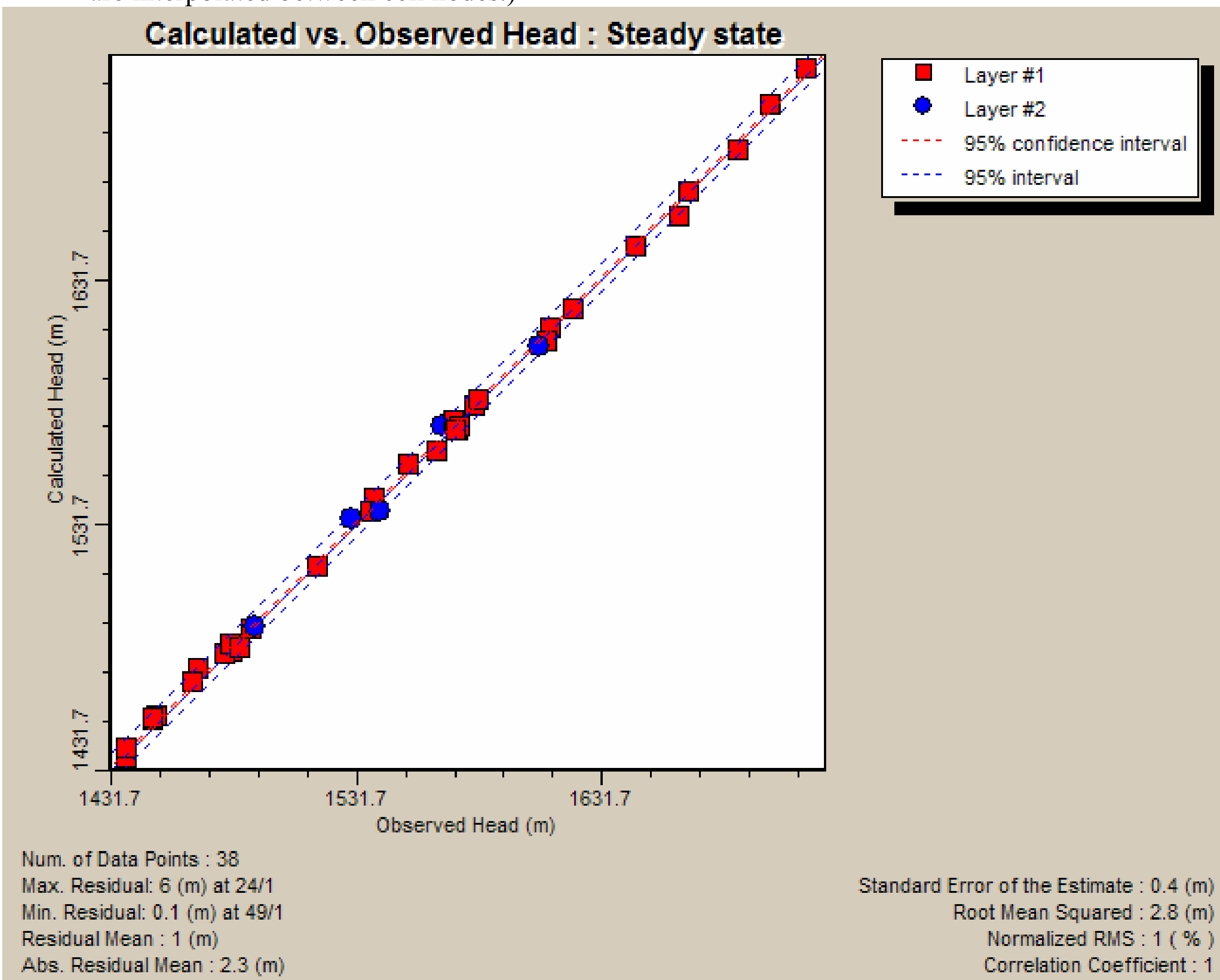


Figure C3: Calibrated model head residuals and equipotential map. (Notes: Head residuals shown on this map are not interpolated between cell nodes. Lighter equipotential lines are from field data; darker lines are modeled.)