State of Washington Water Research Center
Annual Technical Report
FY 2012

Introduction

Since its inception in 1964, the overarching strategic mission of the State of Washington Water Research Center (SWWRC) has been to: i) facilitate, coordinate, conduct, and administer water-related research important to the State of Washington and the region, ii) educate and train engineers, scientists, and other professionals through participation in research and outreach projects, and iii) disseminate information on water-related issues through technical publications, newsletters, reports, sponsorship of seminars, workshops, conferences as well as other outreach and educational activities. While specific emphasis areas have evolved over time, with the competition for water resources continuing to grow, this mission is still vital to the State of Washington today.

The SWWRC has developed a multi-faceted, interdisciplinary approach to accomplish these goals. To promote multidisciplinary research and outreach, the SWWRC has been organized into five program areas: Watershed Management, Groundwater Systems, Environmental Limnology, Vadose Zone Processes, and Outreach and Education. These programs have helped prepare several multidisciplinary research proposals and provide better links between faculty and the SWWRC. These are in addition to the Director's primary research interests in surface-groundwater interaction, remote sensing, and stormwater. The SWWRC is also heavily involved in international research and education activities.

Important lessons learned from the research and outreach components are disseminated to faculty and stakeholders and used by the Director to shape and enhance the education goals. Research projects are also used as a mechanism to fund graduate and undergraduate students as training the next generation of water professionals is an essential role for universities to fill.

The SWWRC is continuing its extensive efforts to reach out to agencies, organizations, and faculty throughout the State. Activities include presentations to watershed groups, discussions with state agencies, participation in regional water quality meetings, and personal contacts. A dynamic web page has been created and is continually updated to share information with stakeholders.

It is within this overall context that the USGS-funded project activities reported in this document must be inserted. These include the internally funded projects as well as the national proposals awarded to the SWWRC. These projects provide a solid core to the diverse efforts of the SWWRC. Water quantity and quality issues continue to be a major concern in the State of Washington due to the endangered species act, population growth, industrial requirements, and agricultural activities. Emerging issues such as water resources management in the face of global warming, water reuse, energy-related water quantity and quality considerations, ecological water demands, the potential renegotiation of the Columbia River Treaty between the US and Canada, and storm water runoff regulations are also beginning to raise concerns.

Research Program Introduction

In accordance with its mission, the SWWRC facilitates, coordinates, conducts, and administers water-related research important to the State of Washington and the region. Research priorities within the 104(b) program for the State of Washington are established by a Joint Scientific Committee which includes representatives from water resource professionals at state agencies, universities, and the local USGS office. The SWWRC supports competitively awarded internal (within the State of Washington) grants involving water projects evaluated by the Joint Scientific Committee. The SWWRC also actively seeks multidisciplinary research at local, state, and national levels. Meetings between stakeholder groups, potential funding agencies, and research faculty are arranged as opportunities arise. Faculty are notified of any opportunities for individual or collaborative endeavors. The SWWRC also submits proposals to various local, state, and federal agencies on its own behalf.

During FY 2012, three locally-relevant research projects were selected for funding by the Center: (1) Response of River Runoff to Black Carbon in Snow and Ice in Washington State, awarded to Susan Kaspari, Assistant Professor, Central Washington University, (2) Climate change, land-water transfer, and in-stream fate of nitrogen in an agricultural setting, awarded to Cailin Orr, Assistant Professor, Washington State University and John Harrison, Associate Professor at WSU-Vancouver, and (3) Progress towards assessing the large-scale impacts of forest fires on runoff erosion across the Pacific Northwest, awarded to Jennifer Adam, Assistant Professor, Washington State University. These projects were competitively awarded based on review and recommendation by the SWWRCs Joint Scientific Board. As described below, these projects address important state issues but are also relevant to national interests.

Response of River Runoff to Black Carbon in Snow and Ice in Washington State

Basic Information

Title:	Response of River Runoff to Black Carbon in Snow and Ice in Washington State
Project Number:	2012WA344B
Start Date:	3/1/2012
End Date:	2/28/2013
Funding Source:	104B
Congressional District:	Washington, 4
Research Category:	Climate and Hydrologic Processes
Focus Category:	Climatological Processes, Hydrology, Management and Planning
Descriptors:	None
Principal Investigators:	Susan Kaspari, Carey Alice Gazis

Publications

1. Delaney, Ian, 2013, Black Carbon Deposition on Snow and Glaciers in Washington State: Implications for Accelerated Snowmelt, MS Thesis, Department of Geological Sciences, Central Washington University, Ellensburg, WA.
2. Kaspari, Susan, Ian Delaney, McKenzie Skiles, Daniel A. Dixon, 2012, Abstract C41D-04, Black Carbon and Dust in Snow and Ice on Snow Dome, Mt. Olympus, American Geophysical Union Fall Meeting, San Francisco, CA.
3. Delaney, Ian, Susan Kaspari, Michael Larrabee, 2012, Abstract C53C-0862. Black Carbon Deposition on Glaciers and in the Seasonal Snowpack in Western Washington s Mountainous Regions, American Geophysical Union Fall Meeting, San Francisco, CA.

I. PROBLEM AND RESEARCH OBJECTIVES

Importance of snow and glacier melt

More than one-sixth of the global population relies on melt water from snow packs and glaciers for their water supply (Barnett et al., 2005). Water derived from snow and glacier melt refills reservoirs and supplies crucial summer flows to rivers used for fisheries, hydropower, irrigation, navigation, recreation and drinking water (Painter et al., 2007) and drives downstream processes such as groundwater recharge and ecological interactions (Bales et al., 2006).

In the Western United States, melt water from mountain regions accounts for more than 70% of annual stream flow. In the Cascade Mountains of Washington State, most of the annual precipitation falls during the winter-spring and is stored in the snowpack (Elsner et al., 2010; Vano et al., 2010). The majority of runoff is derived from the melting snowpack, transferring water from the relatively wet winter season to the typically dry summers (Mote et al., 2005). The timing and availability of water resources is thus strongly related to the duration of mountain snow cover.

Glacier melt water also provides essential water resources in Washington, particularly for watersheds that have a large concentration of glaciers. In some watersheds glacier melt water can account for nearly 50% of the May-September runoff. Glacier melt is variable from year to year, with glacier melt contributing a greater amount of water during years when the snow pack is minimal. Glaciers thus provide an important water resource that can act as a buffer during drought years (Riedel and Larrabee, 2011).

Reduction in seasonal snowpack and glacier retreat in the Cascade Mountains

Spring snowpack levels (snow water equivalent and spatial extent) in the Western United States have declined considerably since the 1950s. The largest decreases occur where winter temperatures are mild, with the Cascade Mountains having experienced some of the largest decreases (as great as 80% decrease since the 1950s). Previous studies suggest that climate change, particularly warming, is the dominant factor inducing earlier snowmelt-fed runoff (Mote et al., 2005). Regions with maritime climates, which have snow season temperatures in the range -5°C to 5°C, are particularly susceptible to warming. Because these regions lie close to 0°C, a slight warming can accelerate the melting rate of the snow pack, and change precipitation from falling in the form of rain rather than snow, preventing water from being stored in the snowpack. This in turn affects the timing and magnitude of water resources available during the comparatively dry summer months. Similar to the snowpack changes, glaciers in Washington State are also retreating. For example, in the North Cascades, glacier area is estimated to have declined ~40% over the past 150 years (Riedel and Larrabee, 2011).

While warming temperatures are a well-recognized factor leading to the reduction in the snowpack and glacier retreat, another cause of accelerated melt is the deposition of impurities onto the snow and glacier surfaces. Snow has the highest albedo (i.e., reflectivity) of any naturally occurring surface on Earth. When impurities such as black carbon (BC, described further below) or dust are present, the snow surface is darkened and snow albedo decreases (Conway et al., 1996; Warren and Wiscombe, 1980), resulting in greater absorption of solar energy and accelerated snow and ice melt (Flanner et al., 2009; Hansen and Nazarenko, 2004; Ramanathan and Carmichael, 2008). This in turn causes peak runoff to occur earlier, reducing water availability during the summer when water demands are highest.

BC (often referred to as soot) is a dark absorptive particle produced by the incomplete combustion of biomass, coal and diesel fuels. In the atmosphere, BC absorbs light and causes atmospheric heating. BC deposited on snow and ice affects climate and water resources by reducing the albedo of snow and ice surfaces and accelerating snow and ice melt (Hansen and Nazarenko, 2004; Ramanathan and Carmichael, 2008). BC has a short residence time in the atmosphere (days to weeks), resulting in regional variations in BC concentrations in the atmosphere and snow/ice. BC emissions have increased globally in recent decades, but emission trends vary regionally. The main sources of BC in the Pacific Northwest are from the transportation sector, residential bio-fuel combustion (primarily wood burning stoves for heating) (Bond et al., 2004), forest fires, and long-range transport from Asia.

The role of absorbing impurities in accelerating snow and glacier melt is an emerging research topic, and few studies have taken place investigating the impacts of absorbing impurities on snow and ice melt in the Pacific Northwest. Two early studies investigated the BC content in snow in Washington. (Grenfell et al., 1981) measured the snow albedo and impurity content in the snowpack in the Cascade Mountains, and determined that impurities were reducing the snow albedo. Similarly, (Clarke and Noone, 1985) collected old and new surface samples of snow on the Olympic peninsula. BC concentrations were slightly lower than those reported by Grenfell. (Conway et al., 1996) applied manufactured BC on the snowpack at Snow Dome on Blue Glacier in the Olympic Mountains, and found that BC applied in high enough concentrations to cause a 30% reduction in albedo increased melting by 50%.

More recently, (Qian et al., 2009) conducted a modeling study that simulated the deposition of BC on snow, and the resulting impact on the snowpack and hydrological cycle in the western United States. Their results suggest that the majority of BC deposited on the snowpack in the Western US is transported from populated metropolitan regions west of the mountains, leading to a decrease in spring snow water equivalent and a shift to earlier peak runoff in the spring. The authors note that more BC-in-snow measurements are necessary in order to improve the accuracy of their models.

Research Objectives

The primary objectives of this study were to further characterize the spatial and temporal variability of BC deposited in Washington snow and glacier ice, and to begin to assess the potential role of BC in accelerating snow and glacier melt.

II. METHODOLOGY

Study Sites

Snow samples were collected from a seasonal snow study site at Blewett Pass, Washington during the winter months of 2013 and from glaciers in the Cascade and Olympic mountains during the spring and summer of 2012 (Figure 1). Kaspari and MS student Matt Jenkins established the Blewett Pass seasonal snow site in 2009, and the snowpack has been sampled every winter since then.

At Blewett Pass during the 2010-2012 winters we had sampled the snowpack at high temporal and spatial resolution. During the winter of 2013 our primary objectives were to characterize surface spatial variability in BC concentrations, dry deposition processes, and interannual variability (by comparing BC concentrations from the winter of 2013 to the prior three winters). To meet these objectives, once or twice a month during the winter we collected snow samples from an established snow pit along with numerous surface samples. Additionally, to investigate BC from dry deposition, surface snow samples were collected daily over a five-day period in February 2013 during which no precipitation fell. MS student Ian Delaney conducted the field sampling at Blewett Pass during the 2012 and 2013 winters.

Seven glaciers were selected for snow sampling during the study period (Figure 1, Table 1). These glaciers were chosen because they are monitored for annual mass balance by the University of Washington (Blue Glacier) and National Park Service (NPS, all other glaciers) and geographically they represent much of the glaciated regions of Washington State.

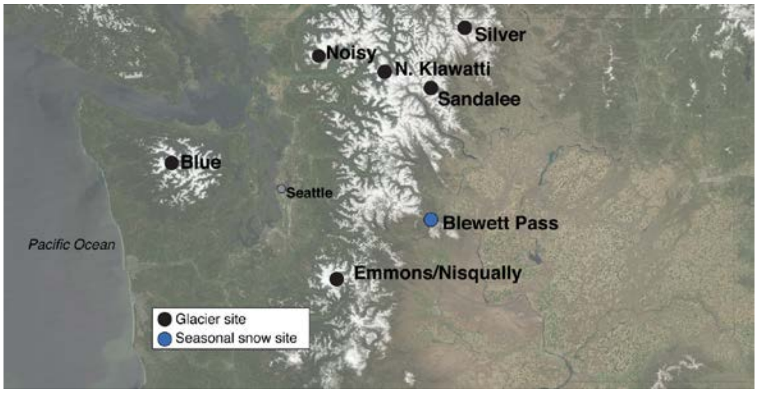


Figure 1. Map of Washington State showing location of sites where snow samples were collected during 2012-2013. Black circles identify glacier sites, and the blue circle identifies the seasonal snow study site at Blewett Pass established in 2009.

Date Sampled	Region	Glacier	Elevation Sampled (m)	Aspect
4/15/12	Rainier	Nisqually	1778-2175	S
4/21/12	Rainier	Emmons	1580-1970	NE
4/23/12	N. Cascades	Sandalee	1996	NE
4/23/12	N. Cascades	N. Klawatti	2080	E
5/10/12	Rainier	Nisqually	2974	S
5/11/12	Rainier	Nisqually	3382	E
6/21/12	Rainier	Emmons	2810-3118	NE
7/19/12	N. Cascades	Noisy	1820-1740	W
7/24/12	N. Cascades	Sandalee	2254-1996	NE
8/24/12	Olympics	Blue	2040	--
8/30/12	N. Cascades	Silver	2290-2075	NW
9/18/12	Rainier	Emmons	2810-3118	NE

Table 1. Date sampled, region, glacier, elevation range, and aspect of snow samples collected from Washington State during the study period. This table does not include sampling conducted at Blewett Pass between December 2012-April 2013.

Sample Collection

Polypropylene gloves were worn at all times during sample collection and care was taken to ensure that clothing fibers did not come in contact with the sample. All samples were collected into either 50 mL polypropylene vials or Whirlpak bags.

At Blewett Pass, snowpits were dug with a shovel, and a clean plexi-glass scraper was used to remove the outer 5 cm of the snowpit wall to provide a clean and uncontaminated surface from which to collect samples. Snowpit samples were collected continuously at 5 cm depth resolution, and numerous spatially distributed surface samples were collected from the top 1-2 cm of the snow surface, where light absorbing impurities (i.e., BC, dust) influence albedo the greatest (Painter et al., 2012b).

At the glacier sites surface snow samples were collected, in addition to subsurface snow samples. MS student Ian Delaney accompanied the National Park Service Glacier Monitoring Program to several of the glacier sites in spring-early summer. The NPS used a manually operated snow-coring device to remove snow cores (up to 4.5 m total depth) for mass-balance purposes. Delaney sampled these snow cores at 10-30 cm increments. During July through September, Delaney sampled snow from shallow (60 cm) snow pits at 5-10 cm resolution. During August of 2012 Kaspari and Delaney collected a 7 m ice core from Snowdome on the Blue Glacier, Mt. Olympus using a 5 cm diameter electromechanical ice drill.

Samples were prevented from melting in the field and during transport to the lab using frozen ice packs or dry ice. Because of the remote location of the study sites, the logistics of maintaining the snow and ice at below freezing temperatures is challenging. In the case of the ice core collected from the Blue Glacier (a 20 mile approach up the Hoh valley), a helicopter was used to fly the ice out to Port Angeles, WA where a freezer was staged. All samples were maintained frozen until prior to analysis. The NPS Glacier Monitoring Group and Bill Baccus at Olympic National Park provided invaluable logistical support in enabling this research to occur.

Sample Analysis

Samples were melted just prior to analysis, sonicated for 20 minutes, and stirred with a magnetic stirrer during analysis. The liquid sample is pumped using a peristaltic pump, nebulized using a CETAC U-5000 AT+ ultrasonic nebulizer, and the resultant dry aerosol is coupled to the sample inlet on a Single Particle Soot Photometer (SP2). The SP2 uses laser-induced incandescence to determine the mass of refractory BC in individual particles (Schwarz et al., 2006; Stephens et al., 2003). Monitoring of liquid sample flow rate pumped into the nebulizer, fraction of liquid sample nebulized and purge airflow rate allows BC mass concentrations in the liquid sample to be determined. Because BC is not nebulized with 100% efficiency, Aquadag standards were used to correct the measured BC concentrations. Nebulization efficiency for the CETAC nebulizer drops at particle sizes greater than 500nm (Schwarz et al., 2012, results confirmed in our laboratory). The concentrations reported herein predominantly represent the mass of BC particles 500 nm and smaller, which corresponds to the size range where the mass absorption cross section of BC particles is greater relative to larger particles, meaning that smaller BC particles absorb light and reduce albedo more efficiently (Schwarz et al., 2013).

To facilitate inter-method comparison, select snow samples were also analyzed for elemental carbon using a Sunset Thermal-Optical Analyzer. This work is still ongoing, thus we don't present results here. Additionally, select snow samples were also filtered through a pre-weighed 0.45μm Millipore filter (Millipore Membrane Filter, Lot no. ROHA46035) using a vacuum pump. The dry mass on the filter is used to determine the total impurity load, which is assumed to largely reflect the dust mass of the sample.

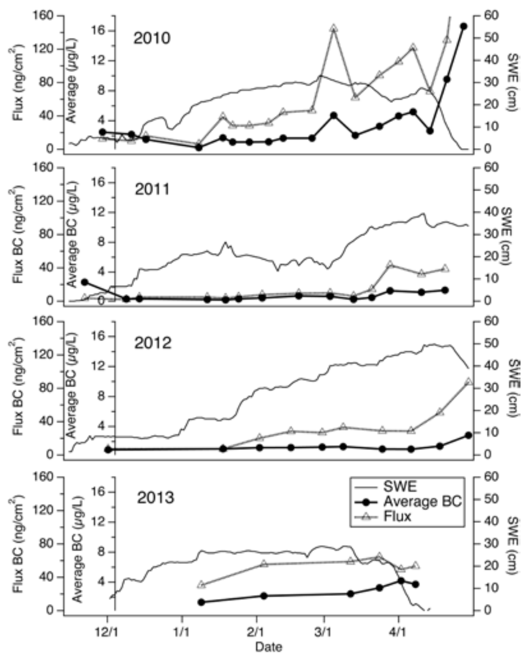


Figure 2. Average BC, BC flux (BC corrected for differences in snow water equivalent (SWE), and SWE from Blewett Pass from the 2010-2013 winters.

III PRINCIPLE FINDINGS AND SIGNIFICANCE

Blewett Pass

Repeat sampling of snowpits at Blewett Pass over the past four winters have shown that BC concentrations are relatively low during the winter accumulation season, and rapidly increase during the spring melt in late March-April (Figure 2). The higher BC concentrations during the spring may be due to higher atmospheric concentrations during spring, or post-depositional melt processes. Analysis of atmospheric BC data from nearby IMPROVE (Interagency Monitoring of Protected Visual Environments) sites indicate that atmospheric BC concentrations rise in the spring-summer relative to the winter (data not shown here), however our results indicate that the higher atmospheric concentrations do not account for the higher BC concentrations observed in the spring snowpack. Rather, the higher BC concentrations during the spring occur primarily due to melt processes that cause BC and dust to become mechanically trapped at the snow surface (Conway et al., 1996; Painter et al., 2012a). The highest BC concentrations at Blewett Pass were observed in surface snow samples, with subsurface concentrations also elevated during the melt period, consistent with results reported from the Sierras (Sterle et al., 2013). Analysis of surface snow samples collected over a five-day dry period in February 2013 demonstrated the role of melt in causing surface enrichment. Higher temperatures on February 13 coincided with 1 cm of snowmelt, and a doubling in surface snow BC concentrations (Figure 3).

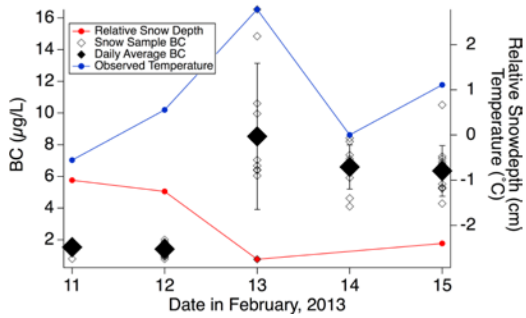


Figure 3. BC concentrations in surface snow samples, snow height and temperature over a five-day period in February 2013.

While BC concentrations were observed to increase during spring each year, average BC concentrations varied interannually. This can be due to differences in BC concentrations in the atmosphere, the processes that integrate BC into the snowpack (dry and wet deposition), or snow accumulation. Regional atmospheric BC data from the IMPROVE network is not yet available from the most recent two winters, thus we don't yet have a means to investigate interannual variations in the atmospheric BC load. Average BC concentrations were highest during 2010 and 2013 (Figure 4), both years with relatively lower snow accumulation at Blewett Pass (Figure 3). We can apply a flux correction, in which the snow water equivalent of the snowpack is multiplied by the average BC concentration in the snowpack, allowing any dilution effect from more or less accumulated snow to be corrected for. That BC Flux is higher in 2010 and 2013 indicates that the higher BC snow concentrations are not due solely to lower snow accumulation. 2013 concentrations were likely higher due to BC related to recent fire activity (discussed below), but we have not yet determined why BC concentrations were elevated during 2010. This is a focus of continued interpretation by MS student Delaney.

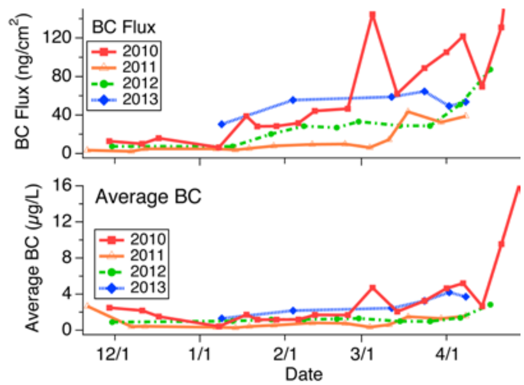


Figure 4. Interannual comparison of BC Flux (BC concentrations corrected for snow accumulation) and Average BC during the 2010-2013 winters.

Two forest fires occurred in the vicinity of Blewett pass in August and September of 2012. The Taylor Bridge fire (August 13 to August 28, 2012; 95 km^2) burned in the Kittitas Valley southeast of Blewett Pass, and the Table Mountain Fire (September 8 to October 5, 2012; 170 km^2) burned in the vicinity of Blewett Pass. These fires left charred material from burned snags that likely introduced an additional and proximal source of black carbon to the snowpack during the 2013 winter (Figure 5), contributing to the elevated BC concentrations during 2013. Spatial sampling in the region surrounding the Table Mountain Fire indicates that BC concentrations in snow are highest in heavily burned areas. This additional source of BC has the potential to remain for years to come.

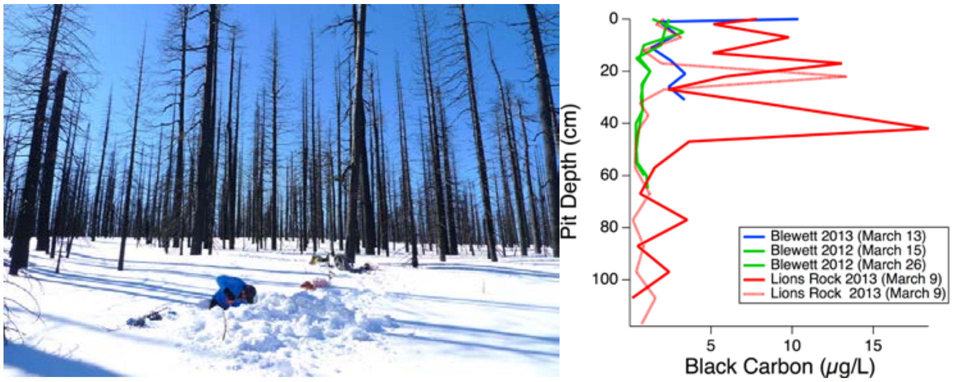


Figure 5. Left: MS student Ian Delaney digging a snowpit in front of burned snags near Lion's Rock. Right: BC concentration in snowpits near Lion's Rock within the Table Mountain burn area, and at Blewett Pass, 9 km away. The highest BC concentrations from the 2013 pits correspond to a relatively dry period during February. Snowpit profiles from 2012 at Blewett Pass are showed for comparison. Note that depths between pits do not correlate to the same time.

Snow Samples from Glaciers in Washington State

We hypothesized that BC concentrations in Washington's snow would show regional differences due to varying proximity to major BC emission sources. However, BC concentrations from subsurface snow samples from around Washington State used as a proxy for BC concentrations in winter snow were fairly uniform (Figure 6). This is due to either relatively uniform BC deposition in snow around the state, or potentially post depositional processes that could move BC to deeper in the snowpack. Previous studies have documented that under strong melt conditions BC can be transported through the snowpack (Xu et al., 2012).

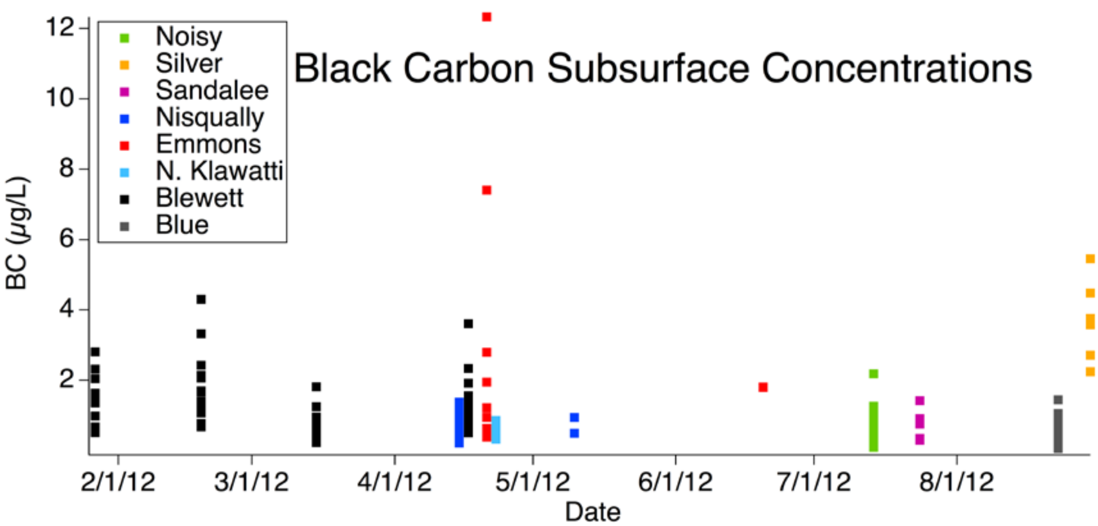


Figure 6. Subsurface BC concentrations, used as a proxy for BC concentrations in winter snow, from the eight study sites.

A significant relationship between elevation and BC concentration does exist, with BC concentrations higher at lower elevations (Figure 7). This is likely due to higher snow accumulation rates at higher elevations that dilute BC in the snowpack, and potentially lower atmospheric BC concentrations at higher altitudes. That BC concentrations are generally lower at higher elevations has been confirmed by analyzing BC data from the IMPROVE network, however the BC concentrations observed in snow are complicated by a combination of wet, dry and post-deposition processes.

Similar to the increase in BC concentrations observed during the melt season at Blewett Pass, surface snow samples from the glacier sites also demonstrate a strong trend towards higher BC concentrations over the summer (Figure 8). Because the glacier sites are at higher elevations (~2000-3000 m, Table 1) relative to Blewett Pass (1295 m), melt commences much later in the year (late summer for high elevation sites relative to spring at Blewett Pass). The higher BC concentrations later in the season are due to both melt-induced enrichment at the glacier surface, and dry deposition during the relatively dry summer months.

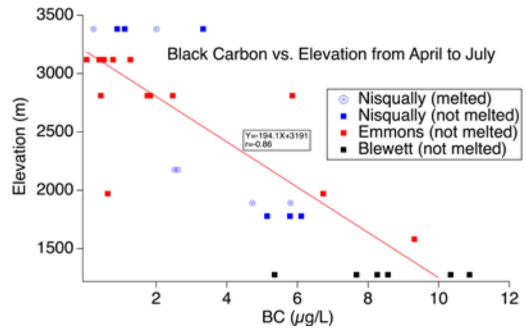


Figure 7. BC concentration vs. Elevation. These samples represent surface snow from prior to the onset of melt at various elevations. A few of the Nisqually snow samples partially melted during transport from the field as noted by 'melted.'

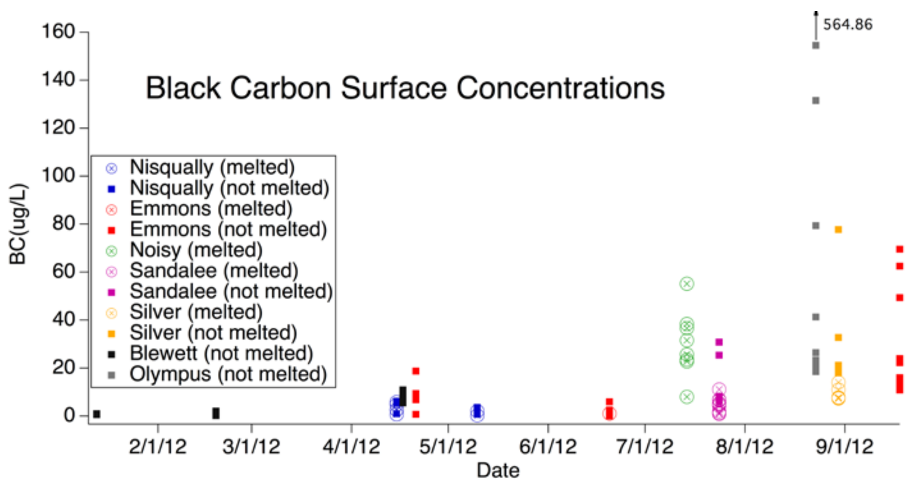
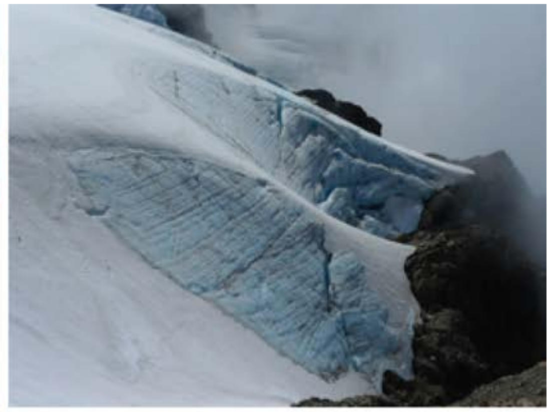


Figure 8. BC concentrations in surface snow during 2012, showing a substantial increasing trend over the summer.

Shallow Ice Core and Surface Snow Samples Retrieved from Snowdome, Mt. Olympus

We initially targeted Snowdome on the Blue glacier of Mt. Olympus as a prime sampling site because it potentially represented background BC concentrations in snow since it was upwind of major regional BC sources in the Puget Sound area, and because Snowdome is a favorable site in regards to glacier dynamics for collecting a shallow ice core. The largest BC sources upwind of Snowdome likely come from ocean shipping traffic, and potentially from trans-Pacific transport of Asian emissions (Hadley et al., 2010). In addition to PI Kaspari and MS student Delaney, PhD student McKenzie Skiles from UCLA/JPL and Dr. Daniel Dixon from the University of Maine participated in fieldwork at Snowdome. Skiles brought a field spectrometer to measure snow surface albedo, and Dixon assisted with ice core drilling. Figure 9 shows pictures from Snowdome fieldwork.

Snowdome, Mt. Olympus. Surface variations in light absorbing impurities are visible on the snow surface.



Side of Snowdome showing distinct layering in light absorbing impurities with depth.

MS student Delaney making measurements on an ice core.

High impurity layer in the ice core corresponding to the 2011 summer layer.

Kaspari operating the electromechanical ice core drill.

Figure 9. Pictures from fieldwork on Snowdome.

Measured BC and dust concentrations from Snowdome surface snow samples that appeared relatively 'clean' and 'dirty' were compared to the spectral albedo as measured using the field spectrometer. The relatively clean snow had lower BC and dust concentrations, corresponding to a maximum albedo of .95 in the visible (Figure 10). BC and dust concentrations were markedly higher in the visibly dirtier snow, which was reflected by a maximum measured albedo of .82 in the visible. We used the Snow, Ice, and Aerosol Radiation (SNICAR) model (Flanner et al., 2007) to estimate albedo reductions due to the measured BC and dust concentrations. The measured and modeled albedos agree well for relatively clean snow, but there was less agreement for the dirtier snow. This may be due to differences in the modeled and actual optical properties of the absorbing impurities in the snowpack. This portion of the study documented that BC and dust concentrations can be highly variable over small spatial scales, and that impurities are present in high enough concentrations to reduce albedo by at least 13%.

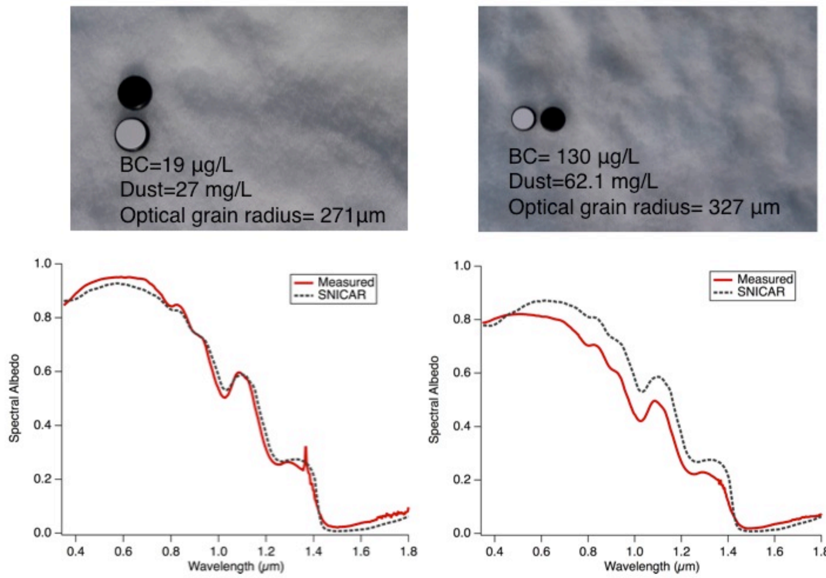


Figure 10. Top: pictures of the snow surface with Spectralon reference panels, and measured BC and dust concentrations. Bottom: Measured and SNICAR modeled spectral albedo. The snow on the left was relatively clean, whereas the snow on the right appeared visibly darker.

The 7.8m shallow ice core collected from Snowdome using an electromechanical drill allowed us to sample through the 2011 summer snow layer at 6.8 m depth (Figure 11). As we observed on other glaciers in Washington, BC concentrations were elevated at the surface of the snowpack. The finding that was not anticipated was the extremely high BC concentration layer (3000 µg/L) found at the 2011 summer layer, which is considerably higher than any other surface snow sample collected during this study. Potential explanations for the formation of this high concentration layer include: 1) Melt and dry deposition during the 2011 summer season. An additional month of melt and dry deposition would have occurred at the 2011 summer surface compared to the surface snow samples that were collected from Snowdome in 2012, or 2) melt and dry deposition during the 2011 summer season combined with percolation through the snow accumulated during the 2012 winter, which coalesced at the 2011 summer horizon. Xu et al. (2012) monitored BC concentrations in the snowpack above the superimposed ice layer on a Tien Shan glacier over a year, and found that relative to freshly fallen snow, during the melt season BC was enriched in the surface snow and to an even greater extent in the snow/firn directly above the superimposed ice layer. Xu et al. propose that meltwater can flush BC through the upper ~1m of the snowpack. This results in lower BC concentrations in the snow

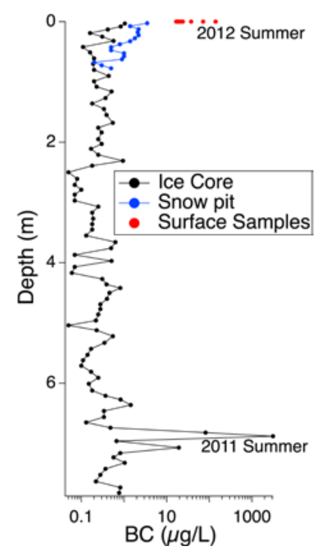


Figure 11. BC concentrations vs. depth from the 7 m snow core retrieved from Snowdome on Mt. Olympus in August 2012.

between the surface and bottom depths of snow/firn, with BC unable to flow below the superimposed ice layer, resulting in build-up of BC at this horizon. The snow accumulation rate at Snowdome was 6.8m (uncorrected for water equivalence), which is much larger than the 1 m Xu et al. discussed. However, there was a clear superimposed ice layer below the 3000 μg/L BC layer that would have prevented BC flow to deeper in the glacier. Lastly, this is one measurement of BC concentrations from the 2011 summer horizon. Due to concerns that our ice core drill would become stuck in the glacier, we were not able to drill additional shallow cores to characterize the spatial variability in the 2011 horizon. The entire 2011 summer horizon potentially may not have BC concentrations as high as 3000 μg/L, but concentrations are likely considerably higher than measured in the surface snow during 2012. The clearly visible horizons of light absorbing impurities on the side of Snowdome indicate that these features are widespread (Figure 9, top right).

Significance

BC concentrations in Washington's winter snowpack were found to be relatively low, with BC concentrations increasing in spring and summer (Figures 2, 4, 6, 8). The timing of the increase in BC at the spatially distributed sites coincided with the onset of melt conditions (i.e., melt commenced earlier at lower elevation sites). In addition to melt resulting in concentration of BC at the snow surface, increased dry deposition associated with an increase in the planetary boundary layer height and minimal precipitation likely contributed to the higher spring-summer BC concentrations. The significance of this finding is that the effects of BC deposition onto the Washington snowpack are greatest during spring to summer. This means that BC induced melt could accelerate the timing of spring snowmelt at lower elevations, however BC induced melt is likely largest at relatively high elevations where the snowpack persists into the summer months when BC concentrations were observed to be highest.

Results from sampling the snowpack at Blewett Pass over four winters demonstrated that BC concentrations are highest during years with low snow accumulation (Figure 2). This has important implications for Washington's snowpack under a changing climate. The winter snowpack is already decreasing (Mote et al., 2005), and is projected to continue to decline in both spatial extent and temporal duration as temperatures continue to increase (Elsner et al., 2010). Our findings indicate that as long as there is not a substantial reduction in BC emissions, a shallower snowpack will result in higher BC concentrations in snow, accelerating snowmelt.

The 2012 forest fires in the vicinity of Blewett Pass likely contributed to the higher BC concentrations observed in the 2013 snowpack, with BC concentrations highest in the vicinity of areas heavily burned (Figures 4, 5). These findings provide insight into an effect of forest fires on the landscape and water resources that has not previously been studied, and could potentially have policy implications for forest fire management practices in regards to prescribed fires vs. wildfires.

The extremely high BC concentration measured in the 2011 summer horizon from Snowdome (Figures 9, 11) suggests that much higher impurity layers may reside below the most recent year's snow accumulation on Washington's glaciers. In a study based on Tibetan glaciers, Xu et al. (2012) noted that due to the coupled impacts of greenhouse-gas warming and BC enrichment in surface snow, dirty ice that can at present form in the accumulation zone underlying the snowpack can be exposed in the future as the glacier equilibrium line altitude (ELA) increases. If the high BC concentration 2011 summer layer was exposed at the glacier surface, albedo would greatly decrease and melt would be substantially accelerated. This is a concern for the future of Washington's snowpack and glaciers, as glacier ELAs will continue to rise and the winter snow accumulation will continue to decrease as temperatures rise.

This study predominantly focused on characterizing the spatial and temporal variability of BC concentrations in Washington's snowpack. In the future, we would like to expand upon this work to characterize other light absorbing impurities (e.g., dust, colored organic material) in the snowpack, and begin to assess the absorption and albedo reduction of BC relative to these other impurities. While BC has the highest efficacy at absorbing radiation, other absorbing impurities (most notably dust) can be present in considerably higher concentrations. To conduct this work, the optical properties of the other light absorbing impurities would need to be characterized. In addition, substantial work is needed to tie the impurity concentrations to albedo reductions and melt rates. As part of this study we documented a 13% reduction in visible albedo due to the presence of absorbing impurities (Figure 10), but more detailed work is needed.

REFERENCES

Bales, R., Molotch, N., Painter, T. H., Dettinger, M. D., Rice, R., and Dozier, J.: Mountain Hydrology of the western United States, Water Resources Research, 42, 10.1029/2005WR004387, 2006.

Barnett, T. P., Adam, J. C., and Lettenmaier, D. P.: Potential impacts of a warming climate on water availability in snow-dominated regions, Nature, 438, 303-309, 10.1038/nature04141, 2005.

Bond, T. C., Streets, D. G., Yarber, K. F., Nelson, S. M., Woo, J. H., and Klimont, Z.: A technology-based global inventory of black and organic carbon emissions from combustion, Journal of Geophysical Research-Atmospheres, 109, 2004.

Clarke, A. D., and Noone, K. J.: Soot in the Arctic Snowpack: A cause for perturbations in radiative transfer, Atmospheric Environment, 19, 2045-2053, 1985.

Conway, H., Gades, A., and Raymond, C. F.: Albedo of dirty snow during conditions of melt, Water Resources Research, 32, 1713-1718, 1996.

Elsner, M. M., Cuo, L., Voisin, N., Deems, J. S., Hamlet, A. F., Vano, J. A., Mickelson, K. E. B., Lee, S. Y., and Lettenmaier, D. P.: Implications of 21st century climate change for the hydrology of Washington State, Climatic Change, 102, 225-260, 10.1007/s10584-010-9855-0, 2010.

Flanner, M. G., Zender, C. S., Randerson, J. T., and Rasch, P. J.: Present-day climate forcing and response from black carbon in snow, Journal of Geophysical Research, 112, doi:10.1029/2006JD008003, 2007.

Flanner, M. G., Zender, C. S., Hess, P. G., Mahowald, N., Painter, T. H., Ramanathan, V., and Rasch, P. J.: Springtime warming and reduced snow cover from carbonaceous particles, Atmospheric Chemistry and Physics, 9, 2481-2497, 2009.

Grenfell, T. C., Perovich, D. K., and Ogren, J. A.: Spectral albedos of an alpine snowpack, Cold Regions Science and Technology, 4, 121-127, 1981.

Hadley, O. L., Corrigan, C. E., Kirchstetter, T. W., Cliff, S. S., and Ramanathan, V.: Measured black carbon deposition on the Sierra Nevada snow pack and implication for snow pack retreat, ATmospheric Chemistry and Physics, 10, 7505-7513, 10.5194/acp-10-7505-2010, 2010.

Hansen, J., and Nazarenko, L.: Soot climate forcing via snow and ice albedos, Proceedings of the National Academy of Sciences, 101, 423-428, 2004.

Mote, P. W., Hamlet, A. F., Clark, M. P., and Lettenmaier, D. P.: Declining mountain snowpack in western north America, Bulletin of the American Meteorological Society, 86, 39-+, 10.1175/bams-86-1-39, 2005.

Painter, T., Barrett, A. P., Landry, C. C., Neff, J. C., Cassidy, M. P., Lawrence, C. R., McBride, K. E., and Farmer, G. L.: Impact of disturbed desert soils on duration of mountain snow cover, Geophysical Research Letters, doi:10.1029/2007GL030284R., 2007.

Painter, T., Skiles, S. M., Deems, J., Bryant, A., and Landry, C.: Dust radiative forcing in snow of the Upper Colorado River Basin: 1. A 6 year record of energy balance, radiation, and dust concentrations, Water Resources Research, 48, 10.1029/2012WR011986, 2012a.

Painter, T. H., Skiles, S. M. K., Deems, J. S., Bryant, A. C., and Landry, C. C.: Dust radiative forcing in snow of the Upper Colorado River Basin: 1. A 6 year record of energy balance, radiation, and dust concentrations, Water Resources Research, 48, W07521, 2012b.

Qian, Y., Gustafson, W., Leung, L. R., and Ghan, S.: Effects of soot-induced snow albedo change on snowpack and hydrological cycle in western United States based on Weather Research and

Forecasting chemistry and regional climate simulations, Journal of Geophysical Research, 114, doi:10.1029/2008JD011039, 2009.
Ramanathan, V., and Carmichael, G. R.: Global and regional climate changes due to black carbon, Nature Geoscience, 1, 221-227, 2008.
Riedel, J., and Larrabee, M. A.: North Cascades National Park Complex glacier mass balance monitoring annual report, water year 2009: North Coast and Cascades Network, National Park Service, Fort Collins, 2011.
Schwarz, J. P., Doherty, S., Li, F., Ruggiero, S. T., Tanner, C. E., Perring, A. E., Gao, R. S., and Fahey, D. W.: Assessing recent measurement techniques for quantifying black carbon concentration in snow, Atmospheric Measurement Techniques Discussions, 5, 3771-3795, 2012.
Schwarz, J. P., Gao, R. S., Perring, A. E., Spackman, J. R., and Fahey, D. W.: Black carbon aerosol size in snow, Nature Scientific Reports, 3, 1356, doi:10.1038/srep01356, 2013.
Sterle, K. M., McConnell, J., Dozier, J., Edwards, R. L., and Flanner, M. G.: Retention and radiative forcing of black carbon in eastern Sierra Nevada snow, The Cryosphere, 7, 365-374, 2013.
Vano, J. A., Scott, M. J., Voisin, N., Stockle, C. O., Hamlet, A. F., Mickelson, K. E. B., Elsner, M. M., and Lettenmaier, D. P.: Climate change impacts on water management and irrigated agriculture in the Yakima River Basin, Washington, USA, Climatic Change, 102, 287-317, 10.1007/s10584-010-9856-z, 2010.
Warren, S., and Wiscombe, W.: A model for the spectral albedo of snow II. Snow containing atmospheric aerosols, Journal of Atmospheric Sciences, 37, 1980.
Xu, B., Cao, J., Joswiak, D., Liu, X., Zhao, H., and He, J.: Post-depositional enrichment of black soot in snow-pack and accelerated melting of Tibetan glaciers, Environmental Research Letters, 7, 014022, 2012.

Climate change, land-water transfer, and in-stream fate of nitrogen in an agricultural setting

Basic Information

Title:	Climate change, land-water transfer, and in-stream fate of nitrogen in an agricultural setting
Project Number:	2012WA346B
Start Date:	3/1/2012
End Date:	2/28/2012
Funding Source:	104B
Congressional District:	3rd
Research Category:	Water Quality
Focus Category:	Hydrogeochemistry, Solute Transport, Water Quality
Descriptors:	None
Principal Investigators:	Cailin Huyck Orr, John Harrison

Publications

1. Schwarz, J.E., C.H. Orr, C.K. Keller, C.J. Kelley, J.L. Smith, Magnitude and Timing of Stream Nitrate Export in a Semi-Arid Dryland Agricultural Watershed. (In Preparation)
2. Martin, R.A., J.A. Harrison, and J.A. Needoba, Hydrologic control of dissolved organic matter quality in a semi-arid agricultural catchment. (In Preparation)
3. Schwarz, J.E., 2013, Magnitude and Timing of Stream Nitrate Export in a Semi-Arid Dryland Agricultural Watershed, MS Dissertation, Geology, School of Earth & Environmental Sciences, Washington State University, Pullman, WA.
4. Orr, C.H., L. Moon-Neilsen, J. Schwarz, C.J. Kelley. C.K. Keller, 2013, Seasonality in flux of nitrogen to streams and atmosphere in a semi-arid and dryland agriculture dominated watershed, Ecological Society of America, Minneapolis, MN.
5. Moon-Neilson, L., T.T., Brown, C.J.Kelley, R.A. Martin, S.Waldo, 2013, Budget Approach Provides Insights into Transformations and Fate of Nitrogen Applied to an Eastern Washington Agroecosystem, NSF IGERT Poster and Video Competition.
6. Martin, R. A., and J. A. Harrison, 2012, Hydrologic variability controls the quantity and quality of dissolved organic carbon and nitrogen exported from agricultural soils, poster, WSU Vancouver Research Showcase, Vancouver WA.
7. Martin RA and JA Harrison, 2012,Flow paths control the delivery of dissolved organic carbon and nitrogen from agricultural soils to surface water, poster, Washington State University Academic Showcase, Pullman, WA.

RESEARCH PROBLEM AND OBJECTIVES

In recent decades, climate change has significantly altered Washington precipitation and streamflow (Fu et al., 2010), and climate effects on hydrologic fluxes are predicted to intensify in coming decades (Mote and Salathé, 2010). Two robust historic and projected trends in eastern Washington precipitation include increasing fall precipitation and increases in heavy precipitation events (Meehl et al., 2005, Salathe et al., 2010). In the Palouse, these changes, along with higher predicted winter temperatures (leading to more rain, less snow) are likely to intensify pulsed hydrologic events (rainstorms and associated runoff) that carry solutes and sediments into streams. The effects of such changes on water fluxes and quality are certain to be important but are currently poorly understood.

Soluble, reactive nutrients such as nitrate (NO_3^-) and dissolved organic matter (DOM) are of particular concern. NO_3^- at high concentrations is a contaminant regulated by the Clean Water Act and is a common pollutant in agricultural watersheds. While DOM can be an important source of carbon and nitrogen to stream ecosystems at low levels, both excess NO_3^- and DOM can lead to poor water quality, contributing to low dissolved oxygen conditions (Jassby and Van Nieuwenhuyse, 2005), harmful algal blooms (Glibert et al., 2010), and drinking water contamination (Chow et al., 2007; Thouin et al., 2009). NO_3^- and DOM-related water quality problems are widespread in Washington, with over 700 water bodies currently listed as impaired with respect to dissolved oxygen (WA DoE), and more than 40 sites listed as impaired by high total phosphorus or total nitrogen levels (Washington DoE 2008).

Agricultural land is an important source of NO_3^-, dissolved organic carbon (DOC) and nitrogen (DON), to surface water (Royer et al., 2006, Warrner et al., 2009, Pellerin et al., 2006). Agriculturally-derived DOC and DON can be more bioavailable than DOC and DON derived from natural systems (Wiegner and Seitzinger, 2004, Warrner et al., 2009), and bioavailable DOC may stimulate in-stream nitrogen removal via denitrification in these NO_3^--rich streams (Jansson et al., 1994). Hence, Washington's more than 2.4 million hectares of cultivated land (USDA/NRCS) are likely to have a strong influence on the state's surface water quality, and understanding the controls over NO_3^-, DOC and DON loss from these systems is critical for understanding and anticipating the effects of climate change on water quality. Current understanding of dissolved nitrogen dynamics is based almost entirely on warm-season studies of transport and in-stream processing. However, in the semi-arid region of eastern Washington, higher precipitation and discharge occur in the cooler winter months, causing the majority of NO_3^- and DOM mass flux from watersheds to occur during periods that have thus far been largely unstudied.

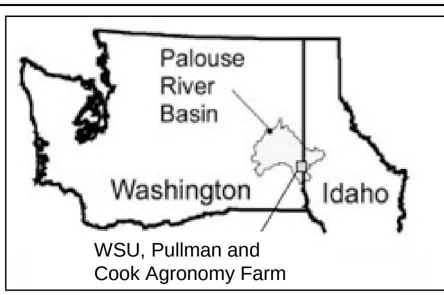


Figure 1. Location of Cook Agronomy Farm at Washington State University, after Keller et al. 2008.

Both NO_3^- and DOM are likely to be quite sensitive to changes in the frequency and intensity of hydrologic pulse events. Such events not only increase water discharged from soils to streams, but are also often associated with increased in-stream NO_3^- and DOM concentrations (e.g. Harrison and Matson, 2003; Martin and Harrison, 2011), such that high flow events of relatively short duration (days to weeks) can account for the majority of annual DOM and NO_3^- flux through streams (e.g. Inamdar et al., 2006, Bernal et al., 2005). Increased DOM and NO_3^- concentrations during hydrologic pulses can result from many factors, including: 1) changing hydrologic flow paths that mobilize novel sources of NO_3^- (e.g. Tiemeyer et al., 2008) and DOM (e.g. Hagedorn et al., 2000, Martin and Harrison 2011), and 2) the bypassing of environments (e.g. wetlands or riparian zones) or processes (e.g. plant or microbial uptake) which would otherwise retain these compounds. Much remains to be learned about interactions between hydrologic pulse events and land-to-water transport of DOM. Similarly, the relationship between such pulse events and in-stream DOM and NO_3^- processing is poorly characterized, especially in agricultural settings, and interactions between delivery of DOM and in-stream NO_3^- and DOM processing are almost completely uncharacterized despite the fact that they are likely to be important, and of increasing importance as the frequency and intensity of pulse events increases with climate change.

It is in this context that we proposed work that would enhance the ability of scientists, policy makers and stakeholders to understand, predict, respond to and/or mitigate climate change impacts on water quality. Specifically, we proposed to use WSU's Cook Agronomy Farm (Figure 1) as a study system to: 1) understand how hydrologic variability affects a) nitrate and DOM transport from agricultural fields to surface water and b) in-stream fate of nitrogen, and 2) use this information to develop, apply, test, and iteratively refine a model that utilizes a dynamic representation of hydrologic flow paths and organic matter source pools to predict terrestrial-to-aquatic NO_3^- and DOM transport, under current and anticipated future climate.

Specifically, we addressed the following questions:

- **Q1.) How do hydrologic conditions affect nitrate and DOM transport from agricultural land to surface drainage waters?**
- **Q2.) How do hydrologic conditions affect in-stream nitrogen retention?**

METHODOLOGY

Soil water and tile drain sampling

Nests of shallow and deep wells and lysimeters were installed along a transect that approximated the buried tile drain route and upslope of the tile line. Shallow and deep lysimeters were placed at 0.5 and 1 m depths, respectively, and shallow and deep wells were screened at depths above and below an argyllic layer, respectively, that is approximately 1 meter below the surface and intermittent across the basin.

Soil water, tile drain, and surface water samples were collected weekly to bi-weekly during the 2012 water year, and tile drain samples were collected more frequently during high flow events to capture temporal variability associated with rapid changes in discharge. During the dry season, most shallow lysimeters and wells did not yield water samples. Specific electrical conductivity was measured in the field using an Orion Model 115 with Conductivity Cell 014016 probe. Samples for DOM, nutrients, and major cations were filtered in the field, transported on ice to the laboratory, and frozen until analysis. Samples to be analyzed for absorbance and fluorescence spectra were stored at $4^{o}C$ and analyzed within 5 days. Discharge from the tile drain was monitored every 15 minutes in a receiving flume equipped with a pressure transducer, and electrical conductivity was measured simultaneously with a Campbell Scientific Temperature and Conductivity probe (CS547A-L).

Surface Water Monitoring

Surface water has been sampled at 3 locations along Missouri Flat Creek since 2000 to capture nitrate export dynamics at a range of catchment sizes [660 (site 1), 3800 (site 2), 6300 (site 3) ha; the smaller catchments are nested within the larger catchments]. Samples were taken approximately twice a month all three sites, with more frequent sampling during the winter/spring runoff period. Stream discharge has been monitored at sites 1 (2000-present) and 2 (2000-2010) using digital pressure transducers in combination with rating curves. Discharge at site 3 was measured with digital pressure transducer prior to 2004 and modeled for years 2005-2010 based on discharge data at site 2 during that period and the relationship between discharge at sites 2 and 3 prior to 2004.

Laboratory Nutrient and Dissolved Organic Matter Analyses

Samples were analyzed for nitrate and ammonium according to the standard EPA methods (353.2 and 350.1, respectively) using a continuous flow analyzer (Model RFA300, Alpkem/OI Analytical) or discrete nutrient analyzer (WestCo Smartchem) and for dissolved organic carbon (DOC) and total dissolved nitrogen (TDN) on a Lachat TOC-TN analyzer (IL 550 TON-TN) equipped with electrochemical (ECD) NO and non-dispersive infrared absorption (NDIR) detectors. Dissolved organic nitrogen (DON) was calculated as the difference between TDN and total inorganic nitrogen (NO_3^- + NH_4^+). Samples were analyzed for major cations (Na^+, K^+, Mg^{2+}, Ca^{2+}) using a Shimadzu atomic absorption emission spectrophotometer (model AA-6601F). Absorbance spectra of bulk DOM samples were analyzed using a J&M TIDAS spectrophotometer (World Precision Instruments) from 200-700 nm. Fluorescence excitation-emission matrices (ex 240-450 nm, em 300-600 nm) were generated with a Horiba Jobin Yvon Flouromax-4 spectrofluorometer. Corrections for the instrument, water matrix (background), internal absorbance, and Raman signal were applied to fluorescence data to permit comparisons across sampling dates.

Estimating DOM and nitrate export via tile drainage

Annual DOC, DON, and nitrate loads were estimated by assuming that concentrations were constant from halfway between the previous measurement to halfway to the subsequent measurement. Concentrations were multiplied by water discharge, and summed for the year. The hydrograph was then characterized as either baseflow or event-flow, and export during each of these conditions was calculated. An event were defined as the time from the start of rapid discharge increase to the time when steady base flow returned.

DOM Export Modeling Approach

The spatial and temporal variability of the soil water data was too great to constrain shallow and deep soil water end members and meaningfully estimate the contribution of each flow path to tile drain discharge over time, with spatio-temporal variability of individual end members exceeding temporal variability of tile drain chemistry. Despite this, a mixing model was developed to qualitatively assess whether our hypotheses were broadly supported, whether various soil water compartments may be contributing preferentially to tile drain discharge over time (e.g. due to varying hillslope hydrologic connectivity), or other processes might be important for controlling DOM export via the tile drain (e.g. DOM removal via adsorption or decomposition). Further analyses, including sensitivity of DOM export to changing climate, were not conducted because we could not obtain statistically meaningful results.

Shallow and deep soil water end members were characterized by both specific electrical conductivity (EC) and magnesium concentrations ([Mg]), which were significantly correlated with [DOC] in tile drain discharge ($p < 0.001$ for both relationships). Median EC and [Mg] values for each sampling instrument during dry and wet seasons were calculated; medians were then averaged to define the end members. During the wet season (Jan through May), EC was significantly greater in deep than shallow soil water (T-test, one-tailed, df = 9, $p < 0.05$), and [Mg] was borderline significantly greater in deep than shallow soil water (T-test, one-tailed, df = 9, $p = 0.05$) (Table 1). One shallow sampler yielded soil water during the dry season, and the median EC and [Mg] values of this sampler were within the range defined by the mean +/- 2 sd of the deep samplers during the dry period. Additionally, EC and [Mg] values for tile drain samples were outside the end-member range during the dry season, so further assessment of dry-season dynamics was not conducted.

For tile drain sampling time points during the wet season, the fractions of discharge derived from shallow and deep soil water were calculated using a linear mixing model with shallow and deep end members defined by EC and [Mg], yielding two estimates for shallow and deep fractions per sampling time, which were then averaged. Average end member fractions were used to predict [DOC] and [DON] for each sampling time, and Nash-Sutcliffe Efficiency and model error in relation to tile drain discharge were examined to assess model performance.

Nitrate fluxes in nested catchments

Nitrate mass discharge (nitrate-N mass/time) and fluxes (nitrate-N mass/area/time) were calculated for the period 2000-2010 by dividing nitrate-N concentrations by water discharge (nitrate-N mass discharge) and dividing the nitrate-N mass discharge by watershed area (nitrate-N flux). At least one nitrate measurement per month and eight months per year were required to generate annual nitrate loads and fluxes for each location. Since the majority of nitrate is exported during high flows, low estimates for nitrate loads and fluxes during some years may be due to sampling bias during low flow periods, when high flow periods were not captured in the sampling regime.

PRINCIPLE FINDINGS AND SIGNIFICANCE

Annual and event dissolved N export via tile drainage

Dissolved N leaching is a concern in many agricultural regions, but particularly in the Pacific Northwest, where the Mediterranean climate results in the most hydrologic transport during cold periods when vegetation and microbial activity are limited. We found that the annual nitrate flux from the tile drain during the 2012 water year was 13.2 kg N ha^{-1} y^{-1}, with 84% occurring during winter/spring runoff events. The annual DON flux was 0.7 kg N ha^{-1} y^{-1}, with 71% occurring during high flow events. Combined, total dissolved N losses account for ~10% of the average N fertilizer applied to the tile-drained area (T. Brown, personal communication). Tile drain discharge and [NO_3^-] were significantly, positively correlated ($p < 0.001$; discharge ln-transformed for normality), and the time-averaged nitrate concentration was 9.02 mg N L^{-1} – near the EPA limit for drinking water standards and ten times higher than reference conditions described by EPA for this ecoregion. These results highlight the importance of high flow events for delivery of dissolved N to streams in quantities that can cause environmental degradation and human harm.

Annual and event DOC export via tile drainage

DOC export from agricultural systems to surface water is poorly understood, particularly outside of the Midwest where most of the limited research has been conducted to date (e.g. Royer and David 2005, Dalzell et al. 2007, Warrner et al. 2009, Dalzell et al. 2011). Potential sources of DOM to the tile drain were characterized throughout the year, including soil water in top-soil (“shallow”) and sub-soil (“deep”). DOC concentrations exhibited high spatial-temporal variability over space and time in the basin; however, patterns were discernable, with DOC concentration generally decreasing with depth (Figure 2). Additionally, the DOM absorbance and fluorescence data indicate that DOM quality varies consistently with soil depth (Figure 2).

In this study, we found that annual DOC flux from the tile drain was 3.5 kg C ha^{-1} y^{-1}, with 87% occurring during high flow events; DOC concentration increased with tile drain discharge, consistent with the hypothesis that shallow soil water is contributing more to tile drain discharge during high flow, although discharge did not explain a large amount of variation in DOC concentration ($p < 0.001$, $r^2 = 0.17$). Additionally, the quality of DOM in tile drain discharge was significantly correlated with flow rate (Figure 3), further supporting the hypothesis that DOM sources to the tile drain shift with hydrologic conditions. Similar to nitrate and DON export, these data emphasize that high flow events dominate export, and increases in precipitation due to climate change may result in a non-linear increase in delivery of DOC to surface waters.

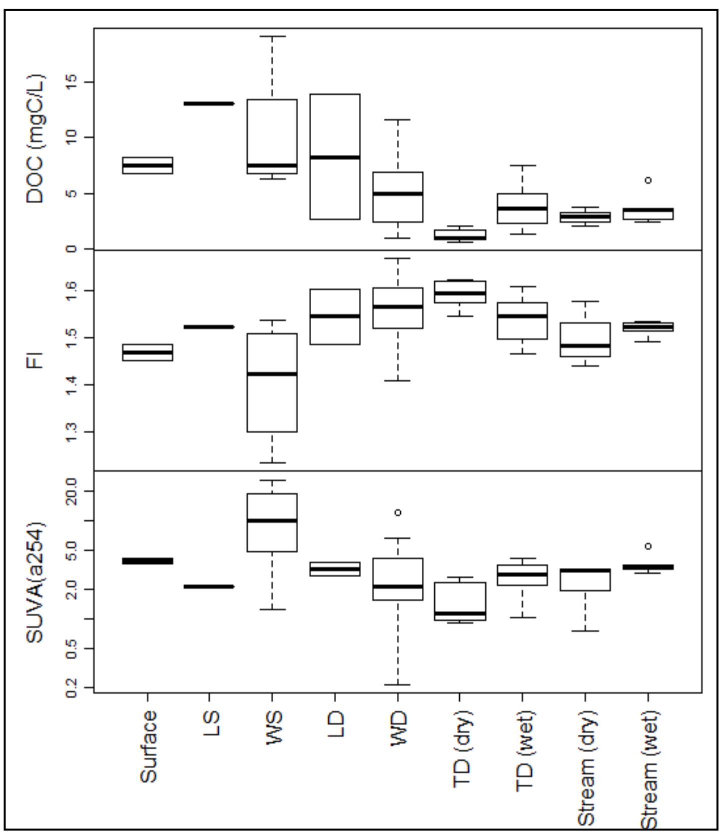


Figure 2. [DOC] and DOM quality indices across a vertical hydrologic gradient (organized from high elevation (left) to low elevation (right) on the x-axis). X-axis labels represent: (Surface) is surface runoff, (LS) shallow lysimeter, (WS) shallow well, (LD) deep lysimeter, (WD) deep well, (TD) tile drain, and (Stream) receiving stream (Missouri Flat Creek) samples. Tile drain and stream samples were separated by wet and dry seasons. (FI) Fluorescence index (FI) is traditionally used to distinguish plant-derived from microbially-derived DOM, with higher values representing microbially-derived DOM. Specific UV absorbance at 254 nm ($SUVA_{254}$) is a proxy for aromaticity of DOM, with higher values indication more aromatic

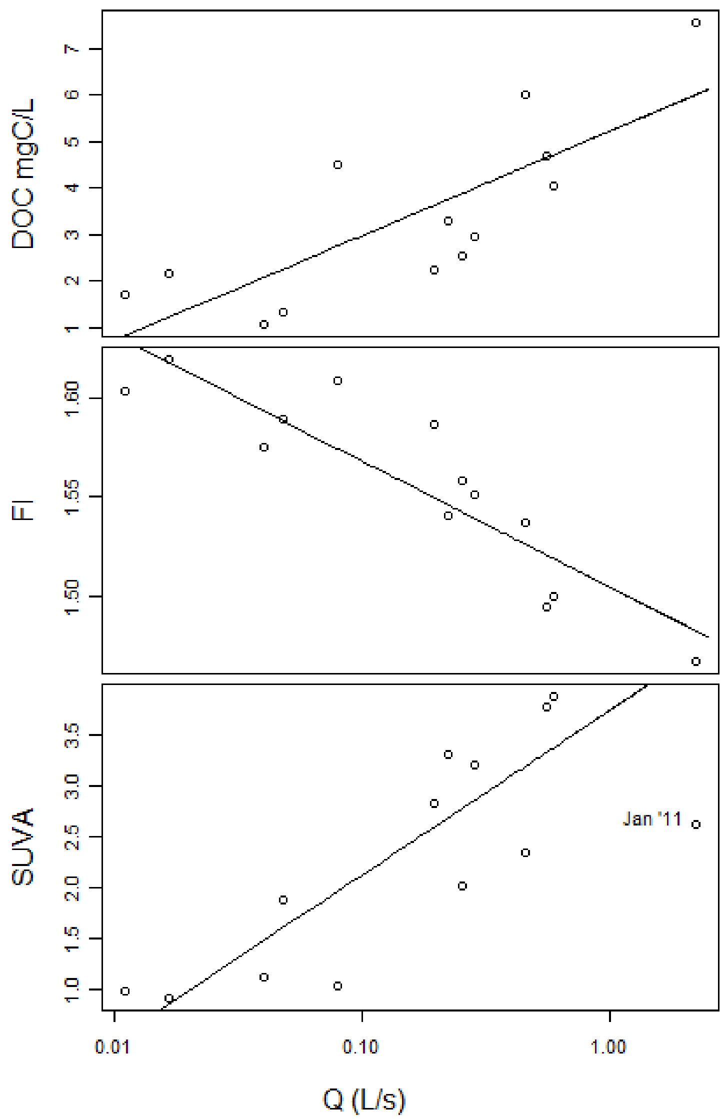


Figure 3. Relationships between DOC concentration (top), Fluorescence Index (FI) (middle), and $SUVA_{a254}$ (bottom) and tile drain discharge (Q) were all significant ($r^2 = 0.61$, $p = .002$ for [DOC]; $r^2 = 0.80$, $p < 0.001$ for FI; $r^2 = 0.74$, $p < 0.001$ for SUVA, with the point from Jan 2011 removed) Discharge was ln-transformed for all regression to meet normality assumption.

The range of DOC concentrations observed in this study is similar to those found in other Midwestern tile drains (e.g. Ruark et al. 2004; Warrner et al. 2009; Dalzell et al. 2011). FI and SUVA values are also similar, although we report a wider range than the others (Warrner et al. 2009; Dalzell et al. 2009). For context, the range in SUVA observed in this one site encompasses the lower 50% of values observed across 12 stream LTER stations, while FI values are more constrained relative to variability observed across LTER stations (Jaffe et al. 2008).

DOM Model Results

The mixing model predicts greater shallow contribution to tile drain flow during high discharge periods (Figure 4), but predicts [DOC] and [DON] in discharge poorly. The Nash-Sutcliffe Efficiency (NSE) for the DOC model is 0.05, indicating that the model is only marginally better at predicting [DOC] than the mean value; NSE for the DON model is -0.03; indicating that the model is worse at predicting [DON] than a simple mean. Additionally, the model systematically over-predicts [DOC] under low flow conditions (Figure 5).

Table 1. Average solute and electrical conductivity of wet season shallow and deep soil water end members. Values in bold are significantly different between deep and shallow sources (one-tailed t-test, $p < 0.05$)

	Shallow	Deep
Mg (mg/L)	**17.3**	**27.6**
EC (uS/cm)	**240**	**409**
DOC (mg C/L)	**6.2**	**4.0**
DON (mgN/L)	0.9	0.4

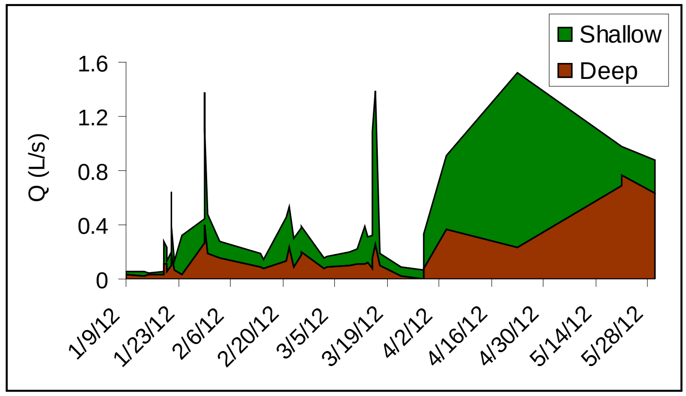


Figure 4. 2012 runoff season hydrograph partitioned into the contributions of deep and shallow soil water to tile drain discharge.

These results point to potential problems with the model and provide a basis for future investigations. First, the model may be missing important end members, as the [DOC] and [DON] values of model end members do not encompass the range of values seen in the tile drain discharge. For instance, an end member with lower "tracer" concentrations and higher [DOC] may be transiently available following precipitation or melt events when event water mobilizes and transports previously disconnected DOM pools to the tile drain rapidly through macropore flow. Second, the model accounts for vertical, but not lateral, expansion and contraction of hydrologic connectivity in the basin. High (lateral) spatial variability observed in soil water chemistry is problematic for interpreting model results; however, soil water chemistry in low-elevation samplers do to correlate better than upslope samplers with tile drain discharge chemistry. Finally, the model assumes no removal of DOM along flow-paths en route to the tile drain outlet. The systematic over-prediction of [DOC] at low discharge suggests that removal processes may be important for controlling export under these conditions. Understanding adsorption and decomposition rates along various flow paths may be critical to effectively predict DOM export to surface water.

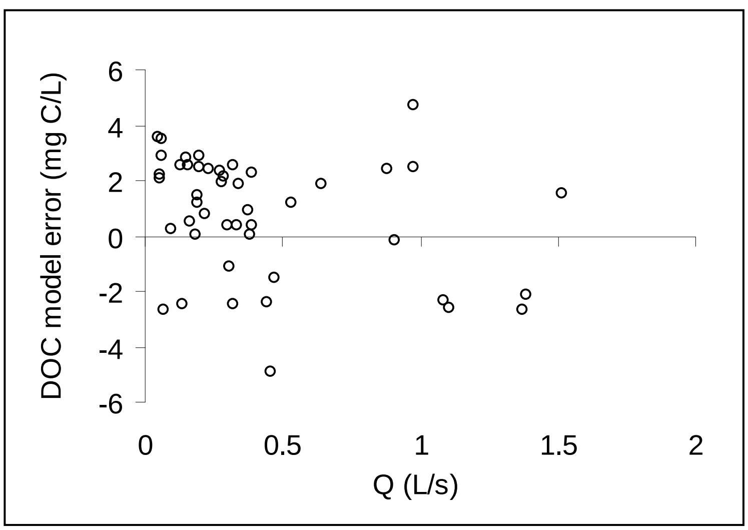


Figure 5. DOC model error relative to tile drain discharge. DOC is systematically over predicted under low discharge conditions.

Controls over nitrate export from and attenuation within the watershed

Annual export of nitrate from the watershed is largely controlled by infrequent (2-7 per year) winter storm events. These large winter storms, which occur with very large instantaneous fluxes (Figure 6), transport 75-99% of the total annual export of nitrate; single large flood events can account for the export of up to 5-10% of the annually applied N fertilizer. Precipitation during the study period is an underrepresentation of long-term precipitation averages in the region, and, given that eastern Washington is projected to see increases in heavy precipitation and fall precipitation (Meehl et al., 2005, Salathe et al., 2010), estimates of nitrate export from the study period may be an underrepresentation of future export from the region. The disproportionate importance of high flow events requires targeted sampling during high winter flows to better constrain the overall nitrogen budget.

Little research has focused on understanding how retention or removal of nitrate from agricultural streams may be affected by DOM availability, for instance, by fueling in-stream denitrification. During summer flows, nitrate concentrations and mass discharges decrease in the downstream direction (Figure 7). Given that there is less mass of nitrate at downstream locations (removal of nitrate mass from the system), this decrease in concentration is likely due to biological uptake. Low flow may be a prime time to further investigate a biological signal. In contrast, we found little evidence of nitrate removal along the stream during winter conditions, suggesting that DOM availability likely has little role in any biological controls of in-stream nitrate dynamics during this period, when low temperatures and fast moving water can mean limited substrate interaction and sub-optimal conditions for biologic activity. Further, evidence from another study at Cook Farm indicates that removal of nitrate by a buffer strip along the stream is minimal during the dormant winter season (C.J. Kelley thesis, 2011). Given that nitrate removal in soils and in the stream channel is likely to be small during the winter, the most feasible means to control N in streams is by managing on-field application timing and rates. Without adjusting management strategies, increases in the quantity and intensity of fall and winter precipitation will lead to larger N export from the system.

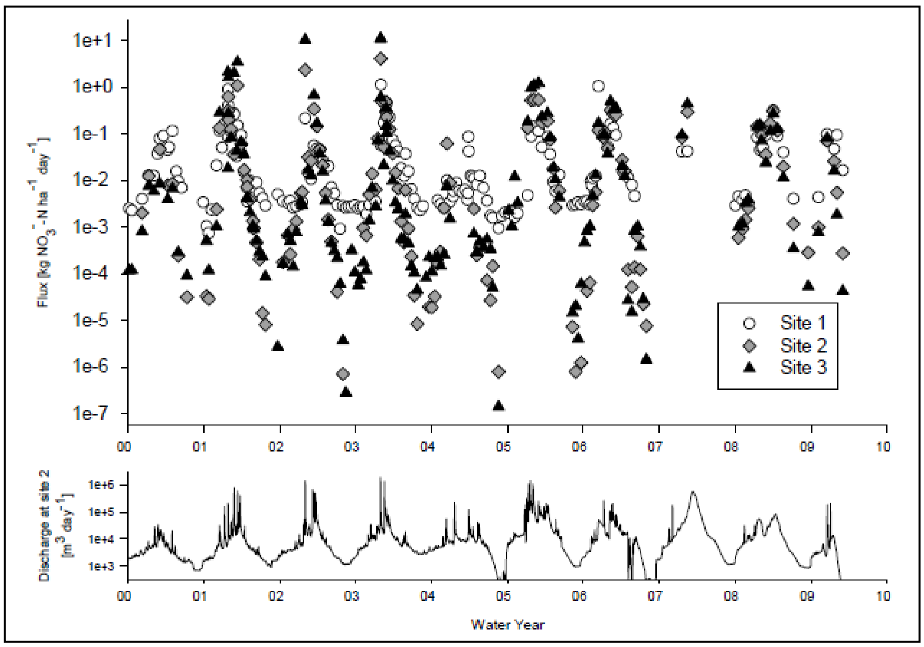


Figure 6. Flux of NO_3^-–N at three sampling sites. Fluxes of NO_3^-–N can be as high as 12 kg NO_3^-–N ha^{-1} d^{-1} during large discharges. In general, site 3 fluxes are larger than sites 1 and 2 during the winter months, and smaller than site 1 during summer months. Out of the three sites, site 1 generally has the largest flux during the summer months, though the overall difference is quite small.

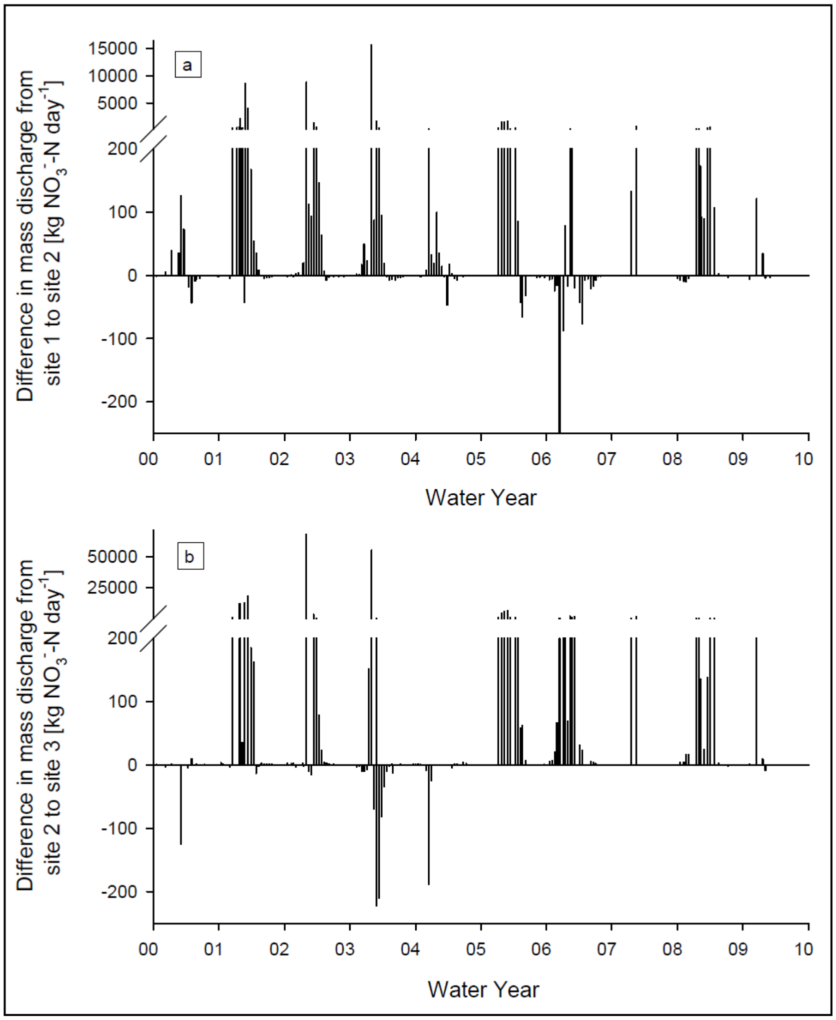


Figure 7. Changes in mass discharge of NO_3^-–N in the downstream direction. The x-axis is in water years, which start on October 1 of the year shown. (a) Mass discharge increases from site 1 to site 2 during most of the year, but differences are negligible or decrease during low flow summer months.
(b) Mass discharge shows increases from site 2 to site 3 during large flows, and some losses during summer months, but the relationship between gains and losses and small and large flows is less clear.

WORKS CITED

Bernal, S, A Butturini, F Sabater. 2005. *Biogeochemistry* 75: 351-372.

Chow, A, RA Dahlgren, and J Harrison. 2007. *Environmental Science & Technology*, 41: 8645-7652.

Dalzell B, T Filley, J Harbor. 2007. *Geochimica et cosmochimica Acta* 71: 1448-1462.

Dalzell B, J King, D Mulla, J Finlay, GR Sands. 2011. *Journal of Geophysical Research* 116: 1-13.

Fu, G, ME Barber, and S Chen. 2010. *Hydrological Processes* 24: 866-878.

Glibert, PM, JI Allen, AF Bouwman, et al. 2010. *J. Marine Systems* 83: 262-275.

Hagedorn F, P Schleppi, P Waldner et al. 2000. *Biogeochemistry* 50:137-161.

Harrison et al. 2003. *Global Biogeochemical Cycles* 17: doi:10.1029/2002GB001991.

Inamdar, SP, N O'Leary, MJ Mitchell et al. 2006. *Hydrological Processes* 20: 3423-3439.

Jaffé, R., D McKnight, N Maie, et al. 2008. *Journal of Geophysical Research* G4:1-15.

Jansson, M, L Leonardson, J Fejes. 1994. *Ambio* 23: 326-331.

Jassby, A, and E van Nieuwenhuyse. 2005. *San Francisco Estuary and Watershed Science* 3:1-33.

Kelley, CJ. 2011. MS Thesis, Washington State University.

Martin, R., and J.A. Harrison. 2011. *Ecosystems* 14: 1328-1338.

Meehl, GA, JM Arblaster, C Tebaldi. 2005. *Geophysical Research Letters* 32: L18719 doi:10.1029/2005GL023680.

Mote, PW and EP Salathé. 2010. *Climate Change* 102: 29-50.

Pellerin BA, SS Kaushal, and WH McDowell. 2006. *Ecosystems* 9:852-864.

Royer, TV, MB David, LE Gentry. 2006. *Environmental Science and Technology* 60: 4126-4131.

Royer T and M David. 2005. *Aquatic Sciences* 67: 465-471/

Ruark M, S Brouder, R Turco. 2004. *Journal of Environmental Quality* 38: 1205-1215.

Salathe, EP, LR Leung, Y Qian, et al. 2010. *Climate Change* 102: 51-75.

Tiemeyer, B, B Lennartz, P Kahle. 2008. *Agriculture, Ecosystems and Environment*. 123: 125-136

Thouin JA, WM Wollheim, CJ Vörosmarty, et al. 2009. *J. North American Benthological Society*, 28:894-907.

Warrner TJ, TV Royer, JL Tank, et al. 2009. *Biogeochemistry* 95: 295-307.

WA DoE (Washington Department of Ecology) 2008 303(d) list

Wiegner TN and SP Seitzinger. 2004. *Limnology and Oceanography* 49:1703-12.

Progress towards assessing the large-scale impacts of forest fires on runoff erosion across the Pacific Northwest

Basic Information

Title:	Progress towards assessing the large-scale impacts of forest fires on runoff erosion across the Pacific Northwest
Project Number:	2012WA353B
Start Date:	3/1/2012
End Date:	2/28/2013
Funding Source:	104B
Congressional District:	Washington5
Research Category:	Climate and Hydrologic Processes
Focus Category:	Hydrology, Sediments, Climatological Processes
Descriptors:	None
Principal Investigators:	Jennifer Adam, Jennifer Adam

Publications

1. Gould, G. K., J. C. Adam, M. Liu, C. Warren, M. E. Barber, P. Robichaud, J. Wagenbrenner, K. Cherkauer, L. Wang. 2012., Large-Scale Simulation of the Effects of Climate Change on Runoff Erosion Following Extreme Wildfire Events. American Geophysical Union Fall Meeting, San Francisco, CA.
2. Gould, G. K., J. C. Adam, M. Liu, C. Warren, M. E. Barber, P. Robichaud, J. Wagenbrenner, K. Cherkauer, L. Wang, Large-Scale Simulation of the Effects of Climate Change on Runoff Erosion Following Extreme Wildfire Events. (In preparation)
3. Gould, G, 2013, Large-Scale Simulation of the Effects of Climate Change on Runoff Erosion Following Extreme Wildfire Events, MS Dissertation, Department of Civil and Environmental Engineering, Washington State University, Pullman, WA.

1 Introduction

1.1 Background

Increasing greenhouse gas concentrations have perturbed the radiative balance of the earth-atmosphere system and led to human-induced global climate change (IPCC 2007). Because of the variability of weather within any climatic condition, no single extreme event can be directly attributed to climate change. However, there is strong scientific evidence indicating climate change is expected to increase the frequency, duration, and intensity of extreme temperature and precipitation events and thus negatively impact associated heat wave, drought, flood, and wildfire phenomena (CCSP 2008). In the western U.S., there is clear concern for increases in wildfire occurrence and severity due to projected climate changes. For example, degradation of water quality occurs in post-fire periods due to water erosion of bare soils (Reneau et al. 2010).

The connection between forest fires and erosion has long been established. Infiltration rates often are reduced by 50% or more as a result of wildfires (Robichaud 2000; Moody and Martin 2001) leading to increases in overland flow rates. Soils can be directly affected by fire, making them water repellent (Doerr et al. 2006) or reducing their aggregate size (DeBano et al. 2005) and thereby making soils more erodible. Further, the burning of surface cover results in a loss of protection to soil surface (Benavides-Solorio and MacDonald 2005; Larsen et al. 2009) and leads to greater erosion rates after high severity fires (Connaughton 1935; Benavides-Solorio and MacDonald 2001; Moody and Martin 2001; Holden et al. 2006; Moody and Martin 2009; Robichaud et al. 2010). Comparing burned to non-burned areas, Johansen et al. (2001) found up to 25 times the erosion rate for burned areas. Fire in a 564-ha forested catchment in central Washington produced dramatically increased sediment volumes due to flow rates, increased overland flow caused by reduced infiltration capacity, and mass soil movement (Helvey 1980). Widespread erosion was reported due to the 1988 Yellowstone fires (Minshall and Brock 1991) and a wildfire in southern Oregon produced 2 to 4 cm of surface soil erosion from steep slopes in a single, intense winter storm (Amaranthus and Trappe 1993). Inbar et al. (1998) used field plots of burnt and undisturbed forests and found sediment yield to be 100,000 times higher in burnt areas the first rainfall season after the fire. This decreased by two orders of magnitude during the second season due to rapid re-vegetation of the area. The increased sediment supply to stream channels often lasts for decades after fires occur (Benda et al. 2003; Moody and Martin 2009). A recent study on the effects of climate change and wildfire on erosion in central Idaho has suggested that sediment yields could potentially increase by 10 folds from observed rates of the 20th century (Goode et al. 2012).

Water erosion is important because excess sediment in streams continues to be a concern for resource managers across the United States. Statistics compiled by the U.S. EPA in 1996 and 1998 indicated that 24% of surface water impairment involved sediments, suspended solids, or turbidity (McCutcheon and Pendergast 1999). Uncontaminated suspended and embedded sediments were identified in 15% of 303(d) listed water (Schubauer-Berigan et al. 2005). Excess sediment adversely impacts aquatic life, navigation, reservoir sedimentation and flood storage, drinking water supply, and aesthetics (Espinosa et al. 1997; Wood and Armitage 1997; Owens et al. 2005; Robertson et al. 2007). In the Pacific Northwest (PNW), Teasdale and Barber (2008) concluded that agricultural production was a primary source of fine sediments but continued research by these and other researchers also found that forest wildfires likely provide a large percent of the coarser sands that settle in navigation channels and in reservoirs (Elliot et al. 2010; Boll et al. 2011).

1.2 Objectives

Our long-term goal is to quantify the adverse water-quality impacts due to extreme wildfires and associated runoff erosion under projected climatic changes across the western U.S. The overarching goal for this particular proposal is to advance our capability to simulate post-fire runoff erosion at scales larger than a single hillslope, in order to examine the relative contribution of sediment being released to larger streams and rivers in response to wildfire. We propose to apply a newly-developed physically-based modeling framework that combines large-scale hydrology with hillslope-scale runoff erosion (VIC-WEPP; Mao et al. 2010). Towards the overarching goal, we propose the follow inter-related specific objectives:

1. ***Implementation and evaluation of model performance* (at experimental sites).**

   To better understand and simulate the large-scale effects of climate change and wildfire on erosion, we applied the Variable Capacity Infiltration-Water Erosion Prediction Project (VIC-WEPP) on a 1/16° (~5-6 km) grid cell spatial resolution (Mao et al. 2010). Sediment yields were compared between VIC-WEPP and Disturbed WEPP (Elliot and Hall 2010) to check if magnitudes of pre- and post-fire rates were similar. Studies show that soil loss rates with one or two orders of magnitude are typical (Jetten 1999; Spigel and Robichaud 2007). A total of 6,368 hillslopes were used in this comparison with no fire and high fire severity conditions. Experience gained under this objective will inform implementation over a larger watershed (objective #2).

2. ***Implement and parameterize model over the Salmon River basin (SRB) of central Idaho.***

   As a proof of concept for large-scale post-fire erosion modeling, VIC-WEPP was implemented over a large watershed that has been relatively undisturbed by human activities. This involved hourly disaggregation of daily precipitation data, downscaling 15 arc-second digital elevation information to 30 meters, identifying burn sites for land cover parameterization and pre and post-fire scenario simulations, and determining soil erodibility and other key soil parameters.

3. ***Run scenario simulations to examine the relative sensitivity of SRB erosion rates to climate versus land cover and soil parameterization.***

   Scenarios were run for both historical and one future climate simulation to examine the sensitivity of SRB runoff erosion rates to climate and wildfire. Scenarios were run with no fire and high fire scenarios with variations in leaf area index (LAI), saturated hydrologic conductivity, interill erodibility, rill erodibility, and critical shear stress.

2 Methods

The Columbia River basin (CRB) is the major watershed of the PNW mostly lies between the Cascades to the west and the Rocky Mountains to the east. The basin has been developed for flood control, hydropower generation (with 14 major hydroelectric dams making it the most hydroelectrically developed river in the U.S.), irrigation, and navigation (Bonneville Power Administration 1991). The river must also be managed for the protection of salmon under the Endangered Species Act; the basin is home to 5 species of salmonids. Within the CRB, the Salmon River basin (SRB) is largely un-impacted by human uses as compared to the other sub-basins within the CRB. As is demonstrated by Figure 1, the basin is primarily forested with some grassland, but with very little croplands or urban areas. It is one of the largest unregulated watersheds in the U.S. (~36,000 km^2). Therefore, any changes that have occurred or will occur in the near-term can be mostly attributed to climate change and associated effects (such as changes in fire severity and frequency) versus direct anthropogenic effects. Climate change in this region is predicted to alter precipitation quantity and timing, vegetation communities, and fire frequencies, all of which are likely to impact water quality and quantity in the basin.

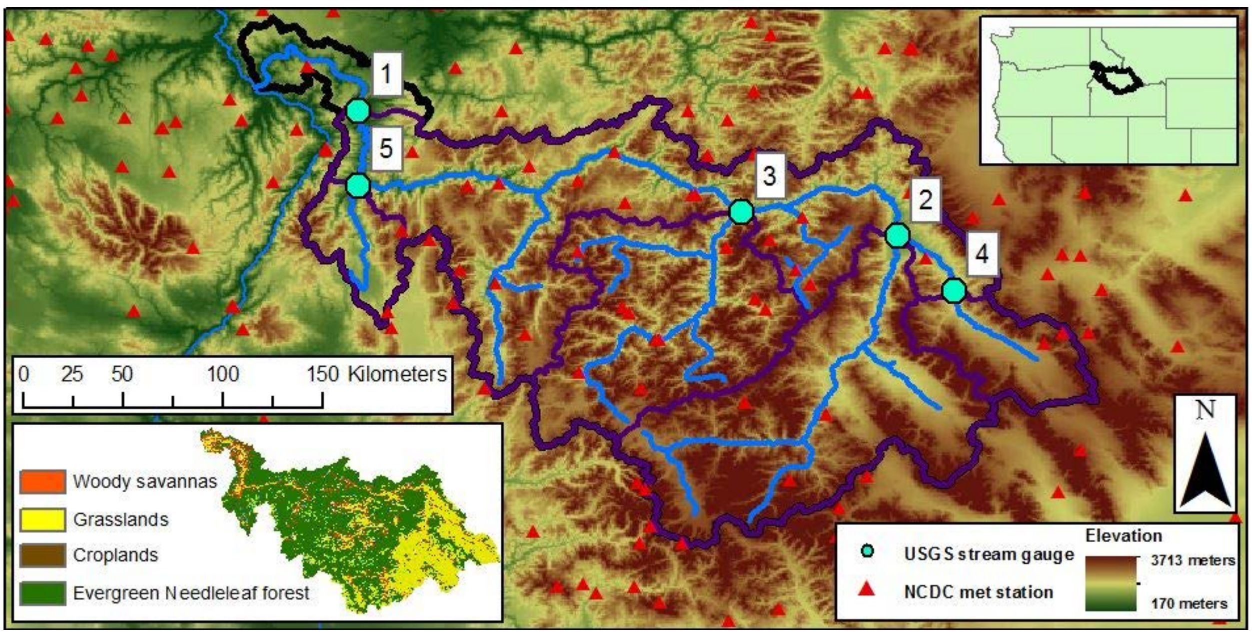


Figure 1. Salmon River basin map showing calibration basins, elevation, and land cover.

Using the Normalized Burn Ratio (NBR), maps compare pre-fire images to next growing season images from the Landsat Thematic Mapper multispectral scanning sensor to determine the severity of wildfires at 30 m spatial resolution. The 207 fires from the Monitoring Trends in Burn Severity (MTBS 2011) online database from 1985 to 2010 in the SRB were translated onto a grid to show the frequency and severity of wildfires on the same grid scale as the VIC model.

2.1 Modeling Framework

2.1.1 Overview of the Modeling Framework

Mao et al. (2010) have recently coupled the VIC model to a soil erosion model, the WEPP-Hillslope Erosion (WEPP-HE) program of Flanagan et al. (2005). WEPP-HE is a stand-alone process-based erosion model that has been extracted from the full WEPP model. Due to the difference in scales between the VIC model (~5-15 km) and the WEPP-HE model (~10-100 m), Mao et al. (2010) distribute each VIC grid cell into a number of slope gradients at the finer (30 m) resolution. For computational feasibility, representative hillslopes are randomly selected from each slope gradient and vegetation classification group within a VIC grid cell and simulated for erosion.

Figure 2 demonstrates the conceptual coupling between VIC and WEPP-HE in which there are 4 groups of information passed to WEPP-HE for simulating runoff erosion at the hillslope scale. (1) For each VIC grid cell, the VIC model passes hydrologic information (runoff depth, peak runoff rate, effective runoff duration, and effective rainfall intensity and duration) to WEPP-HE. (2) In addition, a monofractal scaling method (based on Bowling et al. 2004) is used to downscale digital elevation model (DEM) data to a 30 m resolution for WEPP-HE simulations. This information is used to determine the distribution of slope gradients within each VIC grid cell. (3) Soil information required beyond that needed for VIC modeling includes baseline erodibility, soil particle size classes, size class specific gravity, and organic matter content (Mao et al., 2010). (4) Erodibility adjustments (due to ground cover, canopy effects, live and dead root biomass, and residue) are handled in the coupled model using a variety of relationships that were developed by Mao et al. (2010) by running the full WEPP model for different vegetation types and identifying seasonal values. After WEPP-HE determines erosion and deposition for each representative hillslope and vegetation, total erosion and deposition are calculated for each VIC grid cell by summing across the various hillslope and vegetation classifications.

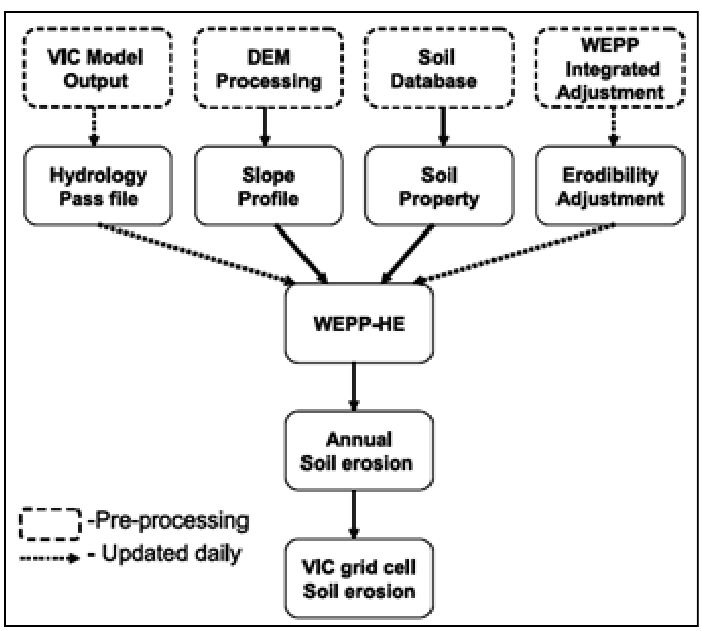


Figure 2. Conceptual integration of the VIC large-scale hydrology model with the hillslope-scale WEPP-HE runoff erosion model (Mao et al. 2010).

2.1.2 Descriptions of Individual Models

2.1.2.1 Variable Infiltration Capacity (VIC) Hydrology Model

The Variable Infiltration Capacity (VIC) model version 4.1.1 is a fully-distributed, physically-based regional-scale model which solves the water and energy budgets at every time step (from 1-24 hours) and for every grid cell (Liang et al. 1994). It was developed for large-scale applications (1/16 - 2°), in which sub-grid variability in land cover and topography is based on statistical relationships. VIC accounts for key moisture and energy fluxes between the land surface and the atmosphere and includes algorithms for shallow subsurface (frozen and unfrozen) moisture, snow, lake, and wetland dynamics (Cherkauer and Lettenmaier 1999; Andreadis et al. 2009; Bowling and Lettenmaier 2010). VIC has been applied over all continental land areas, and has been used extensively over the western U.S. (e.g., Hamlet and Lettenmaier 1999; Maurer et al. 2002; and Elsner et al. 2010).

VIC was calibrated on a daily time step for streamflow and compared to USGS stream gauges at five locations within the SRB. Parameters from the soil input file (variable infiltration curve parameter, maximum velocity of baseflow, fraction of maximum velocity of baseflow where non-linear baseflow begins, fraction of maximum soil moisture where non-linear baseflow occurs, second and third soil layers, and surface roughness of snowpack) were adjusted according to the VIC model technical documentation (Gao 2010). Snow albedo and incoming shortwave radiation are uncertain estimated values calculated in VIC so are available for calibration and were used to match the timing of peak flow for simulated and observations flows.

Along with matching the shape of the average monthly hydrograph, the metrics used in calibration were the average monthly relative bias (RB), monthly RB, monthly Nash-Sutcliffe Efficiency (E), and a peak flow metric (PK) that compares the observed and simulated peak flows (Coulibaly et al. 2001). For each basin, streamflow calibration consisted of the first half of the period of record and evaluation the second half. Table I provides a summary of the period of records, metrics, and percentage of total calibration area for each basin. For a perfect model, RB and PK metrics would show a value of zero and E would have a value of one.

Table I. Calibration summary for VIC. See Figure 1 for basin locations.

Basin	Time Period		E	PK	Percentage of Total Calibration Area
1	Calibration	01/1979 - 12/1994	0.932	0.105	100%
	Evaluation	01/1995 - 12/2010	0.948	0.098	
2	Calibration	01/1979 - 12/1994	0.816	0.120	27.5%
	Evaluation	01/1995 - 12/2010	0.838	0.118	
3	Calibration	10/1993 - 03/2002	0.957	0.118	20.1%
	Evaluation	04/2002 - 09/2010	0.899	0.126	
4	Calibration	01/1979 - 10/1994	0.138	0.169	6.5%
	Evaluation	11/1994 - 08/2010	0.302	0.144	
5	Calibration	01/1979 - 12/1994	0.915	0.101	4.6%
	Evaluation	01/1995 - 12/2010	0.905	0.111	

2.1.2.2 Water Erosion Prediction Project (WEPP)

WEPP is a process based model that was developed in the late 1980s by researchers from four federal agencies striving to create a new and better erosion model (Laflen et al. 1991). At the time, the universal soil loss equation (USLE) was the leading tool to predict and plan for soil erosion (Wischmeier and Smith 1978). As our understanding and knowledge of erosion processes expanded, the USLE erosion tool showed major limitations when applied to different situations than for those it was developed (Laflen et al. 1991). Thus the need of a more comprehensive erosion model became apparent.

The processes in WEPP include erosion, hydrologic and hydraulic, plant growth and residue, water use, and soil processes (Laflen et al. 1991). The erosion processes in WEPP include detachment, transport, and deposition of soil using interrill and rill concepts of detachment. Interrill is the process of raindrops and shallow flows detaching soil particles and transferring them to rill or channel flows. Rill erosion occurs when deeper flowing water detaches soil particles. Erosion in gullies and by other established flows are not included in the WEPP model. The main components of the hydrologic processes in WEPP are climate inputs, infiltration rates, and winter processes. The hydraulic component of WEPP, which determines the hydraulic shearing forces, is important for modeling erosion processes. WEPP uses water balance computations to accurately predict infiltration and runoff rates essential to describe soil erosion volumes. The vegetation processes in WEPP are also critical for accurate erosion rates since plant growth and decomposition greatly impact the soil water content, amount of runoff, and erosion. WEPP has improved the soil erodibility values over those of its predecessor, the USLE, increasing the accuracy of soil erosion volumes. The scale of application for WEPP is typically in the range from a hillslope (tens of meters) to a small watershed (hundreds of meters) (Flanagan and Nearing 1995).

2.1.3 Coupling VIC and WEPP

2.1.3.1 Rainfall Disaggregation

The need for fine resolution precipitation data to simulate soil erosion is important to reduce the amount of uncertainty in predicting soil loss (Kandel et al. 2004). Since most available precipitation data occurs on daily time scales, a process was developed to disaggregate daily precipitation data into hourly time steps (Mao et al. 2010). Using monthly precipitation statistics from the National Soil Erosion Laboratory (NSERL) and daily precipitation, rainfall duration, relative time to peak, and relative peak rainfall intensity are produced using CLIGEN, a stochastic weather generator, for use in the final disaggregation and erosion calculations (Nicks et al. 1995, Mao et al. 2010). Finally, a WEPP model subroutine called DISAG is used to disaggregate daily precipitation into hourly following a double exponential function while conserving total daily precipitation amounts (Flanagan et al 1987).

2.1.3.2 Hydrologic Input Calculations

The disaggregated precipitation is used as input to VIC (with daily data of maximum and minimum temperature and average wind speed) to hourly energy and water fluxes. VIC runoff is then used to calculate the total runoff depth, peak runoff rate, effective runoff duration, effective rainfall intensity, and the effective rainfall duration. These five parameters are needed as hydrologic inputs to the WEPP-HE program. Mao et al. 2010 found that VIC produced many small runoff events that greatly overestimated erosion compared to the full WEPP model. They found that a fraction of saturation area in a VIC grid cell of 7.5% was the minimum value that the full WEPP model would produce soil loss. Thus, when the VIC model generated runoff when the saturated area was below 7.5% of the VIC grid cell area, the runoff was not passed to WEPP-HE program for soil loss calculations.

2.1.3.3 Spatial Downscaling of the Slope Profile

Due to the discrepancy in spatial scales of VIC and WEPP, a process is used to downscale DEM to 30 meter slopes. Mao et al. 2010 implemented VIC at 1/8° resolution which used 30 arc second DEM data. This study applies VIC on a 1/16° resolution so 15 arc second DEM data was used. A monofractal scaling method derives 30 meter slopes from the DEM (Bowling et al. 2004). Mao et al. 2010 demonstrated that this downscaling method provide improved results comparing the coarse DEM slopes to the derived 30 meter slopes.

2.1.3.4 Soil Characteristics

Soil parameters included in VIC are based on the State Soil Geographic Data Base (STATSGO) and are gridded to 1/16° resolution with three soil layers and also contain parameters for soil frost (Maurer et al. 2002; Mao and Cherkauer 2009). Although soil properties change from cell to cell, each gird cell is consistent throughout its domain. The additional soil inputs required to run the WEPP-HE program include baseline erodibility, soil particle size classes, size class specific gravity, and fraction of sediment, diameter, and organic matter content (Mao et al., 2010). Three baseline erodibility factors were estimated for use in WEPP-HE: (1) interill erodibility which measures the soil rate transfer to rills, (2) rill erodibility which describes how vulnerable soil is to detachment, and (3) critical shear stress that determines the shear stress at which no erosion occurs (Mao et al. 2010; Elliot et al. 1989; Flanagan and Nearing 1995). After including organic matter in the VIC soil database, size distributions, fractions, and specific gravities were calculated using a WEPP subroutine (Mao et al. 2010).

2.1.3.5 Erodibility Adjustments

The baseline erodibility factors described above are adjusted to account for ground cover, canopy effects, root biomass, and soil freeze and thaw cycles (Mao et al. 2010). An interpolation scheme to calculate the adjusted erodibilty factors was developed from running the full WEPP model with varying vegetation types to identify typical seasonal values for a range of rainfall amounts and slope gradients (Mao et al. 2010). Mao et al. 2010 integrated these seasonal values by interpolating between the actual rainfall, slope, and day of year to determine unique erodibility adjustments.

2.1.3.6 Hillslope Erosion and Slope Sampling

A sampling scheme based on Park and Van de Giesen (2004) and Thompson et al. (2006) was developed to select hillslopes within a VIC grid cell to reduce computation time and without creating major errors (Mao et al. 2010). First, a set of hillslopes is generated within a VIC grid cell which is then grouped into similar slope ranges. Hillslopes are sampled randomly from each slope range but proportionally based on the number of slopes in each range and the total number of slopes. Each slope range is divided further into different vegetation types according to the fractional area of vegetation in each VIC grid cell. This process reduces the number of times WEPP-HE is run while conserving the variations in slope and vegetation differences for hillslopes within the VIC modeled grid cell. Total erosion for each VIC grid cell is the sum of all vegetation and hillslope groups multiplied by the fractional area of each within the hillslope (for vegetation) and the VIC grid cell (for hillslopes).

2.1.3.7 Post-fire Adjustments

To account for vegetation, soil, and erodibility changes induced by wildfire, adjustments were made to five parameters: leaf area index (LAI), saturated hydrologic conductivity, interill erodibility, rill erodibility, and critical shear stress. LAI was adjusted for low and high severity fires according to Parson et al. (2010) and implemented into the VIC source code. The remaining four parameters were adjusted based on WEPP soil database values (Frankenberger et al. 2011) and pre- and post-fire values from Robichaud et al. (2007). A summary of the post-fire adjustments is listed in Table II that show the adjustment factors as an average of clay loam, silt loam, sandy loam, and loam soil textures and an average of values from the two sources. Critical shear stress fluctuates with soil texture but on an average over all soil textures the critical shear stress is constant for different fire severities.

Table II. Adjustment factors for key post-fire erosion parameters for low, moderate, and high fire severity conditions implemented in VIC-WEPP code.

Parameter	No Fire	Low Fire	Moderate Fire	High Fire
LAI	1.00	0.60	0.25	0.05
Hydrologic conductivity	1.00	0.90	0.75	0.65
Interill erodibility	1.00	1.87	2.60	3.33
Rill erodibility	1.00	13.06	21.71	30.36
Critical shear stress	1.00	1.64	1.64	1.64

2.2 Data

Historical model simulations were driven by gridded daily precipitation, air temperature, and wind speed from Abatzoglou (2011) which used the North American Land Data Assimilation System Phase 2 (NLDAS-2, Mitchell et al. 2004) and the Parameter-elevation Regressions on Independent Slopes Model (PRISM, Daly et al. 2008) to create a high-resolution, 4-km gridded dataset from 1979 to 2010. This dataset was aggregated to VIC's 1/16th degree scale. For future climate, daily downscaled CMIP5 data using the method by Abatzoglou and Brown (2011) from 2039 to 2070 was used. The historical and future daily precipitation data described above was disaggregated to hourly using the methodology of Mao et al. (2010), who use the CLIGEN weather generator (Zhang and Garbrecht 2003) and precipitation statistics from the National Soil Erosion Laboratory (NSERL) to determine storm pattern parameters.

An ensemble of 24 future climate simulations was run to examine the sensitivity of climate on streamflow in the SRB. Only one model and scenario (bcc-csm1-1 model with future scenario RCP45) was used in future erosion simulations over the entire Salmon basin. An area of interest (AOI) was determined that included ranges of annual precipitation, average slope, land cover, and fire severity. Plotting the difference in precipitation and temperature for all future scenarios, the most extreme (the corners) and the center scenarios were used in additional simulations over the AOI. Figure 3 shows the selection of the five future scenarios.

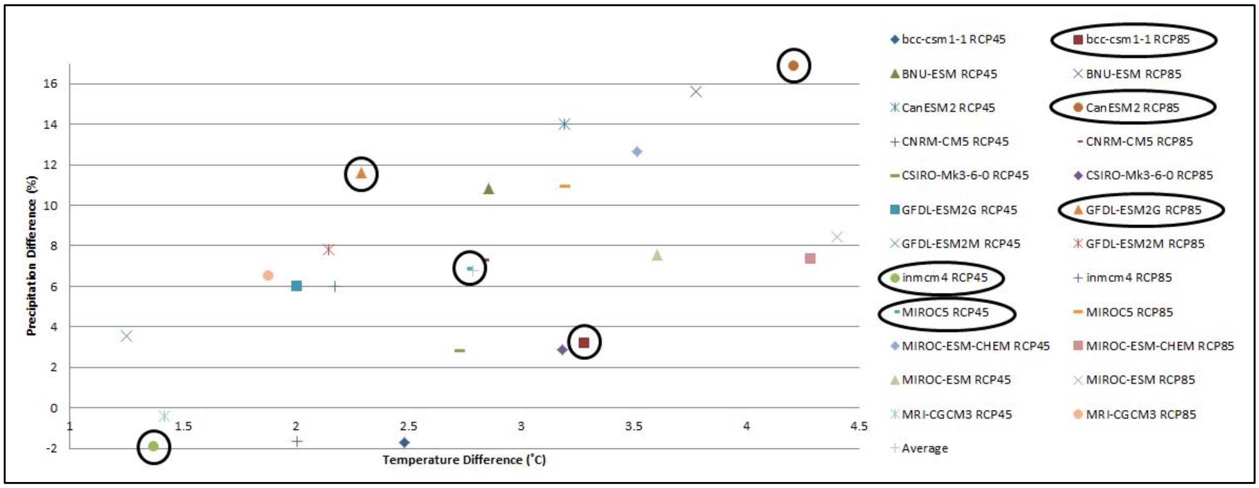


Figure 3. Selection of the five future scenarios.

The historical and future daily precipitation data described above was disaggregated to hourly using the methodology of Mao et al. (2010), who use the CLIGEN weather generator (Zhang and Garbrecht 2003) and precipitation statistics from the National Soil Erosion Laboratory (NSERL) to determine storm pattern parameters.

The soil and vegetation information was taken from the Maurer et al. (2002) VIC implementation. Soil data was originally derived from the State Soil Geographic (STATSGO) Data Base. The land cover used was reclassified from MODIS MOD 12Q1 data with 500-meter resolution. Digital elevation model (DEM) data at 500 meters was used from the Global Multi-resolution Terrain Elevation Data 2010 (GMTED2010; Danielson and Gesch 2011).

3 Results

3.1 Streamflow and Sediment Yield Results

Initial future streamflow results for the largest basin (Basin 1; see Figure 1) show a decrease in peak flow and a shift in timing one month earlier compared to historical streamflow. Snowmelt is the main source of streamflow in the SRB and Figure 4.b shows a basin average temperature increase of 2.48 degrees Celsius which may be a factor in snowmelt depletion earlier in the season. The spatial distribution of precipitation change is also provided in Figure 4 with a basin average decrease by 1.74 percent.

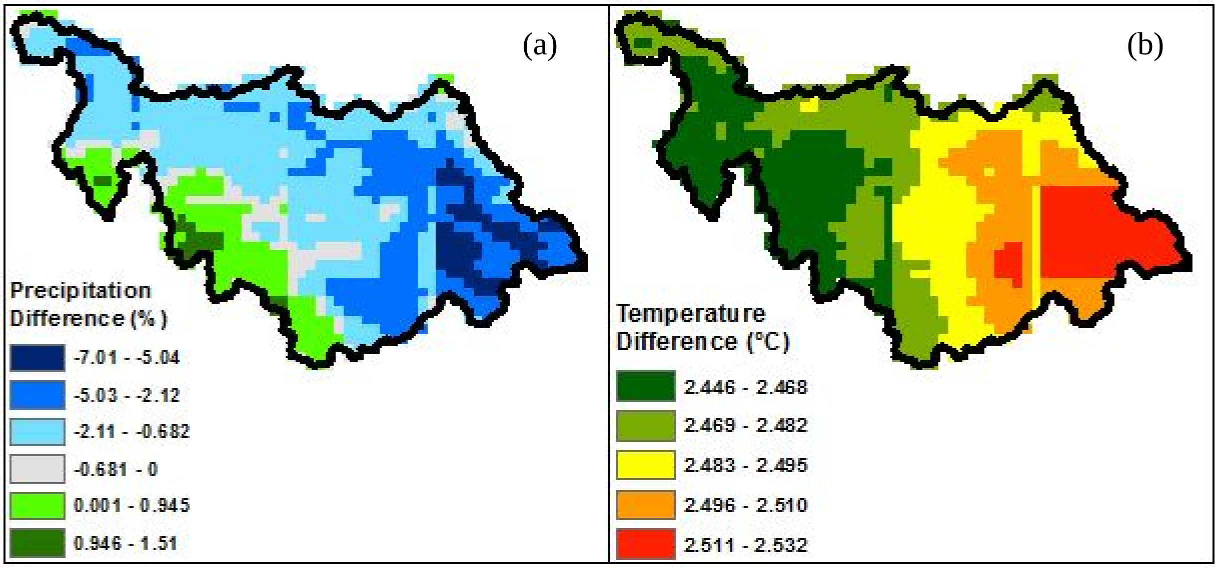


Figure 4. Precipitation (a) and temperature (b) differences for future scenario bcc-csm1-1 RCP45 compared with historical climate.

The next three figures show how erosion changes with fire severity and climate. Each panel (a, b, c, and d) for Figures 5, 6, and 7 represent a specific combination of severity and climate which is shown in Table III. For Figures 5 and 6, there are two different scales for average annual yield; one for the top and another for the bottom panels. Average annual yield is the total yield from for one year over one grid cell averaged from 30 years. Figure 7 shows the center of timing of the average annual yield curve. The center of timing is the Julian day at which half of the erosion has occurred in the year.

Table III. Equations for producing maps in each panel for Figures 5, 6, and 7.

Panel	Erosion Calculation for Figures 5 and 7	Erosion Calculation for Figure 6
a	(historical high fire) – (historical no fire)	$\frac{((\text{historical high fire}) - (\text{historical no fire}))}{(\text{historical no fire})}$
b	(future high fire) – (future no fire)	$\frac{(\text{future high fire}) - (\text{future no fire})}{(\text{future no fire})}$
c	(future no fire) – (historical no fire)	$\frac{(\text{future no fire}) - (\text{historical no fire})}{(\text{historical no fire})}$
d	(future high fire) – (historical high fire)	$\frac{(\text{future high fire}) - (\text{historical high fire})}{(\text{historical high fire})}$

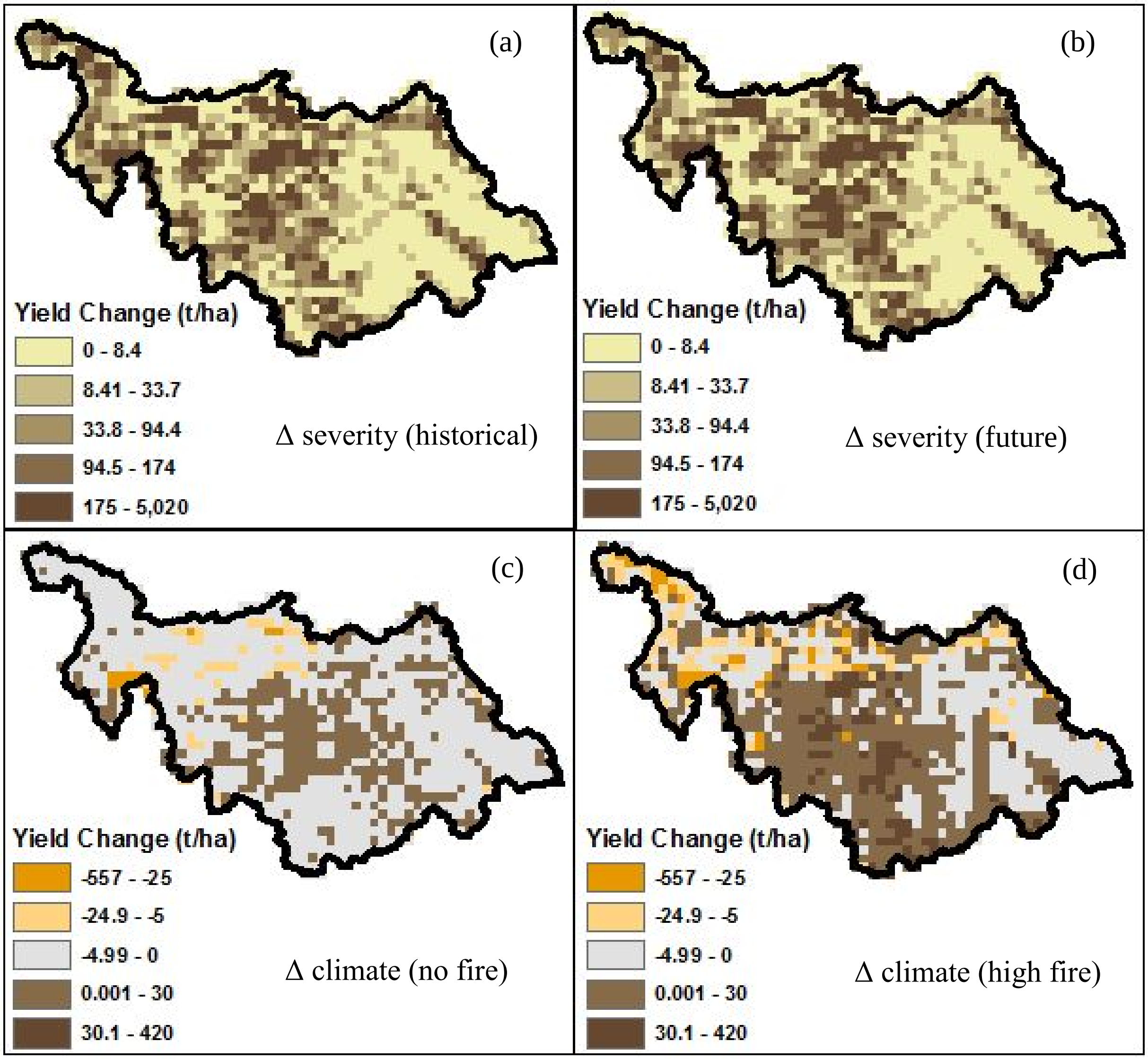


Figure 5. Average annual sediment yield changes for high and no severity for constant historical (a) and future (b) climate and changes for historical and future climate for constant no fire (c) and high fire (d) severity.

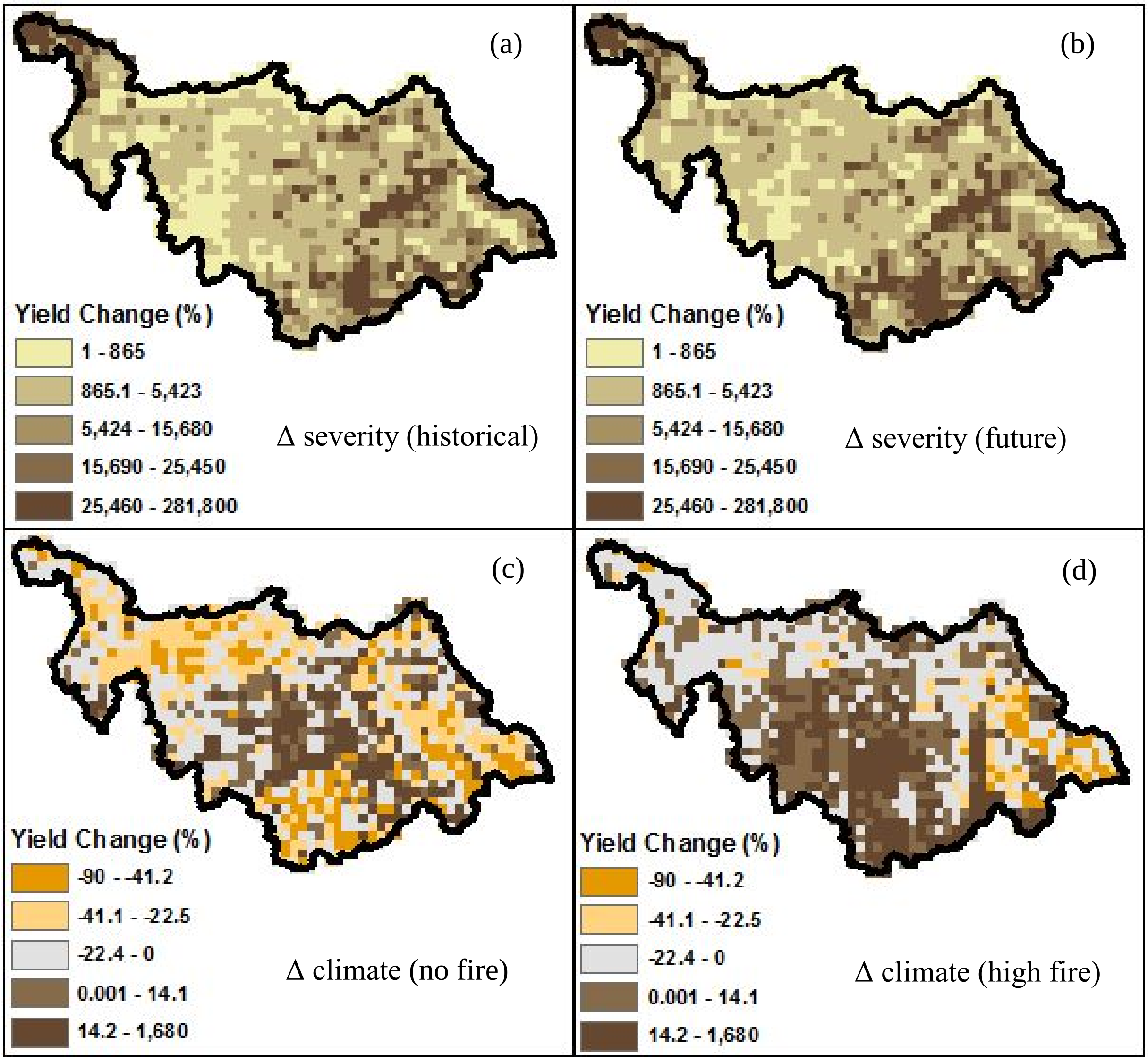


Figure 6. Average annual sediment yield percent difference changes for high and no severity for constant historical (a) and future (b) climate and changes for historical and future climate for constant no fire (c) and high fire (d) severity.

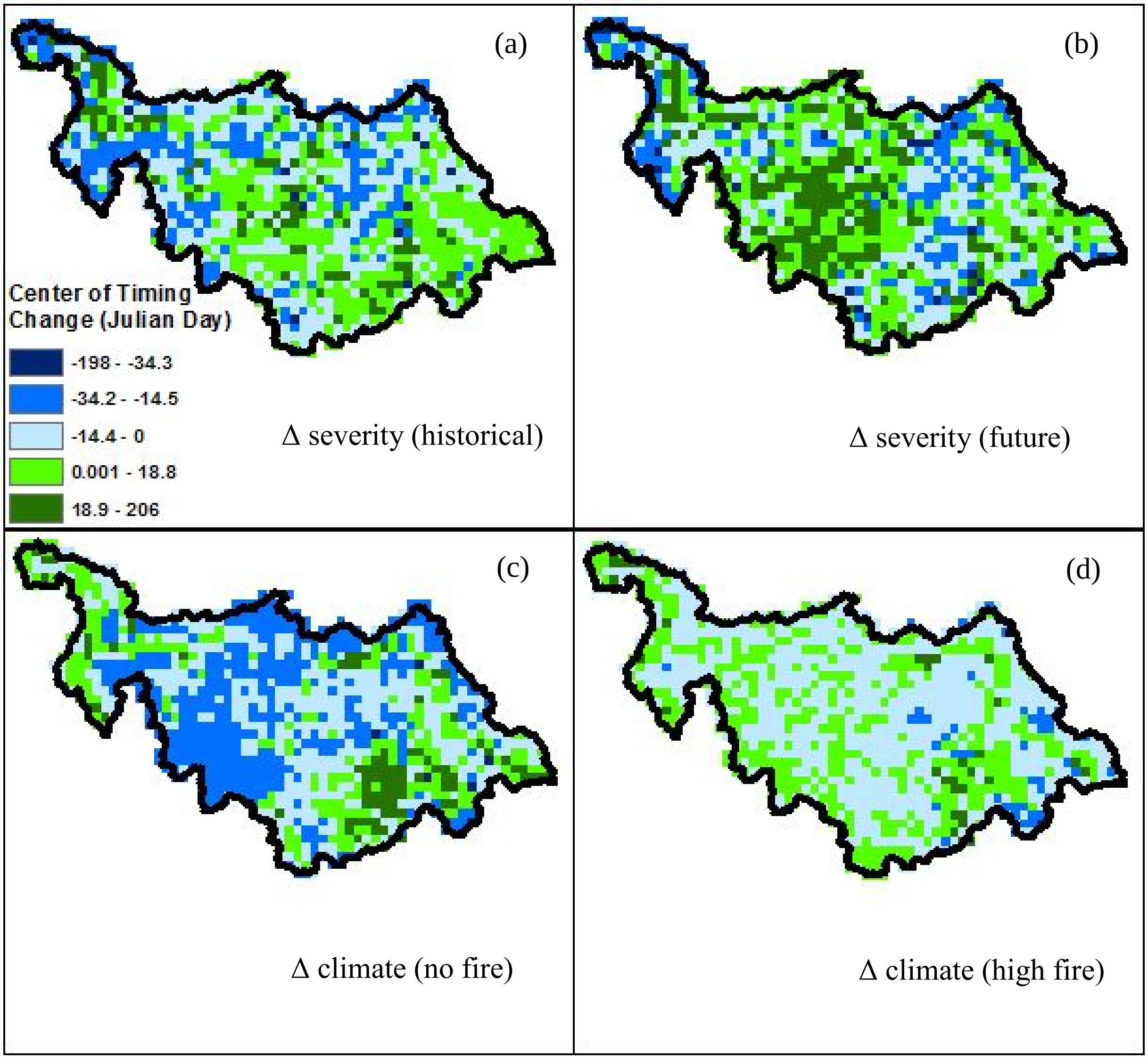


Figure 7. Center of timing changes for high and no severity for constant historical (a) and future (b) climate and changes for historical and future climate for constant no fire (c) and high fire (d) severity.

3.2 Communication of Results

Dissemination of findings through this research will be through journal article publication. Currently, an open-source journal is being considered to allow for more widespread distribution.

4 Discussion and Conclusions

In all three Figures 5, 6, and 7, panels a and b are more similar then panels c and d, showing that for this future scenario climate has less impact and control on sediment yield then changes in fire severity. For example, in Figure 5 the average yield change is 112 t/ha, 118 t/ha, -1.39 t/ha, and 4.50 t/ha for panels a, b, c, and d, respectively. First, the magnitudes of panels a and b (each representing change in yield when fire severity is changed) are much larger than panels c and d (each representing change in yield when climate is changed) indicating fire has more influence on yield. Second, the percent difference between panels a and b is near 5 percent compared to -30 percent for panels c and d. Comparing panels a and b will provide insight on climate whereas comparing panels c and d will give information about fire severity. For example, if climate had no effect on yield, panels a and b would be exactly the same and similarly if fire severity had no effect on yield panels c and d would be the same. This means, the larger the average yield changes are apart the more the factor (severity or climate) has on yield. Since the percent difference in comparing panels c and d is much larger, fire severity has a greater influence on yield.

Figure 6 gives insight into what areas of the SRB are seeing the largest increase in yield relative to previous conditions. This is different than Figure 5 which shows absolute changes in yield. Although the greatest absolute changes in yield are dominant in the central and western parts of the SRB (Figure 5 a and b), the greatest relative increases are mainly in the eastern part (Figure 6 a and b). This is because the no fire severity results in very little yield in the eastern part but after the high fire is overlaid the increase is relatively large compared to the central area which does provide larger yield in the no fire condition. Between Figures 5 and 6, panels c and d following generally similar patterns.

Changes in the center of timing (Figure 7) are not obvious and need more analysis to determine what mechanisms are controlling the timing or yield and similarly runoff in the SRB. Factors such as slope, land cover, and annual precipitation or a combination of factors may be controlling the patterns in Figure 7.

For one future scenario, sediment yield in the SRB is mainly controlled by fire severity with lesser contributions from climate. The average increase in yield for the future scenario with high fire conditions is 118 t/ha (11.8 kg/m^2).

References

Abatzoglou, J. T. 2011. Development of Gridded Surface Meteorological Data For Ecological Applications And Modeling. International Journal of Climatology. DOI: 10.1002/joc.3413.

Abatzoglou J.T. and T. J. Brown. 2011. A Comparison of Statistical Downscaling Methods Suited for Wildfire Applications. International Journal of Climatology. DOI: 10.1002/joc.2312.

Amaranthus, M.P. and J.M. Trappe. 1993. Effects of Erosion on Ecto- and VA-mycorrhizal Inoculum Potential of Soil Following Forest Fire in Southwest Oregon. Plant and Soil, Vol. 140, No. 1, pp 41-49.

Andreadis, K., P. Storck, and D.P. Lettenmaier. 2009. Modeling snow accumulation and ablation processes in forested environments. Water Resources Research, 45, W05429, doi:10.1029/2008WR007042.

Benavides-Solorio, J.D. and L.H. MacDonald. 2001. Post-fire runoff and erosion from simulated rainfall on small plots, Colorado Front Range. Hydrological Processes, Vol. 15, No. 15, pp 2931-2952.

Benavides-Solorio, J.D., and L.H. MacDonald. 2005. Measurement and prediction of post-fire erosion at the hillslope scale, Colorado Front Range. International Journal of Wildland Fire, Vol. 14, pp 457-474.

Benda, L., D. Miller, P. Bigelow, and K. Andras. 2003. Effects of post-wildfire erosion on channel environments, Boise River, Idaho. Forest Ecology and Management, Vol. 178, pp. 105-119.

Boll, J., E. Brooks, J. McAtty, M. Barber, J. Ullman, D. McCool, X. Lu, A. Lawler, and J. Ryan. 2011. Evaluation of sediment yield reduction potential in agricultural and mixed-use watersheds of the Lower Snake River basin. Technical Report, submitted to US Army Corps of Engineers by State of Washington Water Research Center, Pullman, WA.

Bonneville Power Administration. 2001. “The Columbia River System Inside Story.”

Bowling, L.C. and D.P. Lettenmaier. 2010. Modeling the effects of lakes and wetlands on the water balance of Arctic environments. Journal of Hydrometeorology, Vol. 11, pp 276-295.

Bowling, L.C., J.W. Pomeroy, and D.P. Lettenmaier. 2004. Parameterization of blowing-snow sublimation in a macroscale hydrology model, J. Hydrometeorology, Vol. 5, pp 745-762.

Climate Change Science Program (CCSP) 2008. Weather and climate extremes in a changing climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research. [Thomas R. Karl, Gerald A. Meehl, Christopher D. Miller, Susan J. Hassol, Anne M.

Waple, and William L. Murray (eds.)]. Department of Commerce, NOAA's National Climatic Data Center, Washington, D.C.

Cherkauer, K.A. and D.P. Lettenmaier. 1999. Hydrologic effects of frozen soils in the upper Mississippi River basin. Journal of Geophysical Research-Atmospheres, Vol. 104, pp 19599-19610.

Connaughton, C.A. 1935. Forest fires and accelerated erosion. Intermountain Forest and Range Experiment Station. Journal of Forestry, Vol. 33, No. 8, pp 751-752.

Coulibaly, P., B. Bobee, and F. Anctil. 2001. Improving extreme hydrologic events forecasting using a new criterion for artificial neural network selection. Hydrological Processes. 15: 1533-1536.

Daly C., M. Halbeib, J. I. Smith, W. P. Gibson, M. K. Doggett, G. H. Taylor, J. Curtis, and P. A. Pasteris. 2008. Physiographically-sensitive mapping of temperature and precipitation across the conterminous United States. International Journal of Climatology DOI: 10.1002/joc.1688.

Danielson, J.J., and D. B. Gesch. 2011. Global multi-resolution terrain elevation data 2010 (GMTED2010): U.S. Geological Survey Open-File Report 2011–1073, 26 p.

DeBano, L.F., D.G. Neary, and P.F. Folliott. 2005. Soil physical properties. In: D.G. Neary, K.C. Ryan, and L.F. DeBano, eds., Wildland fire in ecosystems: effects of fire on soils and water. U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station: General Technical Report, RMRS-GTR-42, pp 29-51. Ogden, UT

Doerr, S.H., R.A. Shakesby, W.H. Blake, C.J. Chafer, G.S. Humphreys, and P.J. Walbrink. 2006. Effects of differing wildfire severities on soil wettability and implications for hydrological response. Journal of Hydrology, Vol. 319, pp 295-311.

Elliot, W.J. and D. E. Hall. 2010. Disturbed WEPP Model 2.0. Ver. 2011.11.22. Moscow, ID: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. Online at <http://forest.moscowfsl.wsu.edu/fswepp>.

Elliot, W. J., A. M. Liebenow, J. M. Laflen, and K. D. Kohl. 1989. A compendium of soil erodibility data from WEPP cropland soil field erodibility experiments 1987 & 1988, NSERL Report no.3, USDA-ARS National Soil Erosion Research Laboratory, West Lafayette, IN.

Elliot, W.J., R.B. Foltz, and S. Miller. 2010. Upland erosion processes in northern Idaho forests. U.S. Forest Service, Rocky Mountain Research Station, Moscow, ID.

Elsner, M., L. Cuo, N. Voisin, J. Deems, A. Hamlet, J. Vano, K. Mickelson, S.-Y. Lee, and D. Lettenmaier. 2010. Implications of 21st century climate change for the hydrology of Washington State, Climatic Change, 102(1), 225-260.

Espinosa, F.A., J.J. Rhodes, and D.A. McCullough. 1997. The failure of existing plans to protect samon habiata in the Clearwater National Forest in Idaho. Journal of Environmental Management, Vol. 49, pp 205-230.

Flanagan, D.C. and M.A. Nearing. 1995. USDA – Water Erosion Prediction Project: Hillslope Profile and Watershed Model Documentation, West Lafayette, IN.

Flanagan, D. C., G. R. Foster, and W. C. Moldenhauer. 1987. How storm patterns affect infiltration, paper presented at Proc. International Conference on Infiltration Development and Application, University of Hawaii, Water Resources Research Center, Honolulu, HI, Jan 6–9, 1987.

Flanagan, D.C., J.C. Ascough, W.F. Geter, and O. David. 2005. Development of a hillslope erosion module for the object modeling system, paper presented at ASAE Annual International Meeting, Tampa, FL., ASAE, St. Joseph, MI, July 17-20.

Frankenberger, J. R., S. Dun, D. C. Flanagan, J.Q. Wu, W. J. Elliot. 2011. Development of a GIS Interface for WEPP Model Application to Great Lakes Forested Watersheds. Presented at the *International Symposium on Erosion and Landscape Evolution*. Anchorage, Alaska. ISELE Paper Number 11139.

Gao, H., Q. Tang, X. Shi, C. Zhu, T. J. Bohn, F. Su, J. Sheffield, M. Pan, D. P. Lettenmaier, and E. F. Wood. 2010. Water Budget Record from Variable Infiltration Capacity (VIC) Model. In Algorithm Theoretical Basis Document for Terrestrial Water Cycle Data Records (in review).

Goode, J. R., C. H. Luce, and J. M. Buffington. 2012. Enhanced sediment delivery in a changing climate in semi-arid mountain basins: Implications for water resource management and aquatic habitat in the northern Rocky Mountains. Geomorphology 139-140: 1-15.

Hamlet, A.F. and D.P. Lettenmaier. 1999. Effects of climate change on hydrology and water resources in the Columbia river basin, Journal of the American Water Resources Association, Vol. 35, pp 1597-1623.

Helvey, J.D. 1980. Effects of a north central Washington wildfire on runoff and sediment production. Journal of the American Water Resources Association, Vol. 16, No. 4, pp 627-634.

Holden, Z.A., P. Morgan, C. Luce, M. Crimmins, and E. Heyerdahl. 2006. Sensitivity of recent wildfire extent and severity to annual streamflow distribution and timing in the Pacific Northwest USA (1984-2005). The Journal of the Association for Fire Ecology.

Inbar, M., M. Tamir, and L. Wittenberg. 1998. Runoff and erosion processes after a forest fire in Mount Carmel, a Mediterranean area. Geomorphology, Vol. 24, No. 1, pp 17-33.

International Panel on Climate Change (IPCC). 2007. Climate Change 2007: Impacts, adaptation, and vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Parry, Martin L., Canziani, Osvaldo

F., Palutikof, Jean P., van der Linden, Paul J., and Hanson, Clair E. (eds.)]. Cambridge University Press, Cambridge, United Kingdom, 1000 pp.

Jetten, V., A. de Roo, and D. Favis-Mortlock. 1999. Evaluation of field-scale and catchment-scale soil erosion models. Catena 37 (3-4), 521-541.

Johansen, M.P., T.E. Hokonson, and D.D. Breshears. 2001. Post-fire runoff and erosion from rainfall simulation: contrasting forests with shrublands and grasslands. Hydrological Processes, Vol. 15, No. 15, pp 2953-2965.

Kandel, D. D., A. W. Western, R. B. Grayson, and H. N. Turral (2004), Process parameterization and temporal scaling in surface runoff and erosion modeling, Hydrologic Processes, 18, 1423–1446.

Laflen, G. A., J. M. Elliot, W. J. Simanton, J. R. Holzhey, C. S. and K. D. Kohl. 1991. WEPP: Soil erodibility experiments for rangeland and cropland soils. Journal of Soil and Water Conservation, 46(1), 39-44.

Larsen, I.J., L.H. MacDonald, E. Brown, D. Rough, M.J. Welsh, J.H. Pietraszek, Z. Libohova, J.D. Benavides-Solorio, and K. Schaffrath. 2009. Causes of post-fire runoff and erosion: water repellency, cover, or soil sealing? Soil Science Society of America Journal, Vol. 73, pp 1393-1407.

Liang, X., D.P. Lettenmaier, E. Wood, and S.J. Burges. 1994. A simple hydrologically based model of land-surface water and energy fluxes for general-circulation models. Journal of Geophysical Research-Atmospheres, Vol. 99, pp 14415-14428.

Mao, D. and K. A. Cherkauer. 2009. Impacts of land use change on hydrologic responses in the Great Lakes region, Journal of Hydrology, 374, 71–82.

Mao, D., K.A. Cherkauer, and D.C. Flanagan. 2010. Development of a coupled soil erosion and large-scale hydrology modeling system. Water Resources Research, 46, W08543, doi: 10.1029/2009WR008268.

Maurer, E.P., A.W. Wood, J.C. Adam, D.P. Lettenmaier, and B. Nijssen. 2002. A long-term hydrologically-based data set of land surface fluxes and states for the conterminous United States. Journal of Climate, Vol. 15, pp 3237-3251.

McCutcheon, S.C. and J. Pendergast. 1999. Overview of total maximum daily load (TMDL) problem and supporting model development. Presented at Appalachian Rivers II Conference, Morgantown, WV.

Minshall, G.W. and J.T. Brock.1991. Observed and anticipated effects of forest fire on Yellowstone stream ecosystems. Chapter 10 in The Greater Yellowstone Ecosystem, edited by R.B. Keiter and M.S. Boyce, Yale University.

Mitchell, K. E., D. Lohmann, P. R. Houser, E. F. Wood, J. C. Schaake, A. Robock, B. A. Cosgrove, J. Sheffield, Q. Duan, L. Luo, R. W. Higgins, R. T. Pinker, J. D. Tarpley, D. P. Lettenmaier, C. H. Marshall, J. K. Entin, M. Pan, W. Shi, V. Koren, J. Meng, B. H. Ramsay, and A. A. Bailey. 2004. The multi-institution North American Land Data Assimilation System (NLDAS): Utilizing multiple GCIP products and partners in a continental distributed hydrological modeling system. Journal of Geophysical Research, Vol. 109: D07S90, DOI: 10.1029/2003JD003823.

Moody, J.A. and D.A. Martin. 2001. Initial hydrologic and geomorphic response following a wildfire in the Colorado Front Range. Earth Surface Processes and Landforms, Vol. 26, pp 1049-1070.

Moody, J.A. and D.A. Martin. 2009. Synthesis of sediment yields after wildland fire in different rainfall regimes in the western United States. International Journal of Wildland Fire, Vol. 18, pp 96-115.

MTBS. "Monitoring Trends in Burn Severity.". Wildland Fire Leadership Council, December 2011. Web. 27 Feb 2012. <http://mtbs.gov/index.html>.

Nicks, A. D., L. J. Lane, and G. A. Gander. 1995. Weather generator, in USDA-Water Erosion Prediction Project: Hillslope Profile and Watershed Model Documentation, edited by D. C. Flanagan and M. A. Nearing, pp. 2.1–2.2, USDA-ARS National Soil Erosion Research Laboratory, West Lafayette, IN.

Owens, P.N., R.J. Batalla, A.J. Collins, B. Gomez, D.M. Hicks, A.J. Horowitz, G.M. Kondolf, M. Marden, M.J. Page, D.H. Peacock, E.L. Petticrew, W. Salomons, and N.A. Trustrum. 2005. Fine-grained sediment in river systems: environmental significance and management issues. River Research and Applications, Vol. 21, No. 7, pp 693-717.

Park, S. J., and N. Van de Giesen. 2004. Soil-landscape delineation to define spatial sampling domains for hillslope hydrology. Journal of Hydrology, 295, 28–46.

Parson A., P. R. Robichaud, S. A. Lewis, C. Napper, and J. T. Clark. 2010. Field guide for mapping post-fire soil burn severity. Gen. Tech. Rep. RMRS-GTR-243. Fort Collins, CO. U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. pp 49.

Reneau, S.L., D. Katzman, G.A. Kuyumjian, A. Lavine, and D.V. Malmon. 2010. Sediment delivery after a wildfire. Geology, Vol. 35, pp. 151-154.

Robertson, M.J., D.A. Scruton, and K.D. Clarke. 2007. Seasonal effects of suspended sediment on the behavior of juvenile Atlantic salmon. Transactions of the American Fisheries Society, Vol. 136, pp 822-828.

Robichaud, P.R. 2000. Fire effects on infiltration rates after prescribed fire in northern Rocky Mountain forests, USA. Journal of Hydrology, Vol. 231-232, pp 220–229.

Robichaud, P.R., W.J. Elliot, F.B. Pierson, D.E. Hill, and C.A. Moffet, 2007. Predicting postfire erosion and mitigation effectiveness with a web-based probabilistic erosion model. Catena 71 (2), 229-241.

Robichaud, P.R., J.W. Wagenbrenner and R.E. Brown. 2010. Rill erosion in natural and disturbed forests: 1. Measurements. Water Resources Research, Vol. 46, W10506.

Schubauer-Berigan, J.P., S. Minamyer, and E. Hartzell. 2005. Proceedings of a workshop on suspended sediments and solids. EPA/600/R-06/025, U.S. EPA Office of Research and Development, Cincinnati, OH.

Spigel, K.M. and Robichaud, P.R. 2007. First-year post-fire erosion rates in Bitterroot National Forest, Montana. Hydrological Processes. 21: 998-1005.

Teasdale, G.N. and M.E. Barber. 2008. Aerial Assessment of Ephemeral Gully Erosion from Agricultural Regions in the Pacific Northwest. ASCE Journal of Irrigation and Drainage Engineering, Vol. 134, No. 6, pp 807-814.

Thompson, J. A., E. M. Pena-Yewtukhiw, and J. H. Grove. 2006. Soil landscape modeling across a physiographic region: Topographic patterns and model transportability, Geoderma, 133, 57–70.

Wischmeier, W. H., and D. D. Smith. 1978. Predicting rainfall erosion losses - a guide to conservation planning, Agr. Handbk. No. 537. U.S. Dept. Agr., Washington, D.C.

Wood, P.J. and P.D. Armitage. 1997. Biological effects of fine sediment in the lotic environment. Environmental Management, Vol. 21, No. 2. Pp. 203-217.

Zhang, X. C., and J. D. Garbrecht (2003), Evaluation of CLIGEN precipitation parameters and their implication on WEPP runoff and erosion prediction. Transactions of the American Society of Agricultural Engineers, 46(2), 311–320.

Information Transfer Program Introduction

Public Outreach and Education are critically important components of the State of Washington Water Research Center mission. As agency and stakeholders struggle to comprehend important decisions facing water resources, it is essential that they receive unbiased scientific information. The primary outreach goal is to facilitate information exchange by providing opportunities for combining the academic work of research universities in the state with potential users and water stakeholders. The education goal is typically achieved through faculty and student involvement in public presentations and gatherings to promote and build better two-way understanding of water issues and possible solutions and provide unique educational experiences for tomorrow's water resources professionals. These processes occur through a variety of activities, formal and informal, that raise the visibility of university research results throughout the Pacific Northwest. Federal, state and local agencies, non-governmental organizations, watershed groups, and concerned citizens are in need of interpreted science that can be applied to solving the regions' water problems. The SWWRC makes substantial efforts to facilitate this process. The items described in the following Information Transfer Report constitute the core of the technology transfer activities.

Information Transfer

Basic Information

Title:	Information Transfer
Project Number:	2012WA366B
Start Date:	3/1/2012
End Date:	2/28/2013
Funding Source:	104B
Congressional District:	WA 5th
Research Category:	Not Applicable
Focus Category:	Education, Management and Planning, None
Descriptors:	None
Principal Investigators:	Michael Ernest Barber, Michael Ernest Barber

Publications

1. Barber, M.E., J.C. Adam, M. Brady, K. Chinnayakanahalli, K. Rajagopalan, S. Dinesh, C. Kruger, C. Stockle, and G. Yorgey, 2012, Global Change Implications on Long-term Water Supply and Demand Forecasts in the Columbia River Basin, Sustainable Irrigation 2012, 4th International Conference on Sustainable Irrigation and Drainage: Management, Technologies and Policies, Adelaide, Australia.
2. Orr, C., J.C. Adam, A. Beall, M.E. Barber, T. Nguyen, 2012, Using Linked Models to Study Interactions between Water Use Decisions and Climate Change-driven Watershed Processes in the Pacific Northwest Region, American Geophysical Union Fall Meeting, San Francisco, CA.
3. Gould, G., J.C. Adam, C. Warren, M. Barber, J. Wagenbrenner, P. Robichaud, and K. Cherkauer, 2012, Large-Scale Simulation of the Effects of Climate Change on Runoff Erosion Following Extreme Wildfire Events, American Geophysical Union Fall Meeting, San Francisco, CA.
4. Barber, M.E., J.C. Adam, M. Brady, K. Chinnayakanahalli, S. Dinesh, C. Kruger, T. Peters, K. Rajagopalan, C. Stockle, J. Yoder, and G. Yorgey, 2012, Climate Change Impacts on 2030 Water Supply and Demand in the Columbia River Basin, 3rd Annual Pacific Northwest Climate Science Conference, Boise, ID.
5. Barber, M.E. and M. McDonald, 2012, Post-2024 Expectations for Tributary Headwaters Management: Libby Dam Operations in a Changing Climate, 2012 American Water Resources Association - Washington Section, Ellensburg, WA.
6. Barber, M.E., L. Dilley, and M. McDonald, 2012, Integration Challenges and Opportunities Associated with Hydropower and Renewable Energy in the Pacific Northwest, 2012 UCOWR/NIWR Annual Conference, Planning for Tomorrow s Water: Managing Water, Energy, and Food in an Uncertain World, Oral Presentation, Santa Fe, NM.
7. Adam, J., M.E. Barber, K. Ragagopalan, M. Brady, C. Stockle, C. Kruger, R. Nelson, K. Chinnayakanahalli, K. Malek, S. Dinesh, and G. Yorgey, 2012, 2011 Technical Report for the Columbia River Basin Long-Term Water Supply and Demand Forecast, Washington State Department of Ecology, Office of Columbia River, Yakima, WA.

To achieve the goals outlined in the introduction, the following information transfer activities were conducted. It is important to recognize that several of these activities are highly leveraged with activities related to other research projects being conducted by the SWWRC. Nevertheless, without support from the program, these activities would not be possible, or as frequent.

Continued funding for a USDA-CSREES grant was received. The project helps to coordinate research and extension activities of the Water Research Institutes and Extension Services in Alaska, Idaho, Oregon, and Washington with US EPA Region 10 and the NRCS. Six meetings are held each year and communication between researchers, extension faculty, and government agencies is improved considerably by the activity. This project also provides some of the funding that the SWWRC leverages for support of a biennial water conference related to an emerging theme as identified by a regional steering committee.

SWWRC co-sponsored the Palouse Basin Water Summit; a local event attracting stakeholders and concerned citizens from the bi-state watershed (ID and WA). Participants learn about water conservation, efforts to quantify groundwater resources, and other critical aspects of local watershed planning and management. It was also co-sponsored the Spokane River Forum conference. The Spokane River Forum serves as a clearinghouse and information exchange for all things Spokane River. Its annual conference event is attended by 300-400 stakeholders with presentations regarding aquifer storage options to mitigate surface/groundwater depletion occurring in the region.

The SWWRC lead a multidisciplinary collaborative effort for the State of Washington resulting in a legislative summary and technical report aimed at predicting future water supply and demand in the Columbia River basin. These reports were submitted to the Governor and the State Legislature as required by law. The SWWRC team, including the Director, also presented several technical sessions related to the forecast to the Columbia River Policy Advisory Group, the State of Washington Department of Natural Resources, and watershed groups throughout the state.

Director Michael Barber attended the annual NIWR meeting in Washington, DC to interact with other directors from around the country and engage in dialog concerning regional water issues. One outcome of these discussions was that regional institute directors from Idaho, Oregon, and Washington agreed to initiate a collaborative water initiative for the Columbia River Basin. In addition, the SWWRC continued preliminary discussions with Institutes in Idaho and Oregon to conduct a regional “Geothermal Energy” conference late 2013 or early 2014. Dr. Barber also attended the UCOWR/NIWR 2012 conference in Santa Fe, NM, attended a UCOWR Board meeting, and presented an oral presentation on research conducted at WSU.

The SWWRC is also engaged at helping develop a long-term vision for water and agriculture at the national level. As a consequence of Director Barber attending a meeting of 40 water professionals in Monterey, CA last year to help develop a strategic roadmap, he is continuing to serve on the team developing the strategic plan.

Maintaining and updating our web site is a continuous process. This is an important avenue for us to present information about the activities of the Center and the research faculty in the state as well as news and events, research reports, and opportunities for research funding. We currently have all our research reports available for download via PDF format allowing for greater access and utilization of study results.

USGS Summer Intern Program

None.

Student Support					
Category	**Section 104 Base Grant**	**Section 104 NCGP Award**	**NIWR-USGS Internship**	**Supplemental Awards**	**Total**
Undergraduate	1	0	0	0	1
Masters	4	0	0	0	4
Ph.D.	2	0	0	0	2
Post-Doc.	0	0	0	0	0
Total	7	0	0	0	7

Notable Awards and Achievements