Peer-reviewed Research Article

Potential Channel Changes Following Wildfire: Applying the UBC Regime Model to Fishtrap Creek, BC

Brett Eaton

Introduction

Channel response to forest fire involves both exogenous (i.e., originating outside the study reach) and endogenous (i.e., originating within the reach) changes in the conditions governing channel morphology. The exogenous impacts that commonly occur following a forest fire include changes in the timing and magnitude of the peak flows, and volume and character of sediment delivered from the hillslopes (Wondzell and King 2003). To some extent the exogenous changes are stochastic, driven by, for example, heavy precipitation events or melting of the accumulated winter snowpack. These events are determined by the sequence of weather systems that happen to occur. The endogenous changes are more predictable because they are typically related to changes in the bank strength and large woody debris volumes over time. As a result, it is not possible to define a single trajectory for the channel state following wildfire, since channel morphology will respond to both randomly occurring and predictable changes in the governing conditions.

To investigate the full range of potential channel trajectories following disturbance, it is necessary to define the entire suite of potential channel states. This can be achieved by conducting a sensitivity analysis using a simplified model of reach-scale channel conditions, and then varying one governing condition at a time over the potential post-disturbance range. Researchers at the University of British Columbia (UBC) developed the model used for this analysis, called the UBC Regime Model (UBCRM). The model and a user's manual are available online to the general public.[1]

The goals of this article are to give an overview of the UBCRM, to briefly discuss the data collection and calibration procedures, and then to use the model to evaluate how a particular case study stream might respond following a forest fire. The study stream is Fishtrap Creek, which was burned by the McLure fire in August 2003. Additional information about the study site and the changes that have been observed there since the fire are available in several other articles in this issue. This article explores the potential future response of the stream channel over several decades and does not focus on the changes observed to date.

The UBC Regime Model

The UBC Regime Model (UBCRM) has evolved over many years as a result of collaboration between researchers in the Department of Civil Engineering and the Department of Geography at UBC. The model is based on the understanding that a simple model with modest data requirements is more likely to be useful than a data-intensive, numerically demanding one, especially for environmental practitioners. While simple, the model does consider the relevant controlling factors, the most important of which are the nature and erodibility of the channel banks.

Rational regime theories, such as those that underpin the UBCRM, have a long history (e.g., Chang 1979; White *et al.* 1982). There are two main impediments to the general acceptance of rational regime models: (1) the development of a scientifically reasonable understanding of the assumptions necessary to make a unique prediction using these models (called extremal hypotheses); and (2) the incorporation of a bank stability analysis in the model. Researchers at UBC (including M. Church, B. Eaton, R. Millar, and M. Quick) have made significant progress on these two issues. We have been able to reformulate the extremal hypotheses to make the underlying principle more easily understood (Eaton *et al.* 2004; Millar 2005). We have tested this principle against observed channel adjustments in the laboratory and in the field, where we have observed behaviour that is consistent with our generalized extremal hypothesis. We have also incorporated various bank strength formulations into the regime model, which results in a general agreement between model predictions and observed channel dimensions, overcoming the long-standing criticism that regime models consistently underpredict channel width. (The details are in various papers by the author and others, listed on the UBCRM Webpage.) We believe that the UBCRM has numerous practical applications, including the assessment of potential impacts due to landscape disturbances, land use changes, and climate change, as well as preliminary assessment of channel rehabilitation designs.

> *The regime model is predicated upon the idea that channel morphology is related to the flows carried by the stream, averaged over some suitably long time scale.*

Data Requirements and Model Calibration

To set up and run the model, it is necessary to have information on the stream channel under consideration: this usually requires field measurements. Detailed descriptions of the data requirements and collection procedures, as well as the program installation and operation procedures, are in the user's manual, which is available on the Web Site listed above. The key issues are summarized below:

The model requires an estimate of the formative discharge (Q) for the channel; the Manning's flow resistance parameter (n) at the formative discharge; the reach-average channel slope (S); the median surface grain size (D_{50}); and a measure of the largest commonly occurring stones found on the streambed (usually the D_{95}, or 95th percentile of the surface grain size distribution). It also requires knowledge of the relative erodibilility of the channel banks. The typical bankfull flow channel dimensions should also be documented, since the model calibration procedure involves varying the least certain model parameter to fit model predictions to the existing conditions.

The regime model is predicated upon the idea that channel morphology is related to the flows carried by the stream, averaged over some suitably long time scale (see Eaton *et al.* 2004). Gravel streambeds only ever mobilize their bed material during periods of relatively high flow. The bankfull flow (which is the discharge that just fills the channel up to the level of the floodplain surface) is the best representative of the formative discharge for two reasons: flows less than bankfull are not capable of doing much geomorphic work, and flows greater than bankfull spill out onto the floodplain, contributing little to the flow acting directly upon the stream channel. For streams that are gauged, estimating the formative discharge is relatively straightforward. In Canada, the Water Survey of Canada (WSC) collects streamflow data for selected streams, and estimates the peak flows in each year of record. Since channel morphology changes over relatively long time periods, it is probably most appropriate to use a 5- to 10-year average of the peak discharge values immediately before the period of interest. When there are no stream discharge measurements on the stream of interest, it is sometimes possible to estimate the formative flows using regional hydrology analyses. For example, Eaton *et al.* (2002) present a regional analysis of the peak flows for British Columbia. In other cases, it is necessary to estimate the bankfull flow from the observed channel dimensions, assuming a value for flow resistance: a description of exactly how this is done is presented in the UBCRM user's manual.

The flow resistance parameter, n, relates the channel width (W), mean water depth (d), and channel slope (S), to the discharge (Q). The model predictions are quite sensitive to the value of n, so getting a reasonable value is critical. Several references can be used to estimate flow resistance. For example, Cowan's (1956) method is a useful approach, because it attempts to attribute flow resistance to various components of the river form. His equation takes the form:

$$n \qquad n_0 \qquad n_1 \qquad n_2 \qquad n_3 \qquad n_4 \quad m_5$$

The parameters and typical values are given in Table 1. This is a reasonable technique to apply to large rivers (c.f. Church 1992), provided the channel gradient is not too high. However, it tends to perform poorly in steep mountain streams (Marcus *et al.* 1992). According to Chow (1959), mountain streams with gravel-cobble boundaries have Manning's n values that average 0.040 (ranging from 0.030 to 0.050). Mountain streams with cobble-boulder boundaries have higher Manning's n values (mean of 0.050, ranging from 0.040 to 0.070).

The bankfull channel dimensions (i.e., the bankfull width, W, and depth, d), as well as the reach average bed gradient (S), can be estimated by surveying the stream channel. In the field, the bankfull stage must be carefully identified, and at least 10 channel cross-sections should be surveyed to determine the average values for W and d. S can be estimated from a longitudinal profile along the channel centreline that plots elevation against distance along the channel: typically, it is reasonable to fit a linear regression to the data and use the regression coefficient as an estimate of S.

Accurate information on the surface texture is also required to run the model. The median grain size (D_{50}) is used to estimate the sediment transport rate, and the 95th percentile of the grain size distribution (i.e., the D_{95})

Continued on page 30

Continued from page 29

is used to assess channel stability. The best means of determining the surface grain size distribution is by conducting a surface Wolman sample, described in the user's manual. Samples should ideally be taken on the coarsest part of the bar heads in the study reach: the intent is to characterize the bed texture along the thalweg of the stream channel.

Unfortunately, we can only draw on limited work to accurately estimate the bank strength as a function of characteristics that can be measured in the field. The bank stability analysis used herein invokes a characteristic riparian rooting depth, H, which produces a vertical upper bank section, above a cohesionless gravel toe (Eaton 2006). An analysis of Hey and Thorne's (1986) data, in which channels are classified by riparian vegetation type, indicates that the value of H varies systematically with the density of vegetation on the floodplain. For type I, channels in their data set (grass, no trees or shrubs), $H = 0.36$ m; for type II channels (1–5% shrub or tree cover), $H = 0.53$ m; for type III channels (>5–50% shrub or tree cover), $H = 0.89$ m; and for type IV channels (>50% tree or shrub cover), $H = 1.07$ m. This approach is equivalent to invoking an effective root cohesion term, which Eaton (2006) demonstrates is consistent with studies of debris slides (Buchanan and Savigny 1990). As long as the surveyed cross-sections are sufficiently detailed, it should be possible to estimate a reach-average value for H from them to confirm the choice of H used in the model. Note that this approach applies only to gravel streambeds having a coarse, gravelly lower stratum in the channel bank associated with deposition of bed material, generally as channel bars, and an upper stratum of finer, overbank deposits reinforced by riparian vegetation root systems. Where this general sedimentological model does not apply, H should not be used to parameterize bank strength. Other bank strength approaches have been developed (see references on the UBCRM Webpage), but are not used for the sensitivity analysis presented herein.

Table 1. Cowan's (1956) method for estimating Manning's *n*

Parameter		Values
Sediment type, n_o		
	Earth	0.020
	Rock cut	0.025
	Fine gravel	0.024
	Coarse gravel	0.028
Degree of cross-section irregularity, n_1		
	Smooth	0.000
	Minor	0.005
	phaModerate	0.010
	Severe	0.020
Downstream variations in cross-section shape, n_2 (e.g., thalweg shifts from side to side)		
	Gradual	0.000
	Alternating occasionally	0.005
	Alternating frequently	0.010–0.015
Relative effect of obstructions, n_3 (e.g., logs, boulders)		
	Negligible	0.000
	Minor	0.010–0.015
	Appreciable	0.020–0.050
	Severe	0.040–0.060
Vegetation, n^4		
	Low	0.005–0.010
	Medium	0.010–0.025
	High	0.025–0.050
	Very high	0.050–0.100
Degree of meandering, m_5		
	Minor (sinuosity 1.0–1.2)	1.00
	Appreciable (sinuosity 1.2–1.5)	1.15
	Severe (sinuosity > 1.5)	1.30

Once the input parameters have been determined, the model is run and the results compared with the known channel dimensions. If there are significant differences between the model predictions and the observed channel dimensions, then the input parameters are re-evaluated and adjusted where appropriate, based on consideration of the field observations. Even if the model predictions agree well with the observed conditions, we recommend varying the inputs to determine how sensitive the model predictions are to the selected values. After the model has been successfully calibrated, it can be used to evaluate channel response to changes in the governing conditions due, for example, to land use changes, natural disturbance in the watershed, or direct human modification of the stream channel.

Calibration of the model to Fishtrap Creek involved setting $Q = 7.4$ m^3/s (which, based on the WSC data for Fishtrap Creek, is the mean annual daily peak flow), $S = 0.019$ m/m (based on the channel survey), $n = 0.06$ (back-calculated from a direct field measurement of the bankfull flow), $D_{50} = 40$ mm, and $D_{95} = 181$ mm (from Wolman samples taken in the field). Notably, the estimate of slope, S, does not include the steep sections associated with drops over individual logs in the stream or over larger log jams. The bank strength parameter, H, was varied until the predicted and observed channel dimensions matched: the calibrated H value was 0.50 m, which is consistent with the observed rooting depth of the (now dead) trees on the floodplain. The predicted width for these governing conditions is 9.3 m, which compares favourably with the measured value of 9.5 m.

Sensitivity Analysis

The long-term average discharge carried by a stream can vary with climate indicators such as the Pacific Decadal Oscillation index (Moore 1996), or in response to long-term climate change. The formative flow can also vary due to changes in land use and/or forest cover within a basin. At Fishtrap Creek, the post-fire flows have been very close to the long-term average: however, some evidence indicates that runoff production may have been desynchronized within the basin, resulting in a decrease in the peak flows. Alternatively, the disturbance of the forest may have resulted in an increase in the seasonal snowpack and an increase in the rate of melt, which could produce higher peak flows. Our sensitivity analysis assumes that peak flows could either increase or decrease by a maximum 50%. This range is generous and any real changes are almost certain to fall within this range. The analysis was relatively straightforward: keeping all other input parameters the same, the stream discharge was varied and the change in the predicted channel dimensions and sediment transport capacity were documented.

The results of the analysis are presented in Figure 1A. The discharge values used in the sensitivity analysis have been normalized by the original value (Q_o), such that the *x* variate (Q/Q_o) ranges from 0.5 to 1.5. The normalized widths (W/W_o), depths (d/d_o), and sediment transport capacities (Q_b/Q_{bo}) associated with each run are plotted against Q/Q_o. The original channel configuration is indicated on the figure by the dashed lines that cross at the centre of the figure. Despite the large range of discharge values tested, the channel dimensions do not change by much. The depth changes by a maximum of about ± 10%, while the width varies by about ± 30%. The sediment transport capacity varies almost linearly with the discharge, and has a range of about ± 60%. However, a change in transport capacity will likely produce a change in bed sediment texture (Dietrich *et al.* 1989; Buffington and Montgomery 1999), which could be sufficient to reduce the transport capacity to its original value (i.e., to Q_{bo}) (see Eaton and Church 2008). Therefore, it is possible that the channel could remain relatively stable even with fairly large changes in *Q*.

A similar analysis of the channel sensitivity was conducted considering the typical large woody debris (LWD) loads in the stream (Figure 1B). It has been assumed that accelerated bank erosion and windthrow of dead riparian trees could initially increase the volume of LWD in the stream, but that accelerated decay of the burned pieces and a long-term reduction of LWD input as the forest canopy regrows would likely reduce the volume of LWD in the stream several decades after the fire. The effect of LWD is to trap and store sediment and to dissipate energy at vertical drops or steps, the largest of which are associated with jams comprising numerous individual LWD pieces. Adding LWD to the stream effectively removes from the system some of the potential energy with which to transport sediment and to maintain the channel morphology, while removing LWD increases the available potential energy. This has been modeled by adjusting the reach average slope, since this represents the total potential energy available. Currently, the existing LWD in the study reach dissipates about 3.0 m of potential drop: removing all the LWD would increase the effective slope from 0.02 to 0.03 m/m. It was also assumed that post-fire inputs of LWD could double the volume of instream LWD, reducing the effective slope to 0.01 m/m, giving a total range for the sensitivity analysis of 0.01–0.03 m/m.

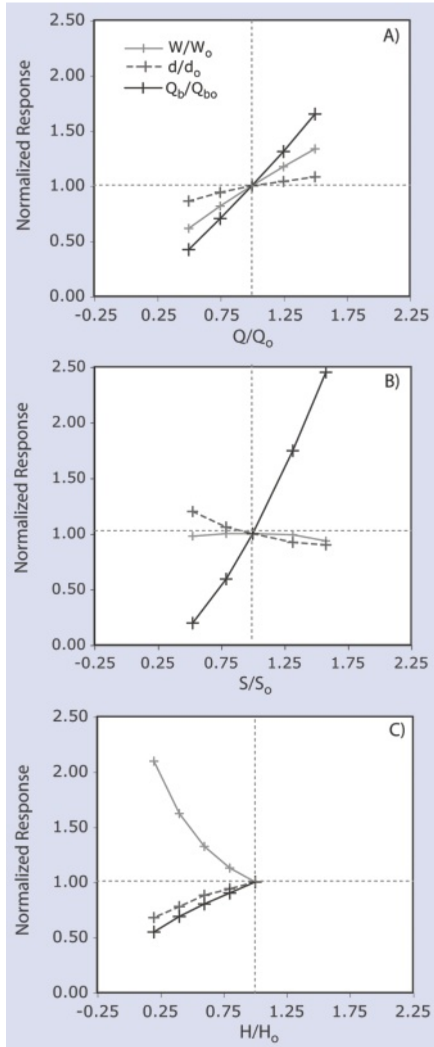

Figure 1. Stability analysis of Fishtrap Creek using the UBCRM. A) The effect of varying the formative discharge (Q) on the width (W), depth (d), and sediment transport capacity (Q_b), relative to the original values (W_o, d_o, Q_{bo}). B) The effect of varying the reach-average effective slope (S). C) The effect of varying the bank strength, expressed as a characteristic rooting depth for riparian vegetation (H).

The predicted channel geometry (W/W_o and d/d_o) is nearly constant over that range of slopes, but the relative transport capacity varies non-linearly with normalized slope. Over the range of slopes analyzed, it increases approximately with the square of S/S_o. It seems very likely, then, that changes in LWD loading could produce changes in the transport capacity that could not be accommodated by changes in the bed surface texture. Furthermore, the expected temporal pattern—involving an initial increase in LWD volumes as banks erode and dead trees are blown over, followed by a gradual but persistent decline in in-stream LWD volumes as LWD decays without being replaced by the immature forest canopy—will push the system one way (i.e., towards aggradation, overbank flooding, and possibly channel avulsions), and then the other (i.e., towards vertical incision, abandonment of secondary channels on the floodplain, and erosion of sediment stored in the stream channel). The effects of disturbance-driven changes in LWD supply to the stream channel are probably long lived, and effects may persist for many decades after the initial disturbance.

The final sensitivity test was conducted on the bank strength. The model calibration suggests that the initial channel morphology is consistent with a characteristic rooting depth of about 50 cm, which is also consistent with observations of the root structure exposed by bank erosion at the study site. However, it is assumed that much of the bank strength due to root cohesion would be lost as the dead root systems decay, returning only when the density and rooting depth of live riparian vegetation become significant. Benda and Dunne (1997) present a model for investigating the rate at which root strength is lost following a

Continued on page 32

Continued from page 31

forest fire, which has been used here to model both the range and the timing of changes in bank strength. After calibrating it to our estimate of the initial bank strength, their model predicts that 5 or 6 years after the fire, *H* will fall as low as 0.10 m (Figure 2). Their model also predicts that substantial recovery is likely to occur by about 20 years after the initial disturbance. Accordingly, *H* could decline from 0.5 to 0.1 m, with the further understanding that the most dramatic loss of bank strength is likely to occur in the first decade following the fire, followed by a more gradual recovery.

The results of the sensitivity analysis (see Figure 1C) demonstrate that both the channel geometry and the transport capacity could be significantly altered. The predicted channel width for *H* = 0.1 m is more than twice the original value, while the transport capacity is just over half the original value. Furthermore, the width-to-depth ratio for this predicted channel state is about 57, which is higher than that usually associated with stable, meandering streams and suggests that the channel could become braided (Fredsoe 1978).

If loss of bank strength results in significant lateral erosion, then that could dramatically increase the LWD volumes in the stream, reducing the effective slope and thereby reducing the sediment transport capacity of the system even further. This sequence of events would produce a laterally unstable, aggrading system, even if the sediment supply from the hillslopes remained unchanged. Any increase in sediment supply due, for example, to debris flows from the burned hillslopes is likely to occur in the same window as is the loss of bank strength, and it would only exaggerate the morphologic instability caused by changes to the boundary conditions (i.e., root cohesion and instream LWD volumes). Since most of the steep, landslide-prone terrain is immediately adjacent to the channel network in Fishtrap Creek (and in many other mountainous watersheds), any significant delay in landslide occurrence and increase in sediment supply to the stream channel are unlikely. The effect of potential increases in peak flows could offset these changes by increasing the sediment transport rate and reducing the rate of aggradation. However, increased peak flows would be more likely to erode the weakened banks, so the net effect of increased peak flows is uncertain. If peak flows were to decrease due to desynchronization of snowmelt in the basin, the transport capacity of the system would be reduced, but bank erosion might still occur. For example, judging from the trees along the banks, Fishtrap Creek had been stable for several decades before the fire, but bank erosion was pervasive at flows no larger than the mean annual peak flow after the fire.

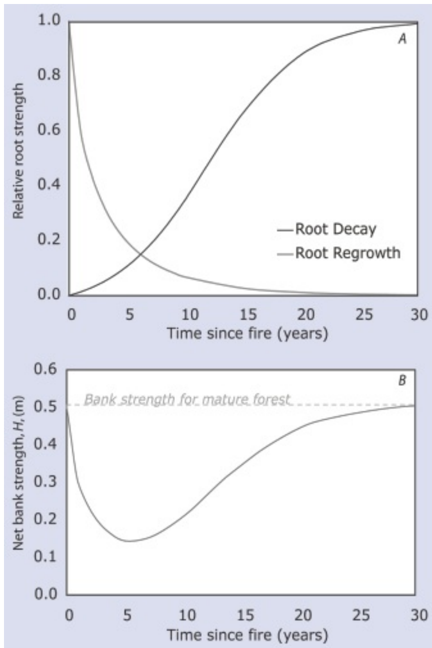

Figure 2. Estimated changes in bank strength over time. A) The predicted relative decline in root strength for dead root systems as they decay and the increase in root strength as vegetation regrows using Benda and Dunne's (1997) model. B) The combined effects of decay and regrowth on the vegetation-related component of bank strength for Fishtrap Creek.

Summary

Our sensitivity analysis indicates that aggradation and widening are likely to occur following a fire: there is field evidence for both aggradation and widening at Fishtrap Creek since the McLure forest fire in 2003. The analysis predicts that this instability can be triggered merely by loss of bank strength and increases in the LWD loading for streams where the riparian area is severely burned. While increases in sediment supply from the hillslopes have not yet been detected at Fishtrap Creek, such increases are likely to exacerbate the level of channel instability. Changes in peak flow are predicted to be second-order effects, since increases could accelerate channel widening and thereby contribute to aggradation, while peak flow decreases could directly contribute to aggradation by reducing the sediment transport capacity of the stream. The period of lateral instability and aggradation that is likely to follow a major fire is predicted to persist for a decade or two, after which substantial root cohesion is likely to return. This would stabilize both the banks and the hillslopes, reduce the channel width, increase the transport capacity, and reduce the frequency (or probability) of debris flow.

The channel is likely then to drift back towards a stable, single-thread channel. However, as LWD jams and steps decay, more potential energy will become available to transport sediment and modify the channel boundary. While coarsening of the channel bed and development of surface structures (Church *et al.* 1998) may prevent detectable channel degradation for a time, the continued loss of instream LWD could produce vertical instability wherein the channel degrades and ultimately becomes disconnected from its floodplain. These sorts of changes are likely to occur after significant decay of the LWD in the channel has occurred but before the riparian vegetation has matured enough to supply appropriate volumes of large, durable LWD to the stream channel. If such instability does manifest itself at Fishtrap Creek, it will do so many decades after the initial disturbance.

For more information, contact:

Brett Eaton
University of British Columbia
Tel: (604) 822-2257
Email: brett.eaton@ubc.ca

References

Benda, L. and T. Dunne. 1997. Stochastic forcing of sediment supply to channel networks from landsliding and debris flow.

Water Resources Research 33(12):2849–2863.
Buchanan, P. and K.W. Savigny. 1990. Factors controlling debris avalanche initiation. Canadian Geotechnical Journal 27(5):659–675.
Buffington, J.M. and D.R. Montgomery. 1999. Effects of sediment supply on surface textures of gravel-bed rivers. Water Resources Research 35(11):3523–3530.
Chang, H.H. 1979. Minimum stream power and river channel patterns. Journal of Hydrology 41:303–327.
Chow, V.T. 1959. Open-channel hydraulics. McGraw-Hill, New York. 680 p.
Church, M. 1992. Channel morphology and typology. In *The Rivers Handbook. P. Callow and G.E. Petts (editors). Blackwell Science, pp. 126–143.*
Church, M., M.A. Hassan, and J.F. Wolcott. 1998. Stabilizing self-organized structures in gravel-bed stream channels: Field and experimental observations. Water Resources Research 34(11):3169–3179.
Cowan, W.I. 1956. Estimating hydraulic roughness coefficients. Agricultural Engineering 37:473–475.
Eaton, B.C. 2006. Bank stability analysis for regime models of vegetated gravel bed rivers. Earth Surface Processes and Landforms 31:1438–1444.
Eaton, B.C., and M. Church. [2008]. Channel stability in bedload dominated streams with non-erodible banks: Inferences from experiments in a sinuous flume. Journal of Geophysical Research – Earth Surface. In press.
Eaton, B.C., M. Church, and D. Ham. 2002. Scaling and regionalization of flood flows in British Columbia, Canada. Hydrological Processes 16(16):3245–3263.
Eaton, B.C., M. Church, and R.G. Millar. 2004. Rational regime model of alluvial channel morphology and response. Earth Surface Processes and Landforms 29(4):511–529.
Fredsoe, J. 1978. Meandering and braiding of rivers. Journal of Fluid Mechanics 84(4):609–624.
Hey, R.D. and C.R. Thorne. 1986. Stable channels with mobile gravel beds. Journal of the Hydraulics Division-ASCE 112(8):671–689.
Marcus, W.A., K. Roberts, L. Harvey, and G. Tackman. 1992. An evaluation of methods for estimating Manning's n in small mountain streams. Mountain Research and Development 12(3):227–239.
Millar, R.G. 2005. Theoretical regime equations for mobile gravel-bed rivers with stable banks. Geomorphology 64:207–220.
Moore, R.D. 1996. Snowpack and runoff response to climatic variability, Southern Coast Mountains, British Columbia. Northwest Science 70(4):321–333.
White, W.R., R. Bettess, and E. Paris. 1982. Analytical approach to river regime. Journal of the Hydraulics Division-ASCE 108(HY10):1179–1193.
Wondzell, S.M. and J.G. King. 2003. Postfire erosional processes in the Pacific Northwest and Rocky Mountain regions. Forest Ecology and Management 178:75–87.