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Stream Power: Origins, Geomorphic Applications, and GIS Procedures

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Abstract

Stream power is a widely used parameter to investigate, engineer, and manage river systems. The varied uses of stream power are increasing as it becomes easier to derive stream power using remotely sensed data coupled with Geographic Information Systems (GIS) and improved computational power and technology. This document was created to provide clarity to researchers and practitioners using stream power in their work. It includes a review of the physical basis, terminology, and applications of stream power; an examination of the many considerations and techniques for computing stream power; and a step by step example workflow to compute stream power using GIS and Microsoft Excel. The 33 steps in this example can be adjusted to suit individual needs of different projects.

1. Introduction

Sediment transport is a central concern for river scientists, engineers, and managers. It determines which areas along rivers are susceptible to erosion hazards, sedimentation, and impaired water quality. Sediment transport also modulates the natural functions of rivers of shaping landscapes and creating riparian habitat.

Stream power, which is essentially the product of stream discharge, stream slope, and the weight of water, has a direct relationship with sediment transport. This relationship has been used to characterize sediment transport and investigate geomorphic questions across a range of time and spatial scales, from the instantaneous rate of sediment transport at a river cross section (Bagnold, 1966) to the evolution of river long profiles across mountain ranges over geologic time (Kirby and Whipple, 2001). Stream power is being used increasingly among river scientists and engineers because it can be computed remotely without extensive field measurements. This allows researchers and practitioners to characterize sediment transport quickly and inexpensively along rivers. However, stream power is sensitive to choices in its computation, especially regarding data sources and the distance over which computations are averaged, commonly called the smoothing distance.

This document was created to provide clarity to researchers and practitioners using stream power in their work. The following sections (1) review stream power terminology, its physical underpinnings, its relationship with sediment transport, and some geomorphic applications; (2) discuss how stream power can vary based on data sources, smoothing distance, and other computational choices, and (3) present a step by step workflow for computing stream power from Digital Elevation Models (DEMs) using ArcGIS and Excel.

2. Stream power derivation and applications

2.1 Definition

The term stream power was originally coined by R.A. Bagnold (1960, 1966) when he used a first principles approach to quantify the rate of energy supply to a unit length of stream to investigate sediment transport in stream channels. Earlier work had introduced the idea that sediment entrainment is related to stream energy, which could be quantified as the product of river discharge, slope, and gravity (Gilbert and Murphy, 1914; Rubey, 1933; Knapp, 1938). However, Bagnold (1966) made a critically important contribution when he eloquently justified the importance of stream power, and, furthermore, developed and provided evidence for quantitative equations that predicted total sediment concentration. He defined stream power as the product of the river discharge, slope, and weight of water,

$$\Omega = \rho\, g\, Q\, S \tag{1}$$

where Ω is stream power, ρ is the density of water (1,000 kg/m^3 at 4 °C), g is gravity (9.8 m/s^2), Q is river discharge, and S is energy gradient, which is equivalent to channel slope in uniform flow.

Bagnold also defined the "mean available power supply to the column of fluid over unit bed area,"

$$\omega = \Omega\, /\, W \tag{2}$$

where W is the flow width.

In the time since Bagnold first defined Ω and ω, researchers have published inconsistent terminology for stream power (Rhoads, 1987), largely because it has been used in many different applications since Bagnold's seminal work. For example, ω has been called "unit stream power," "mean stream power," and "specific stream power." Moreover, "unit stream power" has been used to describe ω and VS ($\omega/\rho gh$ is equivalent to VS, where V is water velocity and h is water depth.) Additional terms, such has "total stream power," have been used inconsistently, sometimes referring to Ω, and sometimes referring to ΩL, where L is channel length.

Rhoads (1987) attempted to remedy this inconsistency by proposing a standard terminology, but subsequent authors have not adhered to the standardized nomenclature. Nevertheless, Eqs. (1) and (2) and the variables Ω and ω have been used consistently in most literature. Many authors have circumvented the issue of inconsistent terminology by referring to the variables directly and by carefully defining the variables with equations. We follow this practice and only use the term "stream power" for general statements that can be applied equally to Ω, ω, and other variants of stream power.

2.2 Physical underpinnings

Bagnold (1966) regarded streams as sediment transporting machines where the rate of work done is the efficiency times available power. He wrote "power—that is, a time rate of energy expenditure—is necessary to maintain [sediment] motion at a given time rate." He derived stream power using a first principles of physics approach, starting at the level of established science without making assumptions of an empirical model or fitting parameters. The derivation of stream power stems from conversion of water's potential energy (PE) into kinetic energy over time,

$$\Omega = \frac{\Delta PE}{\Delta t} = mg\frac{\Delta z}{\Delta t} \tag{3}$$

where t is time, m is water mass, and z is elevation. Stream slope (α, measured from horizontal) and downstream velocity (u) can replace the vertical component of velocity, $\Delta z/\Delta t$, because $\sin(\alpha) \approx \tan(\alpha)$ for small angles,
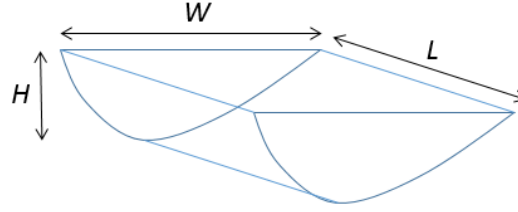
$$\frac{\Delta z}{\Delta t} = V\sin(\alpha) \approx V\tan(\alpha) = VS \tag{4}$$

By definition, mass is the product of density and volume,

$$m = \rho LWH, \tag{5}$$

where L is channel length, W is flow width, H is channel depth. The length term, L, goes to 1 because stream power is defined as per unit channel length.

Figure 1. Abstracted view of a river channel

Discharge (Q) is defined as

$$Q = VWH \tag{6}$$

Through substitution, we see that stream power is the product of the weight of water, discharge, and slope, as shown in Eq. (1),

$$\Omega = \rho g VWHS = \rho g QS \tag{7}$$

2.3 Relationship with shear stress

Shear stress, τ, is another commonly used parameter in sediment transport. For example, τ is the numerator in the computation of the Shields parameter, θ, which is used widely to determine if a given flow has the competence to transport sediment of a given size (Shields, 1936). In addition, many sediment transport equations are of the form $Q_s = a(\tau)^b$, where Q_s is the mass rate of sediment transported by a river, and a and b are fitting parameters (Meyer-Peter and Müller, 1948; Dietrich et al., 2003). Shear stress is defined as the product of river depth, river slope, and water weight,

$$\tau = hS\rho g \tag{8}$$

The relationship between stream power and shear stress is rather uncomplicated; ω is the product of shear stress and stream velocity. Combing Eqs. (2), (7), and (8),

$$\omega = \tau V \tag{9}$$

2.4 Sediment transport relationships

Stream power characterizes the driving force available for sediment transport. As evident in the preceding sections, Ω is derived from analogy of a machine performing work over time, in this case a stream moving sediment. Ω has a likeness to the mechanics of the displacement of sediment by the shearing action of the water, characterized by τ. However, Ω does not characterize all of the elements that affect the physics of moving sediment, for example the kinematic viscosity, grain sheltering and imbrication, roughness from sediment particles and bed forms, lift forces from turbulence, lift forces from pressure gradients, and organic mats (Hickin, 1995; Knighton, 1998). Thus Ω-based formulae represent a correlation, and they are no different than nearly all other sediment transport formulae which express a correlation between rate of sediment flux and driving force of the stream (Martin and Church, 2000).

Here, we describe four important characteristics of the well-established, positive relationship between stream power and sediment flux.

First, many studies have investigated several variations of stream power, showing that different manipulations of the term are preferred for different applications, yet the positive relationship between stream power and sediment flux holds true through all of these studies. For example, Bagnold (1960, 1966) plotted ω versus sediment flux, but later

work showed better agreement between ω/ρgh and sediment flux (ω/ρgh is equivalent to *VS)* and even better agreement between ω/ρghc and sediment flux, where c is particle settling velocity (Yang, 1972, 1974, 2006). In addition to experimental studies, Yang and Molinas (1982) showed theoretically that bedload, suspended load, and total load are directly related to ω/ρgh.

Second, this relationship holds true only above a threshold stream power, also known as a critical stream power, ω_c (Ferguson, 2005; Petit et al., 2005). Critical stream power increases with coarser sediment (see figure below). At very low stream power, sediment flux is minimal to non-existent, but above the threshold for sediment transport there is a clear positive relationship such that higher stream power produces greater sediment flux. This phenomenon is evident in many natural rivers. On most days, there is little to no sediment transport. Typically, sediment is mobilized only a few days per year when flows are high, causing stream power to exceed a critical value to initiate sediment transport. A threshold stream power is incorporated in some transport relationships, but they are not always more accurate than equations without criterion for incipient motion (Yang, 1979; Bagnold, 1980).

Figure 2. Critical stream power and sediment diameter, excerpted from Petit et al. (2005)

Fig. 3. Relations linking the critical specific stream power and the material size (D_{50} in mm) in Belgian gravel bed rivers and mountainous boulder rivers of southern France and Colorado. Bagnold I is the Bagnold's equation (Eq. (4)) with the Ourthe parameters (D_i=55–160 mm and d=2 m) and Bagnold II is Eq. (4) with the Rulles parameters (D_i=5–40 mm and d=1 m).
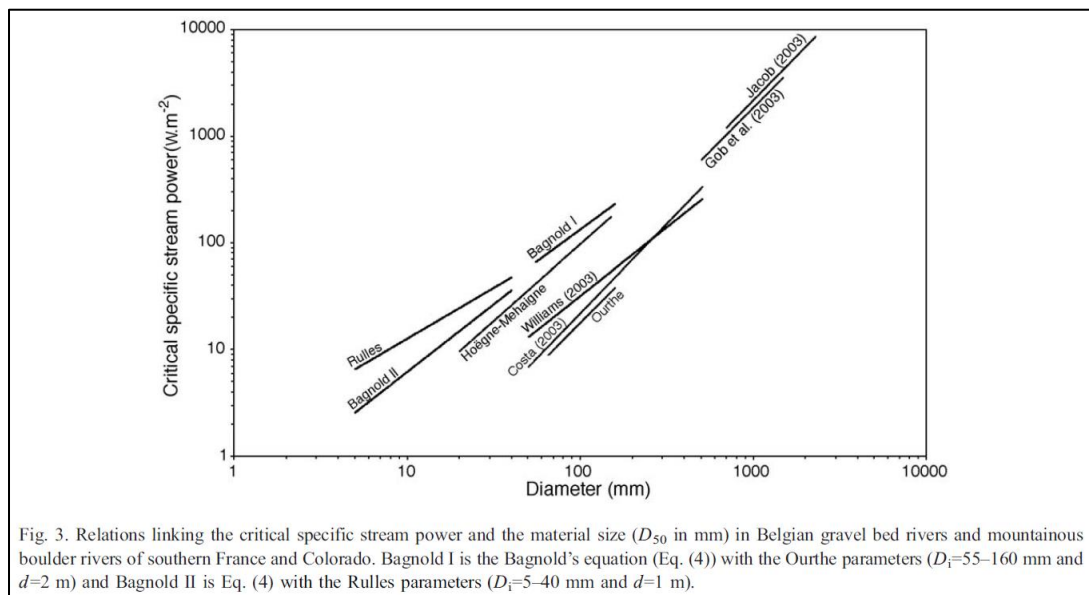
Third, most stream power equations predicting sediment flux apply to capacity limited channels, i.e. alluvial channels with non-cohesive sediment; however, the positive relationship between stream power and sediment flux can be applied broadly to bedrock channels and cohesive silt/clay channels. Studies of sediment transport in cohesive silt/clay channels typically examine τ as the driving force, sometimes including parameters for the thickness of the mobile layer and electrochemical cohesive properties affecting erodibility (Yang, 2006). Nevertheless these studies show that sediment transport increases with shear stress, suggesting that sediment flux also increases with greater stream power. In bedrock channels, *instantaneous sediment transport rates* as a function of stream power have not been studied intensively, largely because the physics of sediment transport in alluvial channels differs from that of bedrock channels, where abrasion, plucking and sliding are the processes required to mobilize material (Lamb et al., 2015). For a given stream power, a bedrock channel may have a lower sediment load than an alluvial channel because non-cohesive sediment is not readily available for transport. In these cases, easily mobile material can be transported from upstream alluvial reaches or recruited by bank erosion. Either way, stream power characterizes the vigor of sediment transport forces in bedrock channels, and therefore has been applied widely to examine *long-term incision rates* into bedrock channels (Dietrich et al., 2003). A rich body of research employs stream power to investigate the evolution of river longitudinal profiles (Royden and Perron, 2013), the interplay between tectonic uplift and river incision (Kirby and Whipple, 2001), and height limits of mountain ranges (Whipple and Tucker, 1999). Note that these studies consider geologic time scales, whereas Bagnold (1966) and Yang (2006) consider instantaneous sediment transport loads.

Fourth, Ω and ω do not account for resisting forces in sediment transport; however, some applications of Ω and ω do. In natural stream channels, some of the resisting factors for the initiation of sediment transport include the weight of sediment particles keeping them on the channel bed (related to particle size and typically quantified by particle diameter), vegetation that dissipates flow energy, roots that hold sediment, and cohesiveness of bed material (e.g. bedrock and clays versus sand and gravel), and grain sheltering and packing. Once sediment is in the water column, the settling velocity is an important factor, with larger particles settling faster than smaller particles. Some sediment transport equations based on stream power account for threshold to initiate sediment mobility, the submerged weight of sediment particles, and the settling velocity of particles (Bagnold, 1966; Engelund and Hansen, 1972; Yang, 1973, 2006). In contrast, other applications of stream power do not incorporate the resisting forces in sediment transport, relying on the stated or implicit assumption that the resisting forces are less important than the driving forces for the questions being asked. For example, studies often do not directly consider resistance to erosion when using the stream power to evaluate channel incision over very long time frames law (Dietrich et al., 2003). Also, stream power has been used to predict locations of erosion and deposition without incorporating a term for resistance to erosion in studies on large flood events, where high flows greatly exceed thresholds for erosion (Gartner, 2015).

2.5 Applications in geomorphology research

The preceding sections have touched upon several applications of stream power. The foundation for these applications is the direct relationship between stream power and sediment transport shown in flumes and natural channels in several studies, many authored by Bagnold (1960, 1966, 1977, 1980) and Yang (1972, 1973, 1979, 1984, 2006). This relationship has been utilized in engineering-style studies to quantify sediment loads and erosion/deposition budgets and also in geomorphology research to explain landscape form and processes. Stream power has been shown to influence channel pattern and form (Chang, 1979; Kondolf et al., 2003; Wohl, 2004), riffle and pool characteristics (Wohl et al., 1993), channel migration (Nanson and Hickin, 1986; Mcewen, 1994), channel aggradation and degradation (Bull, 1979), floodplain dynamics (Nanson and Croke, 1992), extreme geomorphic changes in floods (Magilligan, 1992; Buraas et al., 2014), stream bed grain size (Snyder et al., 2013), abundance of landslides (Larsen and Montgomery, 2012), and bedrock incision rates (Dietrich et al., 2003; Sklar and Dietrich, 2006; Ouimet et al., 2009).

While most of these studies focus on at-a-point magnitudes of stream power, a growing body of research has shown the importance of spatial gradients in stream power, i.e. whether steam power is increasing or decreasing with respect to distance downstream, on channel form (Bizzi and Lerner, 2015) and locations of erosion and deposition (Graf, 1983; Gartner et al., 2015a; Lea and Legleiter, 2016). The aforementioned studies use stream power as the independent variable affecting geomorphic form or processes, yet some studies flip this construct and examine stream power as a dependent variable that provides insight on long term effects of differential uplift (McKeown, 1988; Whipple and Tucker, 1999; Snyder et al., 2000; Kirby and Whipple, 2001) and rainfall (Schlunegger et al., 2011) across mountain ranges. Theoretical work has proposed that stream power should peak at intermediate locations within river basins (Knighton, 1999) because stream power is the product of discharge, which tends to increase with distance downstream, and channel slope, which tends to decrease with distance downstream. However, a variety of studies that show stream power as a function of distance downstream based on DEM analysis (Graf, 1983; Snyder et al., 2013; Bizzi and Lerner, 2015; Gartner et al., 2015a; Lea and Legleiter, 2016) demonstrate several peaks in stream power along single rivers due to geologic controls. This suggests there may be no normal pattern of downstream changes in stream power except potentially in purely alluvial rivers.

This review is far from a complete treatise of the applications of stream power but is meant to provide the reader context for the kinds of applications to which stream power may be applied. At present, Google Scholar shows over 2,000 citations of Bagnold's seminal 1966 paper, which is testimony to the multiple diverse ways that stream power has been used. The applications of stream power will only increase, especially as advances in DEM analysis and computational power and technology allow interrogation of downstream changes in stream power in ways that were not possible when field surveys were the primary method of computing stream power.

3. Computing stream power (Ω and ω), overview of considerations and approaches

Many approaches exist to compute stream power and resulting sediment transport. On one end of the spectrum is a highly detailed accounting of stream power to determine instantaneous rates of sediment transport at the scale of a river cross section or reach. In this scenario, studies often use field surveyed data and programs like HEC-RAS to compute discharge, slope, and sediment transport rates (Brunner, 2002; Gibson et al., 2006). On the other end of the spectrum is a more generalized computation of stream power used at larger scales to shed light on geomorphic form and processes (Wobus et al., 2006). These studies often employ GIS and remotely sensed data. The following sections lean toward the second approach. The sections first describe the methods used to derive Q, S, and w, then show how these are applied to computation of Ω and ω, and finally show by a set of instructions for computing stream power using GIS.

3.1 Computing stream discharge (Q)

The discharge one uses in stream power computations depends on the application of the analysis. Common choices are the peak flow of a single event (Phillips and Slattery, 2007; Gartner, 2015), bankful flow (Fonstad, 2003), mean annual flow (Finnegan et al., 2008; Larsen and Montgomery, 2012), mean annual maximum flow, and flow magnitude that typically occurs at a specific recurrence interval such as the 2-year, 10-year, or 100-year flow (Reinfelds et al., 2004; Phillips and Slattery, 2007; Phillips and Desloges, 2014).

Innumerable ways exist to determine the discharge at points along a river. Methods include direct field measurements, gaging records (requiring a developed stage-discharge relationship), area-discharge proxy, and empirical regressions based primarily on drainage area. In locations with sparse discharge measurements, such as the Himalaya, remotely-sensed precipitation data can be routed through DEMs for broad scale analysis (Finnegan et al., 2008; Larsen and Montgomery, 2012).

The area-discharge proxy works well for many small to mid-sized watersheds. It has the form

$$Q = aA^b \tag{10}$$

where A is drainage area, and a and b are empirically derived. In an analysis of 1659 gaging station across the U.S., Finlayson and Montgomery (2003) found that drainage area (measured in km^2) explains 61% of the variation in mean annual maximum discharge (measured in m^3/s) with the formula

$$Q = 0.92A^{0.7} \tag{11}$$

The area-discharge proxy can also be utilized to determine other flows, such as the peak flow of a recent storm or the 2-year flood, based on area ratio of areas for the point of interest and a reference location with known discharge (often the place of a stream gage or a field measurement) with the formula

$$Q_i = Q_{ref} (A_i/A_{ref})^b \tag{12}$$

where Q_i is discharge at the location i, Q_{ref} is discharge at a nearby reference location, A_i is contributing area at location i, A_{ref} is contributing area at the reference location with known discharge (often the place of a stream gage or a field measurement), and h is the same empirical constant in Eq. (10). Commonly b is set at 1 for simplicity (Gartner, 2015; Lea and Legleiter, 2016), or it can be refined for specific regions, for example 0.8 in New Hampshire and Vermont (Dingman and Palaia, 1999). Values for b from previous studies are in the range of 0.6 to 1.0 (Phillips and Desloges, 2014).

Empirical relationships also exist to determine the discharge for a given recurrence interval at a given location. These are compiled in the National Streamflow Statistics Program and used extensively in the USGS StreamStats Program (Turnipseed and Ries, 2007; Ries et al., 2008). In Massachusetts, for example, flood flows of specific recurrence intervals may be estimated for natural-flow streams with drainage areas between 0.25 square miles and 260 square miles based on drainage area, main-channel slope, mean basin elevation, and the area of swamps, lakes, and ponds (Wandle, 1983). Often the most heavily weighted variable in these regression equations is drainage area, which indicates that a discharge-area proxy equation would describe most of the variability between sites within a

roughly homogenous region. Flow computations may be improved with the added parameters of empirical relationships, but the improvement comes at the expense of computational efficiency.

3.2 Computing slope, or stream energy gradient (S)

There are several ways to derive the energy gradient, S, for stream power computations, and a common concern with all methods is the smoothing distance. The ideal method depends on the application. Since the data sources and data manipulations can affect the results, as with any study, it is critical to state explicitly how S is determined in stream power computations.

For spot measurements of stream power at specific locations, there are five choices to determine S: (1) hydraulic modeling, (2) field surveys of water surface, (3) field surveys of channel bed, (4) topographic map analysis, and (5) DEM analysis. Each have benefits and limitations. The first method, hydraulic modeling, can be a robust method to determine S at a variety of discharges, but it can require substantial computational time and input data. HEC-RAS, a widely used hydraulic modeling program, requires several cross sections from field surveys or high resolution DEMs with channel bathymetry. The second method, field surveying of water surface slope, is effective because the water surface slope is equal to the energy gradient in uniform flow (not accelerating or decelerating). However, field surveys of water surface slope can be difficult, especially in dense riparian vegetation and in very low gradient rivers. Furthermore, the water surface slope can differ between high and low discharges, thus a survey during a low flow period may not replicate the slope during a higher flow event. The third method, field surveys of the channel bed, are equally if not more difficult, especially due to riffles and pools in rivers, and are generally less desirable than surveys of the water surface slope. The fourth method, topographic map analysis, has the advantages of being expedient and replicable, but slopes from topographic maps may differ from the energy gradient. In a study of large magnitude floods in the Upper Mississippi Valley, Magilligan (1988) found that both water surface slope and map slope generally over predicted energy gradients determined from HEC-RAS modeling. With each of these four methods the inherent smoothing distance is a result of the spacing of measurements, whether it is the spacing between cross sections in a hydraulic model, spacing between field survey points of water surface or bed surface slope, or spacing between contour lines on a topographic map.

3.2.1 Sinks, smoothing, and other DEM-derived slope considerations

The fifth method, DEM analysis, is the standard method to determine S for continuous measurements of stream power along river channels. DEMs can also be used for local computations of stream power. A normal workflow (described in detail in Section 3.4) is to (1) prepare the DEM for hydraulic modeling, (2) extract elevation at points along the stream to produce a long profile (elevation versus distance downstream), (3) derive slope from long profiles, and (4) smooth the long profile. DEM analysis is expedient, replicable, and does not require field measurements; however, care must be taken to correct for inaccuracy in the DEMs. Some examples of errors in DEMs come in the form of sinks, peaks, and striping. Revised methods to create DEMs have reduced or eliminated striping, especially in recent DEMs in the U.S. National Elevation Dataset. If striping exists, cubic convolution and bilinear interpolation are preferred methods for filtering or resampling DEMs to produce a smoother appearance.

Sinks are likely to occur in all DEMs. In 30-m DEMs, it is common to have sinks in 1 percent of the cells. Filling sinks is a routine operation in creating hydrological enforced DEMs, but this manipulation of the elevation of some cells can affect local slopes along the river. Nevertheless, sinks should be filled before extracting slope along river profiles.

Another issue with DEM analysis is misalignment of channel location between (a) flow paths derived from the DEM, (b) the GIS data in the National Hydrography Dataset (NHD), and (c) the actual location of the stream. This can result in incorrect stream lengths in slope calculations if using a stream network derived from the contiguous low points in a DEM. Also, inaccurate elevations may result if elevations are simply extracted along a stream line that does not correspond to the contiguous low points in the DEM. This issue can be resolved by "burning" or "etching" the known stream line into the DEM (Maidment and Djokic, 2000) . Note that the NHD stream lines are imperfect, and some research groups have spent considerable time correcting the NHD for their study areas.

The resolution and extent of DEMs can also be a concern. On one hand, very coarse DEMs tend to decrease mean river slope, increase mean drainage area, and reduce channel lengths (Finlayson and Montgomery, 2003). Yet on the other hand, very high resolution DEMs derived from lidar can have accentuated errors in water surface elevation and steam length (Snyder, 2009; Gartner et al., 2015b). Lidar DEMs are often resampled to decrease the resolution (Slovin, 2015). Finally, when performing stream power analysis at the scale of continental mountain ranges, such as across the Himalaya, the curvature of the earth can create a length and area distortion in the two-dimensional plane of a DEM (Finlayson and Montgomery, 2003).

Many of these potential errors boil down the question of what distance should slope be averaged to best represent the river and its geomorphic processes. Smoothing removes the effect of minor errors in the DEM. It also removes the effect of localized non-uniform flow, making the channel slope a better approximation of energy gradient. Different researchers have chosen different smoothing distances and different smoothing techniques depending on whether the analysis focuses on the reach, watershed, or mountain range scale. Some example smoothing distances are 1-10 m (Worthy, 2005), 120 m (Lea and Legleiter, 2016), 200-1,000 m (Gartner et al., 2015a), 0.5 to 2 km (Phillips and Desloges, 2014), and 10 km (Finnegan et al., 2008).

There are no hard fast rules about smoothing distances, but one proposal is 1/10 of the square root of the drainage area for watershed scale analysis (Kasprak et al., 2012). Phillips and Desloges (2014) propose to incrementally increase the smoothing scale until the profile looks acceptable, and assess if the downstream slope variations are truly from topographic change, data artifacts, or imposed variations like dams. They also provide insightful review of the considerations in smoothing DEM-derived slope for stream power analysis. The method presented in the workflow below is to smooth the slope to an amount that is reasonable for the data source, then compute stream power, then smooth the stream power to an amount that fits the scale of the research question (e.g. reach, watershed, or mountain range).

3.3 Computing channel width (*W*)

Width must be evaluated to compute ω. The width parameter is flow dependent. Some common choices are the channel width, the total width of an individual flood, or the 100-yr flow width, and the choice depends on the aim of the study. Four potential data sources for *W* are hydraulic modeling, aerial photography analysis, field measurements, and empirical hydraulic geometry relationships. An advantage of using hydraulic models such as HEC-RAS to simulate channel width and total flow width is that a user can select to compute ω for just the channel area or for the entire width of flow. As mentioned above, the disadvantage of HEC-RAS and other hydraulic models is that they require detailed field surveys or high resolution DEMs with channel bathymetry.

Aerial photograph analysis is commonly used for continuous downstream measurements of *W*. The disadvantage is that *W* can be difficult to discern if the channel edge is among riparian vegetation, especially in small headwater streams. Another disadvantage is that total flow width in a flood may or may not be obvious in imagery.

Some studies have used empirical relationships between channel width and drainage area to estimate *W*. These hydraulic geometry relationships have the form

$$W = aD^b \tag{13}$$

where *D* is drainage area and *a* and *b* are regional-specific parameters for alluvial rivers (Leopold and Maddock Jr, 1953; Bent and Waite, 2013). Finnegan et al. (2005) proposed an alternate, more versatile empirical relationship that can be applied to bedrock, boulder, cobble, or gravel river beds,

$$W = 5 \, [\alpha(\alpha + 2)^{2/3}]^{3/8} Q^{3/8} S^{-3/16} n^{3/8} \tag{14}$$

where α is width-to-depth ratio and *n* is roughness

An advantage of these empirical relationship approaches is that *W* can be computed from a DEM based on flow accumulation area. However, local variances in *W* from the empirical relationship can be critically important for ω. Thus it is desirable to use satellite imagery or field measurements to determine *W* more directly if possible (Montgomery and Gran, 2001; Whittaker et al., 2007).

4. Step-by-Step Procedures for computing stream power using GIS analysis

Here we provide an example workflow for one way to compute Ω and ω using GIS analysis, from obtaining a DEM to creating a plot of distance downstream versus stream power. The method is performed in ESRI ArcGIS including the Spatial Analyst extension and the Arc Hydro Tools toolbar (ESRI, 2011) and Microsoft Excel including the free HydroTools add-in (Renshaw, 2016). The ArcGIS tools (marked in bold type in this documents) can be found efficiently in ArcMap using the search function or using the Arc Hydro Tools toolbar. Performing these steps requires proficiency with ArcGIS and Excel. In an attempt to show all the major steps without excessive text, some details are omitted. Several tutorials, user manuals, technical documents, blogs, and online user forums provide in-depth information on how to perform individual steps, trouble shoot problems, and avoid mistakes. There are several recommended resources for tips on working with DEMs (Frye, 2008), comprehensive treatises on using Arc Hydro Tools (ESRI, 2011; Djokic, 2012; Merwade, 2012), and using the HydroTools add-in (Renshaw, 2016).

General tips:

(1) Some of the ArcGIS tools are found most easily in the Arc Hydro Tools toolbar. Others are found most easily by clicking the Search window icon 🔍, or clicking **Windows > Search** on the main menu.

(2) It is best to disable background processing by clicking **Geoprocessing > Geoprocessing Options** on the main menu and making sure **Enable** is not checked.

(3) Also, make sure that the Spatial Analyst tools are enabled, by clicking **Customize > Extensions** and checking the box for Spatial Analyst.

4.1 Alternate workflows

This document details just one possible workflow relying heavily on ESRI ArcGIS, yet several other options exist both within ArcGIS and in other programs. Phillips and Desloges (2014) and Jain et al. (2006) both present workflows that rely heavily on ESRI ArcGIS and are similar to the one described below. TopoToolbox is a MATLAB program for the analysis of DEMs (Schwanghart and Scherler, 2014). It is free and open source, as long as you have MATLAB. This well-established program has been used in many peer-reviewed publications to analyze stream power, slope, and other parameters along river profiles, typically at the watershed to mountain range scale, rather than at the reach scale. River Tools (http://rivix.com/) and Tau DEM (http://hydrology.usu.edu/taudem/taudem5/index.html) are two additional standalone tools for DEM analysis

4.2 Prepare DEMs

Step A1: Obtain DEM data from USGS National Map. DEMs can be obtained from the USGS National Map at http://viewer.nationalmap.gov/basic/. Many sources for DEM data exist. Here we use 1/3 arc-second DEMs, which are suitable for watershed and reach scale analysis. The 1/3 arc-second DEMs are commonly called 10-m DEMs because they have approximately 10-m grid spacing, depending on latitude. The 1/3 arc-second DEMs are the highest resolution seamless data, currently available across the entire U.S. and its territories, excluding some portions of Alaska. Higher resolution DEMS, including approximately 3-m grid spacing and 1-m grid spacing from lidar data, are available in some locations via the USGS National Map and National Science Foundation's Open Topography websites. However, higher resolution DEMs do not always improve stream power computation. They require more computational time, and the results are nearly always smoothed over scales greater than 10 m. A descriptive list of commonly used DEMs across the globe is available at ESRI mapping Center (http://mappingcenter.esri.com/index.cfm?fa=arcgisResources.more).

Step A2. Combine multiple DEMs using Mosaic to New Raster. Often the project watershed spans multiple DEMs downloaded from the National Map. In ArcMap, use the tool **Mosaic to New Raster** to combine these into one single DEM. "Pixel Type" should be set to 32_BIT_FLOAT. This is critical for current USGS DEMs, and it is not the default for this tool. The "Number of Bands" should be 1 for DEMs. "Spatial Reference" and "Cellsize" can be blank, and "Mosaic Operator" and "Mosaic Colormap Mode" can be left in default positions of "last" and "first," respectively.

Mosaic To New Raster

Input Rasters

grdn43w073_13
grdn43w074_13

Output Location
C:\UMass_GIS\DEM\Deerfield

Raster Dataset Name with Extension
Dfield_UTM_m.img

Spatial Reference for Raster (optional)

Pixel Type (optional)
32_BIT_FLOAT

Cellsize (optional)

Number of Bands
1

Mosaic Operator (optional)
LAST

Mosaic Colormap Mode (optional)
FIRST

Output Location

The path to contain the raster dataset. The path can be to a folder or geodatabase.

OK   Cancel   Environments...   << Hide Help   Tool Help

Step A3. Project DEM data using Project Raster. In ArcMap, use the tool **Project Raster** to change the DEM from a geographic coordinate system to gridded projected coordinate system. The "Input Coordinate System" must be designated. For "Output Coordinate System," choose a gridded projection, such as UTM (the UTM zone depends on project location). For the "Output Raster Dataset," specify the file extension as .img (not essential, but recommended for this workflow). The "Resampling Technique" should be CUBIC. "Geographic Transformation" and "Output Cell Size" can be left empty.

Project Raster

Input Raster
Dfield_UTM_m.img

Input Coordinate System (optional)
GCS_North_American_1983

Output Raster Dataset
C:\UMass_GIS\DEM\Dfield_UTM.img

Output Coordinate System
WGS_1984_UTM_Zone_18N

Geographic Transformation (optional)

WGS_1984_(ITRF00)_To_NAD_1983

Resampling Technique (optional)
CUBIC

Output Cell Size (optional)

X   Y
9.06909748674863   9.06909748674865

Registration Point (optional)
X Coordinate   Y Coordinate

Output Cell Size (optional)

The cell size for the new raster dataset.

The default cell size is the cell size of the selected raster dataset.

OK   Cancel   Environments...   << Hide Help   Tool Help

Step A4. Remove extraneous area of DEM using either Clip or Extract by Mask. To reduce computation time for some of the more complex hydrological analysis tools, it is helpful to remove unnecessary areas of the DEM. One

approach is create a new polygon shapefile, make a polygon that is roughly around the watershed, and then use the shapefile in either the **Clip** or **Extract by Mask** tools. In current versions of ArcMAP, these tools have overlapping functionality. Tip: Add a basemap such as the USGS National Map to help locate the watershed boundary.
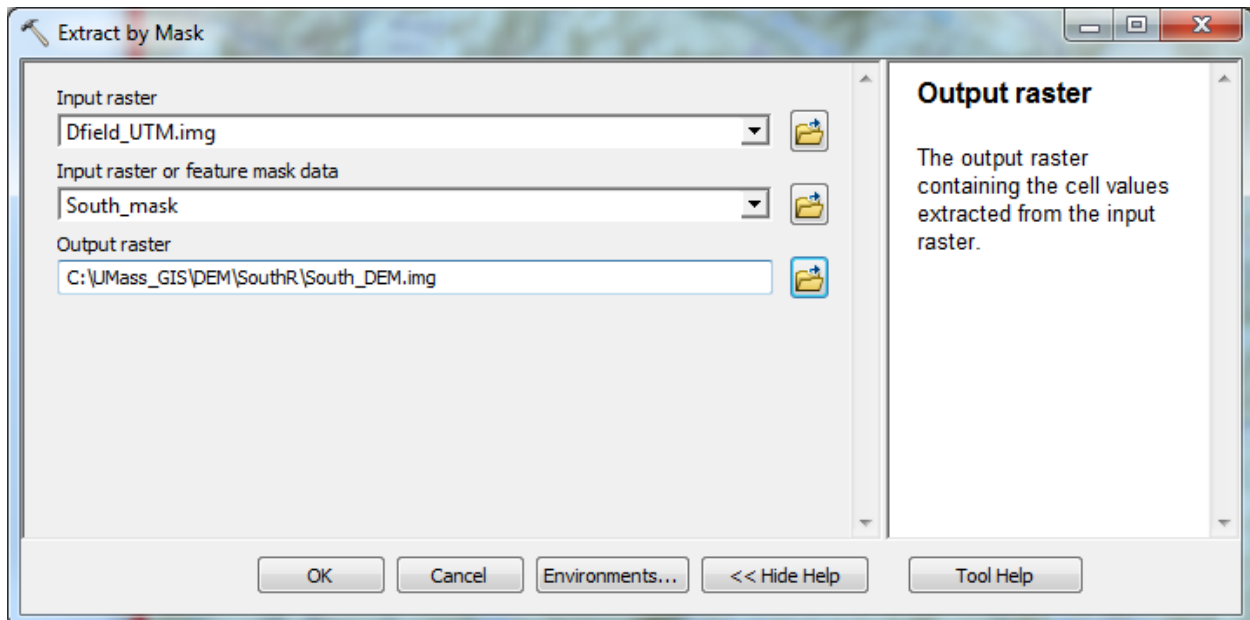
The following screenshot shows the ArcMap project at the end of this DEM preparation steps.
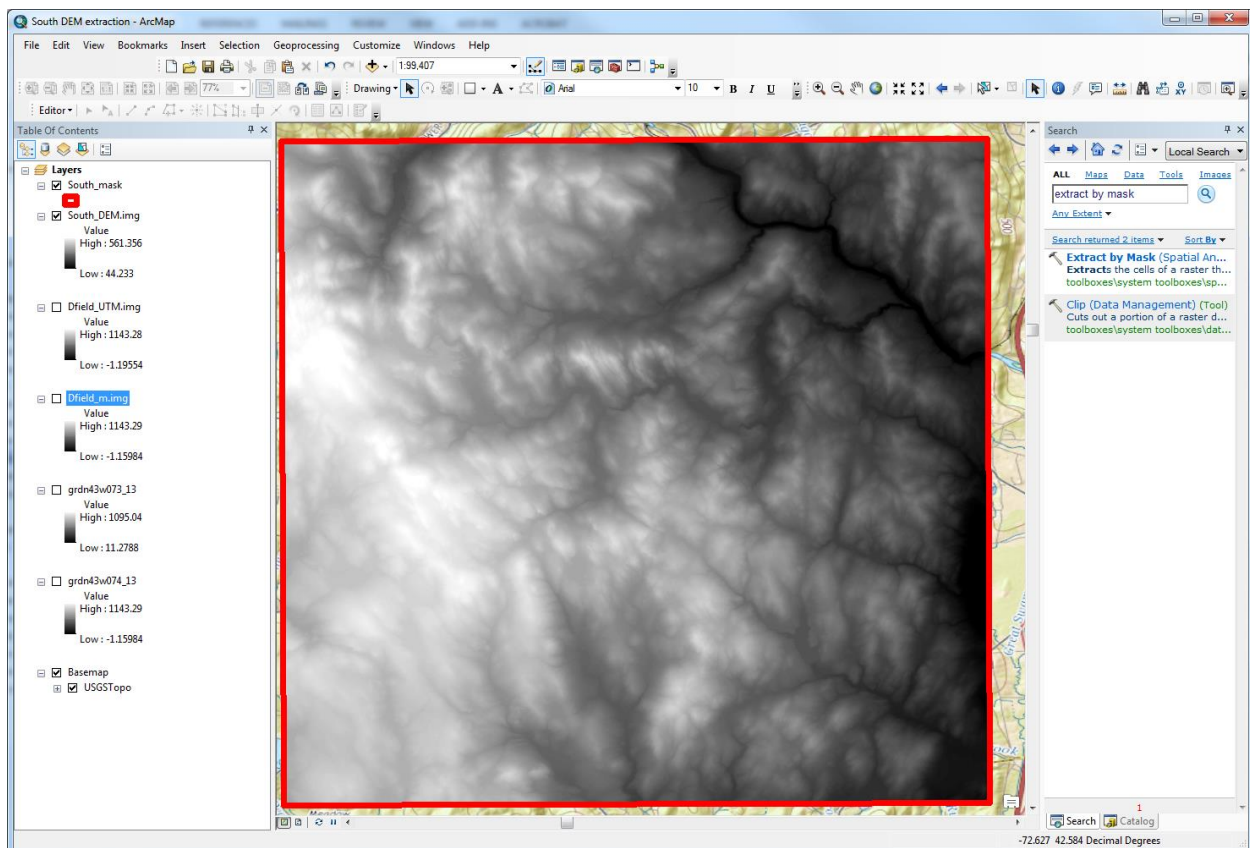
4.3 Hydrological analysis and extraction of long profile

The purpose here is to develop a hydrologically correct DEM to determine flow accumulation (the drainage area) and the elevation at points along the river (to derive slope). These tools are well organized in the Arc Hydro Tools toolbar, and thus they are performed quickly. However, if a misstep is made, it is typically easiest to abandon the project, start a new map in a new folder loaded with the DEM, and redo these steps. Several very comprehensive sources can assist with these steps (ESRI, 2011; Djokic, 2012; Merwade, 2012).

Step B1. Start new ArcMap project. Start a new ArcMap project, and make sure that the first data added is the clipped DEM from step A4 that has UTM coordinates. Save the project in a new folder. Although this may seem a trivial step, it is critically important to establish the projection of the map and location of the geodatabase where the layers generated from these steps are stored.
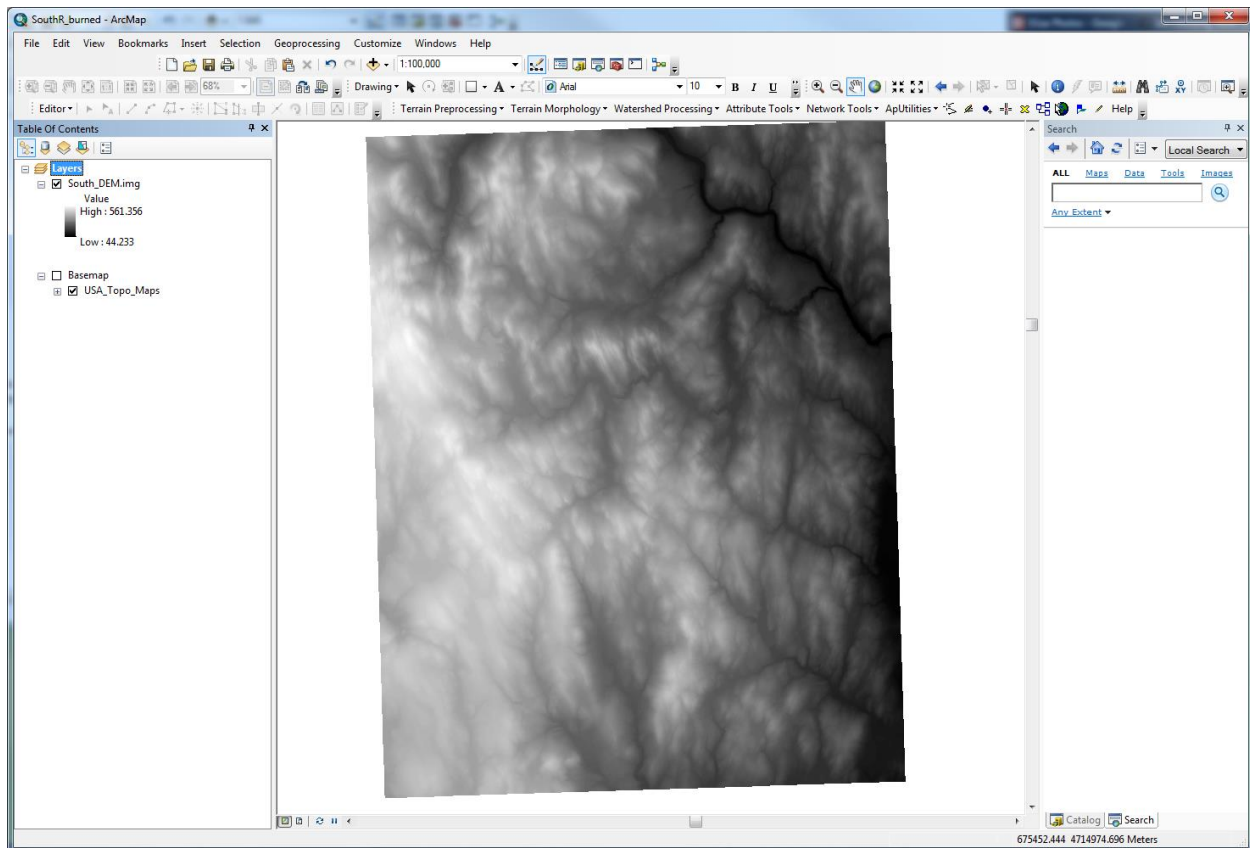
Step B2. Fill sinks. This step fills depressions in the DEM, many of which are artificial artefacts of DEM construction. However, some landscape have natural sinks, as in karst topography. The function is found in the Arc Hydro Toolbar under **Terrain Preprocessing > DEM Manipulation > Fill Sinks**. For standard applications, "Deranged Polygon" is set to Null and the "Fill All" button is checked. This produces a filled DEM with no sinks, and the DEM has the default name "Fil."
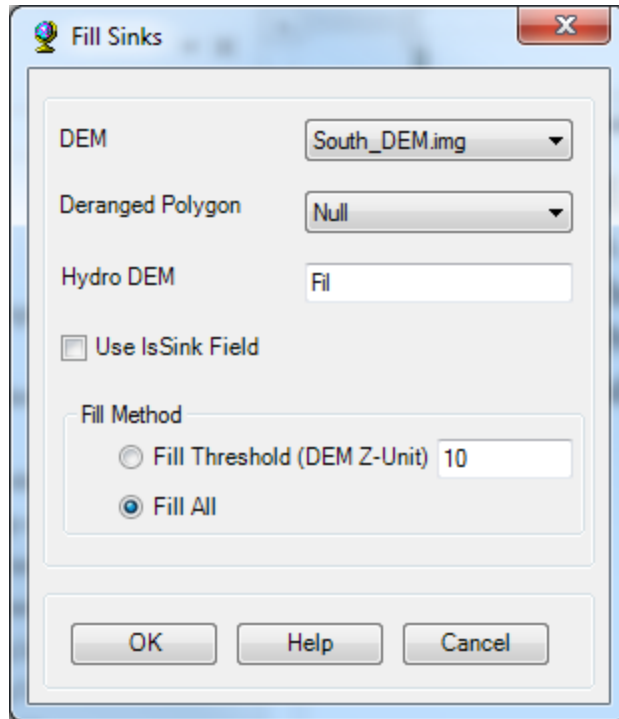
Optional Step B3. Burn streams through DEM reconditioning. Sometimes the synthetic drainage network derived from the DEM is inadequate. For example in a broad floodplain with low relief, the low points on the DEM may not align with the actual location of the river, largely due to subtle inaccuracies in the DEM. This can create a problem because the river length derived from the DEM differs from the actual river length, and therefore the resulting slope (rise/run) may be inaccurate. To force the derived stream network to the proper location, the known location of the streams can be "burned" into the DEM. This artificially lowers the elevations in the DEM at the location of the stream, and ensures that this is the path of the DEM-derived stream. This process should not be taken lightly, and Djokic (2012) explains many of the considerations. Burning streams is performed under "**Terrain Preprocessing > DEM Manipulation > DEM Reconditioning**." In this tool, the "Raw DEM" is the filled DEM from step B1, the "AGREE Stream" is the known stream layer. This layer must be carefully checked. The known streams should have the same scale and time of the DEM, have no braids, be dendritic, extend beyond the lowest point in the watershed, and not be closer to the watershed boundary than the stream buffer. The National Hydrography Dataset (NHD) is one GIS data source that shows the locations (but not elevations) of streams in the U.S.; however, the NHD has several errors that are evident upon close inspection. In this example, we used a GIS layer derived from NHD dataset that was then carefully checked for accuracy of stream locations. This stream network layer was created for a project called Designing Sustainable Landscapes at University of Massachusetts, Amherst (http://www.umass.edu/landeco/research/dsl/dsl.html).

When using the DEM reconditioning tool, the parameter "Sharp Drop/Raise" designates the amount that the DEM will be lowered at the location of the stream network. In this example, this value is chosen arbitrarily as 40 m, which is less than the lowest point on the DEM to insure that negative elevations are not created. Several iterations of the tool may need to be performed to obtain acceptable results by changing the three input parameters of stream buffer, smooth drop/raise, add sharp drop/raise (Djokic, 2012). The resulting raster is named by default "AgreeDEM." The effect of burning in the stream layer is shown in the figures below, evident in the 40 m elevation drop in at station 60.

DEM Reconditioning

Raw DEM	Fil
AGREE Stream	NHDhighRes_Deerfield_UTM
AGREE DEM	AgreeDEM
Stream buffer (number of cells)	5
Smooth drop/raise (DEM Z-unit)	10
Sharp drop/raise (DEM Z-unit)	40

OK   Help   Cancel

Profile Graph Title

Original DEM: Cross stream profile

Profile Graph Title

Burned DEM: Cross stream profile
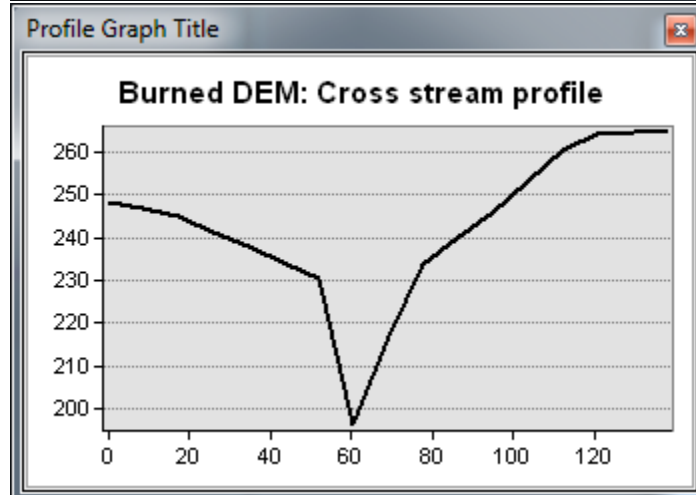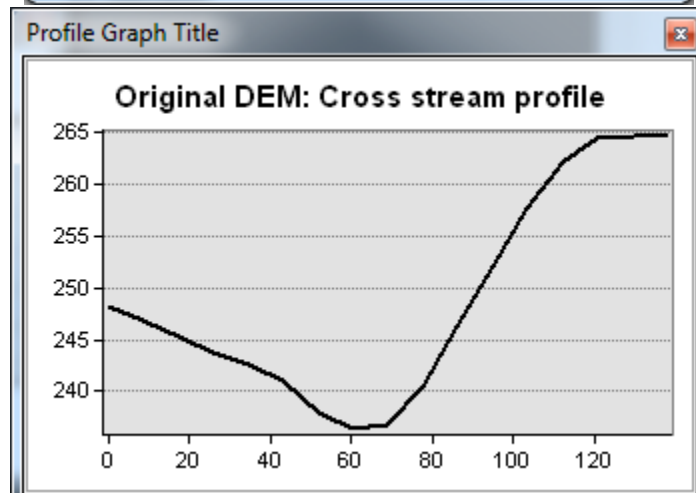
Optional Step B4. **Fill sinks** again. If a known stream network is burned in, then the reconditioned DEM should have the sinks filled again. The resulting raster can be named "Fil_burn."
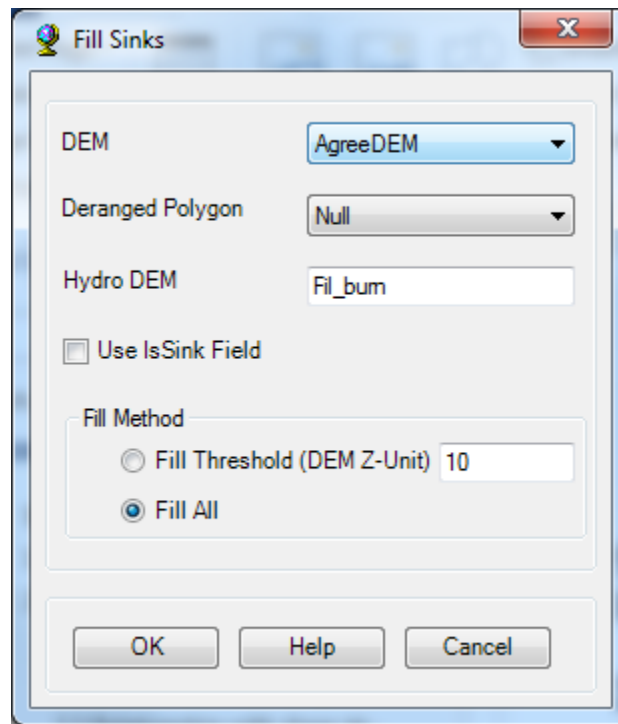
Step B5.   Flow Direction. This and the following steps are performed regardless of whether a known stream network is burned in or not. This step determines the direction that water flows based on the steepest descent direction determined from the 8 neighboring cells (D8 method). The tool is found under **Terrain Preprocessing > Flow Direction**. The input DEM is the filled DEM from the previous step "Fil," and the output is a new raster called "Fdr."
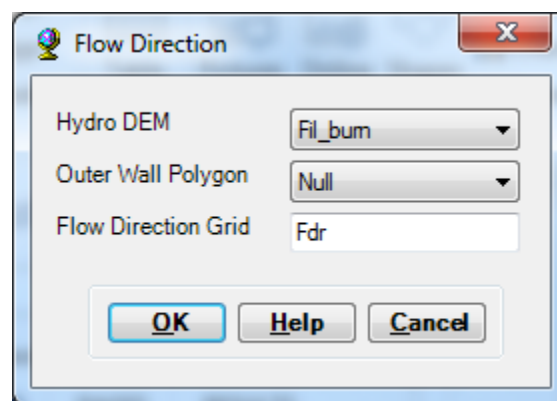
Step B6. Flow Accumulation. This step determines the contributing area for every cell in the filled DEM. This is a key layer in stream power computations because it defines the area used in the area-discharge proxy. The tools is found as **Terrain Preprocessing > Flow Accumulation**.
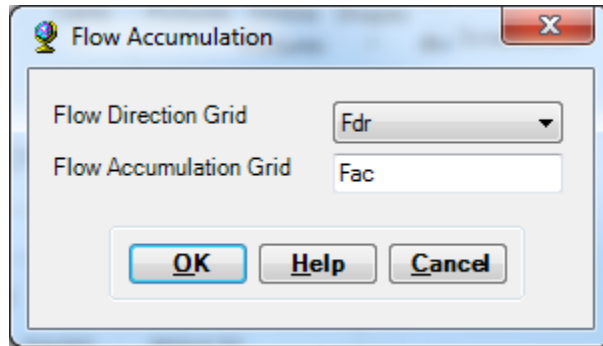
To expedite the process deriving steam power from a DEM, it is possible to skip from here to step B15. The watershed boundaries and synthetic stream network would not be created.

Step B7. Steam Definition. The tools is found as **Terrain Preprocessing > Stream Definition**. In this step, the user determines how much contributing area is required to start a channel. A small area will create a dense stream network with the heads of channels near the drainage divide. Several iterations may be required to obtain a synthetic (or "modeled") stream network that reasonably matches the actual stream network. A representation of the actual stream network can be viewed in base maps from USGS National Map, or from the National Hydrography Dataset. Djokic (2012) recommend keeping the threshold value between 0.5 % and 1 % of the greatest flow accumulation value, but that may not produce a desirable synthetic stream network. The default name for this file is "Str," but it is good practice to include the contributing area in the name, for example "Str_05km" for a 0.5 km^2 threshold. Note that when changing the "Area" parameter, the field in the "Number of cells" parameter is updated automatically.
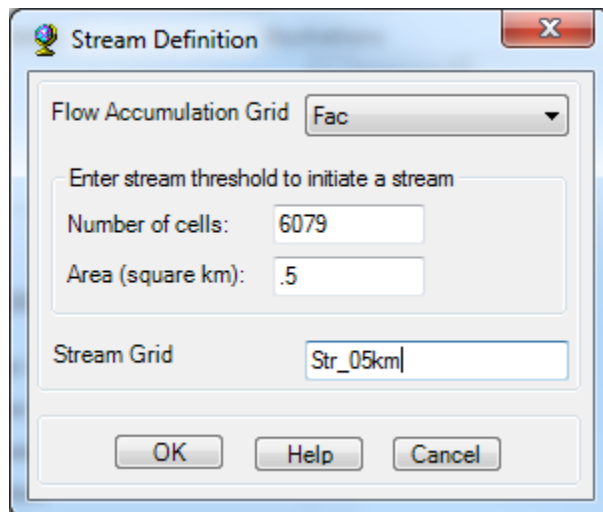
The following three examples show DEM-derived synthetic stream networks a different thresholds compared with the National Hydrography Dataset (blue lines), which are derived primarily from the blue lines on USGS 15-minute topographic maps. The 0.5 km^2 threshold streams (red lines) are less dense and have shorter channels than the NHD, the 0.1 km^2 (black lines) threshold streams are more dense and longer than the NHD. The 0.15 km^2 threshold streams (green lines) achieve a reasonable solution of imitating the NHD streams for this basin. The channel lengths are close to those of the NHD (some longer, some shorter), and the channel network is not exceedingly dense. This threshold will differ across geographic regions. GeoNet (https://sites.google.com/site/geonethome/) is a computational tool for automatic extraction of channel networks and channel heads from high resolution topography that uses steep channel banks to extract channel location and channel heads; therefore it is not affected by the differing drainage area of different channel heads.
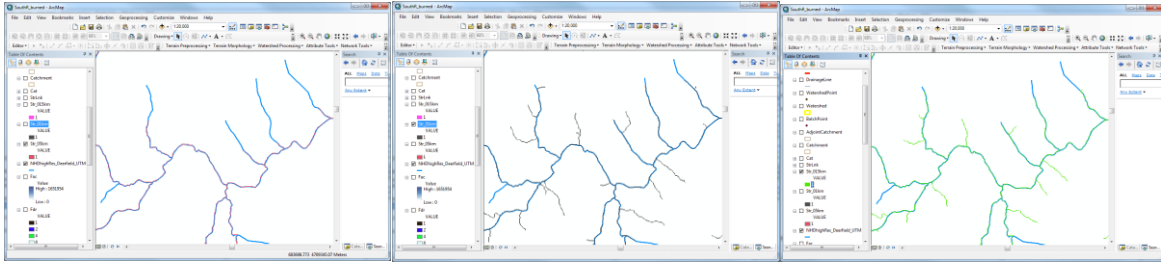
These preceding examples show the difficulty of replicating real stream network, and there are added complications when the known stream network is not burned in, as shown below. The green lines are a synthetic stream network with a 0.15 km^2 threshold, but the channel locations were not burned in to the DEM. The blue lines are the NHD channels, as shown in the previous examples, considered the actual location of channels. There is not only a discrepancy in channel length, but also notable difference in channel location. The synthetic streams do not always occupy the inflection points (crenulations) of contour lines, which are the local low points. Furthermore, this mountainous topography makes a relatively easy terrain to model stream locations compared to regions with broad flat topography. As discussed in section 3.2.1, the inaccurate stream lengths can affect river slope computations. Despite these inaccuracies, there is immense potential in being able to derive a roughly accurate stream network in a few minutes that can be used to for a variety of hydrological analyses.
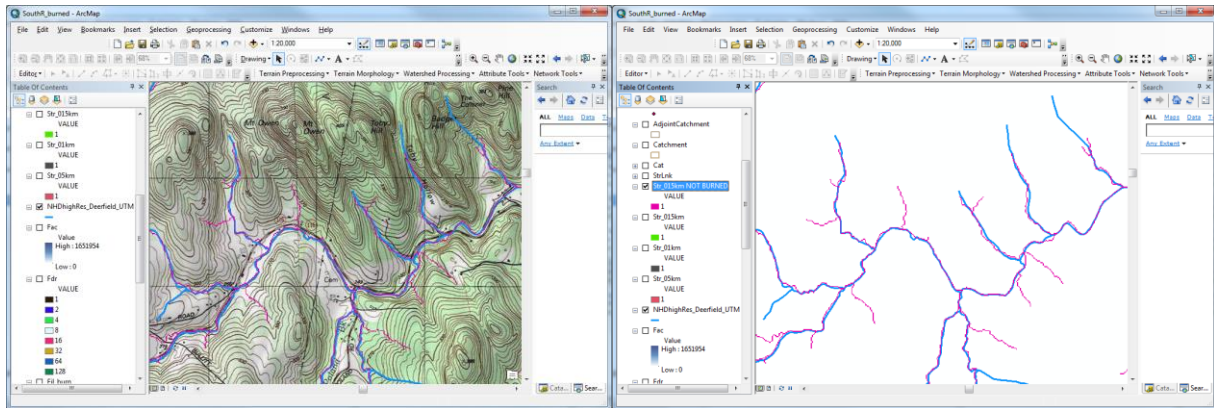
Step B8. Stream Segmentation. This step uniquely numbers stream segments between confluences. The tool is found under as **Terrain Preprocessing > Stream Segmentation.** For "Stream Grid," use the best choice from Step B7, in the example a 0.15 km^2 threshold. The "Sink Watershed Grid" and "Sink Link Grid" should be set to Null so that the whole Dem is processed.
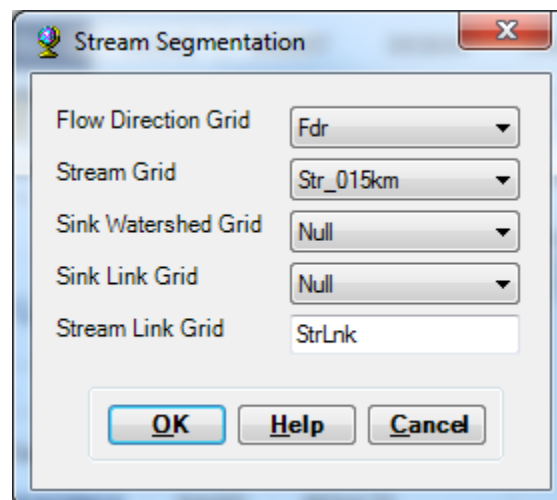
Step B9. Catchment Grid Delineation. This step identifies drainage areas that drain to each stream link. The tool is found under as **Terrain Preprocessing > Catchment Grid Delineation**.
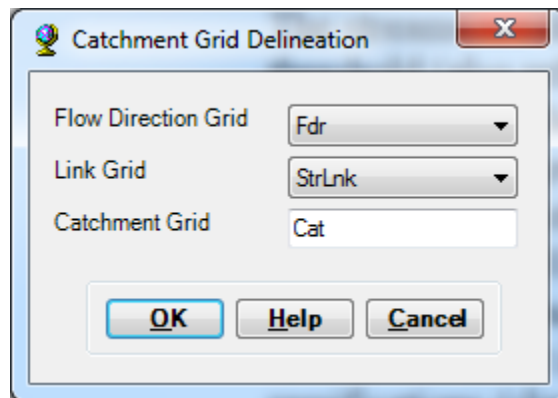
Step B10. Catchment Polygon Processing. This step defines catchments in vector format. The tool is found under as **Terrain Preprocessing > Catchment Polygon Processing**.
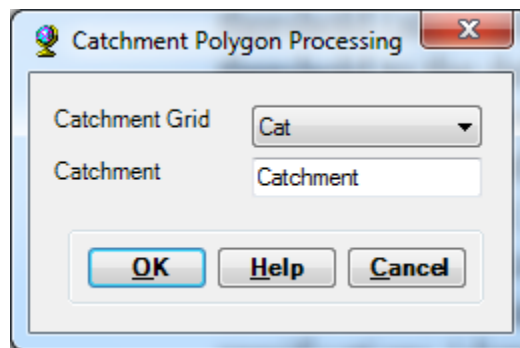
Step B11. Drainage Line Processing. This step defines stream segments in vector format. The tool is found under as **Terrain Preprocessing > Drainage Line Processing**.
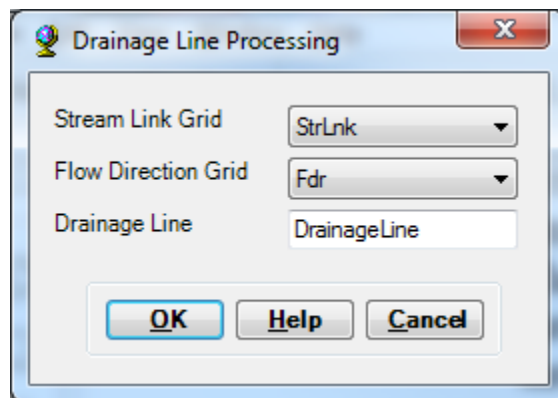
Step B12. Adjoint Catchment Processing. This step determines the cumulative area upstream from a catchment in vector format. The tool is found under as **Terrain Preprocessing > Adjoint Catchment Processing**. At this point, the stream network and catchment units need to delineate watersheds.
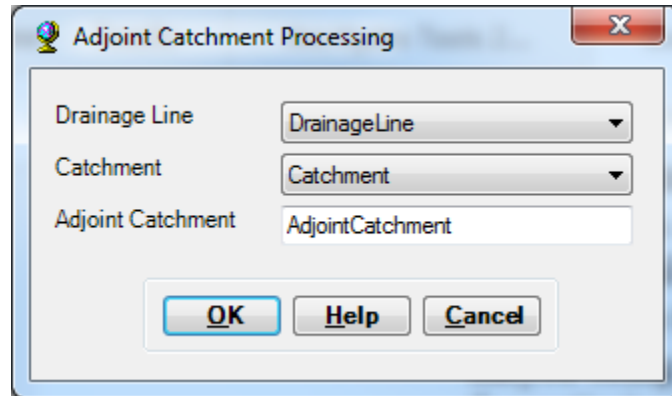
Optional step B13. Identify the outlet of the watershed using Batch Point Generation icon. Before delineating a watershed boundary, one must identify the outlet of the watershed. It is helpful to zoom in on this location, and use a base layer like a topographic map or aerial imagery. From the Arc Hydro Tools toolbar, click on the **Batch Point Generation icon** (looks like a yellow "x"), click "OK" in the dialog box, then click in the map area at the point that is the outlet of the watershed.
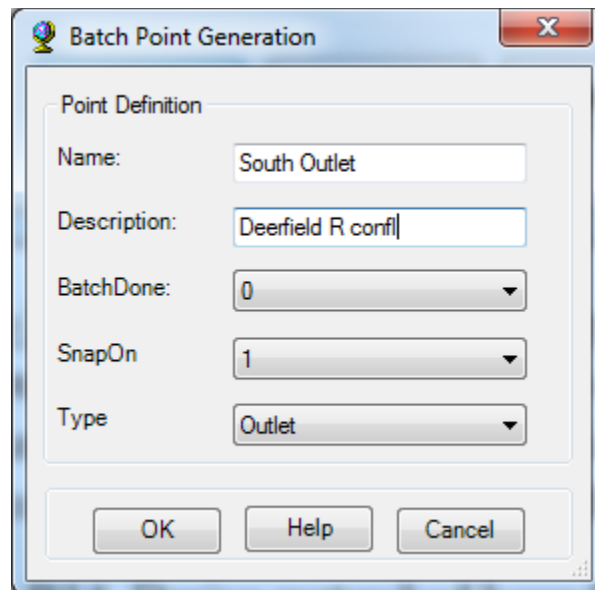
Optional step B14. Derive watershed boundaries using Batch Watershed Delineation. From the Arc Hydro toolbar, choose **Watershed Processing > Batch Watershed Delineation**. This creates a polygon of the watershed. The stream within this watershed can be clipped using the **Clip** tool. Incidentally, if one wants only a watershed polygon shapefile, it can be created in USGS StreamStats, a web-based GIS application without performing any of these steps.
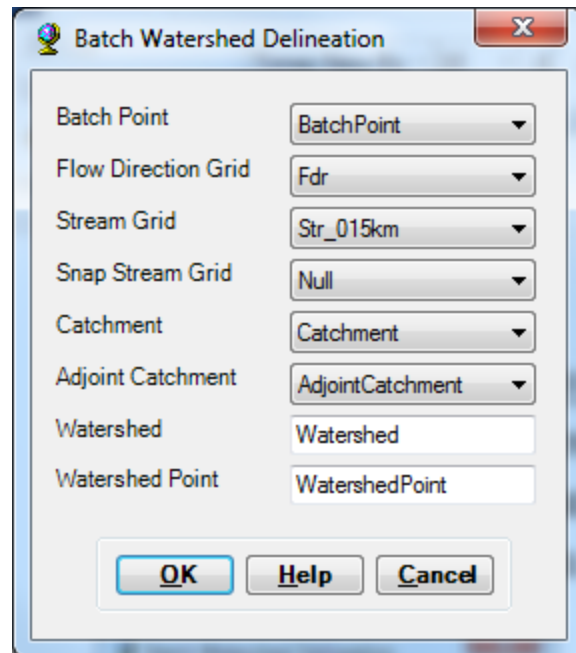
Step B15. Produce a single stream line. Trace a flow path using **Interactive Flow Path Tracing** tool on the Arc Hydro Tools toolbar. Click on the icon that looks like a hand with a pen, ✐ , then click on the upstream point on main channel, then click "yes" in the dialog box. The default name for this layer is "LongestFlowPath."

** To expedite the process of deriving stream power from a DEM, it is possible to skip to Step B15 (Produce a single stream line) directly from Step B6 (Flow accumulation). The watershed boundaries and synthetic stream network would not be created. To find the top end of the stream channel, use a basemap from the "USA Topo Map" under the "Add Data" icon. A watershed polygon shapefile can be created in USGS StreamStats, a web-based GIS application **
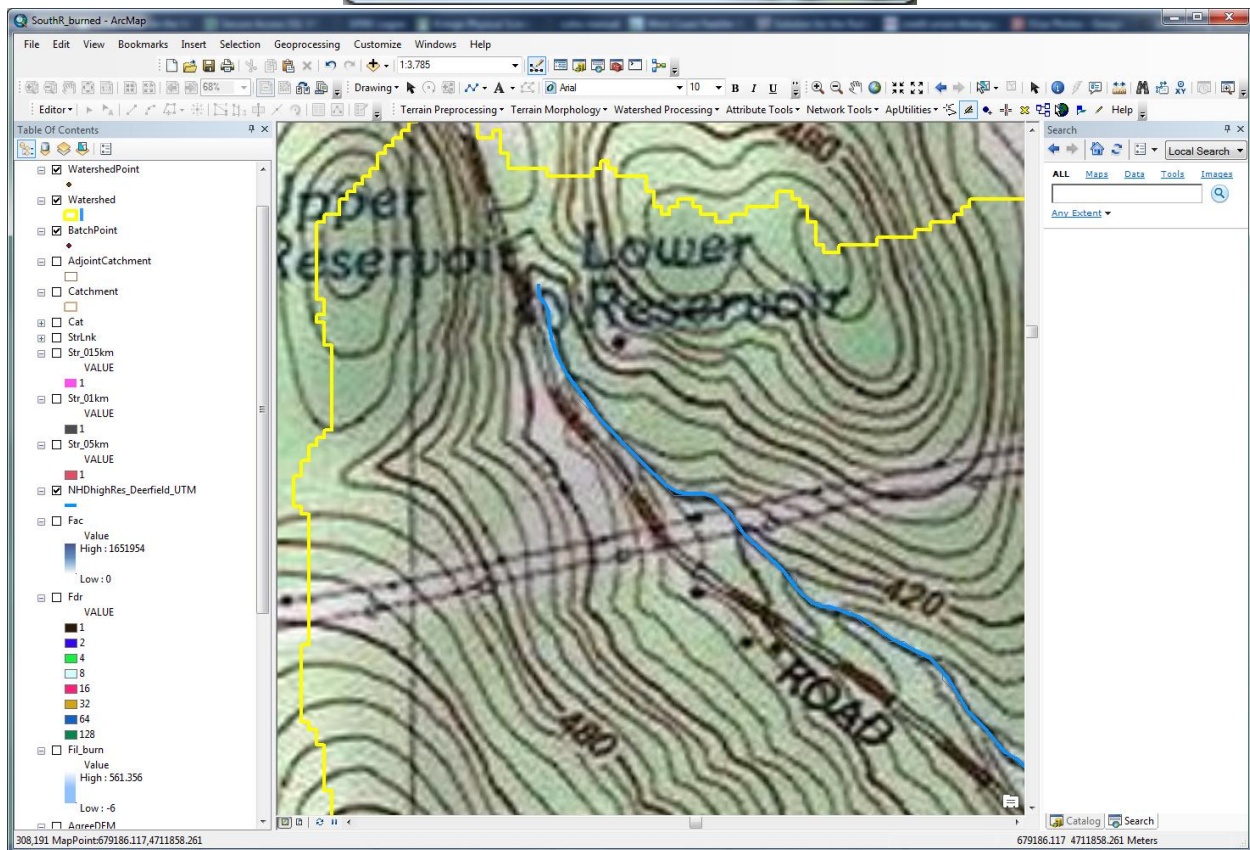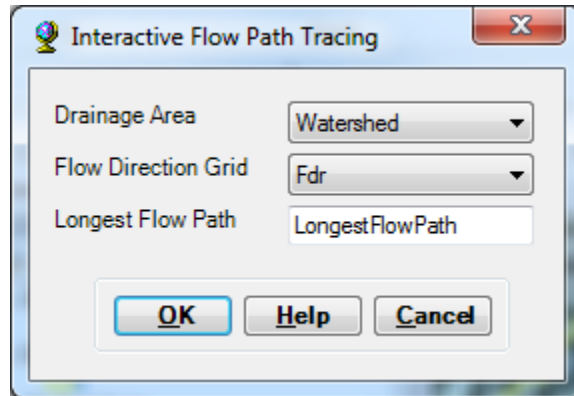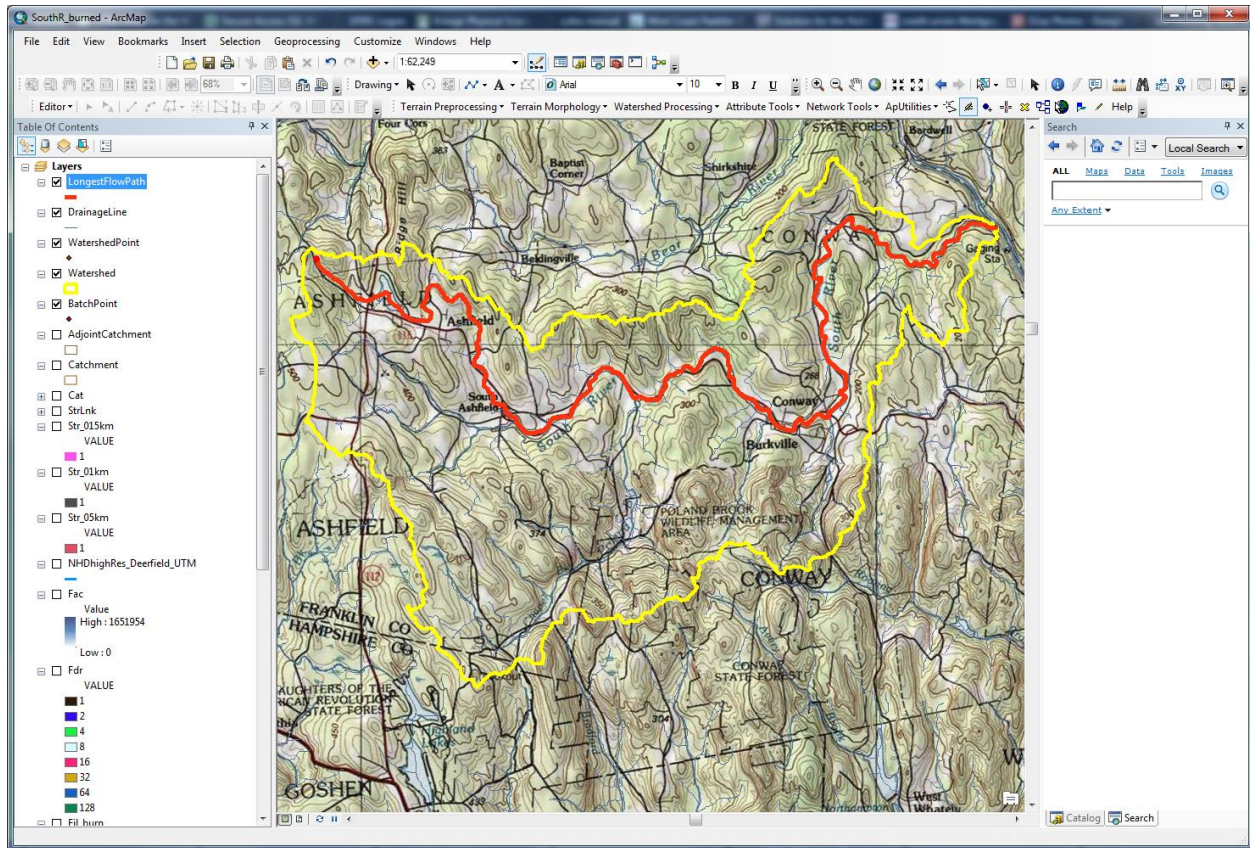
Step B16. Create a polyline shapefile for the single stream line. Right click on the "LongestFlowPath" layer in the Table of Contents, click **Data > Export Data**. Click "yes" to add exported data to the map as a layer.

Step B17. Create a point shapefile and construct points along the single stream line. Create a point shapefile, add this empty shapefile to the map, and start an editing session for this point shapefile. Using the edit tool on the editing toolbar, click on the stream line from the polyline shapefile created in step B16. The stream will be highlighted light blue. On the Editing toolbar choose "**Editor > Construct Points…**" in the tool dialog box, the template is the point shapefile just created. Click the radio button "Distance," and designate the spacing along the line. Since the map projection and shapefile projection are both UTM projections, then the units of this distance are unquestionably meters. The point distance is the user's choice, and here 10 m is used for this medium sized watershed. The arrows on the highlighted stream polyline show the start and end of the feature. Select the "Orientation" such that points are constructed from upstream to downstream along the polyline. For example, the button "From start point of line" should be selected if arrows point in the downstream direction, and "From end point of line" if arrows point in the upstream direction. (The default start point of the polyline may be either the upstream point or downstream point depending on how it was created.) Save and stop editing.
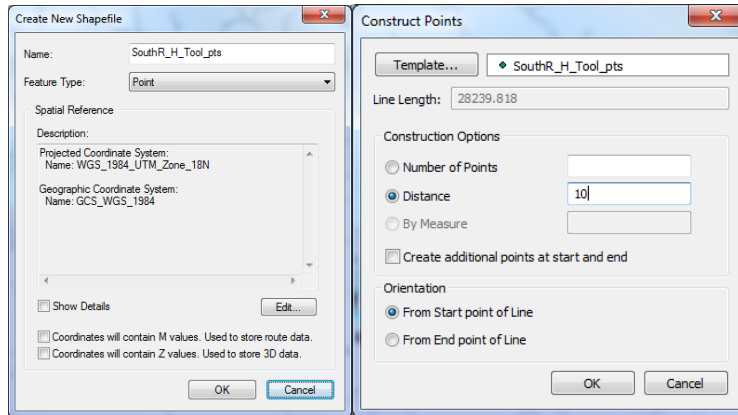
Step B18. Extract multi values from the DEM to the point shapefile. Select the tool **Extract multi values to points**, which adds one or more columns to the attribute table of point shapefile, and records the value from one or more rasters at each point location. This is a spatial analyst tool. The "Input point feature" is the point shapefile created in Step B17. The input rasters are the flow accumulation raster which shows the contributing area at each point, and the filled DEM which shows the elevation at each point. Do <u>not</u> check bilinear interpolation, because this can assign a value to a point that is interpolated between cells that are along and not along the stream path. This would result in some very low discharge values and incorrectly high elevation values at some points. Open the attribute table by right clicking on the point shapefile in the Table of Contents.
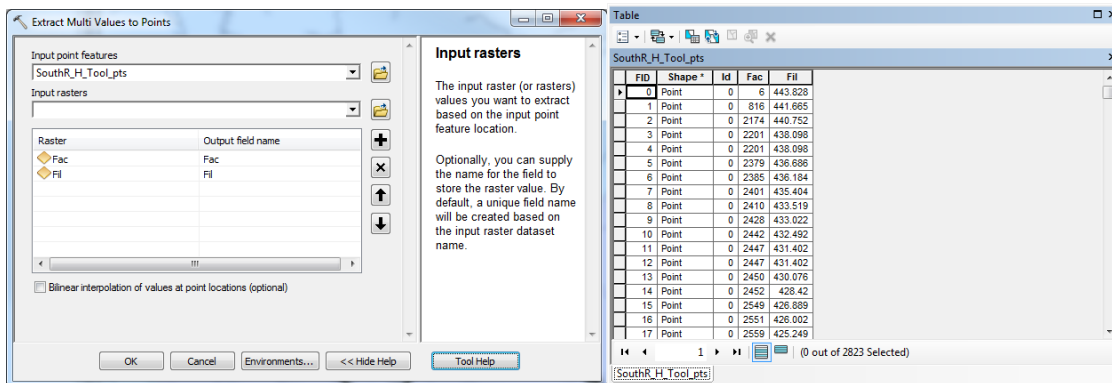
Step B19. Add columns with easting and northing values to attribute table. With the attribute table open, choose "**Table Options > Add Field…**" In the dialog box, give the field name "easting" and, unless sub-meter precision is desired, choose the Long Integer for "Type." Choose Double if submeter accuracy is desired. Next, right click the field header, then choose "**Calculate Geometry**…" In the dialog box, choose X-coordinate for easting (and Y-coordinate for northing). Select the button "Use the coordinates for the data source" and keep the units in meters. Repeat these steps for northing. Make sure no rows are highlighted, or only that data point will be updated.
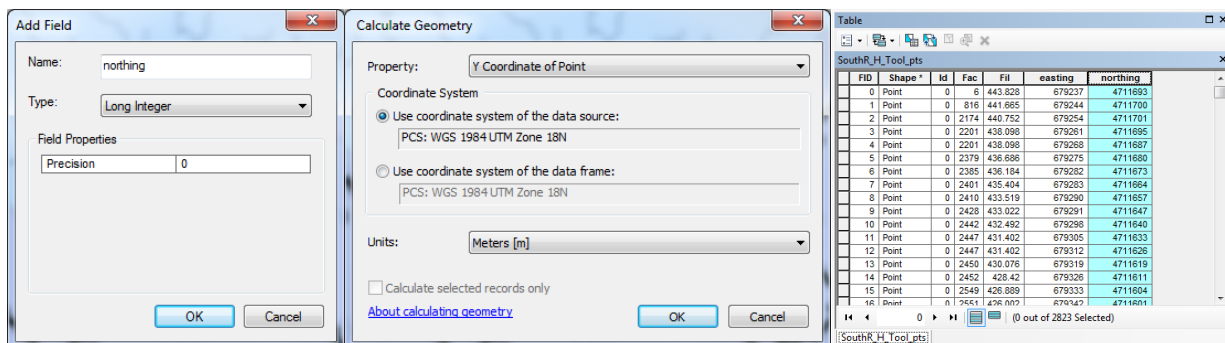
Step B20. Export attributes table. With the attribute table open, choose "**Table Options > Export…**" In the dialog box, make sure "All records" is selected. Click to the browse button next to select the file type and location. For the "Save as type:" choose dBase Table (.dbf) or text file (.txt).
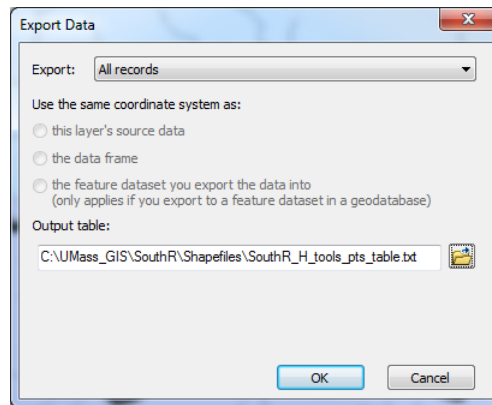
Hoorah! We are done with the GIS portions of this work, unless we want to display stream power on the map later. A simple method to show stream power on the map is to import the same point data with an additional columns for Ω and ω.

4.4 Computation and smoothing in Excel

Here we use Excel to manipulate and graph the data derived from ArcGIS. In this example, we take advantage of functions included in HydroTools, an Excel Add-in created by Carl Renshaw at Dartmouth College (Renshaw, 2016). As with the GIS workflow, we present a set of steps that have worked and been used to prepare data for peer-reviewed publications. Many other programs can be used to manipulate and graph the data, for example, MATLAB and Kaliedograph.

Step C1. Install HydroTools Add-In in Excel. Up-to-date instructions and downloads for installing and using this add-in are found at the HydroTools website, http://www.dimensionengine.com/excel/hydrotools/. Among the many useful functions in HydroTools, this analysis uses the function "ReachAverage." At present, HydroTools works for PC's with Windows operating systems, but not in current Macintosh (Apple) systems.

Step C2. Open GIS data in Excel. Open the GIS data from Step B20 in Excel. The .dbf file can be opened directly in Excel. If the GIS data were exported as a .txt file, then use the "Text to Columns" function under the "Data" tab. Save as an Excel file. In these columns, "FID" shows the downstream order of the points along the stream, "Id" is all zero's, "Fil" is the elevation in meters, and "Easting" and "Northing" show the coordinate location on a grid in units of meters. FAC is the flow accumulation, in the units of number of raster cells. To convert this to a meaningful number, it must be multiplied by the area of each raster cell (in this example 82.25 m^2 for a cell size of 9.069 m by 9.069 m). In this example, columns A through F are the data in the .dbf file. Column H and higher are manipulations of these data in Excel to compute stream power.

Step C3. Compute distance downstream and compute slope. In this example, the points were spaced every 10 m (Step B17), and the distance in meters can be computed as =FID*10. Next compute slope as rise over run. Note that in this step and subsequent steps, the Excel formula for the calculation is shown in the formula bar in the screenshot figures.

Hint

FID	Id	Fac	Fil	easting	northing		Dist (m)	Dist (km)	Slope
0	0	6	443.828	679237	4711693		0	0.0	(H3-H2)
1	0	816	441.665	679244	4711700		10	0.0	0.091
2	0	2174	440.752	679254	4711701		20	0.0	0.265
3	0	2201	438.098	679261	4711695		30	0.0	0.000
4	0	2201	438.098	679268	4711687		40	0.0	0.141
5	0	2379	436.686	679275	4711680		50	0.1	0.050
6	0	2385	436.184	679282	4711673		60	0.1	0.078
7	0	2401	435.404	679283	4711664		70	0.1	0.188
8	0	2410	433.519	679290	4711657		80	0.1	0.050
9	0	2428	433.022	679291	4711647		90	0.1	0.053
10	0	2442	432.492	679298	4711640		100	0.1	0.109
11	0	2447	431.402	679305	4711633		110	0.1	0.000
12								0.1	0.133
13								0.1	0.166
14		1.200						0.1	0.153
15		1.000						0.2	0.089
16								0.2	0.075
17		0.800						0.2	0.170
18		0.600						0.2	0.000
19		0.400						0.2	0.152
20								0.2	0.076
21		0.200						0.2	0.033
22		0.000						0.2	0.034
23		-0.200						0.2	0.090
24								0.2	0.000
25								0.3	0.018
26	0	3570	419.514	679429	4711571		260	0.3	0.012
27	0	3576	419.393	679437	4711564		270	0.3	0.094
28	0	4251	418.454	679444	4711557		280	0.3	0.043
29	0	4311	418.025	679451	4711550		290	0.3	0.000
30	0	4311	418.025	679458	4711543		300	0.3	0.000
31	0	4313	418.025	679465	4711536		310	0.3	0.044
32	0	4315	417.583	679472	4711529		320	0.3	0.274
33	0	4322	414.841	679479	4711522		330	0.3	0.207
34	0	4360	412.768	679486	4711515		340	0.3	0.000
35	0	4360	412.768	679493	4711508		350	0.4	0.086
36	0	5327	411.906	679500	4711501		360	0.4	0.085

Step C4. Smoothing slope. The aim of this step is to lessen the effect of inherent errors in the DEM, and produce a local slope that is reasonable for the river. The **ReachAverage** function in HydroTools provides a convenient way to smooth the slope. In this example, the "Reach-scale" and "Ave-Type" entries are held to single cells (by use of the $), which allows quick adjustment of the smoothing distance. In this example, the reach scale is 200 m, and the smooth type is "0," indicating an average both up and downstream of the location. This distance would likely be larger in a larger river. These values can be change in cells K2 and K4 in this example. For instance, one could enter any value to cell K2 to change the averaging distance to that value. Also, one could enter "1" in this example to take the average only over the preceding (upstream) cells, not the preceding and following cells (upstream and downstream). If one does not use the HydroTools add-in, this smoothing can also be accomplished without the ReachAverage function by using the Average function in Excel with a 200 m averaging window.

FID	Id	Fac	Fil	easting	northing		Dist (m)	Dist (km)	Slope	slope smoothing	Smoothed slope
0	0	6	443.828	679237	4711693		0	0.0	0.216	200	=ReachAverage($H$2:$H$2824,$J$2:$J$2824,H2,$K$2,$K$4)
1	0	816	441.665	679244	4711700		10	0.0	0.091	smooth type	0.1552938
2	0	2174	440.752	679254	4711701		20	0.0	0.265	0	0.1380922
3	0	2201	438.098	679261	4711695		30	0.0	0.000		0.126691
4	0	2201	438.098	679268	4711687		40	0.0	0.141		0.1212828
5	0	2379	436.686	679275	4711680		50	0.1	0.050		0.1184831
6	0	2385	436.184	679282	4711673		60	0.1	0.078		0.1154182
7	0	2401	435.404	679283	4711664		70	0.1	0.188		0.1149496
8	0	2410	433.519	679290	4711657		80	0.1	0.050		0.1098182
9	0	2428	433.022	679291	4711647		90	0.1	0.053		0.110017
10	0	2442	432.492	679298	4711640		100	0.1	0.109		0.1074762
11	0	2447	431.402	679305	4711633		110	0.1	0.000		0.1027033
12	0	2447	431.402	679312	4711626		120	0.1	0.133		0.0960144
13									0.166		0.0876571
14									0.153		0.0876571
15									0.089		0.0817714
16									0.075		0.0799571
17									0.170		0.0807142
18									0.000		0.073781
19									0.152		0.0714143
20									0.076		0.0688905
21									0.033		0.0658048
22									0.034		0.0788619
23									0.090		0.082419
24									0.000		0.0745334
25									0.018		0.0713476
26									0.012		0.0711857
27									0.094		0.0708047
28									0.043		0.0626857
29									0.000		0.0673096
30									0.000		0.0646095
31									0.044		0.0650286
32	0	4315	417.583	679472	4711529		320	0.3	0.274		0.0634763
33	0	4322	414.841	679479	4711522		330	0.3	0.207		0.0683047
34	0	4360	412.768	679486	4711515		340	0.3	0.000		0.0695142
35	0	4360	412.768	679493	4711508		350	0.4	0.086		0.0715905

Step C5. Compute discharge. There are several choices to compute discharge depending on the planned uses of the stream power results, as described in section 3.1. This example uses bankful discharge, derived from bankful discharge relationships for Massachusetts, where $Q_{bankful} = 37.1364\, A^{0.7996}$ (Bent and Waite, 2013). In this empirical relationship, units for Q are ft^3 s^{-1} and units for A are mi^2, but the results are converted to SI in the Excel example shown here.

Formula bar: `=(37.1364*(M2*$N$3)^0.7996)*$N$5`

FID	Id	Fac	Fil	easting	northing		Dist (m)	Dist (km)	Slope	slope smoothing	Smoothed slope	Drainage area (km2)	conversions	Q bankful m3/s (Bent 2013)
0	0	6	443.828	679237	4711693		0	0.0	0.216	200	0.159	0.00	km2 to mi2	=(37.1364*(M2*$N$3)^0.7996)*$N$5
1	0	816	441.665	679244	4711700		10	0.0	0.091	smooth type	0.155	0.07	0.386102	0.06
2	0	2174	440.752	679254	4711701		20	0.0	0.265	0	0.138	0.18	ft3/s to m3/s	0.12
3	0	2201	438.098	679261	4711695		30	0.0	0.000		0.127	0.18	0.0283168	0.13
4	0	2201	438.098	679268	4711687		40	0.0	0.141		0.121	0.18	ft to m	0.13
5	0	2379	436.686	679275	4711680		50	0.1	0.050		0.118	0.20	0.3048	0.13
6	0	2385	436.184	679282	4711673		60	0.1	0.078		0.115	0.20		0.13
7	0	2401	435.404	679283	4711664		70	0.1	0.188		0.115	0.20		0.13
8	0	2410	433.519	679290	4711657		80	0.1	0.050		0.110	0.20		0.13
9	0	2428	433.022	679291	4711647		90	0.1	0.053		0.110	0.20		0.14
10	0	2442	432.492	679298	4711640		100	0.1	0.109		0.107	0.20		0.14
11	0	2447	431.402	679305	4711633		110	0.1	0.000		0.103	0.20		0.14
12	0	2447	431.402	679312	4711626		120	0.1	0.133		0.096	0.20		0.14
32	0	4315	417.583	679472	4711529		320	0.3	0.274		0.065	0.35		0.21
33	0	4322	414.841	679479	4711522		330	0.3	0.207		0.068	0.36		0.21
34	0	4360	412.768	679486	4711515		340	0.3	0.000		0.070	0.36		0.22

Charts: "Slope" (Slope, Smoothed slope) and "Q2 discharge" (Q bankful m3/s (Bent 2013))

Step C6. Compute or enter width. Similar to discharge, there are several considerations in the width used for stream power calculations. Here, width is computed using an empirical relationship from Bent and Waite (2013) derived for Massachusetts, where $W_{bankful} = 15.0418\ A^{0.4038}$. In this empirical relationship, units for w are ft and units for A are mi^2, but the results are converted to SI in the Excel example shown here.

Formula bar: `=(37.1364*(M2*$N$3)^0.7996)*$N$5`

FID	Id	Fac	Fil	easting	northing		Dist (m)	Dist (km)	Slope	slope smoothing	Smoothed slope	Drainage area (km2)	conversions	Q bankful m3/s (Bent 2013)	W bankful (m) Bent 2013
0	0	6	443.828	679237	4711693		0	0.0	0.216	200	0.1594638	0.00	km2 to mi2	=(37.1364*(M2*$N$3)^0.7996)*$N$5	
1	0	816	441.665	679244	4711700		10	0.0	0.091	smooth type	0.1552938	0.07	0.386102	0.06	1.05
2	0	2174	440.752	679254	4711701		20	0.0	0.265	0	0.1380922	0.18	ft3/s to m3/s	0.12	1.56
3	0	2201	438.098	679261	4711695		30	0.0	0.000		0.126691	0.18	0.02831685	0.13	1.57
4	0	2201	438.098	679268	4711687		40	0.0	0.141		0.1212828	0.18	ft to m	0.13	1.57
5	0	2379	436.686	679275	4711680		50	0.1	0.050		0.1184831	0.20	0.3048	0.13	1.62
6	0	2385	436.184	679282	4711673		60	0.1	0.078		0.1154182	0.20		0.13	1.62
7	0	2401	435.404	679283	4711664		70	0.1	0.188		0.1149496	0.20		0.13	1.62
8	0	2410	433.519	679290	4711657		80	0.1	0.050		0.1098182	0.20		0.13	1.62
9	0	2428	433.022	679291	4711647		90	0.1	0.053		0.110017	0.20		0.14	1.63
10	0	2442	432.492	679298	4711640		100	0.1	0.109		0.1074762	0.20		0.14	1.63
11	0	2447	431.402	679305	4711633		110	0.1	0.000		0.1027033	0.20		0.14	1.63
12	0	2447	431.402	679312	4711626		120	0.1	0.133		0.0960144	0.20		0.14	1.63
32	0	4315	417.583	679472	4711529		320	0.3	0.274		0.0646095	0.35		0.21	2.05
33	0	4322	414.841	679479	4711522		330	0.3	0.207		0.0650286	0.35		0.21	2.05
34	0	4360	412.768	679486	4711515		340	0.3	0.000		0.0634763	0.35		0.21	2.05
											0.0683047	0.36		0.21	2.06
											0.0695142	0.36		0.22	2.06

Charts: "Slope" (Slope, Smoothed slope) and "Q2 discharge and width" (Q bankful m3/s (Bent 2013), W bankful (m) Bent 2013)

Step C7. Compute Ω. The product of ρ (998 kg m³ at 20 °C), g (9.8 m s⁻²), Q, and S yields Ω. It is shown in column Q in this example.
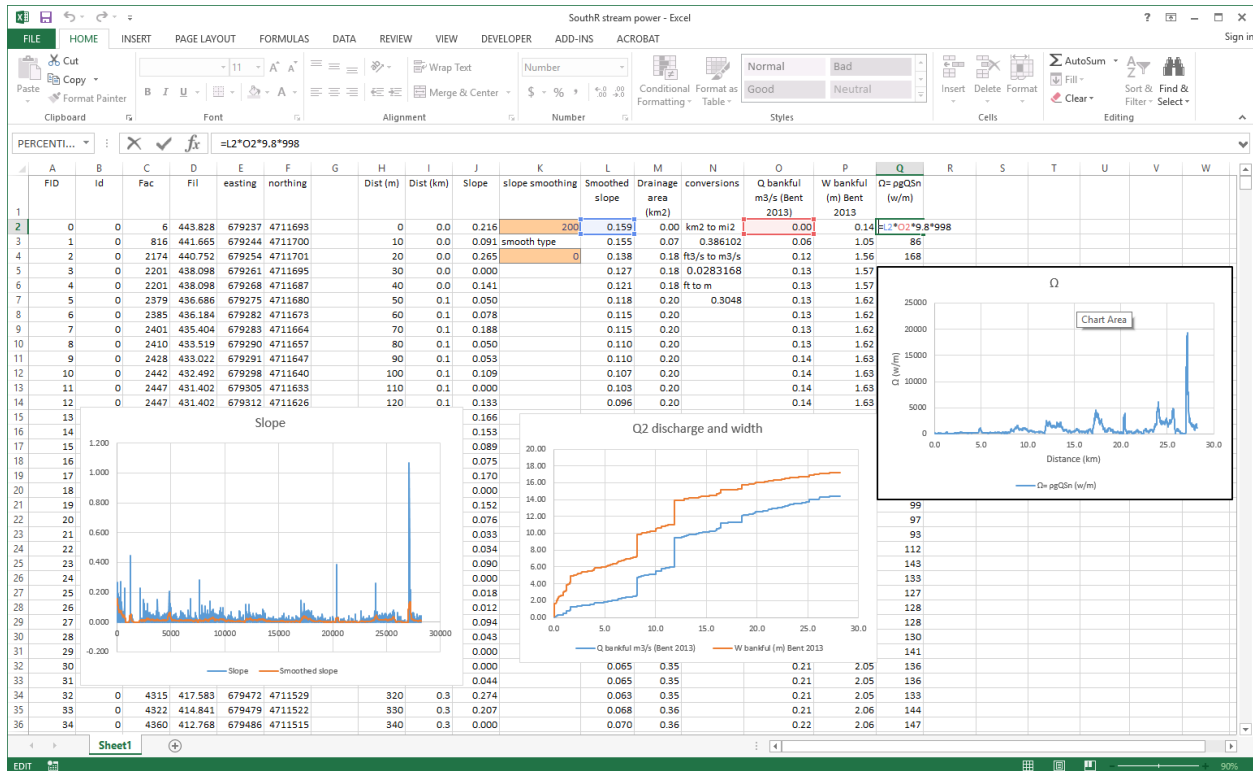
Step C8. Smooth Ω. Stream power, Ω, is smoothed using the **ReachAverage** function in HydroTools over a 1000 m reach, centered at each point along the river. This averaging could also be computed by Excel's Average function using a 1000 m moving average. The smoothing shows the broad trends in stream power. It also lessens the effect of minor inaccuracies in the DEM. This smoothing of stream power in this step is separated from the smoothing of slope (Step C4) to distinguish between the smoothing required to produce a reasonable local slope from a DEM and the smoothing required to depict reach-scale or larger scale trends in stream power. This smoothing distance is a key concern in the stream power computation, and there is no set approach to determine what this smoothing distance should be, as discussed in Section 3.2.1. Users are encouraged to examine the effects of different smoothing distances on their results. Longer smoothing distances often reduce the frequency that stream power increases and decreases with respect to distance downstream, and shorter smoothing distances create a more jagged portrayal of stream power with respect to distance downstream. Playing with the stream power smoothing distance can influence the computed value of stream power a single locations along the river, maybe increasing or maybe decreasing the stream power value if a neighboring locations are including in the average.

FID	Id	Fac	Fil	easting	northing	Dist (m)	Dist (km)	Slope	slope smoothing	Smoothed slope	Drainage area (km2)	conversions	Q bankful m3/s (Bent 2013)	W bankful (m) Bent 2013	Ω= pgQSn (w/m)	Ω smoothing scale	Smoothed Ω (w/m)
0	0	6	443.828	679237	4711693	0	0.0	0.216	200	0.159	0.00	km2 to mi2	0.00	0.14	2	1000	=ReachAverage($H$2:$H$2824,$Q$2:$Q$2824,H2,$R$2,
1	0	816	441.665	679244	4711700	10	0.0	0.091	smooth type	0.155	0.07	0.386102	0.06	1.05	86	smooth typ $R$4)	
2	0	2174	440.752	679254	4711701	20	0.0	0.265	0	0.138	0.18	ft3/s to m3/s	0.12	1.56	168	0	117.8
3	0	2201	438.098	679261	4711695	30	0.0	0.000		0.127	0.18	0.0283168	0.13	1.57			
4	0	2201	438.098	679268	4711687	40	0.0	0.141		0.121	0.18	ft to m	0.13	1.57			
5	0	2379	436.686	679275	4711680	50	0.1	0.050		0.118	0.20	0.3048	0.13	1.62			
6	0	2385	436.184	679282	4711673	60	0.1	0.078		0.115	0.20		0.13	1.62			
7	0	2401	435.404	679283	4711664	70	0.1	0.188		0.115	0.20		0.13	1.62			
8	0	2410	433.519	679290	4711657	80	0.1	0.050		0.110	0.20		0.13	1.62			
9	0	2428	433.022	679291	4711647	90	0.1	0.053		0.110	0.20		0.14	1.63			
10	0	2442	432.492	679298	4711640	100	0.1	0.109		0.107	0.20		0.14	1.63			
11	0	2447	431.402	679305	4711633	110	0.1	0.000		0.103	0.20		0.14	1.63			
12	0	2447	431.402	679312	4711626	120	0.1	0.133		0.096	0.20		0.14	1.63			
13								0.166									
14								0.153									
15		1.200						0.089									25000
16								0.075									
17		1.000						0.170									20000
18								0.000									
19		0.800						0.152									15000
20								0.076									
21		0.600						0.033									10000
22								0.034									
23		0.400						0.090									5000
24								0.000									
25		0.200						0.018									
26								0.012									
27		0.000						0.094						99			131.5
28								0.043						97			131.8
29		-0.200						0.000						93			132.2
30								0.000			0.065	0.35		112	0.21	2.05	132.4
31								0.044			0.065	0.35		143	0.21	2.05	132.7
32	0	4315	417.583	679472	4711529	320	0.3	0.274			0.063	0.35		133	0.21	2.05	132.8
33	0	4322	414.841	679479	4711522	330	0.3	0.207			0.068	0.36		127	0.21	2.06	132.9
34	0	4360	412.768	679486	4711515	340	0.3	0.000			0.070	0.36		128	0.22	2.06	132.9
														128			132.8
														130			132.7
														141			132.4
														136			131.9
														136			131.3
														133			130.5
														144			129.6
														147			128.5

Step C9. Compute ω. To compute ω, divide Ω (column Q) by W (column P), and to compute the smoothed ω, divide the smoothed Ω (column S) by W (column P).

Note that at 27.2 km, ω exceeds 300 w m^{-2} at the bankful flow km, for both the smoothed and unsmoothed values, suggesting that boulders would move (Petit et al., 2005) and flood effects could be catastrophic at the bankful flow (Magilligan, 1992). This is an unlikely condition for a flood that occurs every 1 to 3 years. However, aerial photographs reveal a dam here. The steep slope and high stream power are due to a drop at the face of the dam expressed in the DEM. In addition, the channel width is about 35.5 m based on the aerial photographs, while the empirical relationship predicts $W = 17.1$ m. This elevates the GIS-derived ω compared to the actual ω.

Overall, this location emphasizes that stream power results derived from remotely sensed data should be viewed with caution and ground-truthed when possible.
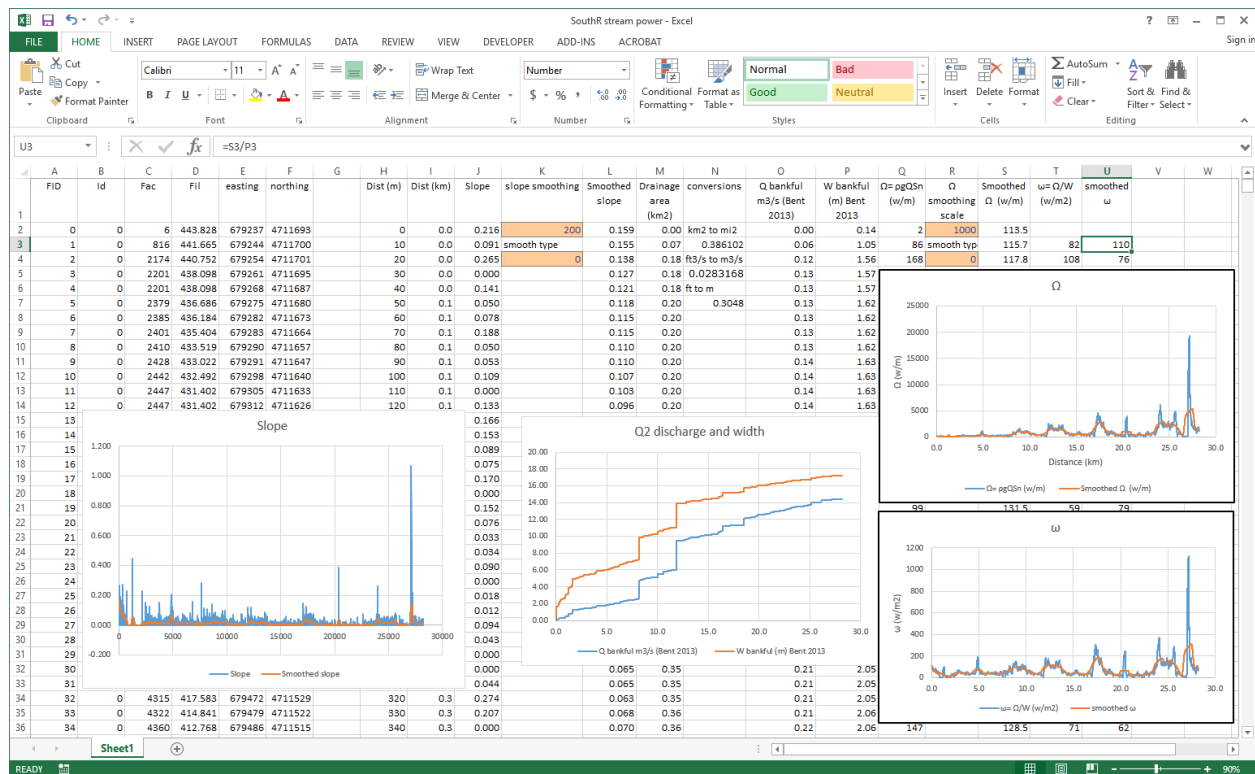
4. Conclusions

This document presents an overview of the origins and terminology of stream power, followed by a discussion of how stream power is used in sediment transport computations and geomorphology research. Next it discusses several consideration in computing stream power, especially the choices in deriving the input parameters of discharge, slope, and width as well as special consideration for smoothing distances and working with DEMs. Finally a workflow for deriving stream power along stream profiles using ArcGIS and Excel is presented. This is just one possible workflow relying heavily on ArcGIS. Other popular options use HEC-RAS for stream power computations at the cross section to reach scale, and TopoToolbox as MATLAB-based alternative to ArcGIS for analyzing DEMs and deriving stream power at reach to watershed scales. Overall, stream power is being used increasingly for basic and applied questions in geomorphology and water resource management, and the tools and application using stream power will continue to develop in coming years.

5. List of variables and acronyms

α	slope, measured from horizontal
Ω	cross sectional stream power
ρ	density of water, characteristically 998 kg/m^3 at 20 °C)
ρ_S	density of sediment, characteristically 2.65 kg/m^3
τ	shear stress
τ^*	shields parameter
ω	mean stream power, a.k.a. specific stream power, unit stream power

ω_c	critical stream power
a	fitting parameter for empirical equations
A	drainage area
b	fitting parameter for empirical equations
c	particle settling velocity
D	grain size of bed material
g	gravity (9.8 m/s^2)
H	depth of water
L	length of stream
m	mass
Q	discharge
Q_s	mass rate of sediment transported by a river
S	slope
t	time
V	water velocity
W	width of channel or flow
Z	elevation
DEM	Digital election model
NHD	National Hydrography Dataset
PE	potential energy
KE	kinetic energy

6. Acknowledgements

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