Chapter 2
HYDROLOGY AND THE SIZE OF A RIVER

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1. River Size and Continuity

To most of us the size of a river is its width. It is what our brain registers in the field: the stretch of water that we see as we look from bank to bank. But the channel width is not the only measure of river size. The depth of flow is also a measure of river size; small rivers can be waded but large rivers we have to swim or cross by boat or bridge. Another measure of river size is the cross-sectional area of the channel that the river has carved out to accommodate its flow. Yet another measure of river size is the flow itself: the volume of water that fully occupies the channel.

These measures of river size are linked by a fundamental definitional connection known as continuity. For a river channel the continuity relation states that

\[ Q = wdv \]  \hspace{1cm} (2.1)

Where \( Q \) = discharge (measured in cubic metres/second, m\(^3\)s\(^{-1}\)),

\( w \) = channel width (measured in metres, m),

\( d \) = the average depth of the channel (measured in metres, m), and

\( v \) = the mean velocity of flow through the channel cross-section (in metres/second, ms\(^{-1}\)).

The cross sectional area of a river channel, \( A \) (measured in square metres, m\(^2\)) is the product of the channel width and mean depth or

\[ A = wd \]  \hspace{1cm} (2.2)

So it follows that the continuity relation can also be expressed as

\[ Q = Av \]  \hspace{1cm} (2.3)
A useful way to think about these relations is that, given the discharge of a river, its size (width and depth or cross-sectional area) will be determined by the velocity that can be achieved by the river. The higher the velocity the smaller the width and depth needs to be in order to carry the flow. Conversely, for a given discharge, the lower the velocity the larger the cross-sectional area needs to be in order to conduct the flow.

**The Equilibrium channel**

So the fundamental control on channel size is the discharge of the river and the velocity of flow. We can rearrange equation (2.1) to state this formally:

\[ \text{wd} = A = \frac{Q}{v} \]  

(2.4)

The relation expressed by equation (2.4) is easily interpreted if we think about a laboratory flume. We pump so much water, \(Q\), through the flume and the channel carved by the flow into the sediment on the flume floor is such (wd) that it accommodates all of the flow. We can think of this as an equilibrium channel regulated by the forces acting on the boundary (see Figure 2.1). Once an equilibrium channel is achieved by the force of the flowing water, it tends to be maintained by adjustments in the cross-sectional area and in the flow velocity. If the sediment being transported from upstream in the flume comes to rest on the bed (deposition) the cross-sectional area will be diminished as the bed level rises. Equation (2.1) tells us that velocity must also increase in order to conduct the given discharge. This increase in velocity results in a larger stress being applied to the bed causing sediment once again to move out of the section (erosion).
and thus the bed level falls. In the event that erosion of the bed overshoots the equilibrium level
the increased cross-sectional area will result in a decline in flow velocity and a return to
deposition. Thus the bed level may fluctuate up and down but is maintained at an average
equilibrium level in the long term. Natural rivers tend to behave in the same self-regulating
manner as our flume channel although there are some very important differences. One of them is
the degree to which we can think of river discharge as ever being constant.

2. Controls on the Discharge of Natural Rivers

Unlike the case of flow in our laboratory flume, the discharge of natural rivers is not constant
over time. Indeed it is extremely variable. To better understand the reasons for this hydrologic
irregularity it will be useful at this point to briefly review the nature of the basin hydrologic cycle
and the water budget.

The basin hydrologic cycle refers to the inputs and outputs of water within a drainage basin that
supply its rivers (Figure 2.2). A river flows within its drainage basin, defined as the land surface
that collects precipitation that potentially will be delivered to the river. A drainage basin is
defined by the drainage divide (or watershed) that marks the location on the ground of the limit
of the source area for water supply to the channel. This drainage divide is also known as the
phreatic divide. Generally the drainage divide is taken as the topographic divide (Figure 2.3: the
locus of the highest points along the ridges defining the basin). We should note, however, that
the phreatic divide and the topographic divide may not always coincide. For example, the
phreatic divide may extend into an adjacent basin if geological structures allow water to travel
from within one basin to another as groundwater. Similarly, a basin may lose water to another
for the same reason.

The simple water budget for a given drainage basin in a given period of time can be expressed by
the equation:

\[ \text{INPUT} = \text{OUTPUT} + \Delta \text{STORAGE} + \text{LOSS} \]  \hspace{1cm} (2.5)

where

INPUT = Precipitation (rain, snow, water vapour, etc) over the drainage basin;
OUTPUT = Streamflow;
\Delta \text{STORAGE} = \text{Changes in the volumes of water stored in lakes, trees, groundwater, soil
moisture, etc};
LOSS = Evapotranspiration and deep groundwater that is completely lost to the basin.
Figure 2.2: The hydrologic cycle describes the circulation of water from precipitation through various transfers and storage to its return to the atmosphere as evaporation and transpiration.

Figure 2.3: The drainage basin and topographic drainage divide.
Our primary interest at the moment is in the output or streamflow component (amount and timing of flow). It is controlled by two sets of factors:

(a) Basin characteristics and
(b) Precipitation characteristics

**Basin characteristics** include geomorphology, geology and vegetation. **Geomorphology** of the basin determines the basin size (drainage area), steepness of slopes, the basin shape (elongated or pear-shaped) and the drainage-network configuration (related to shape). Other things being equal, the size of the drainage basin determines the amount of water that is supplied to its rivers. Small basins generate small flows and large basins generate large flows. If the basin slopes are steep, water from precipitation will tend to be delivered to the rivers more quickly than would be the case for more gentle basin slopes. An elongated basin tends to deliver water to the basin outlet over a prolonged period of time whereas a more compact pear-shaped basin tends to deliver precipitation from all parts of the basin to the outlet in about the same time so that water from a single storm arrives in an abrupt pulse of flow. River channels are much more efficient at conveying water than surface slopes. As a result, if the drainage density is high (many channels per unit drainage area), water tends to be delivered to the basin outlet more quickly than if there are fewer channels to deliver it.

**Geology** exerts a control over streamflow primarily by controlling basin permeability. Lithological control on permeability is affected by porosity and jointing (for example, granite versus limestone).

**Vegetation** is important, not only because it determines evapotranspiration losses from the basin, but also because it can be an important control on slowing runoff (most evident in forested versus non-forested slopes) to the river channels. Vegetation amount and type within the channels may also exercise an important control on the efficiency of streamflow.

**Precipitation characteristics** include precipitation form (snow or rain), distribution in time (climatic regime, precipitation intensity and duration), and distribution in space (storm track in relation to basin; extent (spatially general or focused).
Perhaps the most fundamental distinction among precipitation forms is the phase: solid or liquid. In many parts of the world snow accumulates during the winter and is not released to the rivers until the following spring. Thus snowfall tends to be temporally decoupled from the runoff it produces. In contrast, rain produces runoff relatively rapidly soon after it falls. The amount of runoff to rivers increases as rainfall intensity and duration increase. The spatial distribution of precipitation can also greatly influence the amount of runoff and streamflow. A large storm over an entire drainage basin will have a different effect than a small focused storm that affects just part of the basin. A storm moving down the basin will tend to concentrate the flow at the basin outlet (causing an abrupt flood) whereas a storm moving up the basin will tend to spread the resulting runoff and streamflow out over time (causing a prolonged but lower-magnitude flood at the outlet).

3. Natural River Flow Variability

Having briefly considered the range and variety of potential controls on natural streamflow it hardly is surprising that the overwhelming characteristic of natural streamflow is its variability. Indeed, flows in natural channels vary over a wide range of time scales:

- Hourly fluctuations can occur in small channels as a response to rainstorms;
- Diurnal fluctuations also occur in glacier-fed channels, like Squamish River, as temperature and melt-rate vary over the day;
- Monthly variations relate to seasonal controls such as the nature of the snowmelt freshet.
- Seasonal aspects of climate can cause differences in storminess.

Many BC rivers have high flows that form two distinct populations: the snowmelt-induced seasonal *freshet* that occurs in the spring each year and the rather more erratic rainfall-induced storm floods that commonly occur in the fall. For example, a typical annual hydrograph for Squamish River north of Vancouver is shown in Figure 2.4. Note that the upper graph (of maximum daily discharge) is highly variable with a number of extreme events occurring in the period August through December. The variability of flow captured in the graphs of Figure 2.4 is typical of most rivers. But if natural river flow is so variable does it make sense to think in terms of *the* size of a river and of the application of continuity? Well, as in so much of fluvial
geomorphology, the answer is yes and no! No in the strictest sense because a river continuously varies in size along with its fluctuating flow. But yes in that typical or average or mean states do exist in a statistical sense for rivers just as they do for any population of natural occurrences or objects. We just need to define the averages that make the most sense to us in understanding why rivers look and behave the way they do.

Figure 2.4: Water Survey of Canada flow data for Squamish River (near Brackendale). The upper graph is the maximum daily discharge ever recorded at this station, the middle graph is the mean daily discharge for 2002, and the lower graph is the minimum daily discharge ever recorded at this station. From Water Survey of Canada.

Another way of capturing the day-to-day variability of natural river flow is to summarize the mean daily flow in a frequency histogram. Figure 2.5 is a simple frequency histogram and a cumulative frequency plot of the daily discharge for the Squamish River gauge at Brackendale (constructed in EXCEL using the Data Analysis software). You can see that the most frequent discharges are the smallest and that the very largest discharges are quite few in number. This asymmetry in the distribution of daily discharges is reflected in the cumulative frequency plot: discharges range from about 30 m³/sec to more than 700 m³/sec but more than half of the flows
are less than 160 m$^3$/sec and flows above 630 m$^3$/sec represent just 2.5% (13) of the 365 mean daily discharges.

**Figure 2.5:** An EXCEL frequency histogram and cumulative-frequency plot of mean daily discharge at the Brackendale gauging station on Squamish River.

If we mathematically invert the cumulative-frequency plot in Figure 2.5 so that it shows the percentage time a given discharge is equaled or exceeded, the result is the flow-duration curve (Figure 2.6). This is the conventional way hydrologists present data of this type. The example shown in Figure 2.6 is synthetic so that the underlying calculations are clear. In this case we can say that, for example, a discharge in the lower range, say 100 m$^3$/sec, will be equaled or exceeded more than 50% of the time but a large flow, say 400 m$^3$/sec, will be equaled or exceeded less than 1% of the time.

It is clear from the flow-duration curve that, if the higher discharges are the ones of interest, there is not very much information revealed by this type of analysis. For this reason river scientists have focused on other analytical methods that highlight the statistical characteristics of the flows that on the flow-duration curve appear to be unimportant.
Flow-duration curve
An example

<table>
<thead>
<tr>
<th>Discharge Class (m³/s)</th>
<th># Days</th>
<th>% Time</th>
<th>Cumulative % Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.99</td>
<td>164</td>
<td>100(164/365) = 44.9</td>
<td>55.0 + 44.9 = 100.0</td>
</tr>
<tr>
<td>100-199</td>
<td>152</td>
<td>41.6</td>
<td>13.4 + 41.6 = 55.0</td>
</tr>
<tr>
<td>200-299</td>
<td>35</td>
<td>9.5</td>
<td>3.9 + 9.5 = 13.4</td>
</tr>
<tr>
<td>300-399</td>
<td>13</td>
<td>3.6</td>
<td>0.3 + 3.6 = 3.9</td>
</tr>
<tr>
<td>400-499</td>
<td>1</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>Totals:</td>
<td>365</td>
<td>100</td>
<td></td>
</tr>
</tbody>
</table>

Figure 2.5: A synthetic flow-duration curve and the calculations on which it is based. It is the inverted form of the type of cumulative frequency plot shown in Figure 2.5.

Formative Discharge in Rivers: some Concepts

So, given the natural variability in real river flows, what flows are responsible for forming the channel? If continuity is a useful principle in thinking about rivers what is the appropriate Q in Q = wdv?

River geomorphologists have actually been struggling with these questions for many years and the conclusions they have reached, even today, generate vigorous discussion. There is, nevertheless, a consensus or conventional position on these issues but we need to remain aware that there are also many for whom the idea of a single dominant or formative discharge for rivers is too much of a simplification of reality to be very useful.

The conventional view on the nature of formative discharge in rivers is that, although flow varies widely, a much more limited range of flows actually does most of the work in shaping a river
channel. The notion here is that, for much of the year the lowest flows, although on a daily basis occurring more frequently than larger flows (for example, note Figures 2.5 and 2.6), are simply too low to effectively erode and shape the channel. In other words, at these low discharges the river simply flows through a channel that has been shaped by higher flows. But the highest discharges, so it is argued, also are not very effective at shaping the channel. Although these high flows are the most capable of eroding the channel they occur so infrequently that their morphological impact is small. At the very highest discharges the flow will exceed the capacity of the channel and be conducted outside the channel and across the adjacent floodplain, thus becoming completely ineffective at shaping the channel itself. It is reasoned, therefore, that there is some much smaller middling range of flows that do most of the work shaping the river channel and that some summary value of these intermediate flows represents the formative discharge of the river. Another conceptual step in this line of reasoning completes the hydrologic picture: formative discharge must be the flow that completely wets the boundary of the equilibrium channel. It is the flow that completely fills the channel; it has been termed the bankfull discharge. So the proponents of these morphodynamic generalizations for rivers argue that a fundamental process in rivers is given by the conceptual equation:

\[ \text{Formative or dominant discharge} = \text{bankfull discharge} \]

The notion that rivers essentially are shaped by a dominant or formative discharge is not without its critics and we will revisit this concept briefly after we have considered some of the analytical tools that have been developed to objectively define bankfull discharge.

**Formative Discharge: Some Analytical Tools**

The flow duration curve, such as that shown in Figure 2.6, is one way of graphically simplifying the variable array of daily discharge that most rivers experience during the course of the year. It is a useful summary and graphically draws our attention to the frequency dominance of the low flows and the relative statistical unimportance of the high flows. But this characteristic is also its primary limitation. It does not highlight the very range of flows that most geomorphologists see as the most important in shaping the channel.
So how do we identify that “representative” formative discharge? It should be easy. Right? If formative discharge is bankfull discharge then all we need to do is measure the flow that completely fills the channel and we have the answer! Yes?

Well, yes and no! The answer is “yes” if the channel of interest is an equilibrium channel in which the top of the banks represents the top of the equilibrated channel. But we have already noted that the equilibrium channel is a concept that allows for fluctuations causing the channel to be temporarily oversize or undersize. If we happen upon a river that has just recently experienced a major flood it may have an eroded a channel that is far larger than the normal equilibrium channel and is presently trying (through deposition) to return to what is normal. We say that such a channel is in a transient state; it is a channel type for which defining the morphological bankfull state is exceedingly difficult. Similarly, a river that has experienced a prolonged period of below-average flows may have contracted to something smaller than its normal morphological bankfull state and so it is also in a transient state. In such an undersized channel it will also be exceedingly difficult to measure the normal bankfull channel state.

But there is a further problem that so complicates the field measurement of the normal bankfull state on rivers that river scientists have largely decided to pursue an independent statistical measure of bankfull discharge. The problem relates to incision.

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**Figure 2.7:** The morphological definition of bankfull stage in the field is complicated by the usually unknown degree of channel incision present.
In an unincised channel (A in Figure 2.7) the bed and the banktops represent part of an equilibrium boundary such that the difference in the elevation of these two features is the equilibrium bankfull depth of the channel at bankfull discharge. On the other hand, if a channel occupies a deep canyon (B in Figure 2.7) it is rather obvious to everyone that the channel is incised and that the top of the canyon walls is not the banktop for the river far below. We recognize this as a case where the morphological definition of bankfull discharge simply is not possible. A more difficult case, however, is where the channel is only modestly incised (C in Figure 2.7). How do we know it is incised? What do we look for in the field? If we use the banktop as the level of bankfull discharge we will overestimate the size of the channel relative to an otherwise identical stream that is not incised.

These cases of incised rivers illustrate clearly the virtue of having a hydrologically defined bankfull discharge in order to avoid these complications of transient states and incision in rivers.

So where do we start? As we noted earlier, it is the bankfull flow that seems to be the most significant discharge in terms of shaping the channel. Smaller flows do not have the erosive capacity to do as much geomorphic work and larger flows spill out on to the floodplain where they are no longer relevant to channel formation. Geomorphologists have therefore reasoned that it is not the average flow of a river that shapes the channel but rather it is those relatively high discharges near bankfull stage. These high flows or floods, occur infrequently, perhaps just once every year or so.

The statistical analysis of floods is seen by many river hydrologists to be the key to independently defining the formative bankfull flow. But before we begin this discussion of the statistical properties of floods we need to be aware that the term flood means something different to a river hydrologist. The popular notion of a flood with which we are familiar is a high flow that overtops the river banks and inundates the surrounding land. To a river hydrologist, however, a flood is a high-flow event that may not necessarily overtop the banks. A flood is often characterized by the flood hydrograph as shown in Figure 2.8. The flood hydrograph is a graph of river discharge or stage versus time following some initiating event, usually rainfall from a storm. In the ideal case the rainfall event is concentrated at a point in time and the river
Figure 2.8: The flood hydrograph

Figure 2.9: Flood hydrograph characteristics in relation to basin and rainfall properties.
subsequently responds by rising. The direct runoff from the storm is called quickflow and is in addition to the base flow of the river that it was experiencing before the discharge increased. This part of the flood hydrograph that registers the arrival of the quickflow is called the rising limb. When the river flow peaks at the hydrograph crest the discharge of the river declines more slowly than it rose and this part of the flood hydrograph is termed the falling limb. The time interval from the storm centre to the time of the flood hydrograph crest is known as the basin lag. So typically the flood hydrograph is an asymmetric graph describing the river discharge or stage during the passage of a flood. The shape of the flood hydrograph (for example, whether it is steep and peaked rather than subdued and sustained) depends mainly on the basin and rainfall properties (Figure 2.9).

The statistical analysis of floods is concerned with the magnitude of the flood (the discharge at the flood hydrograph crest) and with the frequency it occurs. There are a variety of analytical techniques available but we will concern ourselves mainly with the way in which the peak annual flood is distributed over time. The peak annual flood is the maximum instantaneous peak flow that a river discharges in a given year. The record of annual peak flows on a river is referred to as the annual series of floods. It turns out that floods of a given magnitude are distributed in time in a remarkably orderly manner and have a distribution that conforms to the theory of extreme events (or extreme-value theory). We can make use of this theoretical magnitude/frequency distribution to assign a recurrence interval to the floods that occur in our rivers. The recurrence interval is simply the average time interval between floods of a given magnitude. Extreme-value theory was pioneered by the German mathematician Emil Julius Gumbel in the 1950s and for this reason the method of analysis outlined here is often referred to as Gumbel analysis.

The extreme-value theory is mathematically complex but mercifully the distribution of event recurrence interval versus event magnitude can be linearized on a special type of graph paper called Gumbel paper and we make use of this in assigning the recurrence interval to floods. The easiest way to develop an understanding of Gumbel analysis is to consider an example of the calculation and the graphed result; this is undertaken in Figure 2.10 and 2.11.
Figure 2.10 is a table of synthetic data of the annual series of peak floods on a river for the period 1970-1978.

<table>
<thead>
<tr>
<th>Year</th>
<th>Qmax (m³/s)</th>
<th>Rank, R</th>
<th>Recurrence Interval, RI (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1970</td>
<td>100</td>
<td>9</td>
<td>((9+1)/9 = 1.1)</td>
</tr>
<tr>
<td>1971</td>
<td>170</td>
<td>3</td>
<td>3.3</td>
</tr>
<tr>
<td>1972</td>
<td>140</td>
<td>7</td>
<td>1.4</td>
</tr>
<tr>
<td>1973</td>
<td>205</td>
<td>1</td>
<td>((9+1)/1 = 10.0)</td>
</tr>
<tr>
<td>1974</td>
<td>120</td>
<td>8</td>
<td>1.3</td>
</tr>
<tr>
<td>1975</td>
<td>150</td>
<td>5</td>
<td>2.0</td>
</tr>
<tr>
<td>1976</td>
<td>150</td>
<td>5</td>
<td>2.0</td>
</tr>
<tr>
<td>1977</td>
<td>160</td>
<td>4</td>
<td>2.5</td>
</tr>
<tr>
<td>1978</td>
<td>180</td>
<td>2</td>
<td>5.0</td>
</tr>
<tr>
<td>Average</td>
<td>153</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Figure 2.10: A synthetic data set of the annual series of maximum floods on a river for the period 1970 to 1978.*

There are three simple steps in determining the recurrence intervals for the annual series of floods listed in Figure 2.10:

1. Sort the annual peak flows by magnitude from the largest to the smallest;
2. Assign a rank to each peak discharge from the largest (1) to the smallest;
3. Calculate the recurrence interval (RI) of each maximum discharge from the formula:

\[
RI = \frac{(N+1)}{R} \quad \text{(where } N = \text{ the number of years in the series and } R = \text{ rank)}.\]

You will notice in Figure 2.10 that two equal discharges (150 m³/s) have the same rank (5) and that there is consequently no rank 6 because the highest rank (9) must equal the number of data points (9).

If the distribution of the data for flood magnitude and recurrence interval conforms to the extreme-value theory the values will plot as a straight line on Gumbel paper, in which case a
straight-line trend can be used to generalize the data and to estimate the magnitude of unmeasured peak flows of given recurrence interval.

Figure 2.11: A magnitude-frequency analysis of floods on the annual series (shown in Figure 2.10) plotted on Gumbel paper. See the text for an explanation.

Figure 2.11 is a Gumbel graph of the data in Figure 2.10. You will notice that the RI scale is close to being logarithmic (but is not exactly so). The data are generalized by a “best-fit” line, in this case fitted by eye. If you look at the data in Figure 2.10 you will see that the average (or mean) of the floods in this synthetic annual series is 153 m$^3$/sec (add them up and divide by 9).
On Gumbel paper this mean annual flood is equal to $Q_{2.33}$ where the subscript is the recurrence interval in years. In other words, the mean annual flood is the discharge that occurs once every 2.33 years (on average, in a long record).

So how does Gumbel analysis help to solve the problem of independently defining the bankfull discharge on rivers? In the second half of last century river scientists, mainly with the US Geological Survey, were able to show that, for a large sample of unincised rivers in the American Midwest, the discharge that filled the channels to the top of the bank (just before overflowing) averaged 1.58 years on the annual series. This average recurrence interval for bankfull discharge has been widely accepted by most fluvial geomorphologists. Part of its appeal as a definition is that it has a theoretical basis in statistics as the most-probable annual flood. That is, $Q_{1.58}$ on the annual series is the peak flood in a given year that is the most likely to occur. Smaller flows are less likely as are larger flows. It is also the flow on a river that early surveyors often mapped (by noting the bank limits of riparian vegetation) as normal high water.

So we can summarize these notions in the following equality:

$$\text{Bankfull discharge} = \text{the most-probable annual flood (} Q_{1.58} \text{)} = \text{normal high water}$$

This equality is extremely useful because it allows us to compare the size of rivers based on a hydrological definition of bankfull discharge, thus obviating the problems of channel incision and what that means for a morphological definition of bankfull stage. It should be recognized, however, that many rivers appear to have a bankfull discharge with a less frequent occurrence than 1.58 years although incision may be a factor here; very little study of bankfull return periods have been conducted in Canada.

Figure 2.11 also shows an estimate of the 100-year flood ($Q_{100}$). This application, of predicting the size and frequency of unrecorded floods, is of great interest to engineers. They are often called upon to estimate how large a flood can be expected on a river, say every 100 or 200 years. River dykes and bridges and buried pipelines that cross rivers, for example, are designed on the basis of these kinds of calculations.
In our example $Q_{100} = 278 \text{ m}^3/\text{sec}$. This is shorthand for the statement “that a discharge of 278 m$^3$/sec will be equaled or exceeded, on average, once every 100 years in a long record of annual floods”. It is important to appreciate that this is a statement about the *statistical behaviour* of extreme events and not a prediction of the absolute timing of a future extreme event. The analysis says nothing about the probability of a given flood occurring in a particular year. For example, if a 100-year flood occurs this year it does not mean that there will not be another flood of this magnitude for another 100 years. Similarly, because a 10-year flood has not occurred in the last 10 years it does not mean that one is now imminent. The recurrence interval refers only to the long-term statistical properties of floods and not to their specific time distribution. The 100-year flood may be equaled or exceeded several years in a row but its long-term average recurrence interval over the long haul will be 100 years. So just because a river at your back door had a 100-year flood last year there is no reason to relax this year!

Statistical analyses other than the Gumbel method are also in use although we won’t consider them in detail. Two other common methods are:

- *Partial-duration analysis*, where floods that exceed a given discharge during a period of record constitute the data set; and
- *Log-Pearson Types I & II* are magnitude-frequency distributions similar to, but differing in detail, from the Gumbel distribution.

Selection of the most appropriate hydrologic model can be complex but the Gumbel distribution is a good robust general model that will serve our purposes well enough. If the data plot as a straight line on Gumbel paper, then they conform to the theoretical distribution of extreme events.

*Two assumptions of Gumbel analysis worth noting* here are that:

- The data series is “stationary.” That is, there is no time-trend to the flood series. If the series is shifting (as might occur in a changing climate, the statistics are no longer valid). Non-stationarity is becoming an increasingly significant problem here in Canada because global warming appears to be associated with increasing storminess which in turn is causing our rivers to generate larger floods in recent decades. Squamish River north of
Vancouver is a well-documented example of a river that now has an annual flood that is increasing over time.

- A second important assumption is that, when the Gumbel line is extrapolated beyond the data range, the projections assume that the Gumbel distribution applies to the entire range. This is the same statistical problem we face whenever we extrapolate a best-fit curve beyond the data on which it is based. For example, predicting the 10 000 year-flood would probably be silly because climatic and hydrologic conditions over that length of time likely will change or physical constraints may begin to limit flood size (for example, rainfall intensity has a physical limit).

Sampling errors can be estimated to yield a standard error and so indicate reliability of flood estimates. Obviously this error becomes larger as the predicted flood RI becomes longer than the length of record. Obviously predicting the 200-year flood from a 5-year record is a rather hazardous business!

Before we leave the concept of formative or dominant discharge and the means of measuring it, we should acknowledge that the notion has its critics. The idea of a dominant discharge is very useful as a means of quantifying the “size” of the hydrologic system forming a channel and it is widely used in numerical models in fluvial geomorphology. Nevertheless, it is a simplification of a complex reality and we should recognize its limitations.

Principal concerns are:

- Channels are formed by all flows so it stands to reason that no single discharge can by itself capture all this hydrologic complexity.
- Dominant discharge may average out to be the most-probable annual flood ($Q_{1.58}$) but many rivers fall well out in the tails of the recurrence interval distribution so that RI values vary between 1 and 30 years. It is not clear why this variation in return interval occurs but some of the reported channels may be incised.
- When you get right down to it, dominant discharge is a rather vague concept. For example, dominant for what aspect of process and morphology? There is no fundamental reason that the dominant discharge for most effectively transporting sediment is the same
discharge that dominates erosion and deposition or the processes that govern river planform (what controls meander wavelength?). For example, in a typical sand-bed river most sediment is moved by relatively frequent small flows but shifts in channel alignment (channel migration and channel avulsion) are related to larger flood flows. Many argue that the concept of a dominant discharge only makes sense if referred to a particular process or morphology.

- Others have observed that bankfull discharge should have a constant recurrence interval along the length of a river but it is common for it to vary considerably. Typically, upstream reaches reflect the effects of large less frequent flows and downstream reaches are more sensitive to flows of higher recurrence interval. That is, some reaches of rivers may be more “alluvial” (more responsive) than others.

- Some river channels never attain an equilibrium condition associated with a dominant discharge because they are always “recovering” from the last significant high flow. This state of transience occurs because relaxation time (after a disturbance) is much longer than the frequency of system perturbations. This is sometimes called “non-equilibrium” behaviour.

The debate continues but meanwhile we make considerable use of this imperfect measure of the size of a river because there is no better alternative available at present.

4. Measuring Flow Velocity and Discharge

Regardless of how we might conceptualize channel formation, a fundamental control on channel size is the discharge and the measurement of discharge depends on the measurement of flow velocity. You will recall that streamflow measurement is based on equation (2.3): the continuity relation \( Q = A \bar{V} \) (discharge, \( Q \), is the product of cross sectional area, A, and mean flow velocity, \( \bar{V} \)). The form of the channel cross-section must be surveyed to determine the cross-sectional area (A) and a flow meter or current meter must be used to determine the average flow velocity (V).
In practice discharge is measured by subdividing the channel cross-section into a number (typically 8-10) subsections or components and determining the discharge in each and summing them. This technique is called the component method (see Figure 2.12).

\[ Q = \sum_{n=1}^{8} A_n \bar{v}_n = \sum_{n=1}^{8} W_n d_n \bar{v}_n \]

Figure 2.12: The Component Method of measuring discharge of a river

A plot of flow velocity versus depth of flow is called the velocity profile (see Figure 2.13). Velocity varies with depth in a non-linear manner so the average velocity for the subsection is not at mid-depth (0.5d) but rather at 0.6d from the water surface. So if you want to determine the average velocity at a vertical, a single measurement can be taken at 3/5 of the depth from the water surface. River measurement technicians often use this method or the slightly more laborious but more accurate method of averaging two flow velocity measurements, one at 0.2d and another at 0.8d.

Flow velocity is measured with a current meter. These come in a variety of types but most consist of a hand-held or cable-suspended meter with a propeller or set of cups that rotate faster as the velocity increases. The rate of rotation is recorded electrically and correlated with the velocity via a rating curve. You will get to use a current meter and learn how to obtain velocity profiles on the field trip. Figure 2.14 shows GEOG 313 students using a hand-held current meter.
Figure 2.13: The velocity profile in a river channel is log-linear (log depth versus velocity). Average velocity for the vertical can be read as a single value at 0.6D from the water surface or as the average of measurements at 0.2D and 0.8D from the water surface.

Figure 2.14: Students using current meters mounted on wading rods to measure flow velocity of the Nicola River on a GEOG 313 field trip.
on a wading rod. On rivers that are too large to wade current meters must be deployed from boats on a cable and winch system such as that being used by the vessel shown in Figure 2.15.

Figure 2.15: A flat-bottomed river boat is an ideal platform from which to winch the cable-supported current meter shown here on an SFU-Geography survey boat working on Beatton River in northern British Columbia. The large lead “fish” seen below the current meter is necessary to stabilize the instrument in the deep fast-flowing water.

The distribution of flow velocity within a channel cross-section is often depicted as an isovel diagram. Isovels are lines joining points of equal flow velocity. They tend to parallel the channel boundary with lowest velocity indicated near the boundary (where friction is greatest) and the highest velocity in the centre of the channel near (but often not quite at) the water surface where frictional effects are least. The highest velocity in a channel cross-section often occurs just below the water surface because of frictional effects (waves) at the air/water interface. Isovel patterns tend to be quite symmetrical in straight channels but are skewed in channel bends where the boundary form is asymmetrical.
Figure 2.16: Isovel diagrams constructed from point velocity measurements. Typical patterns are shown for symmetrical and asymmetrical channel cross sections.

Discharge in Canadian rivers is routinely measured by the Water Survey of Canada and daily records are now posted on the WEB. (http://www.wsc.ec.gc.ca/index.html), including real-time data for a number of rivers in BC (http://scitech.pyr.ec.gc.ca/waterweb/disclaimerB.asp). The discharge is not actually directly measured every day, however, just often enough to establish a reliable discharge rating curve. A discharge rating curve defines the relationship between the discharge and the gauge height at the site. Gauging stations are located in places where the
rating does not “shift” very much over time. *Shifting* is caused by erosion/deposition within the measurement section that changes the relationship between the stage and the discharge. Ideal measurement sections are in bedrock channels but most are located in alluvial reaches of rivers. Gauging stations with shifting rating curves are expensive to maintain because they must be more frequently measured for discharge. Discharge on most rivers takes about a half day to a day to complete and involves at least two technicians.

Gauge heights are measured daily, often automatically by transducers on the channel bed. Flow-depth data are either recorded on a data logger or are telemetered directly to WSC offices in the area (Figure 2.18).

![Figure 2.18: A stage-discharge rating curve is a “best-fit” line through a series of discharge measurements obtained at a variety of river stage.](image)

On days when the discharge is not actually measured in the field the stage and rating curve are used to estimate the discharge on those days.
A question left hanging in this discussion of what determines the size of a river is what determines the velocity in the continuity relationship. The answer to this question comes to us from the fields of fluid mechanics and hydraulics and will be explored in Chapter 3.

5. Some Further Reading
I suggest that revisiting a general textbook on geomorphology to read the chapter(s) on rivers and fluvial geomorphology remains a sound strategy. As noted in Chapter 1, three good sources in the Library are Easterbrook (1999), Ritter et al (2002) and Trenhaile (2006):

Easterbrook, D.J. 1999 Surface Processes and Landforms, Prentice Hall.

I suggest that, at this stage, you limit your further reading (beyond these notes) of more advanced material to the following sections of Knighton (1998):


6. What’s Next?
Now that we have some basic concepts of river hydrology behind us we can focus on the forces that are acting in a river channel to form and shape it. We now know from the continuity relation that the size of a channel, given some discharge, is determined by the flow velocity. But what determines the flow velocity? This fundamental question in fluid mechanics is the subject of Chapter 3.