LECTURE SECTION 4

River Channel Geomorphology

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THE BANKFULL CHANNEL

What is “the bankfull channel”? Fundamentally, it is an intuitive concept: you know it when you see it. And yet, there are complications...

To be most precise, we need to define the bankfull channel with reference to the adjacent floodplain—i.e. the bankfull channel is the feature that is bounded by the floodplain. What is the floodplain? That surface at the elevation of the top of the bankfull channel.

We can resolve this circular definition at either node (channel or floodplain); to start, we will make use an intuitive definition of the bankfull channel and refine it over the next several lectures. So, for now, the bankfull channel is the channel of an alluvial river (recall the first lecture) that has been constructed by, and is in approximate equilibrium with, the river's current hydrologic and sedimentologic regime.

The first lecture (p. 1-13) suggest some ways traditionally used to identify the bankfull channel (or its adjacent floodplain). They are adequate for our purposes, for now.

Hydraulic Geometry of Channels

In its most common definition, the hydraulic geometry refers to the way in which the water’s width, depth, and velocity in a channel change with changes in discharge. Although we might acknowledge that other parameters of channel form or of the water flow can also vary (such as slope, roughness, or degree of meandering), these three parameters have the singular property that

\[ Q = w \cdot d \cdot u \]  

and so a change in Q must be fully reflected by changes in the width, depth, and velocity.
Discharge in a stream system can change in two ways:

(1) The changing dimensions of the flow at a single gauging location as discharge changes during the passage of a flood. This type of change is measured by the *at-a-station hydraulic geometry*.

(2) The general increase in discharge as we move downstream and so collect runoff from a progressively greater drainage area. This is measured by the *downstream hydraulic geometry*. 
By convention, the hydraulic geometry relationships are written with the following symbols:

\[ w = aQ^b \]  \hspace{2cm} (2)  
\[ d = cQ^f \]  \hspace{2cm} (3)  
\[ u = kQ^m \]  \hspace{2cm} (4)  

Multiplying these three equations together,

\[ w \cdot d \cdot u = a \cdot c \cdot k \cdot Q^{(b + f + m)}, \]  \hspace{2cm} (5)  

And because \( Q = w \cdot d \cdot u \),

\[ Q = a \cdot c \cdot k \cdot Q^{(b + f + m)} \]  \hspace{2cm} (6)  

and so

\[ a \cdot c \cdot k = 1 \]  \hspace{2cm} (7)  
\[ b + f + m = 1 \]  \hspace{2cm} (8)  

Although every stream and stream system has differences, repeated measurements of hydraulic geometry yields remarkably similar values.

Values classically reported (data of the 1950’s and 1960’s) for the hydraulic geometry exponents \( b, f, \) and \( m \):

<table>
<thead>
<tr>
<th>TYPE</th>
<th>( b )  (width)</th>
<th>( f )  (depth)</th>
<th>( m )  (velocity)</th>
</tr>
</thead>
<tbody>
<tr>
<td>At-a-Station</td>
<td>0.26</td>
<td>0.40</td>
<td>0.34</td>
</tr>
<tr>
<td>Downstream</td>
<td>0.5</td>
<td>0.4</td>
<td>0.1</td>
</tr>
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</table>
Downstream exponents

<table>
<thead>
<tr>
<th>Channel</th>
<th>$b$ (width)</th>
<th>$f$ (depth)</th>
<th>$m$ (velocity)</th>
<th>Data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand and fine alluvium</td>
<td>0.50</td>
<td>0.40</td>
<td>0.10</td>
<td>Leopold and Maddock (1953)</td>
</tr>
<tr>
<td>Average values for midwestern USA</td>
<td>0.50</td>
<td>0.30</td>
<td>0.20</td>
<td>Leopold et al. (1964, p. 244)</td>
</tr>
<tr>
<td>Ephemeral streams in semi-arid USA</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gravel and cobble</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Green River, Wyoming</td>
<td>0.55</td>
<td>0.35</td>
<td>0.10</td>
<td>Dunne and Leopold (1978, p. 639)</td>
</tr>
<tr>
<td>Upper Salmon River, Idaho</td>
<td>0.54</td>
<td>0.34</td>
<td>0.12</td>
<td>Emmett (1975)</td>
</tr>
<tr>
<td>17 Alaska rivers</td>
<td>0.50</td>
<td>0.35</td>
<td>0.15</td>
<td>Emmett (1972)</td>
</tr>
<tr>
<td>70 Alberta rivers</td>
<td>0.53</td>
<td>0.33</td>
<td>0.14</td>
<td>Bray (1982b)</td>
</tr>
<tr>
<td>62 British rivers</td>
<td>0.52</td>
<td>0.39</td>
<td>0.10</td>
<td>Hey and Thorne (1986)</td>
</tr>
<tr>
<td>6 New Zealand rivers</td>
<td>0.48</td>
<td>0.43</td>
<td>0.11</td>
<td>Griffiths (1980)</td>
</tr>
<tr>
<td>24 Colorado rivers</td>
<td>0.48</td>
<td>0.37</td>
<td>0.14</td>
<td>Andrews (1984)</td>
</tr>
<tr>
<td>Boulder</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bear River, Wyoming</td>
<td>0.62</td>
<td>0.31</td>
<td>0.06</td>
<td>Virmani (1973)</td>
</tr>
<tr>
<td>Western USA rivers</td>
<td>0.53</td>
<td>0.38</td>
<td>0.09</td>
<td>Barnes (1967)</td>
</tr>
<tr>
<td>32 high-gradient western USA rivers</td>
<td>0.505</td>
<td>–</td>
<td>–</td>
<td>Österkamp and Hedman (1977)</td>
</tr>
</tbody>
</table>
Graphically, these exponents equal the slopes of the lines of discharge plotted against each channel parameter on log-log graph paper. Physically, these values indicate that increasing discharge at one location in a channel ("At-a-station") is primarily reflected by increasing **depth and velocity**; relative to the magnitude of the discharge increase, flow width changes least of the three (i.e. b < f or m).

In contrast, the physical dimensions of channels tend to increase sufficiently in the downstream direction ("Downstream") to account for nearly all of the associated increase in discharge--that is, **increasing depth and width**. Notice that f (the velocity exponent) is small but positive, and so average velocity increases downstream as well.

The parameters a, c, and k do not have any such universal values; they depend on the specific size of the channels being measured and also on the units of measurement being used.

**The Theory of Minimum Variance in Hydraulic Geometry Relationships**

One of the earliest efforts to “explain” the consistency in the hydraulic geometry exponents, Langbein (1964) suggested that because the fluvial system is too complex to explain deterministically, it should instead be characterized in terms of its “most probable” state. He postulated that all of the changes within a channel that are possible to accommodate an increase in Q will occur, and they will occur in such a fashion that the net change of all the parameters has been minimized. This is akin to achieving the necessary effort with the minimum amount of work by the system. This concept draws from classical thermodynamics, for example the even partitioning of thermal energy into the full number of available vibrational modes of a molecule.

Here, we are minimizing the variances—so we look to minimize the sum of the squares of the changes.

How will we know if this is the correct explanation for observed changes? We won’t, but if it explains the observations, then perhaps we will believe it.

An example: Imagine a river in a bedrock slot of fixed width. Thus any increase in Q can only be accommodated by increases in velocity and in depth. So plot the changes in d and v (actually, their logarithms) as a consequence of the change in Q between time “0” and time “1”:
Note that

\[ S^2 = (\Delta \log d)^2 + (\Delta \log v)^2 \]  

(9)

or

\[ \left( \frac{S}{\Delta \log Q} \right)^2 = \left( \frac{\Delta \log d}{\Delta \log Q} \right)^2 + \left( \frac{\Delta \log v}{\Delta \log Q} \right)^2 \]  

(10)

Now,

\[ \frac{\Delta \log d}{\Delta \log Q} \]

…is the slope of the line of a (log d)-(log Q) plot, which is exactly the exponent \( f \) in the hydraulic geometry relation plot of \( d = cQ^f \); this is also the case for the second term on the RH side of the equation, which gives the exponent \( m \).
Therefore,

\[
\left( \frac{S}{\Delta \log Q} \right)^2 = f^2 + m^2
\]  

(11)

What we want to minimize, for a given change in \((\log Q)\), is the change in the sum of \((\log d)^2\) and \((\log v)^2\). However, their sum is \(S^2\), and so this problem can be expressed as finding the minimum of \(m^2 + f^2\).

Now,

\[
m + b + f = 1
\]

but \(b = 0\), and so

\[
m = 1 - f
\]  

(12)

So:

\[
m^2 + f^2 = 2f^2 - 2f + 1
\]  

(13)

Find the minimum by taking the derivative and setting it equal to 0:

\[
4f - 2 = 0
\]  

(14)

So by the predictions of minimum variance, the at-a-station exponents have the values \(f = \frac{1}{2}\) and \(m = \frac{1}{2}\) for the case of a bedrock slot.

If we add more relationships, and fewer constraints, the problem gets more complicated rapidly, although we can normally add additional equations to keep pace with the number of variables (minus one). The additional “equation” that is needed to solve the system of equations is the assumption of minimum variance—that’s why we need it. Knighton (p. 182) has another, more complicated example that illustrates the same basic principle.
But—how do you know when you've added enough, but not too many, additional variables and physical relationships?

The minimum-variance framework was quite attractive for several decades, but it lacked a strong theoretical foundation (why should a fluvial system seek to minimize variance?). Ferguson (1986) argued the determinism and hydraulics, not “metaphysics” (his term), should be used to understand these systems. He approached the problem in terms of linear algebra—make sure you have as many equations as unknowns and the system can (at least in theory) be specified.

So, for example, in the at-a-station changes, we have 3 equation…

\[ Q = w \ d \ v \quad (\text{continuity}) \]
\[ V = f(d, \ n \ or \ D, \ S) \quad (\text{Manning’s or Strickler’s equation}) \]
\[ W = f(d) \quad (\text{channel geometry}) \]

…and three unknowns (w, d, and v). So this system can in theory be solved; the complexities arise because width and depth are also functions of Q, and because the variation of width with depth can change as a result of recent large floods. So we have a deterministic, but probably unsolvable, system of equations. This is why the empirical relationships continue to hold such appeal!
DOMINANT DISCHARGE

"Dominant Discharge" and the Size of the Bankfull Channel

What is the flow that determines the size of the bankfull channel? Is it:

- the bankfull flow?
- the flow that moves the most total sediment, on average, over time?
- the flow that moves the most bedload sediment, on average, over time?

Is there one such flow?

Intuitively, flows that move the most sediment should somehow determine the size of the channel in which that sediment is transported. The flows of interest must be "large" flows, because most channels, most of the time, move no sediment--the average daily flow, for example, typically transports no bedload at all.

How large is "large"?

- very large flows are effective but rare; if they only occur once in several decades or even centuries, how significant can they be to the form we see today?
- less large flows are less effective but far more common.

Wolman and Miller (1960, J. Geol.) suggested that the most effective flows would be those that happened relatively frequently, because those that were very large might move more sediment per unit time, but they simply did not occur very often and so their cumulative duration was not great:
The original data set of Wolman and Miller (1960) was dominated by sand-bedded rivers, and it showed a peak in the curve of frequency x transport rate (i.e. curve C in the figure above) at flood stages at about the 1-yr recurrence interval (using the partial duration series—more on this later). If we consider gravel-bed rivers, which have a higher threshold of movement, we might expect a slightly longer recurrence for the peak of the sediment-discharge curve. In any case, we expect by this reasoning that approximately “the annual flood” will fill the channel forms the channel that carries all flows downstream. But even if this is true, we are no closer to understanding how the transport of sediment down the channel should affect the size of that channel—it is simply an intriguing correlation!

So: is the dominant sediment-transporting flow the same as the dominant channel-forming flow?
Evidence for:

1. It’s intuitive.
2. Field data are suggestive:

Sure enough, floods with a recurrence interval of about 1 year on the partial series are the ones that most commonly “just” fill the channel. But not all do, particularly those that have had “extreme” events in their recent past (e.g., hurricane or other very large flood). We might expect a relationship between the “healing time” after a large flood and the recurrence of those large floods. If the former is shorter than the latter (and, barring threshold changes such as channel incision [see below] this is probably true in most cases), we may be ok.
Looking now from the perspective of geomorphic processes, the suite of discharges that move down a channel can be categorized into three fundamental types:

- Those flows that are insufficient to move sediment in any significant quantities, and that do not move “any” bedload (but recall problems with the definition of first motion).

- Those flows that move sediment within a stable channel form: these flows shape the bed, but not the banks, of the channel.

- Those flows that affect the channel form itself.

What discriminates between the two sediment-transporting two types? (The difference between the first one and the others was the subject of Lecture Section 3.)

Using the conceptual framework from Pickup and Warner (1976) and the terminology and data of Carling (1988) from lowland Great Britain streams:

<table>
<thead>
<tr>
<th>$\tau_{\text{critical,bed}}$</th>
<th>$\tau_{\text{critical,bank}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>&quot;Phase 1&quot; (winnowing of fine sediment)</td>
<td>&quot;Phase 2&quot; (bedload mobilization)</td>
</tr>
</tbody>
</table>

--> --> --> --> INCREASING DISCHARGE --> --> --> -->
TRANSPORT TYPE | RANGE OF DISCHARGES (relative to bankfull, $Q_{bkfl}$) | RECURRENCE INTERVAL OF DISCHARGE (RI)
---|---|---
Phase 1 | $Q < 0.6 Q_{bkfl}$ | $RI < 0.35 \text{ yr}$
Phase 2 | $0.6 Q_{bkfl} < Q < 1.3 Q_{bkfl}$ | $0.35 \text{ yr} < RI < 5 \text{ yr}$
Phase 3 | $1.3 Q_{bkfl} < Q$ | $5 \text{ to } 7 \text{ yr} < RI$

Note that $\tau_{critical,bed}$ is largely a function of the interplay of hydraulics and sediment, and so we might expect relatively consistent relationships amongst all alluvial rivers. Field data, world-wide, support this assumption: bedload-transporting events in gravel-bed rivers, for example, are reported as occurring typically 6-12 times per year (i.e. a sub-bankfull, sub-1.5-yr event).

This relationship offers an opportunity to characterize quantitatively the relative “stability” of different channels, by comparing the value of $\tau_{critical,bed}$ to that of the shear stress applied at bankfull discharge, $\tau_{bankfull,bed}$. Normally we would expect this ratio to be less than one ($\tau_{critical,bed} < \tau_{bankfull,bed}$). However, different values of this ratio from different channels provides a mechanism to “rank” channels of different stability (see Olsen et al., 1997, JAWRA); and, in particular, to recognize any channels where these relative values are reversed (i.e. $\tau_{critical,bed} > \tau_{bankfull,bed}$). Under such conditions channel sediment is likely to be quite stable, because depth (and thus shear stress) increases only very slowly with increasing discharge once flow overtops the bankfull channel.

In contrast, $\tau_{critical,bank}$ depends on local, non-alluvial conditions, such as bank vegetation, logs, obstructions in the channel, etc. So this value is likely more variable from one river to another, and perhaps even from one reach to another even on the same river.

There is also a time dependency in channel-altering flows: in humid environments, channel form can be stabilized in part by bank vegetation, and so the rate of channel restabilization depends on the rate of vegetation regrowth. Two large flows in relatively rapid succession may accomplish significantly more work on the channel form than those same two flows with some years of relative quiescence in between.
Determining a Stable Channel Cross-Sectional Form

Perhaps we can learn about the determinants of channel size by considering what controls the shape and width of a channel. No general theory exists for predicting the equilibrium shape of any arbitrary channel (recall Ferguson’s problem), but we can derive a deterministic solution for simplified cases:

Consider a “threshold channel” (i.e. channel boundaries are formed of loose sediment just on the threshold of movement).

1. Simple case: straight banks (trapezoidal channel)
2. Complex case: banks of arbitrary shape

A force balance approach to predicting channel cross-sectional form works for non-cohesive sediment.

With cohesive sediment, the bank can hold vertical slopes banks and the cross-sectional shape simplifies to a trough. This is not very instructive.

Case 1 - Trapezoidal Channel
Simplest geometry for non-cohesive sediment is a trapezoid.

\[ W = \text{submerged weight of particle} \ [i.e., (\rho_s - \rho)g] \]

Slope down-bank = \( \alpha \)
Slope down-channel = \( \theta \)
2 forces act on particles on the bank:

1. Fluid shear: \( a \cdot \tau_s \)
   where \( a \) = drag coefficient and \( \tau_s \) = shear stress on the channel sides

2. Gravity down the bank: \( W \cdot \sin \alpha \) (where \( W \) is the buoyant weight of the sediment particle)

The resultant force oriented obliquely down the bank is given by:

\[
F_D = W \cdot \tan \alpha \cdot \sqrt{\frac{W^2 \sin^2 \alpha + a^2 \tau_s^2}{a^2 \tau_s^2}}
\]  
(15)

The frictional force that resists motion is given by:

\[
F_R = W \cdot \cos \alpha \cdot \tan \phi
\]  
(16)

Where \( \tan \phi \) is the coefficient of internal friction of the sediment (i.e. \( \phi \) is the internal friction angle, which is approximately equal to the “angle of repose”).

At impending motion on the sideslopes the forces in (1) and (2) balance:

\[
F_D = F_R
\]  
(17)

which can be rearranged for \( \tau_s \) - the critical shear stress for the banks

\[
\tau_s = \frac{W}{a} \cdot \cos \alpha \cdot \tan \phi \cdot \sqrt{1 - \frac{\tan^2 \alpha}{\tan^2 \phi}}
\]  
(18)
Similarly, at impending motion on the channel bed, the force arising from the basal shear stress \((a \cdot \tau_b)\) will equal the frictional resistance to downstream motion \((W \cos \theta \tan \phi)\) (the same result comes from equation (3) by setting \(\alpha = 0\)). So:

\[
\tau_b = \frac{W}{a} \cos \theta \tan \phi \tag{19}
\]

Since slopes are low for most channels (and so \(\cos \theta \approx 1\)), the ratio of the sideslope shear stress on the bank to the basal shear stress at the bed is given by the ratio of equation (4) to equation (5). So, at incipient motion:

\[
\frac{\tau_s}{\tau_c} = \frac{\cos \alpha}{\sqrt{1 - \frac{\tan^2 \alpha}{\tan^2 \phi}}} \tag{20}
\]

which in turn simplifies to

\[
\frac{\tau_s}{\tau_c} = \sqrt{1 - \frac{\sin^2 \alpha}{\sin^2 \phi}} \tag{21}
\]

Gary Parker noted a missing factor in this development, namely the lateral momentum exchange due to the effect of wall drag on the velocity distribution across the channel (i.e. flow velocity is slowed as it approaches the wall, and so the shear stress applied to the sidewalls is somewhat less than predicted by the “standard” shear stress equation,

\[
\tau = \rho g H S \tag{22}
\]

How much less?

Field measurements of near bank velocity gradients and numerical modeling have shown that for a wide trapezoidal channel \(\tau_s\) at the base of the bank is approximately given by

\[
\tau_s \approx (0.75) \rho g H S \tag{23}
\]
**Example:** Design a trapezoidal, straight channel (i.e., a canal) that will accommodate a discharge \(Q\) = 2000 m²/s on a gradient of 10⁻⁴ in moderately rounded gravel with \(d_{50} = 50\) mm.

1. Choose some channel sideslope \((\alpha) <\) angle of repose \((\phi)\)

2. Calculate

\[
\frac{\tau_s}{\tau_b} = \sqrt{1 - \frac{\sin^2 \alpha}{\sin^2 \phi}} \tag{24}
\]

Set \(\tau_b = \tau_c\) based on the particle size \((d_{50} = 50\) mm) and Shields equation

3. Calculate \(\tau_s\) required to scour banks from

\[
\tau_s \approx (0.75) \rho g H_{\text{max}} S \tag{25}
\]

where \(H_{\text{max}}\) is the flow depth required to mobilize the bed surface. This approach allows you to calculate the maximum flow depth \((H_{\text{max}})\) that will NOT cause scour of the base of the banks.

4. Having obtained \(H_{\text{max}}\) and specified \(\alpha\) and \(\theta\), and being able to determine Manning’s \(n\) from the channel grain size, you can adjust the width to accommodate the imposed discharge \((Q)\) by iterative solutions.
CHANNEL INCISION

A fundamental condition of channel-forming processes is that of negative feedback—as one variable changes, others adjust to counterbalance the effects of that change. So, for example, larger flows erode the bed of a channel, but the downcutting in turn reduces the channel slope and so makes those flows less competent to move sediment. If this negative feedback is absent (e.g., only well-sorted sediment is available and so coarsening is impossible, or steep channel gradients permit dramatic deepening without corresponding reduction in basal shear stress) then catastrophic incision can occur.

Channel incision is the condition whereby a channel does not expand to flow increases in a proportional fashion. Instead, the channel has crossed a threshold, where the relatively slow expansion of channel width and depth is overtaken by catastrophic downcutting. A bankfull channel may still be defined, but its size and rate of growth is insignificant (and irrelevant) in comparison to the incised valley that now contains it.

Channel incision is the rapid, nearly uncontrolled downcutting of a stream bed. Incision typically progresses upstream as knickpoints, near-vertical steps in the stream profile that can migrate upstream at a rate of up to many meters per year. Although expansion of a channel is damaging under any circumstance, true incision is particularly problematic because the resultant stream is generally devoid of habitat diversity and the eroded sediment can clog the downstream system.

Traditional geomorphology stressed the initiation of channel incision by a change in base level, the minimum elevation of erosive flow. Depending on the location and scale of the channel, base level can be set by sea level, lake level, trunk river level, or even just a local shelf of resistant bedrock in the stream bed. In areas of significant land-use change, however, incision can occur under other circumstances as well. High flows in particular are necessary, but in addition we usually also observe:

- steep channel gradients;
- easily erodible substrate, typically sand; and
- few or only widely spaced controls on the "grade" or bed slope of the channel to anchor the bed elevation, which are most commonly formed by rock outcrops,
large logs lying on the channel bed, or constructed features such as culverts.

Incision represents a loss of geomorphic balance between

(1) The *forces inducing sediment movement* (the moving water) and

(2) The *erosional resistance of the stream bed*, determined by:

• sediment size,

• channel roughness, and

• the action of anchoring debris.
FLOODPLAINS

Recall our early definition of “floodplain”:

The surface that has been built up next to a river channel under the current hydrologic and sedimentological regime. It is composed of alluvium, the sediment carried by the river. An alluvial channel is bounded by a floodplain; conversely, a channel formed within a true floodplain is by definition alluvial.

In contrast, a terrace is also a constructed surface and also underlain by alluvium, but it has not formed under the current regime of the river. Instead it represents floodplain formation at an earlier time when, for whatever reasons, deposition was occurring at a higher elevation. (Note that if the earlier deposition occurred at a lower elevation than at present, the remnant terrace would be buried by the modern floodplain and so we could not see it.)

In general, we also expect floodplains to be relatively smooth and planar (thus their name!). They are part of the river system, and we do well to think of them as part of the “high-flow channel” of the river. Modern civilization’s utilization of floodplains, in this light, becomes difficult to justify (and the physical and ecological consequences of that occupation more easy to understand).

The precise definition of “floodplain,” however, is difficult from a geomorphic perspective: at best, we can say that it is a relatively flat surface occupying much of a valley bottom, normally underlain by unconsolidated sediment—not all of that sediment is necessarily deposited by the river, but it must somehow all “relate” to the activity of the present river. There must also have a hydrologic connection, since the surface must be subject to periodic flooding. (Note that there are also engineering and planning definitions of floodplain that are much more explicit and unambiguous.)
Figure 36.—Relation of flood-damage stage to elevation of flood plain at several river cross sections.
Figure 38.—Map and cross section of a typical point bar of the flood plain of Watts Branch, 1 mile northwest of Rockville, Md.
A floodplain may include the following features (from Leopold et al., 1964):

- The river channel
- Oxbows or oxbow lakes, representing the cutoff portion of meander bends
- Point bars, loci of deposition on the convex (inside) side of river curves
- Meander scrolls, depressions and rises on the convex side of bends formed as the channel migrated laterally downvalley and toward the concave bank
- Sloughs, areas of dead water, formed both in meander-scroll depressions and along the valley walls as flood flows move directly downvalley, scouring adjacent to valley walls
- Natural levees, raised berms or crests above the flood plain surface adjacent to the channel, usually containing coarser materials deposited as floods flow over the top of the channel banks. These are most frequently found at the concave banks. Where most of the load in transit is very fine-grained, natural levees may be absent or nearly imperceptible.
- Backswamp deposits, overbank deposits of finer sediments deposited in slack water ponded between the natural levees and the valley wall or terrace riser.
- Sand splays, deposits of flood debris usually of coarser sand particles in the form of splays or scattered debris.

On major rivers (e.g., Amazon, Mississippi, Mekong) all of these features may be clearly represented. On small rivers many of them may be hard to distinguish or absent where floodplain deposits are subject to rapid removal and alteration.
Processes of Floodplain Formation

3 basic models of floodplain formation:

1. Lateral migration and deposition of point bar material
2. Overbank deposition
3. Patchwork mosaic floodplain formed by local log jams

1. Lateral Migration

Stratigraphy: Bedload overlain with suspended load

Features: Oxbows, scroll bars, sloughs, side channels, vegetation age gradients, asymmetric topography near bends

2. Overbank

Stratigraphy: Suspended load only

Features: Natural levees, backswamps, uniform topography, uniform vegetation age

3. Patchwork Mosaic

Stratigraphy: Bedload overlain by suspended load

Features: Side channels, sloughs, irregular multi-elevation floodplain, vegetation patchwork
1. **Lateral Migration**—Floodplain formed by lateral migration of point bars.
2. **Overbank deposition**—suspended sediment dominates floodplain formation. Occurs in areas where channel migration rates are slow relative to rate of vertical deposition.
Deposition rates highest where close to channel, at bends, and at flow constrictions (determined from $^{136}\text{Ce}$ from nuclear testing).
1950’s-1970’s

Post-Chernobyl

(b)

Deposition Rate
(mm year⁻¹)

River Severn

Mill Avon

0 200m

1

2

3

4

5

6

7
Figure 7.2. Map showing complex distribution of various types of deposits on a portion of the Mississippi River floodplain near Grand Tower, Ill. (Courtesy of S.E. Harris, Jr.)

Figure 7.3. Meander scroll topography formed by point bar deposition in a laterally migrating river. (From Hickin 1974. Used with permission of American Journal of Science)
Figure 7.4. Diagram of two profiles of meander scroll topography that has been preserved for considerable periods of time. Beatton River, Canada. (After Hickin and Nanson, *GSA Bulletin*, 1975, vol. 86, p. 491, fig. 6. Used with permission of the Geological Society of America)

Figure 7.7. Increase in elevation of floodplain with time. Lower curve from empirical data collected on floodplain of the Little Missouri River; upper curve was derived theoretically for Brandywine Creek (Pa.) by Wolman and Leopold (1957). Note different vertical scales. (From Everitt 1968. Used with permission of *American Journal of Science*)

3. Patchwork mosaic floodplain
Queets River:

(a) Floodplain map with contour lines and meters scale. Contour interval is 0.2 meters.

(b) Floodplain map with different grain size categories indicated by shading.
Model of patchwork mosaic floodplain development: a log deposited during the falling limb of a storm modifies the flow (a), enhancing both scour and deposition (b); additional LWD is incorporated and the key member is buried to form a stable island (c); growth of the island can eventually suture it to the adjacent floodplain (d).
Rates of Channel Migration

Wolman and Leopold say that scour is typically 1.75–2 x depth of flood flow (and locally 3–4 times the flow depth), and thus we expect the thickness of deposits to increase downvalley in concert with increasing bankfull dimensions.

Comparative rates of lateral and vertical construction of floodplains: generally, lateral is more important. From curve of elevations (see figure p. 33), most accretion occurs in first 50–100 yrs, and e.g. 3 m accretion >1000 yr. Compare to lateral migration rates (table below):

<table>
<thead>
<tr>
<th>River and location</th>
<th>Approximate size of drainage area (square miles)</th>
<th>Amount of movement (feet)</th>
<th>Period of measurement</th>
<th>Rate of movement (feet per year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tidal creeks in Massachusetts</td>
<td>0</td>
<td>60-75 yr.</td>
<td></td>
<td>0</td>
</tr>
<tr>
<td>Normal Brook near Terre Haute, Ind.</td>
<td>±1</td>
<td>30 1897-1910</td>
<td>2.3</td>
<td></td>
</tr>
<tr>
<td>Watts Branch near Rockville, Md.</td>
<td>4</td>
<td>0-10 1915-55</td>
<td>0-0.25</td>
<td></td>
</tr>
<tr>
<td>Rock Creek near Washington, D.C.</td>
<td>4</td>
<td>6 1953-56</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>Middle River near Bethlehem Church, near Staunton, Va.</td>
<td>7-60</td>
<td>0-20 1915-55</td>
<td>0-0.50</td>
<td></td>
</tr>
<tr>
<td>Tributary to Minnesota River near New Ulm, Minn.</td>
<td>18</td>
<td>25 10-15 yr.</td>
<td>2.5</td>
<td></td>
</tr>
<tr>
<td>North River, Parnassus quadrangle, Va.</td>
<td>10-15</td>
<td>250 1910-38</td>
<td>9</td>
<td></td>
</tr>
<tr>
<td>Seneca Creek at Dawsonville, Md.</td>
<td>50</td>
<td>410 1834-84</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>Laramie River near Ft. Laramie, Wyo.</td>
<td>101</td>
<td>0-10 50-100 yr.</td>
<td>0-0.20</td>
<td></td>
</tr>
<tr>
<td>Minnesota River near New Ulm, Minn.</td>
<td>4,600</td>
<td>100 1851-1954</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>Ramganga River near Shahabad, India</td>
<td>10,000</td>
<td>0 1910-38</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Colorado River near Needles, Calif.</td>
<td>100,000</td>
<td>2,900 1795-1806</td>
<td>264</td>
<td></td>
</tr>
<tr>
<td></td>
<td>100,000</td>
<td>1,050 1806-1883</td>
<td>14</td>
<td></td>
</tr>
<tr>
<td></td>
<td>100,000</td>
<td>790 1883-1945</td>
<td>13</td>
<td></td>
</tr>
<tr>
<td></td>
<td>170,600</td>
<td>20,000 1858-83</td>
<td>800</td>
<td></td>
</tr>
<tr>
<td></td>
<td>170,600</td>
<td>3,000 1883-1903</td>
<td>150</td>
<td></td>
</tr>
<tr>
<td></td>
<td>170,600</td>
<td>4,000 1903-1952</td>
<td>82</td>
<td></td>
</tr>
<tr>
<td></td>
<td>170,600</td>
<td>100 1942-52</td>
<td>10</td>
<td></td>
</tr>
<tr>
<td></td>
<td>170,600</td>
<td>3,800 1903-42</td>
<td>98</td>
<td></td>
</tr>
<tr>
<td></td>
<td>320,000</td>
<td>5,500 170 yr.</td>
<td>32</td>
<td></td>
</tr>
<tr>
<td></td>
<td>320,000</td>
<td>2,400 1896-1916</td>
<td>120</td>
<td></td>
</tr>
<tr>
<td>Kosi River, North Bihar, India</td>
<td>369,000</td>
<td>150 1883-1903</td>
<td>2,460</td>
<td></td>
</tr>
<tr>
<td>Missouri River near Peru, Nebr.</td>
<td>350,000</td>
<td>5,000 1883-1903</td>
<td>250</td>
<td></td>
</tr>
<tr>
<td>Mississippi River near Rosedale, Miss.</td>
<td>1,100,000</td>
<td>2,380 1930-45</td>
<td>158</td>
<td></td>
</tr>
<tr>
<td></td>
<td>1,100,000</td>
<td>9,500 1881-1913</td>
<td>630</td>
<td></td>
</tr>
</tbody>
</table>

From Wolman and Leopold 1957. See this work for data sources for individual rivers.
From USGS Professional Paper 282C:

“This study supports the views of several authors that the flood plain is composed of channel deposits, or point bars, and some overbank deposits. The relative amounts of each vary, but on the average, the proportion of overbank deposits appears to be small. This conclusion is supported by the uniform frequency of flooding and by the small amount of deposition observed in great floods…

“Frequency studies indicate that many flood plains are subject to flooding at approximately yearly intervals. These studies, as well as stratigraphic observations, indicate that the floodplain is also related to the present regimen of the stream flowing within it.

“If neither natural nor man-induced changes take place in the structural, climatic, or physiographic conditions which control the regimen of a natural channel, the channel will not form terraces by gradually building up its own flood plain until flooding no longer occurs. The flood plain can only be transformed into a terrace by some tectonic, climatic, or man-induced change which alters the regimen of the river…

“Lateral migration of a stream across its flood plain takes place with almost no change in channel width. The volume of material deposited tends to be about equal to the volume eroded. Material eroded from a drainage basin is only temporarily stored in the flood plain. Only when the stream erodes laterally into terraces of hillsides higher than the flood plain does the volume eroded exceed the volume deposited. Only in this case can stream-bank protection be expected to reduce the total sediment yield from a drainage basin.”
Nanson and Croke (1992, Geomorphology, volume 4, p. 459-486) proposed a genetic classification of flood plains based on gross differences in specific stream power ($\Omega$ — low, medium, and high) and the nature of the valley fill (cohesive and non-cohesive).

**Stream power** is work per unit channel length - can be considered how much, how fast:

$$\Omega = \rho g Q S$$  \hspace{1cm} (26)

where $\Omega$ is stream power, $\rho$ is density of water, $g$ is gravity, $Q$ is discharge, and $S$ is the channel slope.

Nanson and Croke recognized three categories of flood plains:

- **Class A**  High-energy, noncohesive
- **Class B**  Medium-energy, noncohesive
- **Class C**  Low-energy, cohesive

These different classes of flood plains tend to be formed by different mechanisms and host different flood plain features.

Nanson and Croke also derived 13 specific derivative orders and suborders, illustrated on the next three pages:
Type A: High Energy, Noncohesive

i) Confined Coarse-Textured Floodplain
   $\omega = >1000 \text{Wm}^{-2}$

ii) Confined Vertical-Accretion Sandy Floodplain
   $\omega = 300-1000 \text{Wm}^{-2}$

iii) Cut and Fill Floodplain
     $\omega = \sim 300 \text{Wm}^{-2}$

- Coarse sediment
- Few floodplain features
Type B: Medium Energy, Noncohesive

- Moderately confined channels
- Relatively coarse material
- More abundant floodplain features
Type C: Low Energy, Cohesive

- Well developed levees, finer sediment, complex floodplain features, mainstem wanders (but often slowly) across the floodplain.
CHANNEL PATTERNS

“Channel patterns,” to a geomorphologist, means the appearance of the channel in map view. The classic groupings are meandering, straight, and braided, briefly described in the first lecture section.

In most natural channels the ratio of channel length to straight-line down-valley distance lies between 1.5 and 2. Where this ratio, called the sinuosity, is less than 1.3 the channel is not termed "meandering" but instead is "sinuous" or "straight."
From Church (1992):

- **Decreasing Channel Stability** → **Increasing Sediment Supply**
- Bed material supply dominated channels
  - boulders, cobbles
  - gravel
  - step-pool cascade
  - sand

- **Increasing Channel Gradient** → **Decreasing Channel Calibre**
- Wash material supply dominated channels
  - wandering channels
  - meandering channels
  - braided channels
  - fine sand, silt
  - silt
  - anastomosed channels
Typical Relationships for River Meanders (from Leopold, Wolman, and Miller, 1964; and Williams, 1986)

Using relationships collected from almost 200 rivers world-wide, a number of consistent relationships emerge. These can be used to evaluate the relative stability or degree of disturbance in a river under consideration; they can also be used as design guidelines for constructed channels, assuming that a meandering river is the most stable channel pattern for the conditions of discharge, sediment load, and slope.

We must begin by defining the following variables:

\[ r_c = \text{the radius of curvature of the channel centerline}; \]
\[ \lambda = \text{the downvalley distance of one complete meander bend (i.e. the wavelength);} \]
\[ w_{bf} \text{ and } d_{bf} \text{ are the bankfull width and depth of the channel; and} \]
\[ P = \text{the sinuosity (channel length/wavelength)} \]
The most useful relationships, with all measurements in meters, are then as follows:

- $1.5 < P < 2.0$, for most "true" meandering channels \hspace{1cm} (27)

- $r_c = 1.5 \, w_{bf}^{1.12} \hspace{1cm} (28)$

- $\lambda = 4.53 \, r_c \, (= 6.8 \, w_{bf}^{1.12}) \hspace{1cm} (29)$

- $\lambda = 29.3 \, (w_{bf} \cdot d_{bf})^{0.65} \hspace{1cm} (30)$

Note that because each meander wavelength includes 2 pool-riffle sequences, pools should be spaced about 3.5 bankfull widths apart when measured in the downvalley direction, or about 5 to 7 widths apart when measured along the path of the channel itself.

**STRAIGHT RIVERS** are naturally uncommon because they are inherently unstable: any minor perturbation of the flow, such as caused by a hard projection or a small hollow in the bank, will tend to establish the oscillation of the thalweg that leads to concentrated scour of pools, point-bar formation, and a meandering pattern.
BRAIDED CHANNELS are identified wherever the flow divides into more than one thread. Braided channels are not as common as meandering ones, but they are of special interest because their rates of lateral shifting and of bank erosion are generally very much greater. Two conditions are necessary for braiding:

1. Sediment must be actively and frequently transported, and
2. The banks of the channel must be very easily eroded.

These conditions reflect the frequency at which small concentrations of gravel in the channel can begin to grow into midchannel bars, and the requirement that the concentration of the flow around these growing islands is as likely to erode the opposite bank as it is to sweep away the island itself.

Two other conditions commonly also are associated with braided channels:

3. Rapidly changing ("flashy") discharge, and
4. A heterogeneous (very mixed-size) sediment load.

In combination, these four conditions suggest that irregular but very active transport and deposition of sediment characterize the braided environment.

A number of empirical discriminations have been made between the different types of observed channel patterns, based on flow and channel characteristics. They are descriptive but do not offer any particular insight into why these patterns form!
RIVER PROFILES

So, who cares about river profiles?

• Practical -- how will river profiles (and sediment yields) react to engineering modifications (dams, dredging, straightening, etc...)?

• Theoretical -- essential element of landscape evolution; useful in reconstruction of past landscapes and rates of landform change (e.g., terrace remnants).

Controls on River Profiles

• Early thought regarding river profiles was bound up in a long running debate over whether valleys were formed by rivers, the Deluge, or by ocean currents at a time when the continents were submerged.

• Hutton (1795) recognized the essentially erosional nature of most landscapes and was intrigued by river profiles, which result from fluvial erosion and deposition.

Why do rivers meet at the same elevation? (Why aren’t there waterfalls at every confluence?)

• Playfair (1802) called attention to the accordance of river profiles at their confluences and argued for a “fine adjustment’ of erosion rates in main stem and tributary channels through which they keep pace with each other.
Why are river profiles convex?

G. K. Gilbert (1877) considered the problem of river long profile form and applied the following logic:

• Steeper channels erode faster
• Channels with greater drainage areas erode faster
  • Therefore, for a river system in equilibrium under a uniform uplift rate (i.e., where the erosion rate is uniform along the river) the river profile must progressively decrease in slope as the drainage area increases.
• Therefore, the equilibrium profile of a river system will be convex.

Theory: erosion rate, \(-\frac{dz}{dt}\), is a function of drainage area \((A)\) and slope \((S)\):

\[
-\left(\frac{dz}{dt}\right) = K A^m S^n
\]  

(31)

where \(z\) is height above datum; \(t\) is time; and \(K\), \(m\), and \(n\) are constants.
The Graded River

Mackin (1948) defined a graded river as “one in which ... slope is delicately adjusted to provide, with available discharge and with prevailing channel characteristic, just the velocity required for transportation of the load supplied from the discharge basin (and the river bed).” This definition stems from G. K. Gilbert’s work on what he recognized to be a mutual adjustment between velocity, discharge, slope and load. A graded river is also referred to as being in a state of “dynamic equilibrium” -- constant for even though the landscape may be rising or lowering, i.e. a steady-state morphology.

Recall from our discussion of the downstream hydraulic geometry that we considered how channels adjust their width and depth to accommodate the increase in discharge. Here, the only parameter that is presumed to adjust is the slope. Why? What about channel size, roughness, and channel pattern, all of which can probably adjust much more rapidly, and with no less of an influence on sediment- and water-transporting power, than slope.

Shulits (1941) suggested that the channel profile is graded to pass the supplied bed material, and because in a river system there is downstream attrition of sediment, the slope should also decline in the downstream direction.

Of course, the sediment could be declining in caliber in the downstream direction because the slope is decreasing—i.e. selective transport, and the reduced slope is causing the lower observed sizes, and not the other way around. You will have a chance to explore this conundrum during the sediment-routing exercise.
Alluvial and Strath Terraces

Alluvial terraces: constructional features resulting from episodes of valley filling and erosion -- interpreted as signal of changes in balance between sediment supply and transport capacity.

Strath terraces: erosional features resulting from lateral beveling of the valley floor followed by vertical incision and further planation -- interpreted as an indicator of relative uplift or incision, likely tectonic in origin.