

3 Water, networks and crossings

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3.1 Water balance

3.1.1 Earth

In case all ice would melt

The surface of the Earth is approximately 510,000,000 km² large (see page 23) and there is 1 390 000 000 km³ water. So, if there were no differences in temperature or ground level and water was equally dispersed over the Earth, the planet would be fully covered by a 2.7km deep ocean (Fig. 352)⁸². The 48m upper layer would be ice.

However, there is 148 900 000 km² land and 361 100 000 km³ water. So, 29% is land. It contains 3% of all existing water, and 2/3 of that part is frozen. If all ice would melt by global warming sea level would raise 66m. Water would submerge the most densely populated areas of the Earth. Fortunately the sun still adds snow to the poles.

The case of maximal glaciation

On the other hand, during an age of maximal glaciation the the amount of glacier ice would have been three times larger as the present ice volume. The sea level would have been lowered as much as 140 meters. The continental shelves would have been exposed to the air so man could live there.

The average height of the land is 823 m above sea level. We can calculate the potential mechanical power of the system of the water streaming to the sea over the land. Assuming that 37,000 km³ of runoff water will flow downhill 9 TW (see Fig. 2 and also Fig. 16) would have been produced by the runoff water.

The amounts of water

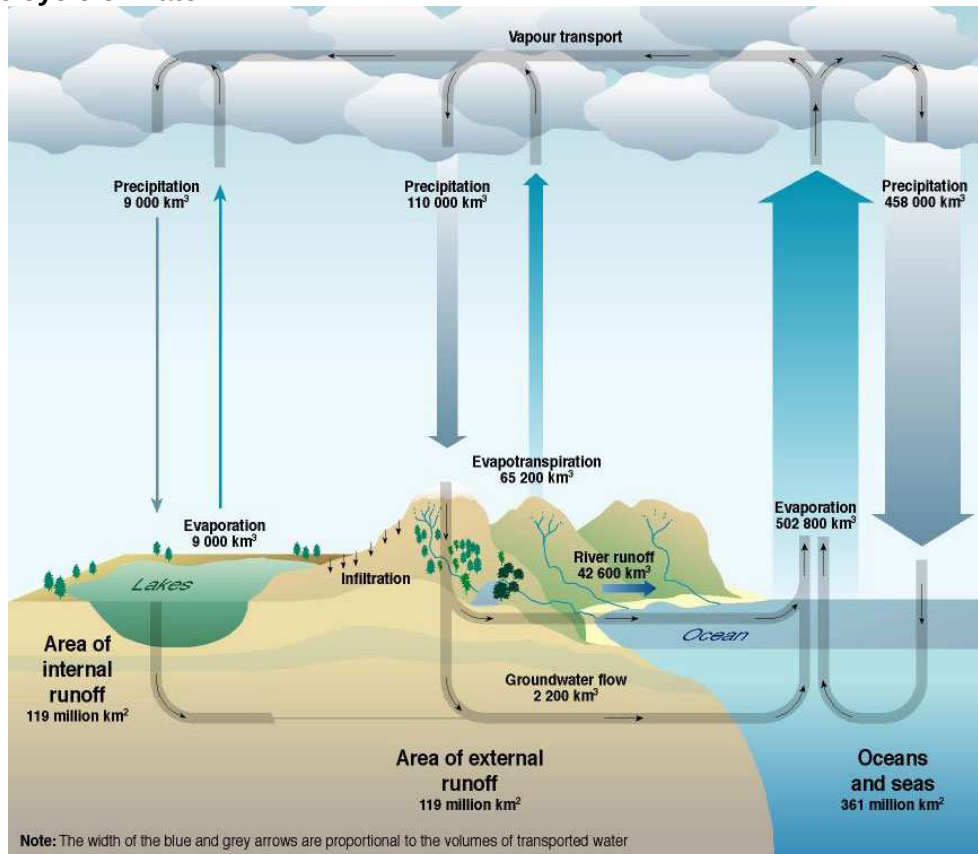
1000 km3	salt	fresh	total	m3/m2	mm
atmosphere		12,9	12,9	0,025	25
sea	1 338 000		1 338 000	2 624	2 624 021
land, from which	12 957	35 004	47 960	94	94 057
snow and ice		24 364	24 364	48	47 782
subterranean	12 870	10 530	23 400	46	45 891
lakes	85,4	91	176,4	0,346	346
soil moisture		16,5	16,5	0,032	32
swamps		2,1	2,1	0,004	4
life	1,1		1,1	0,002	2
total	1 350 957	35 004	1 385 960	2 718	2 718 079

Fig. 352 Total amount of water on Earth(see also **Error! Reference source not found.**)

The amounts of water on the Earth are confined in reservoirs of different size and form. In their order of importance these reservoirs are: oceans, glaciers, groundwater, lakes and rivers, atmosphere and biomass (all living matter man included). In actual fact 97% of all surface water is confined in the oceans and most of the other 3% is fixed in glaciers. So, little water is left over for the other reservoirs.

3.1.2 Evaporation and precipitation

The cycle of water



Igor A. Shiklomanov, State Hydrological Institute (SHI, St. Petersburg) and United Nations Educational, Scientific and Cultural Organisation (UNESCO, Paris), 1999; Max Planck, Institute for Meteorology, Hamburg, 1994; Freeze, Allen, John, Cherry, Groundwater, Prentice-Hall: Englewood Cliffs NJ, 1979.

Fig. 353 The hydrological cycle

Continuously changing the state of water

The sun is the generator or motor of the changes in the state of water. The sun will evaporate water of the oceans and other water reservoirs to the 100% water vapour saturation of the air. The saturation of the air with water vapour is determined by the temperature. The higher the temperature the more vapour the air can contain. The vapour is perceptible by the clouds in the air because of the always present condensation nuclei. The wind will move the clouds from the oceans to the continents and depending the temperature above the continents will happen nothing (temperature $\geq$ temperature in the cloud) or it will rain or snow (in both cases is the temperature $\leq$ temperature in the cloud). Rain, hail and snow is called precipitation.

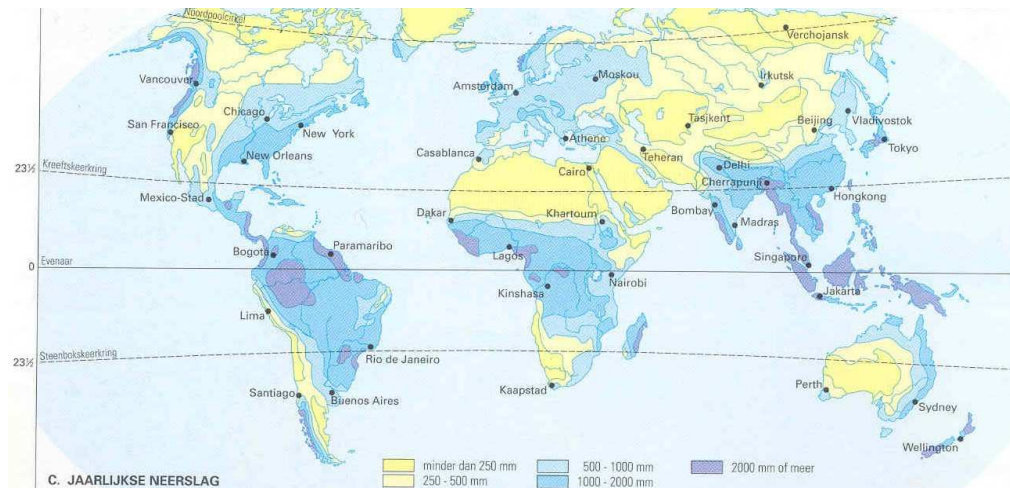
Energy needed for evaporation

You can evaporate 1m³ water by 2.26GJ, 2.26GWs, 630kWh or 72Wa (say 72 m³ natural gas). The Earth's surface receives 81 PW from sun. So the sun could evaporate 1.1 million km³ per year. Actually less than half is evaporated in unsaturated air only (Fig. 354). It falls down discharging its solar heat in the same time as soon as the air becomes saturated in cooler areas by condensation (precipitation). That is nearly 1m³/m² or 1m and more precise 957mm (Fig. 354).

	evaporation	precipitation	runoff	evaporation	precipitation	runoff
	1000 km ³ /a			mm/a		
sea	419	382		1157	1055	
land	69	106	37	467	717	250
total	488	488		957	957	

Fig. 354 Yearly global evaporation, precipitation and runoff

Areas like deserts receive less than 200mm, areas like tropical rain forests more than 2 000mm average per year (Fig. 355).



Wolters-Noordhof (2001) page 181

Fig. 355 Global distribution of precipitation

Europe has the same extremes (Fig. 356).

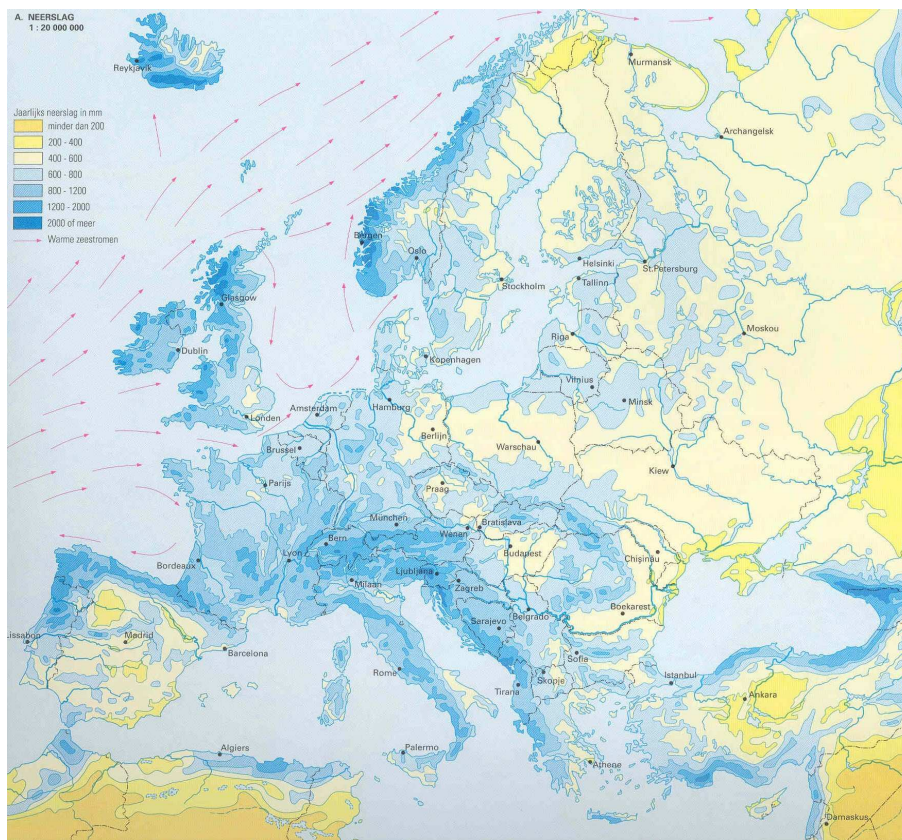


Fig. 356 *European distribution of precipitation* (Wolters-Noordhof (2001) page 61)

3.1.3 Runoff

The Netherlands receives from 700mm in East Brabant to 900mm precipitation in central Veluwe (Fig. 357), but there have been years of 400mm and 1200mm precipitation.

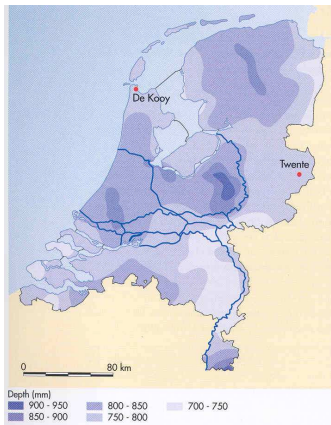


Fig. 357 Distribution of precipitation in the Netherlands (Huisman, Cramer et al., 1998; page 18)

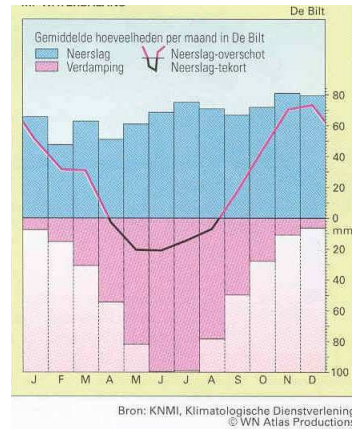


Fig. 358 Precipitation minus evaporation in the Netherlands (Wolters-Noordhof, 2001; page 53)

If precipitation exceeds evaporation lakes and subterranean aquifers fill up. As soon as these cannot be filled up in time, water runs off subterranean or along brooks and rivers (Fig. 359 and Fig. 360).

That part of the precipitation that reaches a stream is called runoff. The water during rainfall will gather into rills and streams down the slope. During and after the rain part of the water will soak into the ground. If the soil is saturated with water the remaining water will stream together in small streams and form a river. The groundwater flows also downhill and where the water bearing layer crops the slope a source will come out. The surface water and the subterranean water feed together a river. When the catchment area is large enough a permanent river will be the result. An estimation is made that $\frac{1}{8}$ of the annual runoff will reach directly overland the sea while the remainder part will go underground.

the Netherlands receive runoff from catchment areas of the Rhine (entering the Netherlands in Lobith), Meuse and Scheldt rivers.

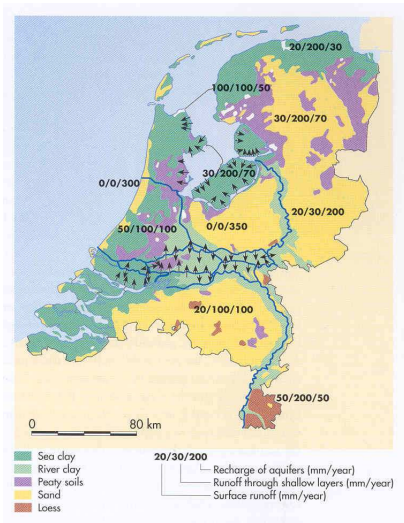


Fig. 359 Major soil types and average annual runoff in the Netherlands (Huisman, Cramer et al., 1998; page 21)

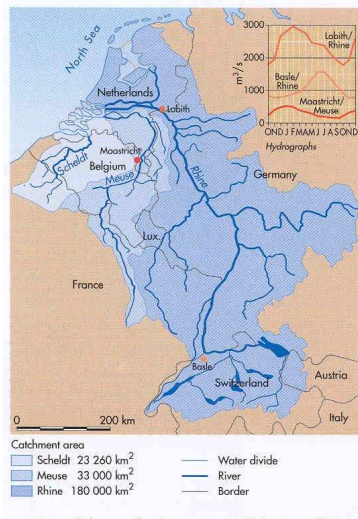


Fig. 360 Received runoff in the Netherlands (Huisman, Cramer et al., 1998; page 13)



Fig. 361 The river basin of the Rhine (Paul Maas, opdrachtgever: Thieme Meulenhoff)

The river Rhine for example

The river Rhine has a catchment area of 160 000km² with an annual average of 1 775mm precipitation minus 1 392mm evaporation in the part of that area as far as Lobith. So, approximately 383mm over an area of 160 000km² produces 61km³/year. So, on average 1942m³/sec of water should run off and enter at Lobith.

Levelling by seasons

Snow and ice in mountains level out seasonal fluctuations of rivers by storing precipitation in winter, releasing it in summer⁸³ (see Fig. 361 and Fig. 363).



Fig. 362 Source of the Rhine
(<http://www.natuurdichtbij.nl/kennismaken/>)

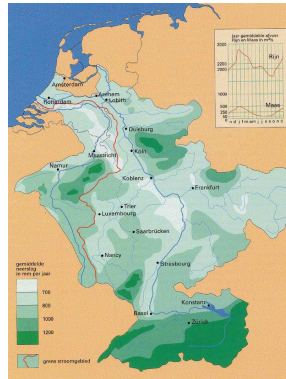


Fig. 363 Precipitation in the basin
(<http://www.natuurdichtbij.nl/kennismaken/>)

Discharge related to catchment area

In Fig. 364 a rough approximation of discharge related to catchment area is shown. A big spot indicates the mentioned values of the river Rhine and a line is drawn for any catchment area producing a discharge in the Rhine circumstances. However, if precipitation is more than the average mentioned the line shifts upward, if evaporation or other reductions are more than mentioned, it shifts downward.

As a rule of thumb the m^3/sec of discharge is $1/100$ of the km^2 catchment area⁸⁴, but any river has its own graph, less regular than suggested here.

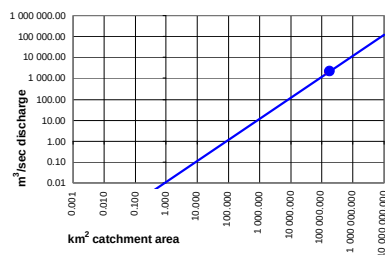


Fig. 364 Discharge Q roughly related to catchment area
(author Jong)

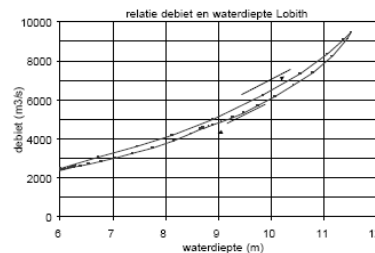


Fig. 365 Discharge Q related to water depth H near Lobith
(<http://www.geog.uu.nl/fg/mkleinhans/teaching/tgrs-hw.pdf#search=%22waterdiepte%20Rijn%22>)

Discharge related to depth

The relation of discharge to the water level near Lobith in Fig. 365 is important for the height of dikes and the draught of ships, but it changes in time because of sedimentation and excavation.

Discharges in time

Because precipitation and evaporation differ much per day, the discharge of the Rhine differs daily (see Fig. 366), as unpredictably as the weather forecast.

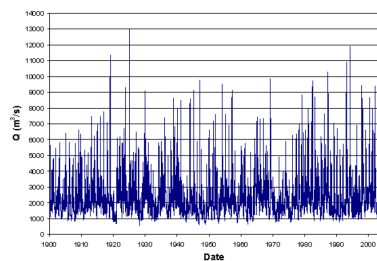


Fig. 366 Daily average discharge of the Rhine at Lobith (Lecture Marc F.P. Bierkens UU Faculty of Geosciences)

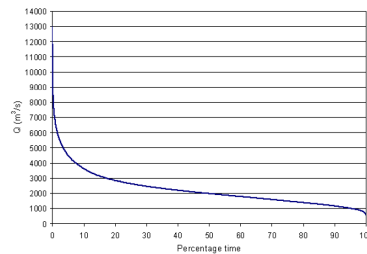


Fig. 367 Duration line of Rhine discharge at Lobith (Lecture Marc F.P. Bierkens UU Faculty of Geosciences)

Ranking Fig. 366 you can derive a 'duration line' as in Fig. 367, indicating how often you can expect a given discharge to be exceeded.⁸⁵ From that figure you can conclude that 50% of the time the discharge of the Rhine did not exceed 2000 m³/sec. The mirrored graph gives the percentages of underspending.

Local impact of rain on discharge

The discharge of a river fed by a catchment area increases some time after the first rainfall (see Fig. 368) and after the last rainfall it continues some time, depending on the size of the area.

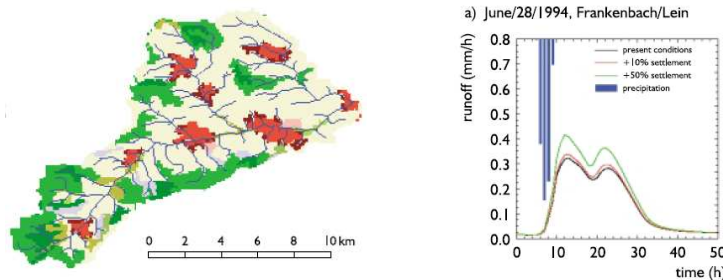


Fig. 368 Local impact of rain in hours $R=10\text{km}$
(<http://www.ncr-web.org/downloads/NCR18nl-2002.pdf>)

Extreme situations

Suppose an unusual system of heavy showers follows the basin around the course of the Rhine and those of its feeding rivers like the Main and Mosel

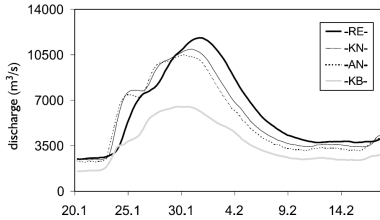


Fig. 369 Flood 1995 (<http://www.ncr-web.org/downloads/NCR18nl-2002.pdf>)

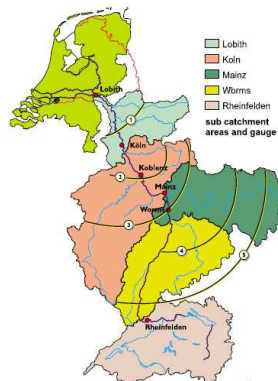


Fig. 370 Impact of rain in days $R=300\text{km}$ (<http://www.ncr-web.org/downloads/NCR18nl-2002.pdf>)

from Switzerland to Lobith and everywhere in the basin drainage is optimal. A wall of water then nears Lobith. How often will that happen, how long will it last? These are the questions to be answered to calculate risks of flooding.

3.1.4 Static balance

Static forces and the potential energy along a slope

The weight W of a bullet on a slope of α degrees can be resolved in factors perpendicular and parallel to the slope (see Fig. 372). The force parallel to the slope equals $W \cdot \sin(\alpha)$.

For example, if $\alpha = 30^\circ$ that force is $\frac{1}{2}W$, because $\sin(30^\circ) = \frac{1}{2}$.

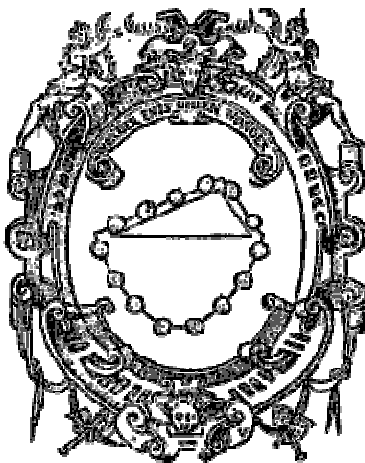


Fig. 371 Stevin: Cloutcrans

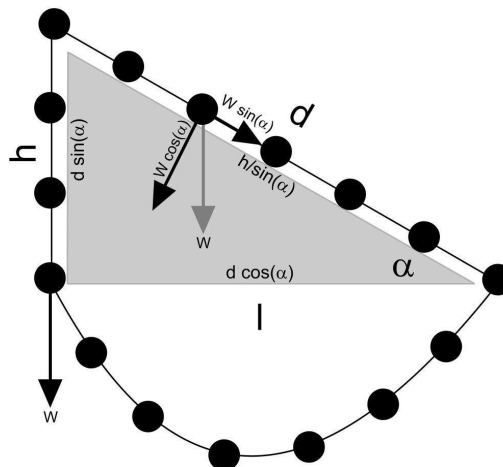


Fig. 372 Balance on different slopes

However, the distance d any bullet has to cover parallel to the slope into the base equals the vertical height *divided* by $\sin(\alpha)$. So, force times distance (potential energy) remains the same at both sides of the summit. For example, if $\alpha = 30^\circ$, the force is $\frac{1}{2}W$, but the distance d to cover is $2 \cdot h$. The 'Cloutcrans' Stevin used as his logo (see Fig. 371) shows the equal potential energy of bullets according to their slope by intuition (count those at the corners in Fig. 372 half).

Potential acceleration

Force is defined as mass times acceleration ($F = m \cdot a$).

At the vertical wall the potential acceleration equals the gravitational acceleration $g = 9.807 \text{ m/sec}^2$.

If the masses of the bullets are the same, but the force F parallel to the slope is reduced by $\sin(\alpha)$ then the acceleration ' a ' parallel to the slope should be reduced by the same factor.

In case $\alpha = 30^\circ$, $a = \frac{1}{2} \cdot g = 4,904 \text{ m/sec}^2$.

3.1.5 Movement ignoring resistance

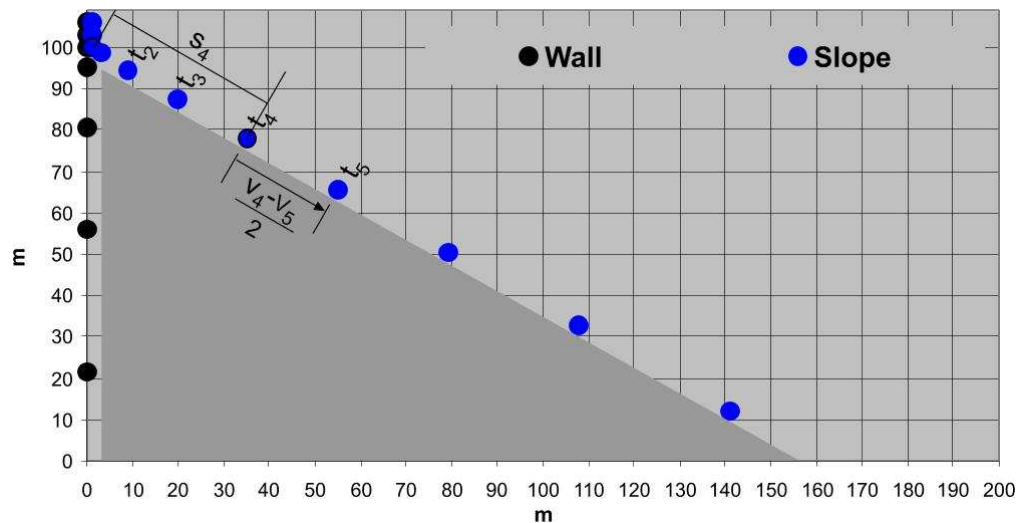
Bullets falling or rolling along a slope

Suppose we disconnect all bullets and supply every second a bullet on the summit at both sides.

Acceleration ' a ' is defined as velocity v divided by time t ($a = v / t$).

As long as there is no resistance the velocity v of any bullet will increase constantly with the time t according to $v = a \cdot t$. But, the covered distance will increase disproportionately, because every next second the bullet has covered a larger distance according to its increased velocity.

So, we can conclude a source distributing an equal amount of bullets per second produces a stream thinning downstream gaining mutual distance by increasing velocity (see Fig. 373).



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Fig. 373 Bullets falling and rolling along a slope every second with a growing distance and velocity

Calculating increasing velocity v and covered distance s along a slope

The growing velocity v and covered distance s shown in Fig. 373 are calculated as follows.

Between any two moments t_p and t_q ($t_p < t_q$) velocity grows from v_p into v_q with a constant acceleration a : $v_q - v_p = a \cdot (t_q - t_p)$. Let the time interval ($t_q - t_p$) near zero. Then $v_q - v_p = a \cdot t$, or $v_q = v_p + a \cdot t$.

At time t half way any t_p and t_q the mean velocity v_m equals $(v_p + v_q)/2$. Here you can substitute v_q .

So, $v_m = (v_p + v_p + a \cdot t)/2$ or $v_m = v_p + \frac{1}{2} \cdot a \cdot t$.

The distance s covered at any moment equals $v_m t$ if you take for v_p the velocity v_0 at the beginning.

So, $s = (v_0 + \frac{1}{2} \cdot a \cdot t) t$ or $s = v_0 \cdot t + \frac{1}{2} \cdot a \cdot t^2$, shortly calculated as a time summing integral of $s/t = v = a \cdot t$:

$$s = \int a \cdot t \, dt = C + \frac{1}{2} \cdot a \cdot t^2$$

Supposed the bullets start in rest ($v_0=0$) and then begin to fall or roll without resistance, then s equals $\frac{1}{2} \cdot a \cdot t^2$ without initial C .

The velocity at the end of the slope is reached at slope length $d = \frac{1}{2} \cdot a \cdot t^2 = \frac{1}{2} \cdot g \cdot \sin(\alpha) \cdot t^2$.

And $d = h/\sin(\alpha)$ (see Fig. 372). So $t^2 = (h/\sin(\alpha))/(\frac{1}{2} \cdot g \cdot \sin(\alpha))$ or $2h/g \cdot \sin(\alpha)^2$.

So, $t_{\text{end}} = \sin(\alpha)^{-1} \cdot \sqrt{(2 \cdot h/g)}$.

At that time $v_{\text{end}} = a \cdot t_{\text{end}} = g \cdot \sin(\alpha) \cdot \sin(\alpha)^{-1} \cdot \sqrt{(2 \cdot h/g)} = \sqrt{(2 \cdot g \cdot h)}$.

So, the velocity at the end of the slope is independent from α : it is the same velocity of a falling bullet at the end of the wall. The average velocity along the slope is half of v_{end} : $v_m := \frac{1}{2} \sqrt{(2 \cdot g \cdot h)}$.

Kinetic energy

If a bullet of mass m [kg] hits you with a velocity of v [m/sec], and you resist its force stepping back slower bringing its velocity back to zero, the bullet has lost $m \cdot v \cdot (v - 0 \text{ m/sec})/2 = \frac{1}{2} m \cdot v^2$ energy.

That kinetic energy E_k could have been built up falling or rolling h [m] with an acceleration a [m/sec²], according to $E_p = F \cdot h = m \cdot a \cdot h$. Falling or rolling, the bullet lost E_p , gaining E_k , while $E_p := E_k$ at last.

So, the process is described as $m \cdot a \cdot h := \frac{1}{2} m \cdot v^2$ [joule].

Running water in a pipe

Suppose running water is a stream of more or less cohesive incompressible drops, flowing downstream in a volume per second of Q [m³/sec] everywhere.

Suppose the bullets of Fig. 373 are cubic metres water forced in a pipe of minimal cross section.

The average velocity will be the velocity at the end of the natural slope $\sqrt{(2 \cdot g \cdot h)}$ divided by two:

$$v_m = \frac{1}{2} \cdot \sqrt{(2 \cdot g \cdot h)}.$$

So, the cross section of a pipe with capacity Q should be at least $A = Q/v_m = 2 \cdot Q/\sqrt{(2 \cdot g \cdot h)}$ [m²].

Its water content is $A \cdot h/\sin(\alpha)$ [m³]. If the mass m [kg] of water relates to its volume [m³] as

ρ (normally 1000 kg/m³) its mass equals $\rho \cdot A \cdot h/\sin(\alpha)$ [kg].

A water ram

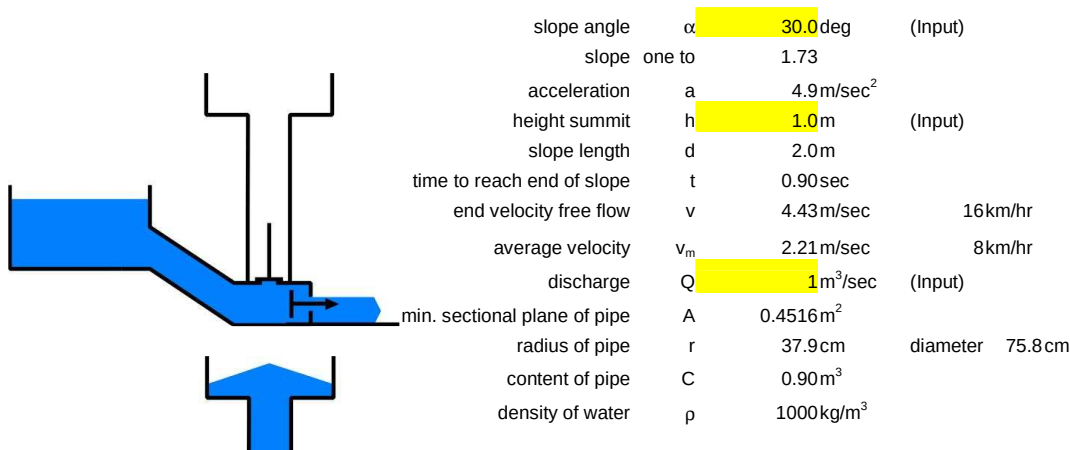
A sudden obstacle at the end of the pipe (like a tap closed at once) shows the large amount of energy built up in flowing water. Such an obstacle has to resist a force F_1 equal to the weight of the water column divided by A [newton/m²] and a force F_2 resulting from kinetic energy $E_k = \frac{1}{2} m \cdot v^2$ divided by some distance s (braking distance) to get the force (energy is force times distance). If that braking distance is very small F_2 increases into infinity, breaking the water pipe.

A water column of height h on a surface A produces a force $F_1 = \rho \cdot h \cdot A \cdot g$ [newton].

A mass $m = \rho \cdot h \cdot A/\sin(\alpha)$ [kg] water with a velocity $v = \frac{1}{2} \sqrt{(2 \cdot g \cdot h)}$ [m/sec] reduced to zero over a distance of s metre (braking distance) produces a force $F_2 = \frac{1}{2} m \cdot v^2/s$:

$$F_2 = \frac{1}{2} \cdot \frac{\rho \cdot h \cdot A}{\sin(\alpha)} \cdot \left(\frac{1}{2} \cdot \sqrt{2 \cdot g \cdot h} \right)^2 \quad \text{or} \quad F_2 = \frac{1}{4} \cdot \rho \cdot A \cdot \frac{h^2}{\sin(\alpha)} \cdot \frac{g}{s}$$

The kinetic force F_2 is many times larger than $F_1 = \rho \cdot h \cdot A \cdot g$ caused by the weight of the water column (the difference is $\frac{1}{4} \cdot h/s \cdot \sin(\alpha)$). In the example of Fig. 374 a kinetic force of flowing water is calculated as 500 times the weight of the water column.



braking distance	s	0.001m	(Input)	
force by weight	F_1	4429 newton	452kgf	
kinetic force	F_2	2213997 newton	225757kgf	226ton
proportion	F_2/F_1	500		
pressure at tap	p	4912447 newton/m ²	500912kgf/m ²	501ton/m ²
m height of rise		501m		

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Fig. 374 Water ram

That force is utilised in a pumping device called 'water ram'. The pressure p built up in the water ram by suddenly closing the tap braking the flow to yield its kinetic force is utilised to push up the water through a valve. Theoretically the water column can be built up until 500 m. However, the pressure falls away shortly after the valve opens, so the procedure has to be repeated often to near that theoretical value.

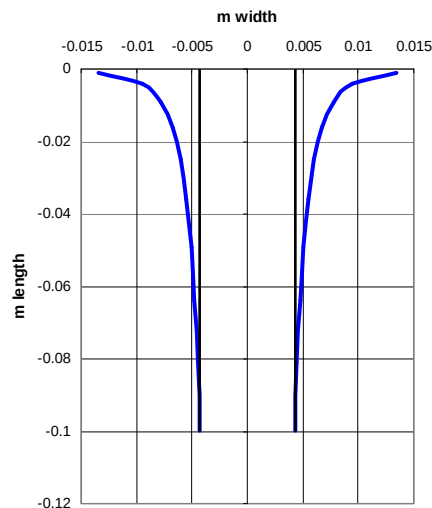
Free flow

The cross section of a free flow $A = Q/v$ will be smaller downstream according to its increasing local velocity $v = \frac{1}{2} \cdot a \cdot t$ (if there are no other sources feeding the stream). You can see that decreasing width already on the tap (see Fig. 375).

Since $s = \frac{1}{2} \cdot a \cdot t^2$ or $t = \sqrt{(2 \cdot s/a)}$ and consequently $v = s/\sqrt{(2 \cdot s/a)} = \sqrt{(s \cdot a/2)}$, the cross section on any distance from the source will be $A = Q/\sqrt{(s \cdot a/2)}$.



Fig. 375 Water flowing from the tap



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Fig. 376 Simulation of 0.00004 m³/sec falling water

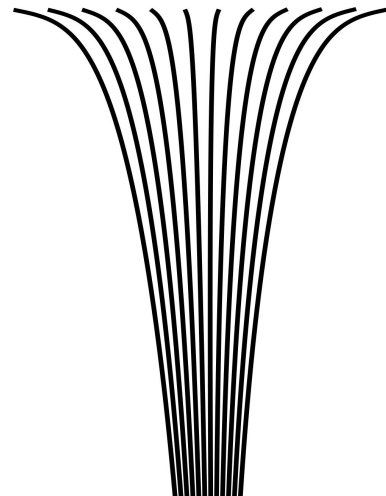


Fig. 377 A river stemming from different separate streams

However, what you see is the diameter, $2 \cdot r$. And $A = \pi \cdot r^2$. So, $2 \cdot r = 2 \cdot \sqrt{(A/\pi)}$.

In Fig. 375, the water from the tap has an initial velocity, perhaps comparable with the 0.02m level of the falling water in Fig. 376. As soon as a critical velocity is passed a continuous flow is falling apart in drops like rain. It shows the limits of water cohesion.

A river

A river, stemming from different separate streams with smaller cross sections (see Fig. 377) will end up flowing faster in the end. Moreover, its resistance reduces because of less contact with its bed, becoming more and more smooth (less rocky) downstream. However, its slope reduces also coming closer to the sea. How do these circumstances balance locally?

3.1.6 Resistance

Until now, we supposed flows, running without resistance.

But, any liquid flowing along a surface encounters a shearing force in the opposite direction dependent on its roughness. That force causes deceleration or even partially flowing back (turbulence).

Force is mass times acceleration. If mass remains the same, the accelerations 'a' of previous paragraph 3.1.5 should be reduced. How much is that reduction in a stream flowing through a landscape?

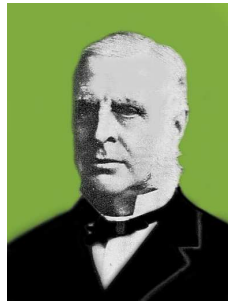
Many parameters play a role, but the result mainly will be that shearing stress reduces the force of water and consequently its acceleration and velocity substantially only if the water level is less than 2m to bottom. However, it always plays an important role in transporting sediments.

So, a river can not adapt its discharge, but rather its form to bring the water most efficiently to the sea. However, that search for the most efficient course may take a very long time, sometimes waiting for a year of extreme rainfall to improve the course, clearing up bottle necks, looking for steeper slopes lessening its tress.

Shearing stress

Manning^a created the formula of *Fig. 378* to calculate the force τ every square metre wetted surface exerts [newton/m²] in opposite direction of the flow ('shearing stress').

^a <http://64.233.183.104/search?q=cache:2qsQymRjhqcJ:manning.sdsu.edu/+Manning+hydrology&hl=nl&gl=nl&ct=clnk&cd=1>



$$\tau := \rho \cdot v^2 \cdot \frac{n^2 \cdot a}{R^3}$$

Fig. 378 Robert Manning and one of his formulas

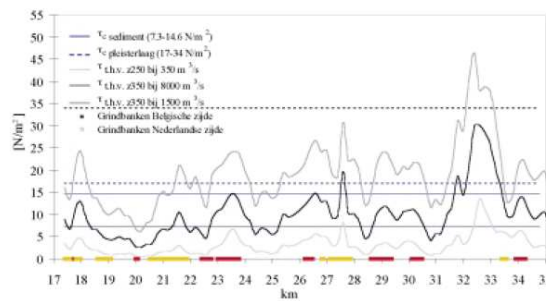


Fig. 379 Shearing stress τ due to different discharge suppositions and local roughnesses and bed forms along 17.5 km of the river Meuse (Grensmaas)^a

Fig. 379 shows τ for different circumstances a part of the river Meuse (Grensmaas) ranging from 1 to 50 newton/m². Fig. 382 shows the studied part in Fig. 379, folded along the boundary of The Netherlands and Belgium within its winter dikes.

The river Meuse for example

Fig. 380 shows a cross section of a river like the river Meuse approximately half way of its 925 kilometres course. Suppose the surface of its cross section $A = 300\text{m}^2$ and its discharge $Q = 600\text{m}^3/\text{sec}$ (often in winter). In that case its water level is 5.7m and it transports a mass $m = 600\,000\text{kg}$ of water per second over 2 metre (so, velocity $v = 2\text{m}/\text{sec}$ or $7.2\text{km}/\text{hr}$). That represents $E_k = 1.2$ million joule kinetic energy over 2 m, and a force $F_2 = 600\,000$ newton equivalent to a weight of approximately 60 tons.

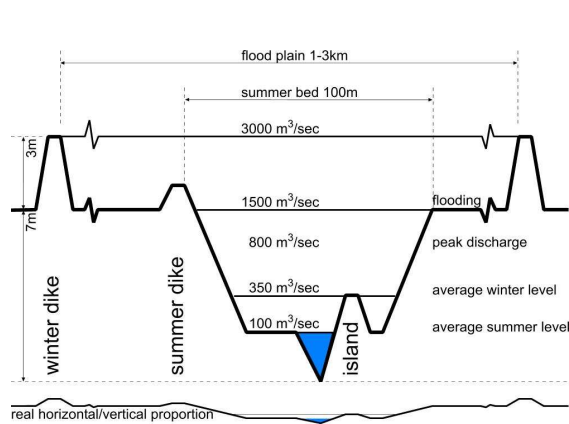


Fig. 380 Typical cross section and wetted surface of a river like the Meuse half way (Grensmaas)^b

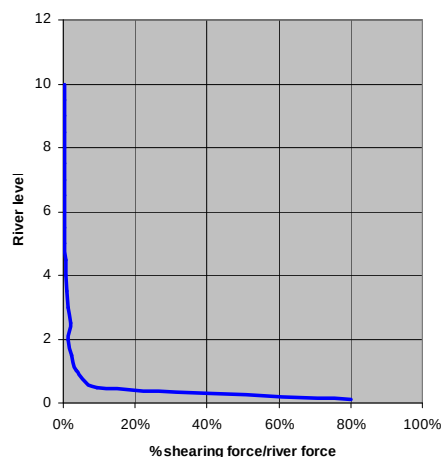


Fig. 381 The influence of shearing force by different water levels in Fig. 380^c

According to Mannings formula in this circumstances the shearing stress τ would be 10 newton/m². If it shears over the river bed taking 100 m in the cross section ('wetted contour'), then perpendicular to that cross section in 1sec over a length of 2m the river has to overcome 2 000 newton resistance. That is only 0.3% of the local force of the river. And in this two metres we did not even count the pushing power of many kilometres moving water coming down upstream.

^a http://viwc.lin.vlaanderen.be/water/ts2003_09_grensmaas.pdf

^b http://viwc.lin.vlaanderen.be/water/ts2003_09_grensmaas.pdf

^c <http://team.bk.tudelft.nl/> > publications 2006 Hydrodynamics.xls

Low shearing stress

So, the influence of the shearing stress τ on velocity and acceleration on a water level of 5.7m is negligible, but in many centuries it has given the actual form to the river by loosening material from the bed. Water with a velocity of 2m/sec could even move stones of 0.5kg, but at the bottom a shearing stress of 10newton/m² will only move some smaller sediment.

High shearing stress

However, at water levels in the same circumstances lower than 2m, τ becomes more than 1%, increasing into 80% on very low water levels (see *Fig. 381*). You can calculate it yourself for different circumstances downloading <http://team.bk.tudelft.nl/> > publications 2006 Hydrodynamics.xls.

So, in small brooks τ will play an important role on the resultant force, acceleration, velocity and kinetic energy.

Kinetic energy per m³ water $\frac{1}{2} \rho \cdot v^2$

In Mannings formula ρ is the mass of 1m³ water (mainly 1000 kg/m³). The kinetic energy reduced by roughness like earlier shown by the water ram (see page 176) is $\frac{1}{2} m \cdot v^2$ (see page 176).

So, $\rho \cdot v^2$ in the formula represents twice the kinetic energy per m³ water.

You can measure the velocity v [m/sec] on different spots in the cross section to calculate the average velocity (see *Fig. 423*).

Kinetic energy [newton·m] per m³ is the same as force per m² like τ [newton/m²].

So, the rest of the formula is a dimensionless factor, but how to calculate it?

Roughness n

The roughness of river beds is expressed in a roughness factor n [sec/m^{1/3}] shown *Fig. 432*, ranging from 0.01 for very smooth concrete until 0.1 sec/m^{1/3} for flooded tight forest.^a

Hydrolic radius R

R [m] in Mannings formula is the 'hydrolic radius', the wet surface 'A' of the cross section divided by the length of its wetted contour 'P' ($R = A/P$). The larger 'A' is (for example increasing by a larger discharge (see *Fig. 380*) the less influence the wetted contour has.

The surface/contour proportion is an important factor in many physical phenomena like roads around an urban island (public investment), volume/surface of buildings or growing animals (insolation). If a volume increases by a third power of distance, a minimal surface containing that volume increases quadratically (slower), while the minimum contour (a circle) containing a surface increases in the same time linear (again slower).

A 'wetted contour' of a river is not a circle, but it increases slower than the contained cross sectional surface also because the horizontal upper surface is ignored.

Fall and acceleration a

Most difficult to estimate is local 'a' in Mannings formula. The total acceleration of a river can be calculated according to page 175 and reduced by varying shearing stress, but that average is locally changed by varying slopes and forced by water masses upstream into increased acceleration in narrow cross sections, partly compensated by higher water levels storing potential energy for accelerations later.

Reduction of acceleration

The part of the river Meuse studied, falls 10m (from 40 to 30 above sea level) over 17.5km length with varying resistance (see *Fig. 383*). However, the total fall of the river Meuse from source to sea is 409m over 925km. That is the tangent of $\alpha = 0.0253$ degree. So, you could expect an average acceleration of $a = g \cdot \sin(\alpha) = 9.807 \cdot \sin(0.0253) = 0.004$ m/sec², partly reduced by a substantial τ in the many feeding brooks at the boundary of the basin (see *Fig. 385*).

^a <http://www.fhwa.dot.gov/bridge/wsp2339.pdf>



Fig. 382 17.5km of Meuse (Grensmaas)^a

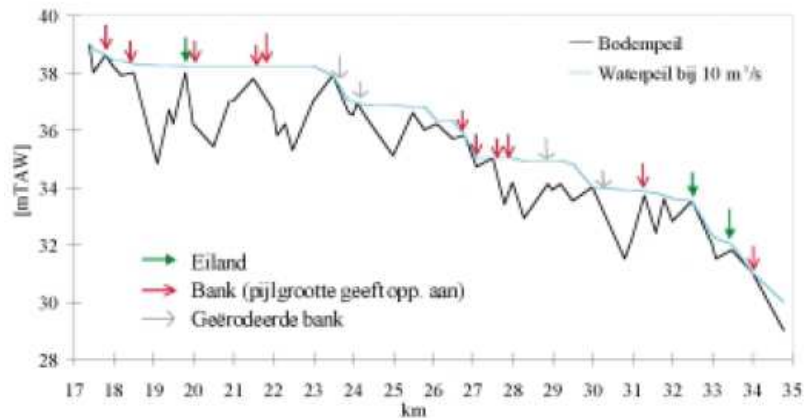


Fig. 383 A fall of 10m along 17.5 km of the river Meuse (Grensmaas)^b

Because $v = a \cdot t$ and consequently $t = v/a$, the distance covered $s = a \cdot t^2/2 = a \cdot (v/a)^2/2$. So, at distance $s = 500\text{km}$ from source the velocity should be $v = \sqrt{(2 \cdot a \cdot s)} = 66\text{m/sec}$. However, we counted $v = 2\text{m/sec}$, to reach $Q = 600\text{m}^3/\text{sec}$ through a cross section (wetted surface) of 300m^2 . So the reduction by τ in all upstream shallow brooks and small rivers of the basin together should be 97%!

Discharge

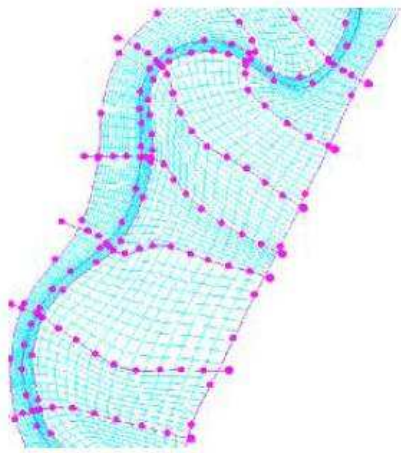


Fig. 384 Successive SOBEK cross-section trajets along the Grensmaas



Fig. 385 Meuse river basin of 36 000 km² through France, Belgium, Germany and The Netherlands

If you measure the cross section 'A' [m²] of a stream and the velocity v [m/sec], the discharge $Q = A \cdot v$. However, you also can measure the rainfall of the Meuse river basin in Belgium and France ($23\,500\text{km}^2$)^c. If in that area at average in a year 1000mm rain has fallen of which 200mm is evaporated or temporarily sunken down into the earth, then $Q = 800\text{mm} \cdot 23500\text{km}^2/\text{yr}$. That is 600 m³/sec of water coming into the Netherlands at the boundary of Belgium averaged over a year (see Fig. 385).

^a http://viwc.lin.vlaanderen.be/water/ts2003_09_grensmaas.pdf

^b The Belgian standard TAW in Fig. 383 means 'above average sea level at ebb-tide on Ostende, 2.426m higher than NAP, the Dutch standard for measuring heights.

^c http://nl.wikipedia.org/wiki/Stroomgebied_van_de_Maas

However, in a concurrence of circumstances like in January 1995, there can be more rainfall (up to 350mm *per day*), less evaporation, no storage in a saturated earth, faster discharge because that earth is frozen, but starting to melt, delivering previously fallen water in the same time. In such a case you can expect floodings.

Velocity and discharge

A river has its largest velocity on its surface, decreasing into the bottom.

The average velocity v is often measured at 0.4-h (see Fig. 386). However, the velocity distribution over the cross section varies substantially (see Fig. 387).

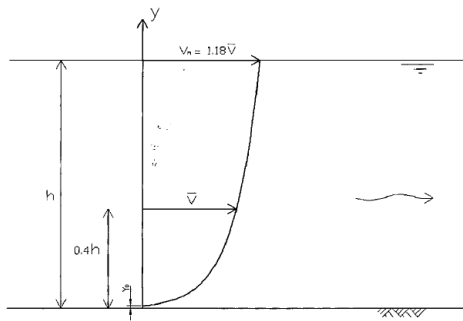


Fig. 386 Velocity in a longitudinal cross section^a

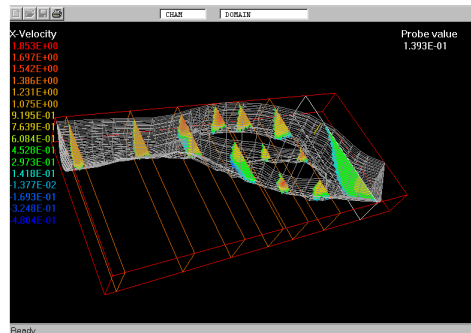


Fig. 387 River cross sections with simulation of velocity profiles^b

Many rivers have a relation $v = k \cdot Q^m$, but 'k' and 'm' differ from river to river (in Fig. 388 Bovenrijn and Waal obey approximately to $v = Q^{0.3}$).

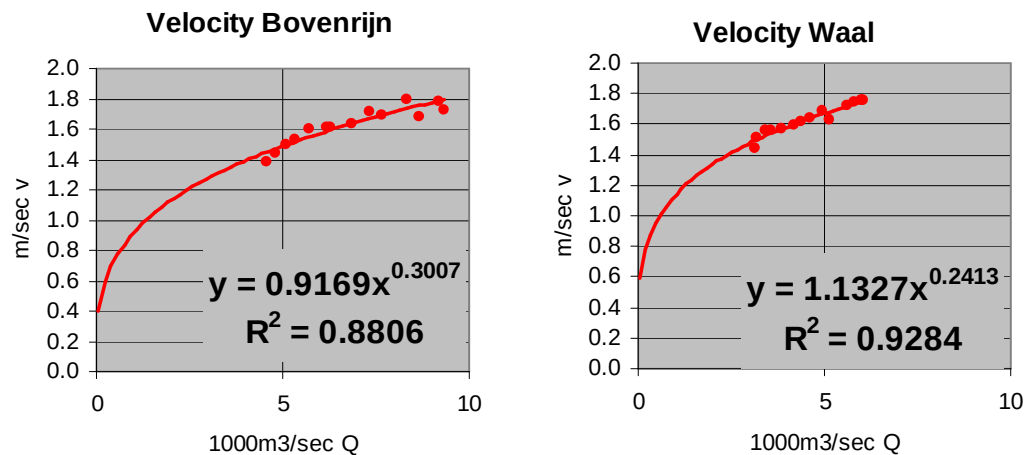


Fig. 388 Flood and velocity in Bovenrijn and Waal^c

Different sensibility of velocity for discharge

Fig. 389 shows that relation for two extremely different American rivers. In a logarithmic representation the measurements fit very well a straight line. An increasing factor 'k' shifts the whole line up, an increasing exponent 'm' makes it steeper (v more sensitive for Q). If the line is horizontal ($m = 0$), there

^a Huub Savenije (2001) Stroming 7 Nummer 4 TU Delft, hsa@ihe.nl

^b <http://www.simuserve.com/cfd-shop/uslib/vrgeom/vrg4.htm>

^c Derived from http://www.engr.colostate.edu/~pierre/ce_old/Projects/Paperspdf/Julien-Klaassenet%20a%20ASCE2002.pdf#search=%22river%20Rhine%20cross%20sections%20Lobith%22

is no relation between v and Q whatsoever. Even if the discharge increases, the velocity will not. These are stoic rivers having other possibilities to give space to their discharge, for example in the lowlands. The steep liners are nervous ones, apparently limited in their cross sections in the highlands.

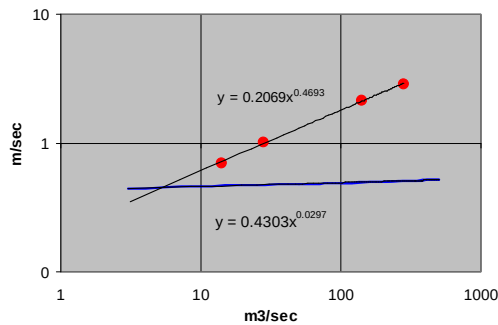


Fig. 389 Different relations between velocity and discharge^a

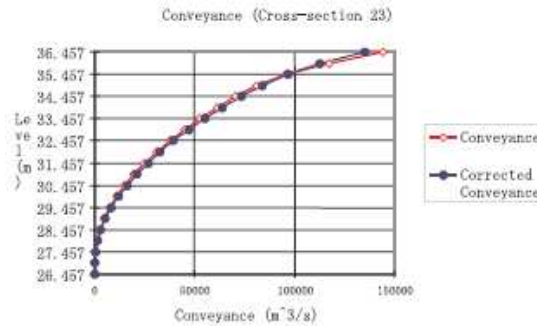


Fig. 390 SOBEK simulation of level related to discharge processed with correction for spatial variations between successive crosssections of Fig. 384^b

Depth related to discharge

Many rivers have a relation depth $D = c \cdot Q^f$, but 'c' and 'f' differ from river to river (in Fig. 391 Bovenrijn and Waal obey approximately to $D = 5 \cdot Q^{0.4}$).

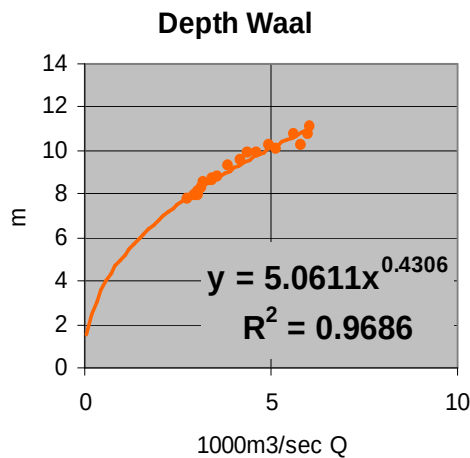
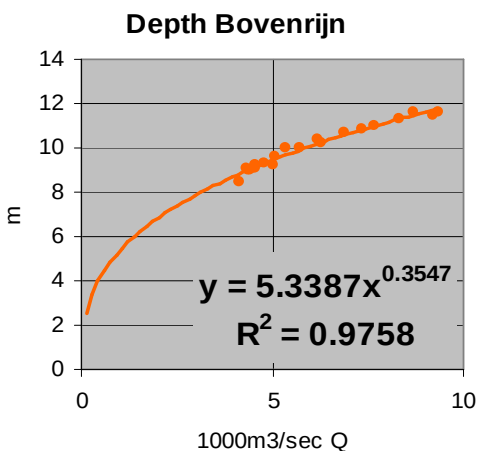


Fig. 391 Depth related to discharge in Bovenrijn and Waal^c

^a Leopold

[http://eps.berkeley.edu/people/lunaleopold/\(043\)%20Downstream%20Change%20of%20Velocity%20in%20Rivers.pdf#search=%22velocity%20rivers%22](http://eps.berkeley.edu/people/lunaleopold/(043)%20Downstream%20Change%20of%20Velocity%20in%20Rivers.pdf#search=%22velocity%20rivers%22)

^b <http://www.wildelft.nl/rnd/pdf/rnd2001.pdf#search=%22river%20Rhine%20cross%20sections%20Lobith%22>

^c http://www.engr.colostate.edu/~pierre/ce_old/Projects/Paperspdf/Julien-Klaassenet%20a%20ASCE2002.pdf#search=%22river%20Rhine%20cross%20sections%20Lobith%22

Width related to discharge

Many rivers can be simulated by an ellipsoid cross section (see).

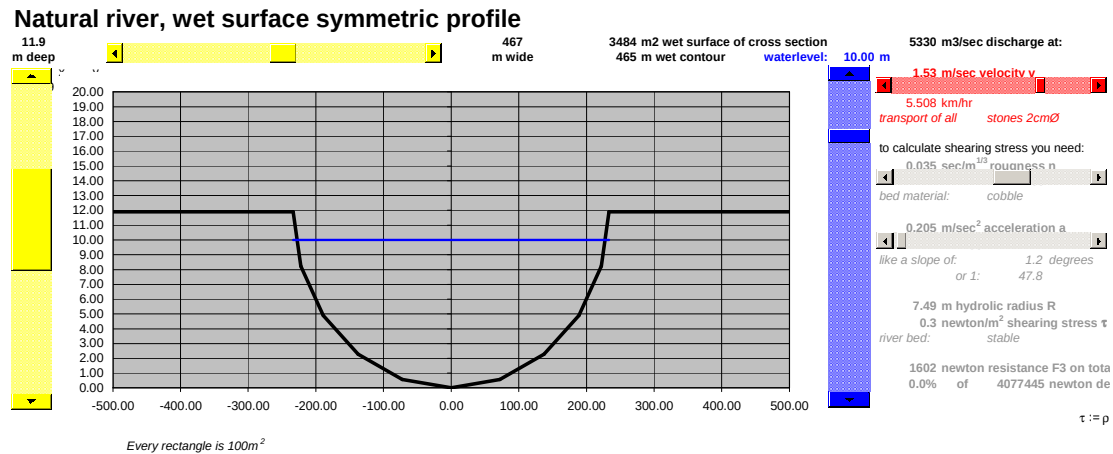


Fig. 392 Simulation of an ellipsoid cross section^a

For a river in weak soil (not forced by artificial measures) it is easier to find space in width than in depth, because sedimentation reduces depth.

3.1.7 Erosion and sedimentation

Material from the river bed (silk, sand and gravel) is transported dependent from the velocity of water.

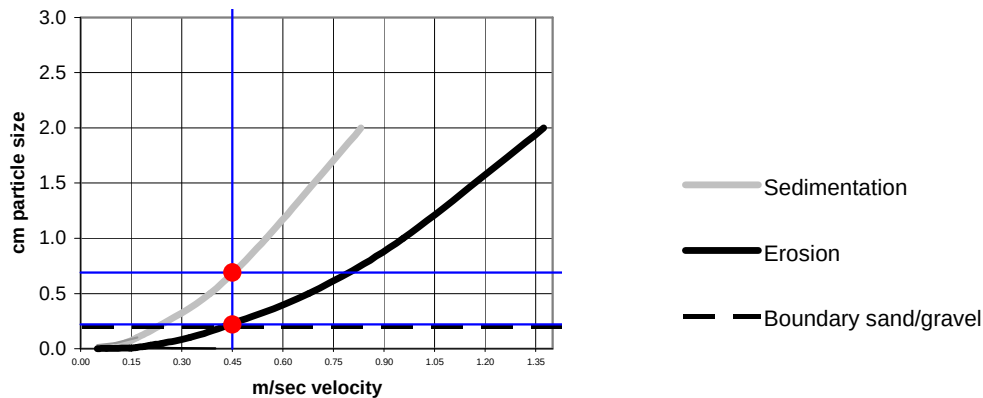


Fig. 393 Erosion and sedimentation dependent from the velocity of water^b

From >0.2 cm particle size we call it gravel (see Fig. 393).

Until <0.2 cm it is named sand or silk (see Fig. 394, an enlargement of Fig. 393).

^a <http://team.bk.tudelft.nl/> > Publications 2006 > Hydrodynamics .xls

^b redrawn according to Pannekoek () Algemene geologie () pag. 225

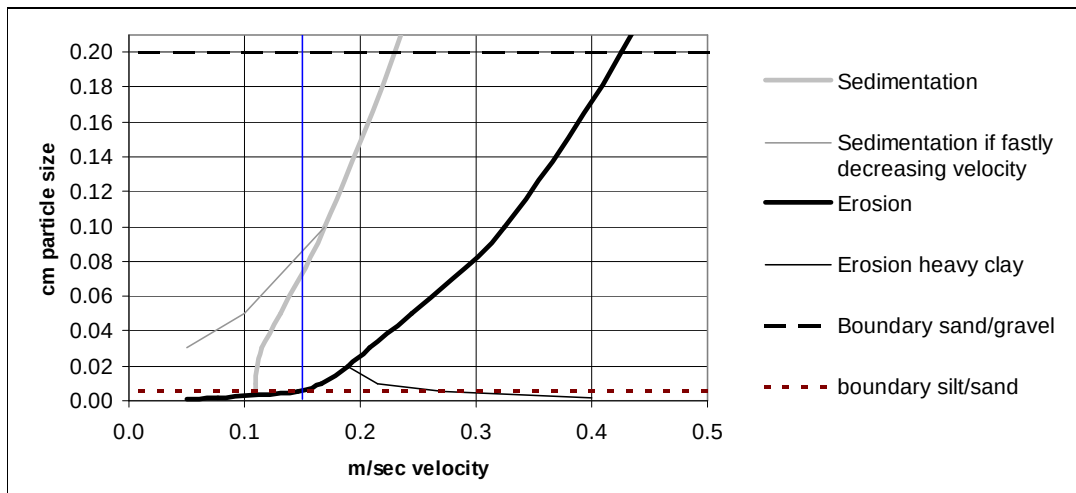


Fig. 394 Erosion and sedimentation at the boundary of silt < 0.005 cm and sand > 0.005 cm dependent from velocity of water, detail from Fig. 393^a

In Fig. 394 until **0.05m/sec** you can conclude that the river bed is stable. Or the reverse: if you see a stable bed, the velocity should be less than 0.05m/sec.

Silt

From a velocity **0.15m/sec** all loose silt is moving. So, if you see silt on the bottom, the velocity of the water should be usually less than 0.15/sec. If you do not see silt, it should usually be more. However, heavy clay densified into a cohesive plaster layer needs a higher velocity to erode than you would expect from their particle size.

Sand

From **0.45m/sec** (ample 1.5km/hr, slowly walking) onwards all sand is moved. So, if you do not see sand, the velocity will be probably more than 0.45m/sec.

Gravel and stones

At higher velocities you have to look at gravel and stones in to estimate the water velocity (see Fig. 393). From **1m/sec** (3.6km/hr) you see stones of 1cm diameter rolling, from **1.45m/sec** (5km/hr) stones of 2cm, from **1.7m/sec** (6km/hr) stones of 3cm, from **1.95m/sec** (7km/hr) stones of 4cm. On that level the diameter of stones moved grows approximately parabolically with the square of velocity. So, stone diameter $\approx v^2$ like $1 \approx 1^2$, $2 \approx 1.45^2$, $3 \approx 1.7^2$ and $4 \approx 1.95^2$. That seems logical, because according to page 176 the kinetic energy of running water ($\frac{1}{2} m \cdot v^2$) is proportional to the square of velocity.

Higher velocities widen passages, lower velocities narrow them.

At the long term wider passages of a river with lower velocities will be filled up with sedimentation and narrow passages with high velocities will be widened by erosion or floodings. So, by an equal discharge Q in older natural rivers the velocity v is equalised as well. However we have artificially narrowed our rivers to save land and to make them deeper for ships.

^a redrawn according to Pannekoek () Algemene geologie () pag. 225

3.1.8 Hydraulic geometry of stream channels

Width (w), depth (d) and stream velocity (v)

The study of the changes of channel width (w) and depth (d), stream velocity (v) and suspended load with a discharge $Q = w \cdot d \cdot v$ is the next step for a better understanding of the behaviour in a landscape. Channel width, depth and current velocity increase during rising water. This is no surprise to anyone familiar with the regime of rivers, but the regular change of each separately is amazing.

With the help of a wide range of streaming conditions it was found experimentally (Leopold and Maddock, 1953) that width, depth, velocity and load increase as simple power functions of discharge. This can be translated in the following equations:

$$w = aQ^b \quad d = cQ^f \quad v = kQ^m \text{ (see page 182)}$$

The numerical values of the arithmetic constants a, c and k are not significant for the hydraulic geometry of streams. On the other hand the numerical values p, q and r are very important. All these values are found by measurements. Leopold and Maddock found that the average for some 20 more or less comparable stations in the United States gave the following values:

$$b = 0.26 \quad f = 0.40 \quad m = 0.34$$

In these cases during a flood the *width* of a channel at a specific cross-section will increase slowest ($w = aQ^{0.26}$), the *depth* (level) fastest ($d = cQ^{0.4}$) and the *velocity* in between ($v = kQ^{0.34}$).

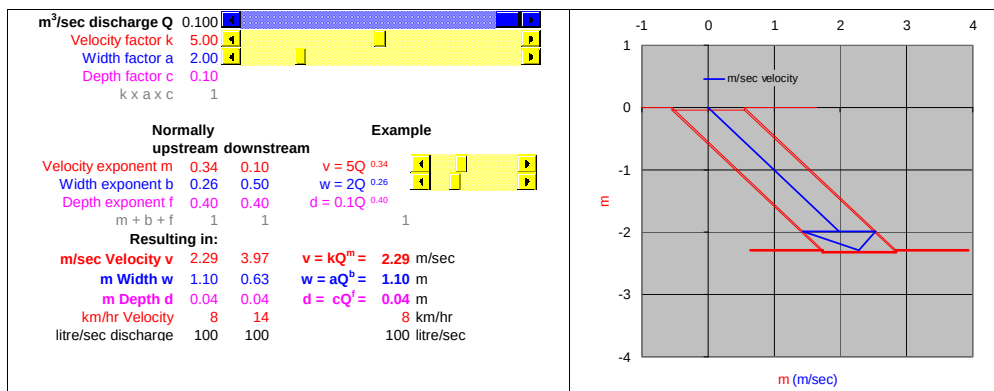


Fig. 395 Calculating v, w and d

Fig. 396 The geometry of the stream^a

Comparing measurements of channel shape and stream velocity in a downstream direction gives surprising results. Normally the discharge of a river downstream increases. The same equations are found to apply at the different downstream cross-sections. Research and measurements proved that:

Width, depth and velocity increase downstream.

According to Fig. 373 this empirical results also reject the idea that streams in the mountains flow wildly and more rapidly than downstream.⁸⁶ These higher streams are characterized by a flow in circular eddies with almost as much backward as forward motion.

The numerical value of the exponents b and m from the equations above are not the same for changes downstream as for changes with discharge passing an upstream cross-section.

In the downstream direction the average values for the exponents become:

$$b = 0.5 \quad f = 0.4 \quad m = 0.1$$

^a <http://team.bk.tudelft.nl/> > Publications 2006 > experiments: Hydrology .xls

Downstream, the *width* of the channel will increase most rapidly, the *depth* a little bit less rapidly, but the mean *velocity* will increase only slightly.

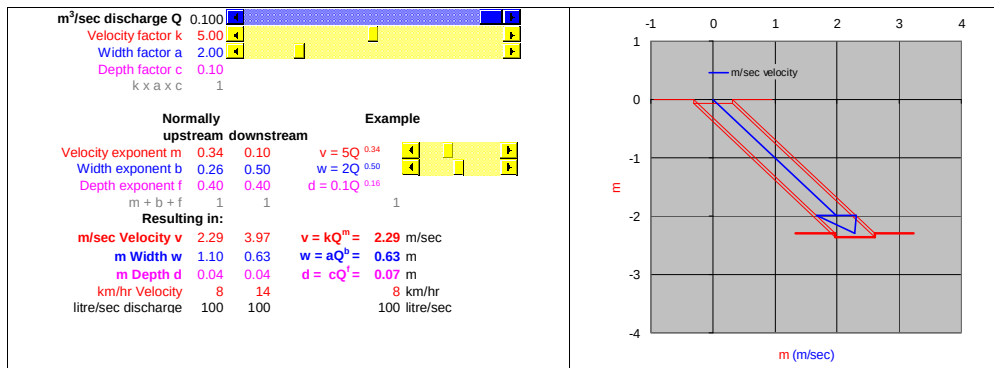


Fig. 397 Calculating v, w and d

Fig. 398 The geometry of the stream^a

It is believed that the increasing depth downwards permits a more efficient flow in a river and so overcompensates the decreasing slope. As a result a slight net increase in velocity at mean annual discharge will take place.

Further mathematical calculations of the hydraulic geometry equations suggests useful applications of the principles.

The discharge is defined as $Q = wdv$

and if $w = aQ^b$ $d = cQ^f$ $v = kQ^m$

then by substitution: $Q = (aQ^b)(cQ^f)(kQ^m)$

or: $Q = ackQ^{b+f+m}$

it follows that: $a \times c \times k = 1.0$ and $b + f + m = 1.0$

As is stated above the arithmetic constants a, c and k are not important. But it is interesting that for all the made measurements and calculations for the different cross-section $b + f + m = 1.0$ agree.

3.1.9 River morphology

The morphology of a river system depends mainly on climate, gravity, height, slope, bedrock, soil type and vegetation. Human impact on the system cannot be neglected and especially not downstreams with all artificial interventions varying from storage reservoirs both for the generation of electricity and for storage purposes of water and for alterations in the system itself and dumping of materials in the system.

^a <http://team.bk.tudelft.nl/> > Publications 2006 > experiments: Hydrology .xls

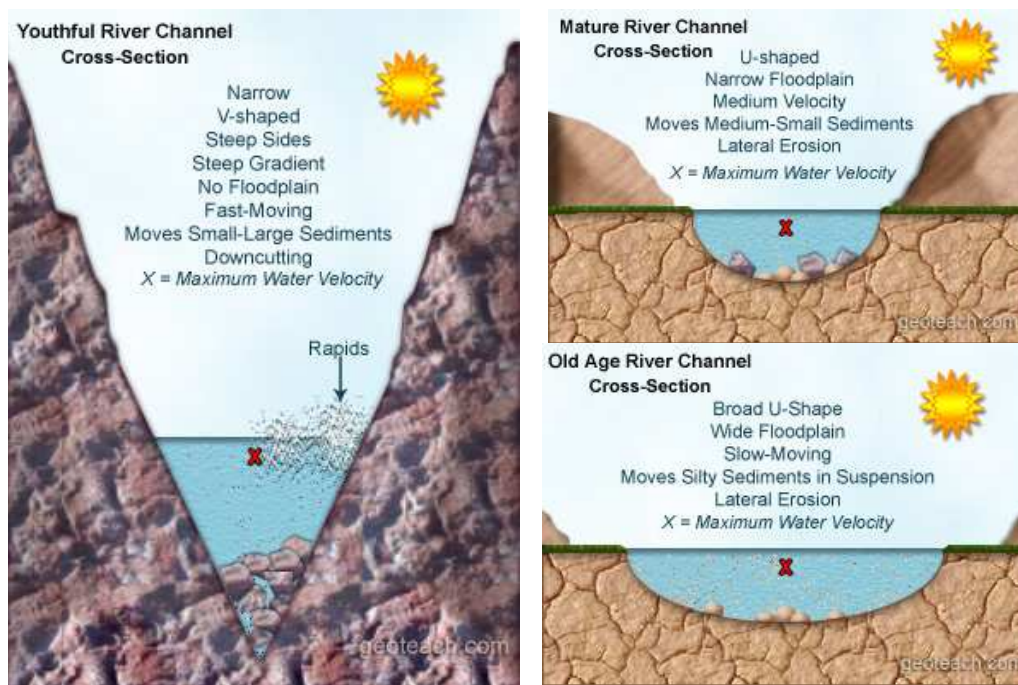


Fig. 399 Development of river beds^a

^a http://geolor.com/geoteach/rivers/Three_Stages_of_River_Development-geoteach.htm

3.1.10 Simulating a simple drainage system

Wind, water and traffic flow along the earth's surface. Some of these flows collect into streams.

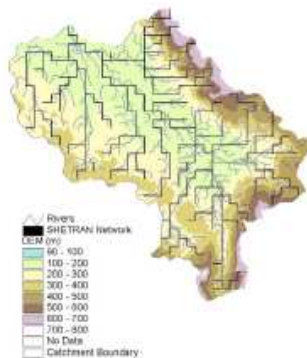


Fig. 400 Schematic of SHETRAN model setup.

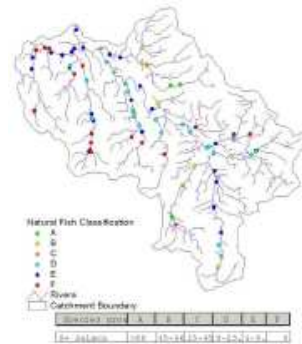


Fig. 401 Salmon Abundance across the Eden catchment^a

Fig. 402 shows a landscape with 24 x 24 squares (sloped mountain areas or a polder with outlets) with 4 possible drainage directions, producing a converging feather or tree like drainage system. Computer programme Jong (2003) 'river(drainage.exe)' (see [http://team.bk.tudelft.nl/publications 2003](http://team.bk.tudelft.nl/publications/2003)), made from the 'random walk' example of Leopold and Wolman cited by Zonneveld (1981), arouses such random landscapes producing drainage systems. The image is built up in columns from upper left to down below. The programme prevents convergent arrows and smallest circuits by changing lowest arrow 90° into right or downward if they occur. So, the runoff tends towards 'South East' as if the landscape has a main slope or a main drainage outlet. Watersheds become visible separating catchment areas. Why do they concentrate into separate basins and converge into main streams?

^a <http://www.ncl.ac.uk/swurve/downloads/2002Synthesis.pdf#search=%22river%20Rhine%20cross%20sections%20Lobith%22>

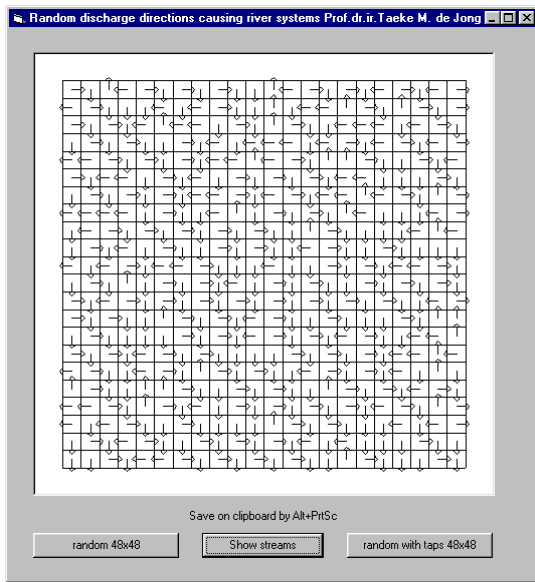
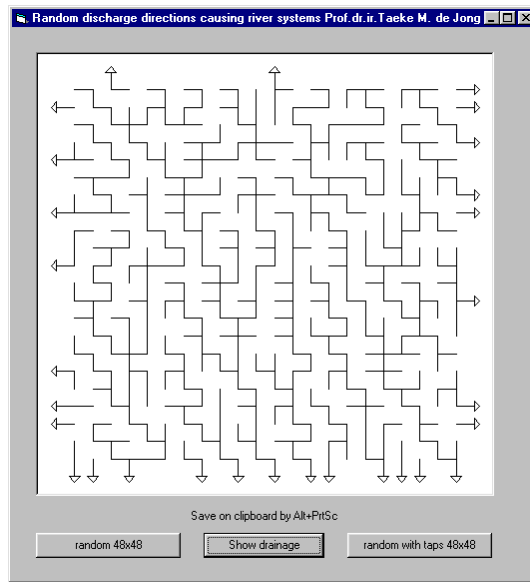


Fig. 402 Directions of drainage in a landscape

Fig. 403 Surface streams caused by Fig. 402^a

Getting a feeling for runoff calculations

Run the program or take Fig. 403, draw the catchment basin of an outlet and calculate the discharge Q for one hour taking arbitrary European precipitation and evaporation values into account. Neglect subterranean flows, width and depth of streams, obstacles or retardations. Suppose surfaces and altitudes, draw the altitude lines and estimate velocities. An exercise like that makes you understand the problems elaborated in next sections.

Truncation orders in river systems

You can divide a river system in different truncation orders from source to the mouth of the system. Fig. 404 shows four methods. All the ordering systems are more or less based on a method starting with the source and going downstream. The first order is called a source river without any tributaries and so on. The differences are more determined by the nomination of the different tributaries than by the differences in system. Strahler (above right) considers small source brooks without tributaries above as first order. Streams collecting water from first order rivers are second order rivers and so on. Try to divide Fig. 403 in such an order system^b.

^a Zonneveld (1981)

^b Mail pattern and calculation to T.M.deJong@bk.tudelft.nl

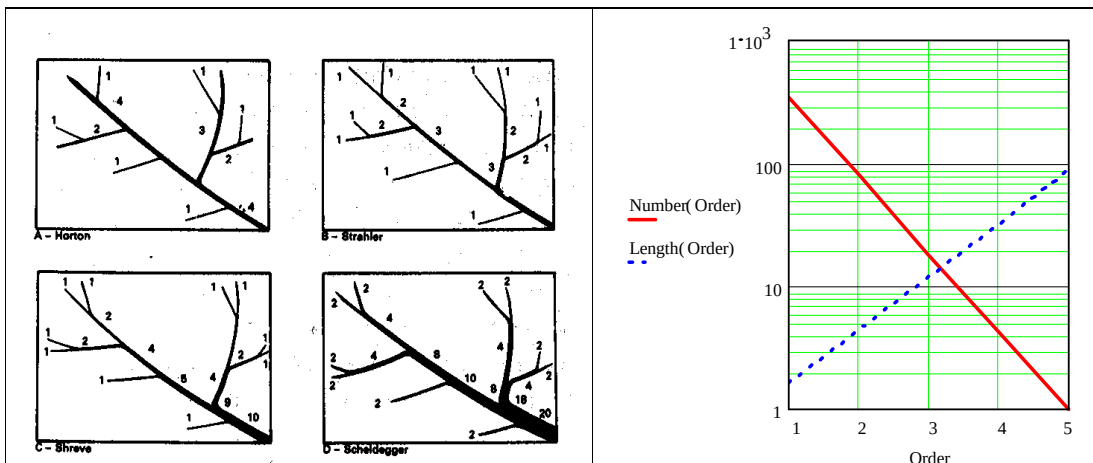


Fig. 404 Four methods to distinguish 'orders' in a feather like drainage pattern^a

Fig. 405 Average number and length of orders in 'random walk rivers'^b

Leopold and Wolman calculated that random rivers have 4.4 upstream branchings of lower Strahler orders according to Strahlers method at average. In practice it varies between 2 and 5. The longer a river is the more orders can be distinguished.

3.1.11 Bifurcation or trunking in traffic networks

This 'bifurcation ratio' plays a rôle in traffic as well, though street patterns and artificial drainage systems in flat lands are not like a tree but like a grid (compare Alexander (1966))⁸⁷. If there are 20km streets per km², you can raise some 7km of them into the order of neighbourhood roads with a higher capacity and transform 2km into district ways with an even higher capacity. So, the optimal proportion between the density of ways and sideways in a grid seems to be approximately a factor three. Do not take it for granted, it is an easy rule of thumb, based on calculations of Nes and Zijpp (2000) indicating factors 2 and 4 are suboptimal in three different types of calculation.

Density of roads and orders of roads

Suppose a metropolis of 30km radius has 60 x 60 = 3600km² surface with 2km/km² district roads (see Fig. 498). There should be 7200km district roads in a grid of average 1x1km. To calculate density from the grid mesh bordered by 4km district roads, you have to count them half because they serve adjacent meshes as well. Many of them would be overloaded by through traffic when you would not raise 1/3 of them into city highways (2400km in a grid of 3x3km, 0.67km/km²) with a capacity of 3000 mv/h and less exits. However, on their turn they would be overloaded. So, this argument produces a semi logarithmic range of orders (Fig. 406).

	km nominal mesh	km/metropolis	km/km ² inclusive	density exclusive	mv/h
district roads	1	72000	2,00	1,33	1000
city highways	3	24000	0,67	0,47	3000
local highways	10	7200	0,20	0,13	10000
regional highways	30	2400	0,07	0,05	30000
national highways	100	720	0,02	0,02	100000
and so on			nearly 3.00	2.00total	

Fig. 406 Theoretical orders of urban traffic infrastructure

The total density of ways is 2km/km². One third of them we have transformed into highways of several orders. So, the density of ways includes the highways. Excluding highways, there are 1.33km/km²

^a Zonneveld (1981) page 179

^b After Zonneveld (1981) page 183

small district ways left. If we would like to reduce the amount of exits of local highways to save velocity, we have to disconnect district ways into dead ends. If we like to connect them mutually with extra parallel service roads along side the city highway we need the inclusive density at least.

If we try to draw a system of highways in a square of 60x60km we firstly draw a grid of 10x10km. There are 14 local highways of 60km, but 6 of them we transform into a higher order. So, their exclusive density is $8 \times 60 / 3600 = 0.13$ indeed. However, we can not fill 10km space between local highways with 3.3 city highways. So we choose 3 highways lowering the inclusive density from 0.67 into 0.60km/km². This causes a raise of exclusive district way density from 1.33 into 1.40, but on this scale we can not draw them anyhow.

Comparing truncing in rivers and roads

For wet connections the same applies when we call city highways supply channels, local highways brooks and regional highways rivers. In Dutch such orders of water ways can be named more precise than in English.

Riviersystems		Road systems	
Dutch	English	Dutch	English
hoofdriever	mainriver	hoofdweg	highway
hoofdader	trunk stream	wijkverzamelweg	trunkroad
zijrivier	tributary	zijweg	sideroad
aftakking	distributary	zijweg	secondary road
beek	brook	buurtweg	tertiairy road
geul	channel	woonstraat	residential road
geultje	rill	woonerf	residential area

Fig. 407 Naming orders of river and road systems according to Moens

Bifurcation ratio, orders and network density

In Fig. 408 left, the bifurcation ratio of brooks before meeting a river is 20. However, the same network density could be reached with a bifurcation ratio 2 and 5 orders (Fig. 408 right).

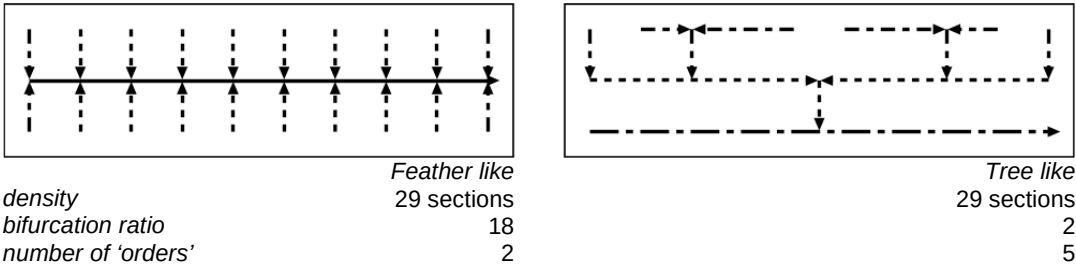


Fig. 408 Feather and tree -like connection patterns

Multiplying and extending these patterns into square surfaces (Fig. 409) tree like connection patterns seem to require a little higher density to open up all parts and consequently higher costs when restricted to bifurcation ratio 2.^a I do not understand why. Is halving the number of outlets responsible for a higher density? If somebody can design a lower density within this boundary conditions or prove its possibility mathematically I will publish it next time. On the other hand, tree like opening up every point of the area makes many variants and greater diversity of locations possible when you have more space to lay out (Fig. 409).

^a Perhaps because this restriction combined with mirroring vertically and horizontally has used all possibilities of external connection by two axes (above and below) counting half. So, vertically opening up the whole area makes more vertical sections necessary.

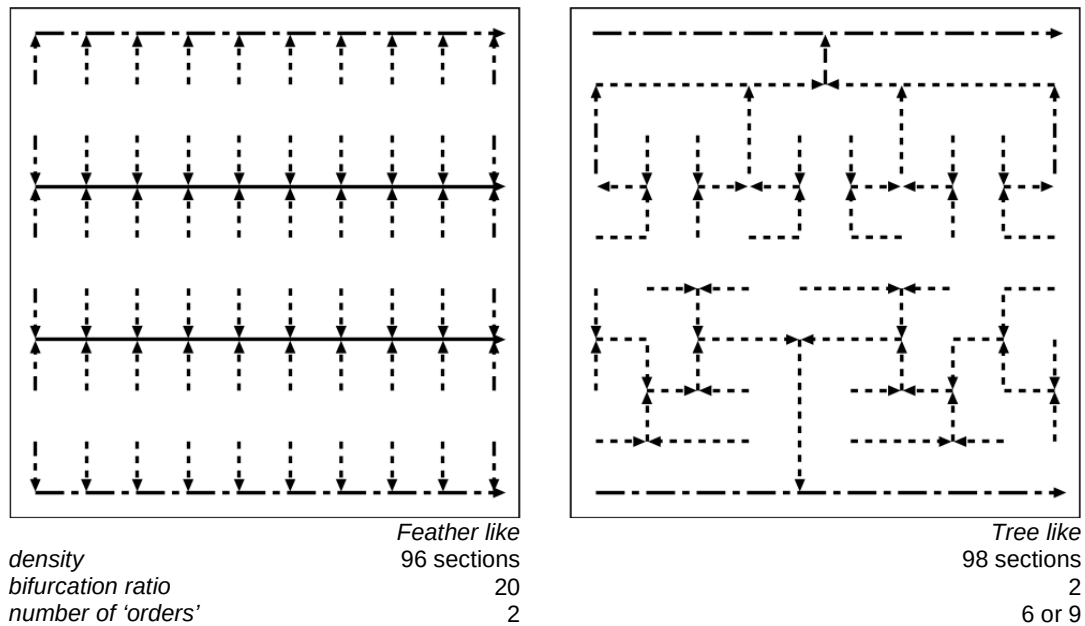


Fig. 409 Feather and tree like connection patterns opening up a square

Perhaps opening up a square in a tree-like way with bifurcation ratio 3 could reach the same or even lower densities and consequently lower costs. Try it. Does it result in less nodes and longer sections, a better readability of the area? The number and characteristics of nodes and the length of sections are important for spatial quality. Which rôle does the length of individual *sections* L play instead of total length per order in Fig. 405?

3.1.12 Catchment area and river length

The average length L of a random walk river *section* is a power its catchment area A ($L=A^{0.64}$). If length L is given the inverse the catchment area is a power of the length ($A=L^{1.563}$, Fig. 410 and Fig. 411). All the figures are experimental, obtained by observing many catchment areas and rivers.

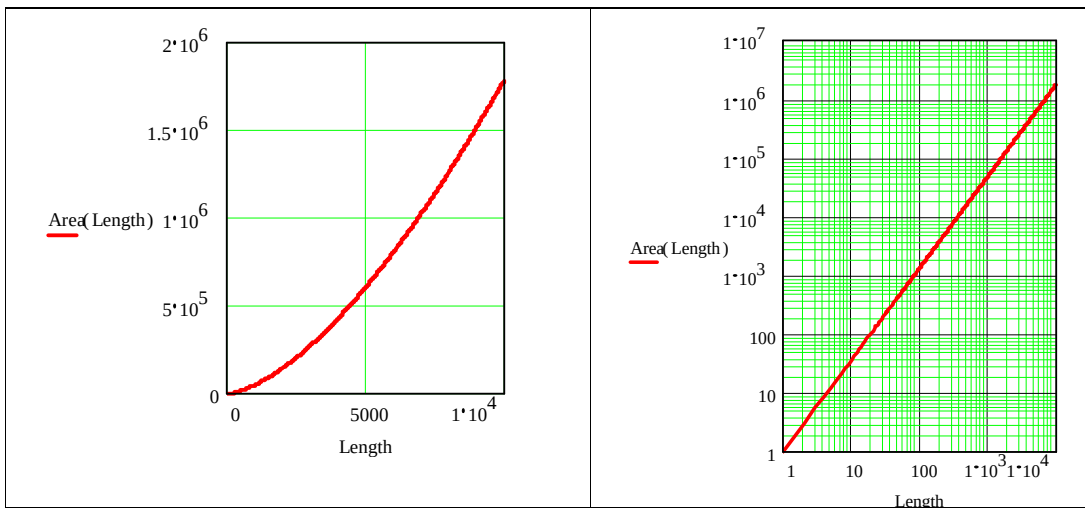


Fig. 410 Catchment area related to the length of a river section

Fig. 411 Logarithmic representation of Fig. 410

Check Fig. 403 by counting the corresponding squares in Fig. 402 of a specified order and its length. Compare your measurements with Fig. 411 and Fig. 405.

3.1.13 Local morphologies

A river can be described by its morphology. It is the credit of William Morris Davis (1912, *die erklärende Beschreibung der Landformen* (Leipzig und Berlin)) that for the first time a system is formulated based on development according to evolution. He describes the evolution of the valleys of the first order rivers as a V-shape without a valleybottom that develops in a wide valley with a valleybottom. This river will develop at the end in a real lowland river as we all know in The Netherlands from the river Rhine. Later scientists built further on his theory and adapted it where it was necessary. Fig. 414 - Fig. 420 show such a development with adaptations.

A classification according to deposits is also developed. The faster the water streams the coarser material can be transported as load. This means that at decreasing velocity of streaming a river will deposit first the coarse material. The slower the stream becomes the finer the sediment will be that will be deposited. Near glaciers coarse material is sedimented and a lowland river will deposit fine material as sediment. Moreover a river in a flat will tend to meander. By doing so the meander curves will move downstream due to the undermining of the outside curve by the streaming water.

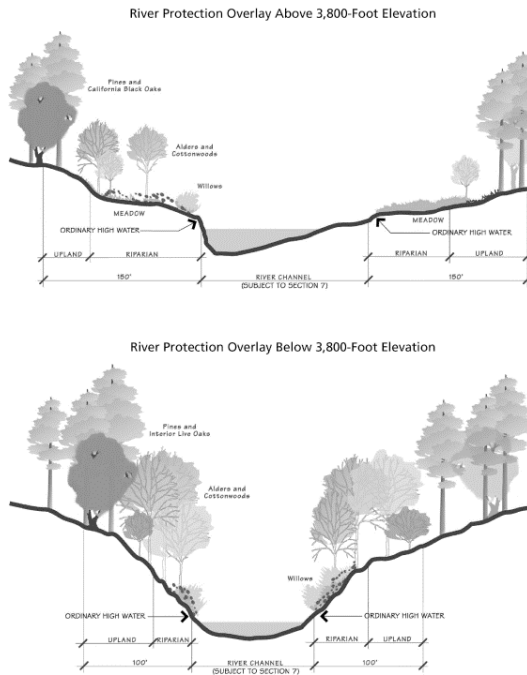


Fig. 412 River curves causing different vegetation^a

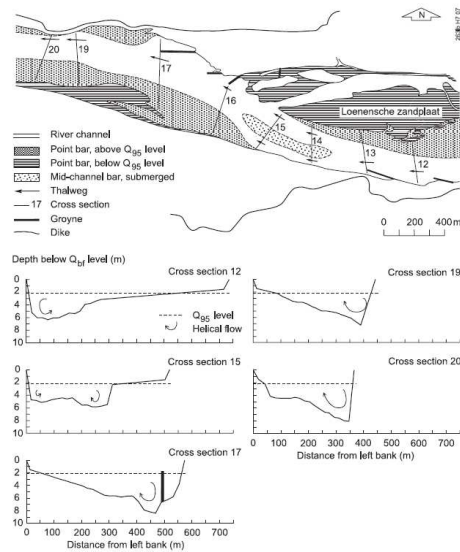
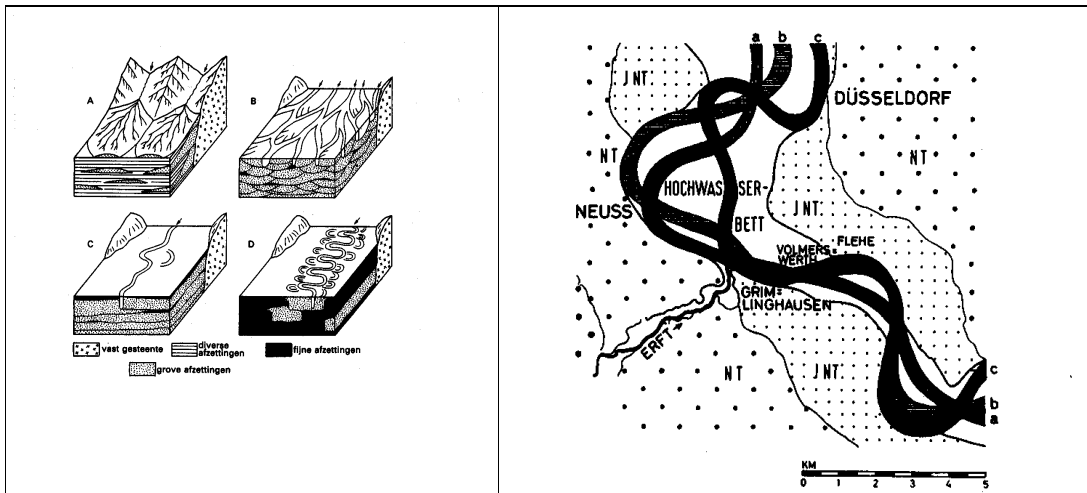


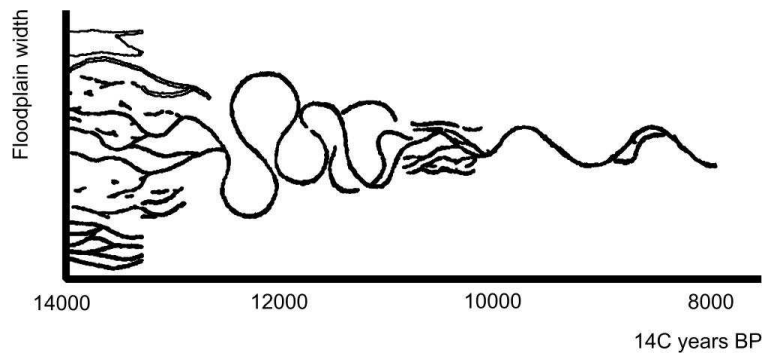
Fig. 413 Water flows causing different river bottoms^b

Every unevenness can cause an alteration the course of the river; many different channels of a river can be recognized in lowland river. So the water takes diverse and changing courses. Lower sections still bear rough material wearing out the outside parts of a bend into meanders, because rough material laid down there in the same time becomes a water barrier until heavy showers force a break through Fig. 415 and Fig. 417.

^a <http://geographyfieldwork.com/RiverEfficiencyCompetency.htm>;
http://www.nps.gov/archive/yose/planning/mrp/html/07_rmrp_ch1.htm
^b <http://igitur-archive.library.uu.nl/dissertations/1983151/c7.pdf>

Fig. 414 Forms of deposit^aFig. 415 Move of Rhine near Neuss from Roman times (a) via Middle Ages (b) until recently^b

In low lands finer deposits raise the bed in calm periods forcing water to find easier courses. A high discharge of a river causes even an river system with many branches in a lowland area. Such a system is called a braided river.

Fig. 416 Change of river behaviour in time^c

The morphology of the braided river is not very stable; it changes often depending the amount of water.

^a Allan cited by Zonneveld (1981) page 148

^b Hoppe cited by Zonneveld (1981) page 149

^c Tebbens et al. (1999), cited by Kroonenberg (2006)

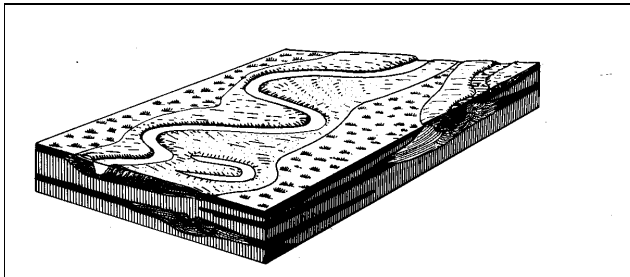


Fig. 417 Meandering river with historical deposits^a

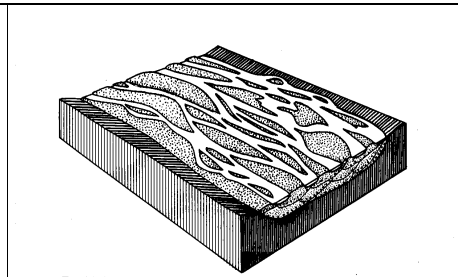


Fig. 418 Twining river^b

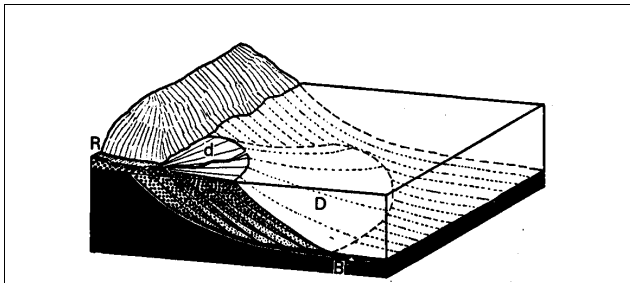


Fig. 419 Delta development with river (R), top-sets (d) and fore-sets (D)^c

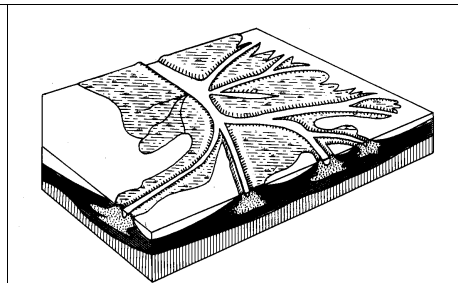


Fig. 420 Mississippi delta^d

The Rhine area downstream of Lobith is formed by both the process of meandering during quiet periods and braiding during periods with large differences of water discharges (Fig. 421).

^a Zonneveld (1981) page 143

^b Zonneveld (1981) page 144

^c Escher 1948 cited by Zonneveld (1981) page 160

^d Zonneveld (1981)page 161

From Lobith Rhine distributes water via Waal, Lower Rhine and IJssel in historically changing proportions.

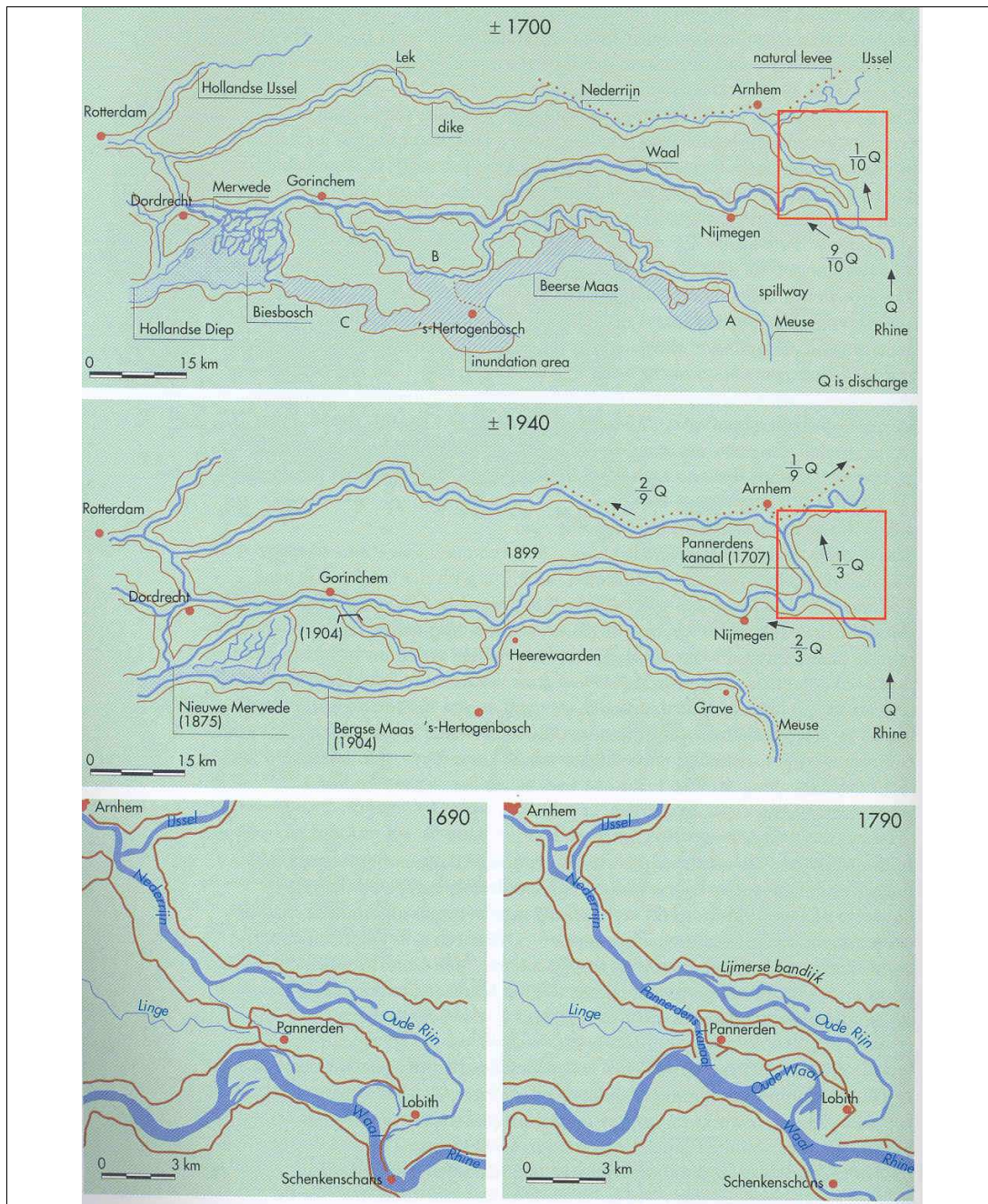


Fig. 421 Historical distribution of Rhine water from Lobith^a

3.1.14 Measuring velocities to get Q

^a Huisman, Cramer et al. (1998) page 38

The velocity v of water in a river can be measured on different depth vertical lines h with mutual distance stretches b of the cross-section B (Fig. 422). You can determine any partial discharge by multiplying $v \times b \times h$. The sum of the outcomes in cross section A for the different stretches b to get $Q = \Sigma(v \cdot b \cdot h)$ is an approach for the discharge. In the equation v is the mean stream velocity of the river and the velocity can easily be measured on site.

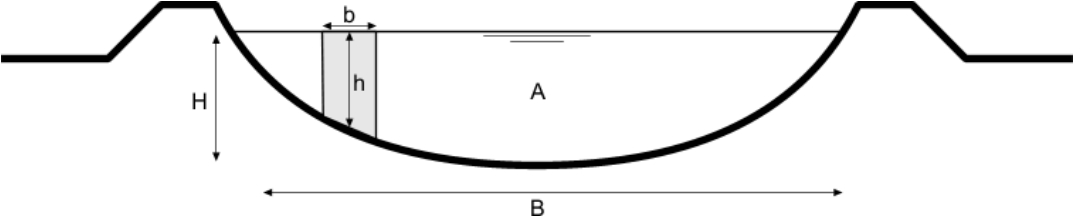


Fig. 422 Profile of a river^a

For example: asked the river drainage Q (Fig. 424), given h_i , b_i and v_i from profile subdivisions (Fig. 423).

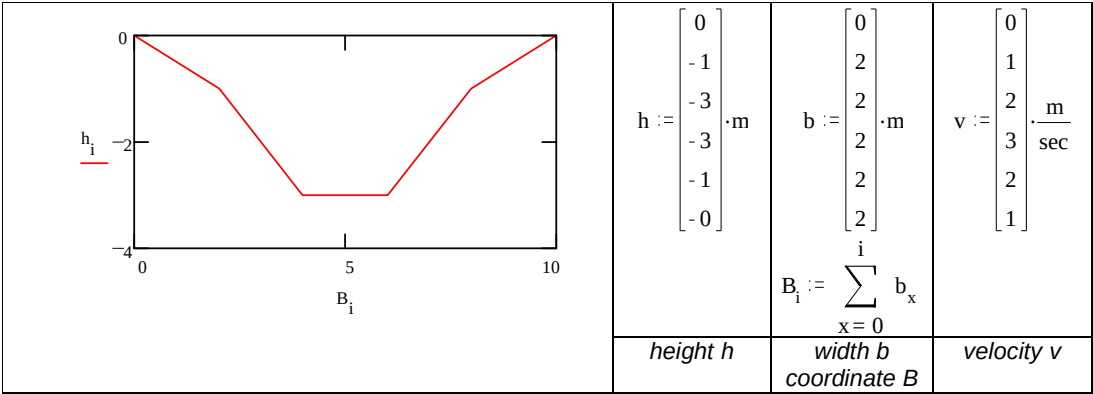


Fig. 423 Data from profile

^a Akker and Boomgaard (2001)

$i := 0..5$ $a_i := b_i \cdot h_i - \frac{1}{2} \cdot b_i \cdot (-h_i - h_{i-1})$ $A := \sum_i a_i$ $A = 16 \cdot m^2$	$Q_i := v_i \cdot a_i$ $Q := \sum_i v_i \cdot a_i$ $Q = 36 \cdot m^3 \cdot sec^{-1}$	v_i <table><tr><td>$0 \cdot m \cdot sec^{-1}$</td></tr><tr><td>$1 \cdot m \cdot sec^{-1}$</td></tr><tr><td>$2 \cdot m \cdot sec^{-1}$</td></tr><tr><td>$3 \cdot m \cdot sec^{-1}$</td></tr><tr><td>$2 \cdot m \cdot sec^{-1}$</td></tr><tr><td>$1 \cdot m \cdot sec^{-1}$</td></tr></table>	$0 \cdot m \cdot sec^{-1}$	$1 \cdot m \cdot sec^{-1}$	$2 \cdot m \cdot sec^{-1}$	$3 \cdot m \cdot sec^{-1}$	$2 \cdot m \cdot sec^{-1}$	$1 \cdot m \cdot sec^{-1}$	a_i <table><tr><td>$0 \cdot m^2$</td></tr><tr><td>$1 \cdot m^2$</td></tr><tr><td>$4 \cdot m^2$</td></tr><tr><td>$6 \cdot m^2$</td></tr><tr><td>$4 \cdot m^2$</td></tr><tr><td>$1 \cdot m^2$</td></tr></table>	$0 \cdot m^2$	$1 \cdot m^2$	$4 \cdot m^2$	$6 \cdot m^2$	$4 \cdot m^2$	$1 \cdot m^2$	Q_i <table><tr><td>$0 \cdot m^3 \cdot sec^{-1}$</td></tr><tr><td>$1 \cdot m^3 \cdot sec^{-1}$</td></tr><tr><td>$8 \cdot m^3 \cdot sec^{-1}$</td></tr><tr><td>$18 \cdot m^3 \cdot sec^{-1}$</td></tr><tr><td>$8 \cdot m^3 \cdot sec^{-1}$</td></tr><tr><td>$1 \cdot m^3 \cdot sec^{-1}$</td></tr></table>	$0 \cdot m^3 \cdot sec^{-1}$	$1 \cdot m^3 \cdot sec^{-1}$	$8 \cdot m^3 \cdot sec^{-1}$	$18 \cdot m^3 \cdot sec^{-1}$	$8 \cdot m^3 \cdot sec^{-1}$	$1 \cdot m^3 \cdot sec^{-1}$
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$8 \cdot m^3 \cdot sec^{-1}$																						
$1 \cdot m^3 \cdot sec^{-1}$																						
profile subdivisions	drainage per subdivision	velocity	surface	drainage																		

Fig. 424 Drainage (profile subdivisions and velocities)

3.1.15 Discharge Q on different water heights

The depth H of the river in a cross-section varies, but it can be measured on site. Then, the drainage Q(H) can be calculated by a practical formula apparently characteristic for the profile concerned. However, periods of high drainage Q or regular floodings in winter change profile and ... the formula. Comparing measurements like in Fig. 423 on different water heights you often find a curve like a parabola, approached by $Q = a \cdot H^b$ or $H = (Q/a)^{1/b}$ (Fig. 425). Parameters 'a' and 'b' should be found non-theoretically by experiment, seem to characterise the profile.

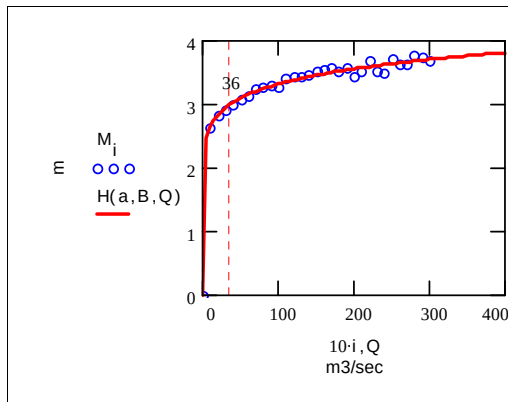


Fig. 425 'Measurements' M_i and $Q(a, B, H) = a \cdot H^B$ or the inverse $H(a, B, Q) = (Q/a)^{1/B}$ to get H on the y-axis

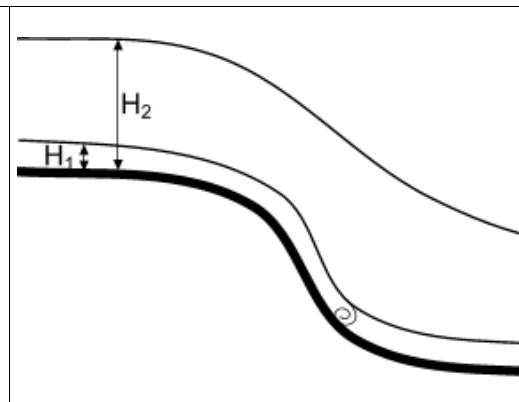


Fig. 426 Change of boundary condition downstream; a 'drowning' waterfall^a

Measurements deviate from the formula because velocity varies. When measurements can not be simulated by a smooth curve, it is probable that conditions downstream are changed by high water levels. Two graphs should then be drawn; one until the point of change, one for the higher values. When for example a waterfall downstream suddenly 'drowns' at increasing water levels (Fig. 426) the slope of the curve will change by the increase of velocity.

3.1.16 Interpolation of experimental data by using Excel

Constants a and b can be found by the least squares method provided by Excel using graphs (see Fig. 427). Enter the data of the measurements of height and drainage calculated according to Fig. 425 in two columns. Make a point graph and select it.

^a Akker and Boomgaard (2001)

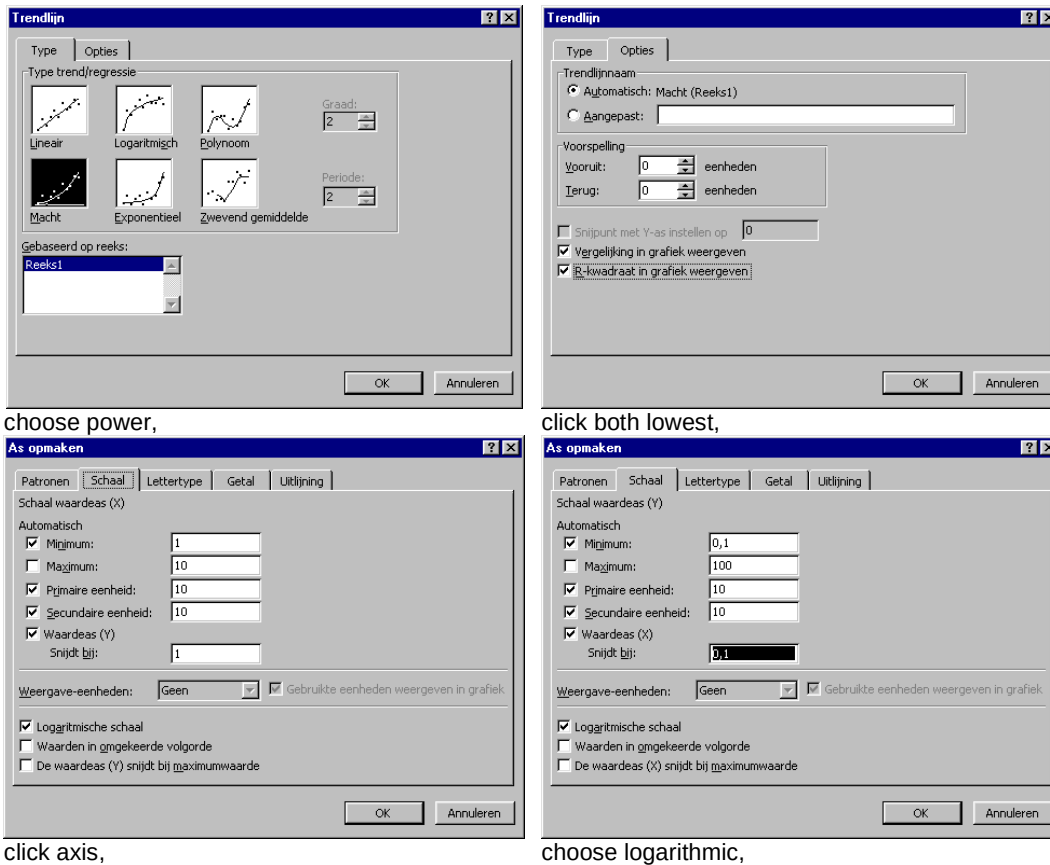


Fig. 427 Adding regression lines in Excel point graphs

Choose 'add trend' in 'graph' from the main Excel window above, and graphs like Fig. 428 and Fig. 429 with power regression line and formula are calculated by the program. With R^2 near to 1 you have a reliable formula. In Fig. 428 we used 'measurements' of Fig. 425 putting the independently variable measurements on the x-axis this time to find $a=0.0003$ and $b=8.7398$.

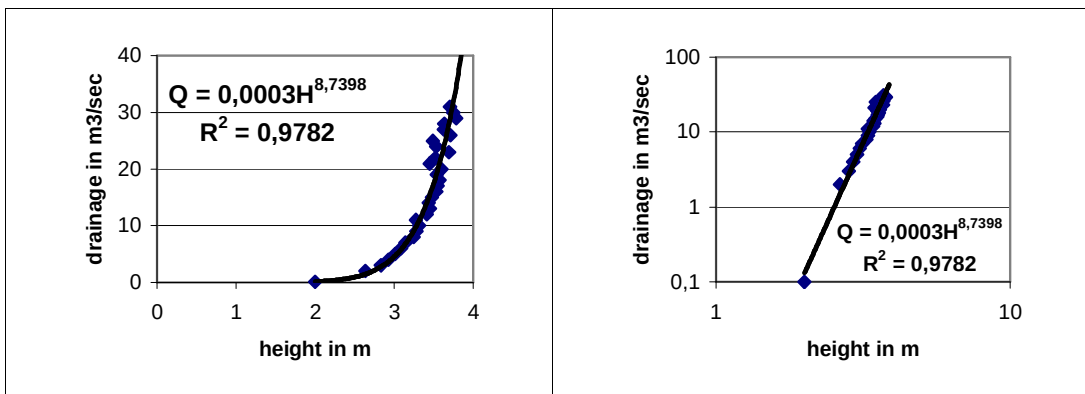


Fig. 428 'Measurements' M_i and $Q(a,b,H) = a \cdot H^b$ Fig. 429 Logarithmic representation of Fig. 428

The logarithmic representation $\log Q = \log a + b \log (H-H_0)$ produces a straight line easy to extrapolate to other heights and drainages. But be careful, there could be jumps in velocity by downstream events.

If you have made graphs before and after the jump because measurements could not be simulated by a smooth curve, each interval in Fig. 429 has different slopes representing different behaviour.

3.1.17 Calculating drainage Q with a rough profile

Just like wind, water slows down by roughness of the bed. The cross length of roughness in a wet profile P (Natte Omtrek) is calculated by summing hypotenuses of triangles according to Pythagoras characterised by the square root of $(b_i)^2 + (h_i - h_{i-1})^2$ (see Fig. 422 and Fig. 431). Considering the profile as a function $H=f(x)$ we can read the waterlevel H from accompanying left border $x_1=l$ and right border $x_2=r$ as values from $f(x)$ (Fig. 430). The length of roughness P within the cross section (Natte Omtrek = wetted contour) and the surface of the wet cross section A are both calculated as a function of H (Fig. 431).

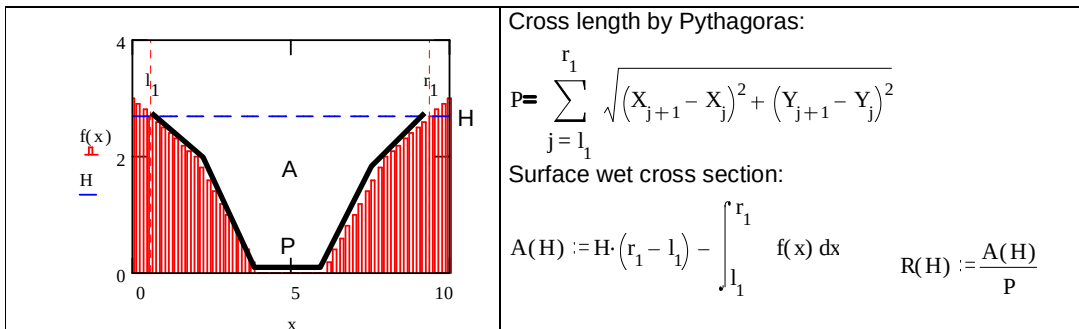


Fig. 430 Profile as a function

Fig. 431 Calculating wet cross section A and cross length of roughness P (NatteOmtrek)

When we divide the surface of the wet cross section A of a stream by this cross length of roughness P we get a measure indicating what part of the flowing water is hindered by roughness called 'hydraulic radius' $R = A/P$ in metres.

Method Chézy

The average velocity of water $v = Q/A$ in m/sec is dependent on this radius R, the roughness C it meets, and the slope of the river as drop of waterline s, in short $v(C,R,s)$.

According to Chézy $v(C,R,s) = C\sqrt{Rs}$ m/sec, and $Q = Av = AC\sqrt{Rs}$ m³/sec. Calculating C is the problem.

Method Strickler-Manning

Instead of $v = C\sqrt{Rs}$, Strickler-Manning used

$$v := \frac{R^{\frac{2}{3}} \cdot s^{\frac{1}{2}}}{n} \cdot \frac{m}{sec}$$

with roughness n taken from Fig. 432.

Characteristics of bottom and slopes	n	
	from	until
Concrete	0.010	0.013
Gravel bed	0.020	0.030
Natural streams:		
Well maintained, straight	0.025	0.030
Well maintained, winding	0.035	0.040
Winding with vegetation	0.040	0.050
Stones and vegetation	0.050	0.060
River forelands:		
Meadow	0.035	
Agriculture	0.040	
Shrubs	0.050	
Tight shrubs	0.070	
Tight forest	0.100	

Fig. 432 Indication of roughness values n according to Strickler-Manning^a

Method Stevens

Instead of $v=C\sqrt{R}$ s Stevens used $v=c\sqrt{R}$ considering Chézy's $C\sqrt{s}$ as a constant c to be calculated from local measurements. So, $Q = Av = cA\sqrt{R}$ m³/sec and c is calculated by $c=(A\sqrt{R})/Q$. When we measure H and Q several times ($H_1, H_2 \dots H_k$ and $Q_1, Q_2 \dots Q_k$), we can show different values of $A(H)\sqrt{R(H)}$ resulting from Fig. 431 as a straight line in a graph (Fig. 433). We can add the corresponding values of Q we found earlier in the same graph related to $A(H)\sqrt{R(H)}$. When we read today on our inspection walk a new water level H_1 on the sounding rod of the profile concerned we can interpolate H_1 between earlier measurements of H and read horizontally an estimated Q_1 between the earlier corresponding values of Q to read Q from graph.

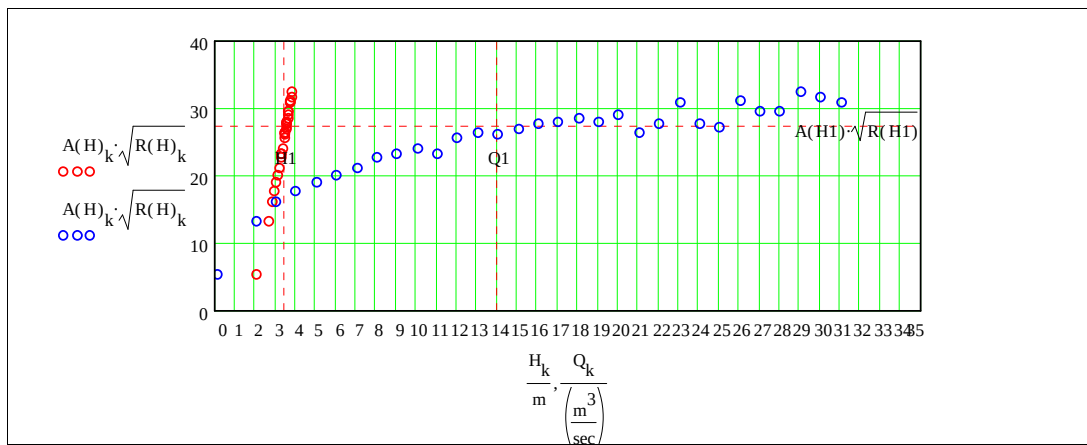


Fig. 433 Graph used according to Stevens with 'measurements' of Fig. 428

However, from these 'measurements' c appears to be not very constant, but the graph remains a practical way to estimate Q from H .

^a Akker and Boomgaard (2001)

3.1.18 Level and discharge regulators

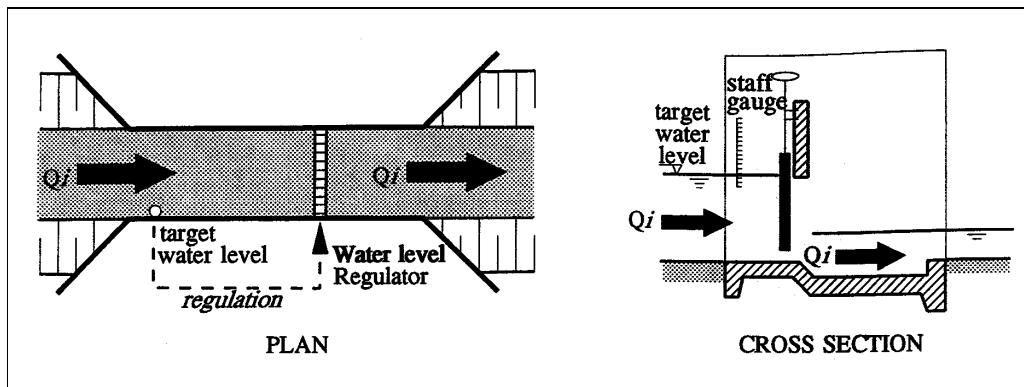


Fig. 434 Level regulator with level as target^a

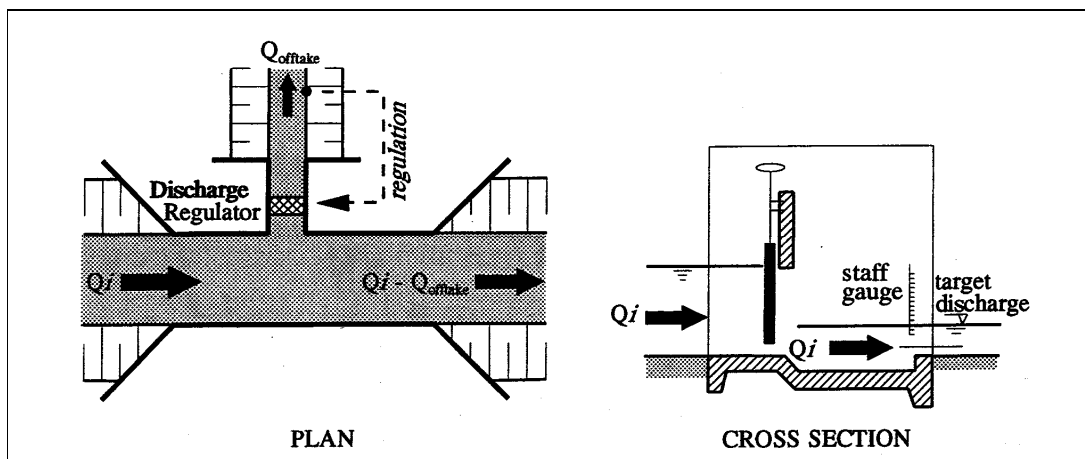


Fig. 435 Discharge regulator with discharge as target^b

^a Ankum (2003) page 156

^b Ankum (2003) page 156

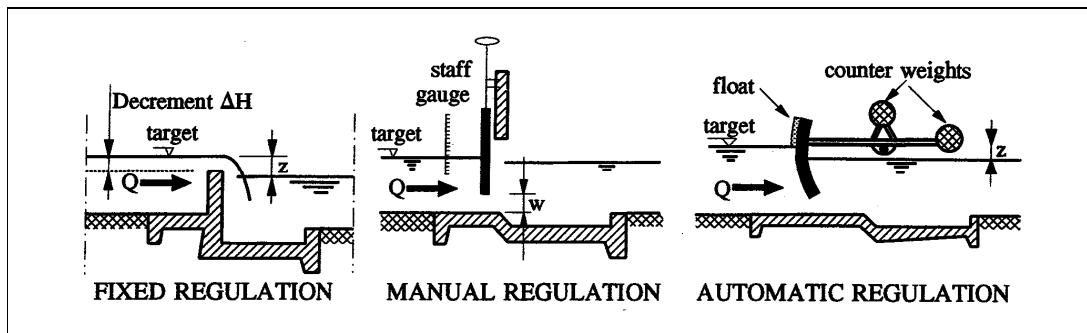


Fig. 436 'Manners' of regulation^a

The fixed regulators are called weirs (stuwen), manual or automatic regulators are called gates (schuiven).

^a Ankum (2003) page 167