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NATIONAL ENERGY AUTHORITY

PROGRESS REPORT OF BASIC RESEARCH

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Reykjavík
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An ever increasing field in the activities of the National Energy Authority, (NEA) is the research related to the exploration of the hydro-electric and geothermal power resources of the country. The National Energy Authority also provides expert assistance in fields related to its own activities.

Following is a series of short articles where the recent developments of scientific interest within the field of hydraulics are briefly discussed.

TIGHTENING OF WATER RESERVOIRS

by Jónas Elíasson.

Abstract.

Since 1964 considerable effort has been diverted to the investigation of reservoir tightening due to sedimentation. Most of the Icelandic rivers are of the glacial type, carrying sediments of grain sizes from fine clay to gravel. In NEA's experimental field, LANGÖLDUVEITA, it has been experimentally verified that lava fields flooded with water from a glacial river leak heavily to begin with, but get quickly tight, apart from very permeable contacts where tightening may take some time.

Carpet tightening.

This is when a sediment layer of thickness s and permeability k_s tightens the reservoir bottom. Let the bottom be a lava stratum of thickness L and permeability k_L overlying an aquifer of constant potential H below the reservoirs, then the overall permeability of sediment layer and lava together is found to be

$$(1) \quad k = (L+s)\,\frac{k_L \cdot k_s}{k_s \cdot L + k_L \cdot s}$$

Now, due to the varying conditions (thickness of lava and sediment) at the bottom, only approximate calculations are justified, so we put

$$(2) \quad s = C \cdot Q \cdot t/A,$$

with Q the mean discharge into the reservoir, C concentration of suspended load, t time and A reservoir area. Furthermore as $L >> s$, $k_s << k_L$ and s growing with time, we can put

$$(L+s)\ \frac{k_L \cdot k_s}{k_s \cdot L + k_L \cdot s} \approx L \cdot \frac{k_L \cdot k_s}{k_L \cdot s}$$ or

$$(2)\quad \frac{k \cdot t}{H} = \frac{L \cdot k_s \cdot A}{C \cdot Q \cdot H} = \alpha$$

α is a dimensionless constant. In the expression for α, H/L is representing the average potential gradient through the reservoir bottom, we hence may put:

$$\frac{k \cdot H}{L} = \frac{Q_L}{A_L}$$

Q_L being the leakage through the bottom A_L the lava area, now we obtain

$$(3)\quad \frac{Q_L \cdot t}{H \cdot A_L} = \frac{k_s \cdot A}{C \cdot Q}$$

which also can be found more directly. k_s varies within wide limits, but eq. (3) can serve as a definition of its average value. But in general, carpet tightening should follow the equation

$$(4)\quad \frac{Q_L \cdot t}{D \cdot A} = \beta$$

where β is a constant, caracteristic for the particular reservoir. D is some caracteristic depth. Field experiments in Langölduveita gave a β value of $\beta = 50$

Tightening of contacts.

Where lava fronts are submerged in water, water can be expected to flow through the highly pervious lava contact, provided there exists a potential drop along the contact layer. But if sediments of whatever grainsize that can get in are carried into the aquifer, it obviously tightens, as the porosity gradually decreases.

Various relations between the coefficient of permeability k, and the porosity n exist, we prefer:

$$(5) \quad k = \frac{k_1 \cdot n^2}{(1 - n)^3}$$

given by Frank Engelund 1. in a slightly different form. Denoting the sediment flow vector by S the equation of sediment continuity yields

$$(6) \quad \frac{\partial n}{\partial t} = \text{div } S$$

Now:

$$(7) \quad S = c \cdot V$$

c concentration, and V the water velocity. For small concentrations the equation of water continuity gives div V = o and (6) and (7) yield:

$$(8) \quad \frac{\partial n}{\partial t} = V \cdot \text{grad } c$$

Now recalling Darcy's law we get:

$$(9) \quad \frac{\partial n}{\partial t} = \div k \; \{\text{gradh} \cdot \text{gradc}\}$$

The expression in the square brackets is a scalar having the dimension of $(\text{length})^{\div 1}$, but no general expression can be found for its value for obvious reasons. The purpose is, to find a qualitative expression for the variation of k with time. The only way to integrate (9) is to produce a relationship between c and h, but no such relationship is known. We therefore proceed by putting:

$$\{\text{gradh} \cdot \text{gradc}\} = \frac{H \, C_o}{L^2} = \text{constant}$$

The idea is of course that H/L should represent the "average" hydraulic gradient and C_o/L the "average" concentration gradient.

Eq. (9) is now easily integrated. In wiev of the simplifications already introduced the use of:

$$(10) \quad k = k_o \cdot \left(\frac{n}{n_o} \right)^2$$

instead of the more ponderous (1) is easily justified. k_o and n_o denote initial values at time $t = o$, and we arrive to:

$$(11) \quad \sqrt{\frac{k_o}{k}} = 1 + \alpha \cdot \frac{k_o \quad t}{H}$$

where $\alpha = H^2 C_o / L^2 \, n_o$, which is dimensionless must be treated as a constant, caracteristic for each particular case. By using (5) instead of (10), eq. (11) gains considerably in length, gains in accuracy being more doubtful. Fig. 1. shows the result from a field experiment, a 20 m deep borehole penetrates a 1 m thick contact, water with average sediment concentration 0.115 o/oo (volume) is pumped into the hole under a constant pressure of 21 m over original GWT. The experiment lasted 2 weeks, and agreement with theory is fair, giving an α of 1.37. This high value of α suggests that L, which here represents "radius of the cylinder under tightening", is rather low, it is found to be of the order 0,4 m, which also explains why any agreement with theory at all is found, where there in fact should be none, because the special features of radial flow are not accounted for, in the derivation of forumula (11). The low value of L also suggests that the tightening could be washed out by applying higher pressure. A later test also showed that a pressure increase of 100% washed out the tightening.

Application of the formulas

Carpet tightening is described by formula (4), which can be used once β is know. When it is not, estimates can be based upon formula (3), especially if data from the same river (same c and k_s) are known. Contact tightening is more complicated as can be seen from formula (11) and the composition of α, which shows that for small gradients, the tightening process may take very long time. Research is now being carried out in order to find the best way to accelerate this process.

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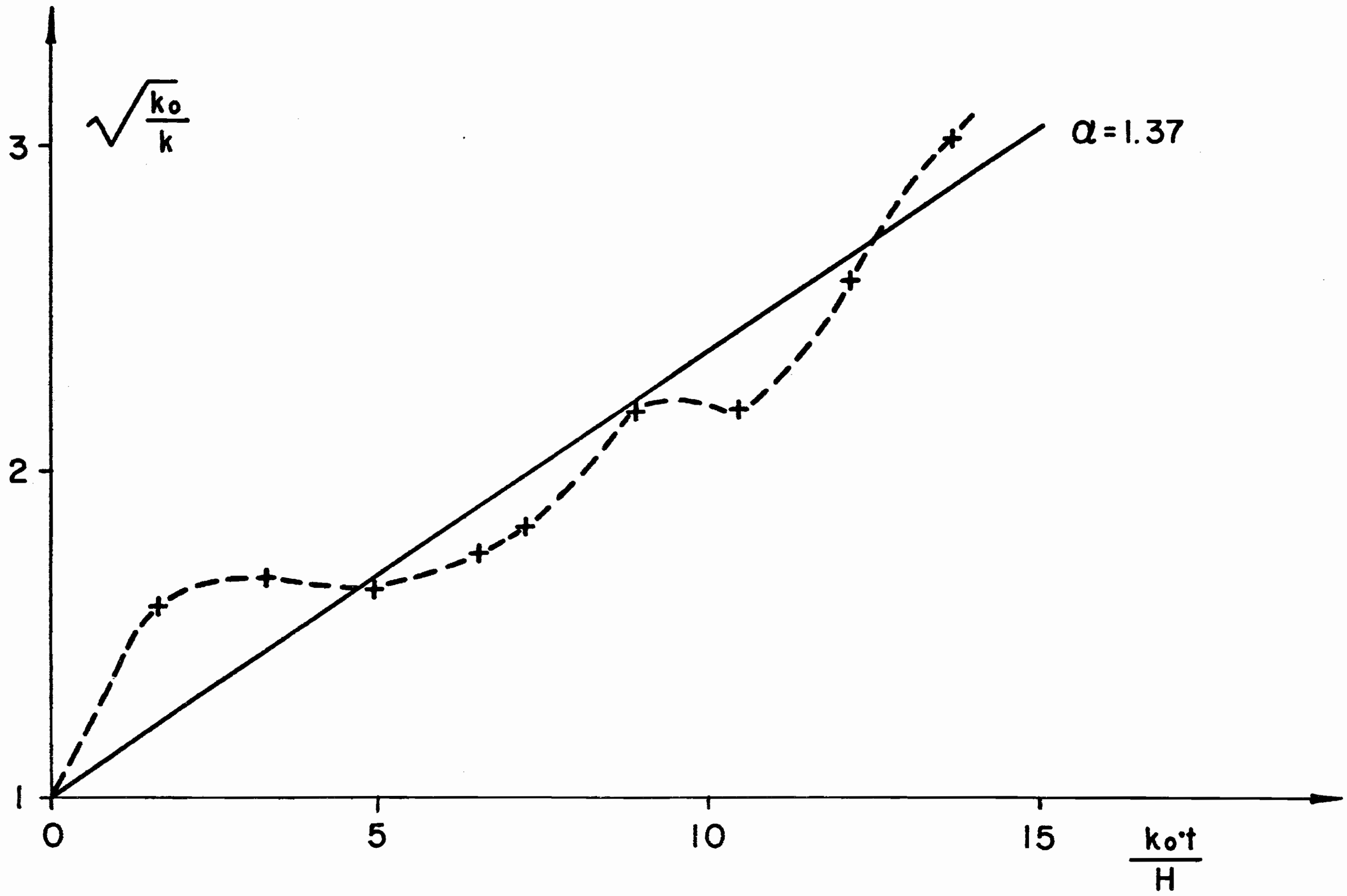

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GROUNDWATER FLOW IN THIN AQUIFERS

By Jónas Elíasson.

Abstract.

Under a lava flow a highly permeable scorious contact is usually formed. These contacts come to the surface along almost any lava front. If a lava front is submerged as a result of a hydroelectric development, the interbed may give rise to a serious leakage.

Basic equations.

In a cartesian co-ordinate system x, y, z, with the z axis vertically upwards, Darcy's law is used to connect the percolation velocity vector $V = (u, v, w)$, the coefficient of permeability k, and the energy $h = z + p/\gamma$ (p pressure, γ unit weight)

$$(1) \quad V = \div k \text{ gradh}$$

continuity provides:

$$(2) \quad \text{div } V = o$$

resulting in the well known Laplace equation:

$$(3) \quad \text{divgradh} = \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = o$$

Suppose that in a particular point x,y, z the coordinate system is orientated thus that $v = 0$, then $\partial h/\partial y = o$, and if now the aquifer thickness is considered uniform and small in relation to its horizontal dimensions we have in this particular point:

$$w \sim i \cdot u$$

i being the inclination of the aquifer in the direction of the x - axis. This immediately gives

$$\frac{\partial h}{\partial z} \sim i \frac{\partial h}{\partial x}$$

and from this we derive, as $i = i\,(x,y)$

$$\frac{\partial^2 h}{\partial z^2} \sim i \cdot \frac{\partial^2 h}{\partial z \partial x} = \cdot\, i \frac{\partial}{\partial x} \left(\frac{\partial h}{\partial z} \right)$$

$$\sim i \cdot \frac{\partial}{\partial x} \left(i \frac{\partial h}{\partial x}\right)$$

Usually we have, $i \ll 1$
and in that case we can put

$$\frac{\partial^2 h}{\partial z^2} \sim i^2 \cdot \frac{\partial^2 h}{\partial x^2}$$

which shows that:

$$(4) \quad \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = 0$$

can be considered valid everywhere. On the boundaries h is either known or the boundary is impervious.

Numerical procedure

When equation (4) is established, an immense simplification from the more gbneral (3) is obtained and h is easily found by relaxation processes:

$$h_{i,j} = h_n = 1/4\cdot(h_{i,j+1} + h_{i,j-1} + h_{i-1,j} + h_{i+1,j})$$

valid away from impervious boundaries.

$$h_{I,j} = h_n = 1/3\ (h_{I,j+1} + h_{I,j-1} + h_{I-1,j})$$

,valid for i = I+1,is animpervious boundary,a similar equation for j = J + 1 impervious is easily found. For i+1+j+1 - K an impervious boundary we have:

$$h_{i,j} = h_n = 1/2\ (h_{i,j-1} + h_{i-1,j});\ j = K - i - 2$$

This iteration process can be started from almost any initial value, it is unconditionally stable, but rather slow.

To speed up the process it is convenient to use an acceleration factor $1 < \alpha < 2$

$$h_{i,j} = h_{i,j} + \alpha(h_n - h_{i,j})$$

Usually α - 1,7 a 1,8 renders processes that converge about 25 times faster than α=1,0. When h is known, any derived quantity, - such as the discharge through any particular cross-section - is quickly found, so the method is a convenient one, to calculate the flow through thin aquifers of moderate inclination and any planar shape.

STORM WAVE GROWTH

by Jónas Elíasson

Abstract

In the following, qualitative formulas for the growth of ocean waves are derived. An expression for the wave steepness is obtained, with combined use of the concept of transported energy and the concept of wave thrust, originally set forth by H. Lundgren.

The analysis, which are mainly dimensional in character, are based upon the use of simplified expressions for the transfer of wind energy to wave energy.

Dimensionless expressions.

The symbols used are:

H	Mean wave height	t_w	Wind duration
T	" " period	F	Free fetch
C	" " velocity	ρ	Unit mass
L	" " length	x	Length
U	" wind velocity		
g	Acceleration of gravity		

$$\theta = \frac{C}{U} \;,\; h = \frac{g \cdot H}{U^2} \;,\; f = \frac{g \cdot X}{U^2} \;,\; s = \frac{H}{L} \;,\; t = \frac{g \cdot t_w}{U}$$

Assumptions.

It is assumed that the wind shear stress, τ_W can be divided into a "current generating" part and a "wave generating part" τ in such way, that

$$(1) \qquad \lim_{C \to U} \tau = 0$$

(1) can be formulated in several different ways, and many arguments favour it, the most obvious being the fact, that a boundary layer current is formed in the surface region of the water, as the waves grow bigger, and the waves don't grow indefinitely.

Deep water and constant wind velocity is assumed, it is also assumed, that the wave energy is conserved within groups of waves.

Basic equations.

The average transported wave energy is denoted E, and the average wave thrust P, with x orientated in the direction of the wind, we get:

$$(1) \quad \frac{dE}{dx} \sim \tau \cdot C$$

by averaging in time the equation of energy transfer, and we get:

$$(2) \quad \frac{dP}{dx} = \tau$$

by averaging in time the momentum equation, the proportionality sign in (1) is because the time average of the work done by the shear stress with average value τ, on a water profile moving with average velocity C, is not equal to $\tau \cdot C$.

If we now introduce the usual deep water expressions for E and P, known from the theory of waves of small amplitude, we get:

(3) $\frac{d}{dx} \left(\frac{1}{16} \rho gH^2C\right) \sim \tau \cdot C$

(4) $\frac{d}{dx} \left(\frac{1}{16} \rho gH^2\right) \sim \tau$

The proportionality sign in (4) is due to the ratio between the time average of the squared wave height and H^2, if Rayleigh's distribution is assumed, this factor is $\frac{\pi}{4}$.

By eliminating τ, (3) and (4) give:

(5) $d(h^2\theta) = k\, dh^2_\theta$, k =constant, (5) is easily integrated, so we obtain:

(6) $h = b\,\theta^a$

where b and a are constants.

By using (1) b is found to be:

$$b = h_{max}$$

h_{max} is the value h approaches, if the storm proceeds indefinitely.

The constants a and b can only be found by observations. If now (6) is fitted to C. Bretschneiders diagrams 1., the constants should be:

$$a = 1.5, \; h_{max} = 0.145$$

Furthermore, if we put by definition

(7) $h = 2\pi s\theta^2$

in accordance with the theory of waves of small amplitude, we get:

$$(8)\quad s = s_o \cdot \theta^{-\frac{1}{2}}$$

where

$$s_o = \frac{h_{max}}{2\pi} = 0.023$$

is the mean wave steepness of fully developed storm waves.

A functional relationship between h and f cannot be theoretically calculated as science has not yet brought about a reliable expression for τ. An empirical relation though exists, in the form of the above mentioned diagrams. Using this relationship and (6), together with:

$$(9)\quad df = \frac{1}{2}\,\theta\, dt$$

a relationship between f and t is easily found by numerical integration, thus relating the wind duration to the equivalent free fetch.

Discussion of the results.

The derivation of the formulas is based upon two things mainly, the existance of the "wave generating" τ , and the use of the formulas of the theory of waves of small amplitude. To speak of the latter first, these formulas can be used, provided they are accurate enough for other technical use in coastal engineering, and if the ratio between the actual value and the mean value of any physical parameter is a stochastic variable, independent of x and t_W, but the different variables cannot be independent of each other ($k = 1$ in (5) is impossible).

A theoretical solution of the problem therefore does not rest so much upon proofing (1), but finding a reliable expression for the wind shear stress. But here it is not sufficient to find its mean value, its variation with the variable wave height must also be found.

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2) LUNDGREN H. Wave Thrust and Wave Energy Level. IAHR - 10th Con. No. 1.20

3) WILSON B.W. Graphical appraoch to the forecasting of Waves in moving fetches. Technical memorandum No. 73 B.E.B. 1955.

SIMULATION OF SEDIMENT LOAD

by Jónas Elíasson

Abstract.

Since 1963 considerable effort has been made to find the amount of sediment load transported by the Icelandic big rivers. The research work is headed by the NEA Sediment Research Committee, and covers e.g. regular sampling from 5 sites, and irregular sampling from many others. Analysis of the samples show, that the transported quantity follows the water discharge, but the deviation from the expected value is a stochastic variable.

Analysis of sediment discharge

The sites are usually so chosen, that all but the largest particles are suspended in the stream. At all sites the samples are always taken at the same place, using the integrated sample technique, the sample is divided at grain size 0.02 mm, and the fine, coarse and total sediment discharges, analysed seperately. Detailed description of the laboratory analysis is given in ref. 1. and 2.

Now, a linear regression of the logarithms of simultaneous sediment and water discharges, defines a "Sediment Transport Rating Curve" of the form:

$$(1) \quad q_s = A \cdot Q_v^B \; .$$

q_s is sediment discharge in kg/sec and Q_v water discharge in m^3/sec, so the coefficient A is not dimensionless. Fig. 1 shows the result of one such regression. The line is well defined, with the individual points scattering around it. This scatter was analysed in three ways, firstly, investigation of the variable $x = q_p - q_s$, q_p being the sample value, gave no definite results. Secondly, investigations of the variable $x = \log q_p - \log q_s$ showed x to be approximately normally distributed with the mean $\mu = 0$ and standard deviation $\sigma < 1$. Thirdly, the serial correlation coefficient $\rho\ (x_{i+1})$, where i is the number of the day in the year, was found to be 0.3 or lower. But it must be pointed out, that information on ρ is rather little, as the regular sampling program does not include sampling at the same site on 2 consecutive days. Nevertheless, it is clear that x_i can be treated as a normally distributed stochastic variable, with sufficient accuracy for all practical purposes.

Simulation

$$(2)\quad q_i = q_{si} \cdot \exp(\sigma \cdot y_i)$$

where y_i is a normally distributed stochastic variable with mean 0 and standard deviation 1, is a statistical model of the sediment discharge. Several results can immediately be dervived from (2),

e.g.:

$$\overline{q_i} = \overline{q_{si} \cdot \exp(\sigma \cdot y_i)} = \overline{q_{si}} \cdot \overline{\exp(\sigma \cdot y_i)}$$

$$= \overline{q_{si}} \cdot \exp(\sigma^2/2)$$

or, the average sediment discharge is the factor $\exp(\sigma^2/2)$ higher than the rating curve value, which again is the median (50%) value. Also, if we denote by

$$Q_j = \sum_i q_{sj,i} \cdot \exp(\sigma \cdot y_{j,i})$$

the total sediment discharge in the year j, we find

$$(3) \quad \overline{Q} = \exp(\sigma^2/2)\cdot\sum_i q_{si} = \exp(\sigma^2/2)\cdot\overline{Q}_s$$

To find

$$\sigma^2(Q) = \overline{Q_j^2} - \overline{Q}^2$$

we insert (2), and (3), and after som calculation we arrive to

$$(4) \quad \sigma^2(Q) = \exp(\sigma^2)\cdot(\overline{Q_s^2} - \overline{Q}^2)$$
$$+ \exp(\sigma^2)\cdot(\exp(\sigma^2) - 1)\cdot\overline{\sum_i q_{si}^2}$$

The first term is $\sigma^2(Q)$ as it would be, if the sediment discharge at a given water discharge was constant, equal to the mean value $\overline{q}_i$. The last term therefore originates from the fluctuations about this mean value.

(2) is ideal for simulation of the sediment discharge on a computer. q_{si} is a function of the water discharge only, and can be calculated aforehand, for any series Q_{vi}, observed or simulated, and the factor $\exp(\sigma\cdot y_i)$ is easily derived with the aid of any random variable. If e.g. $0 \leq x_i \leq 1$ is uniformly distributed then the distribution of

$$y_i = \sum_n x_{i,n}$$

converges in probability to the normal distribution as n increases.

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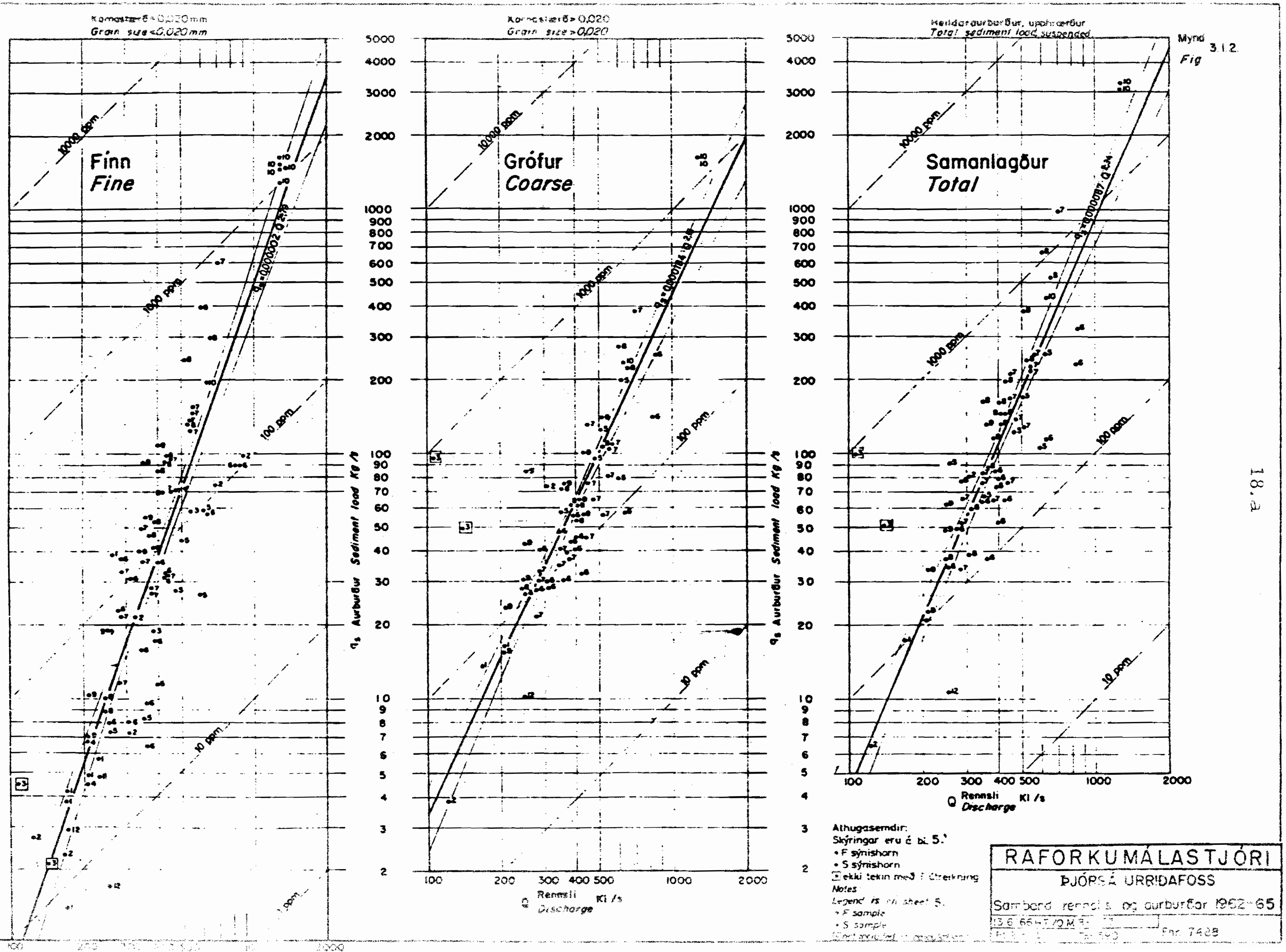

Mynd 3.1.2.
Fig
Kornastærð < 0,020 mm
Grain size < 0,020 mm
Finn
Fine
Kornastærð > 0,020
Grain size > 0,020
Grófur
Coarse
Heildaraurburður, uppitærður
Total sediment load, suspended
Samanlagður
Total
10000 ppm
1000 ppm
100 ppm
10 ppm
q_s = 0,0000002 Q 2,78
q_s = 0,00000784 Q 2,18
q_s = 0,0000087 Q 2,34
q_s Aurburður Sediment load Kg/s
Q Rennsli Kl/s
Discharge
Athugasemdir:
Skýringar eru á bl. 5.
• F sýnishorn
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Legend is on sheet 5.
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RAFORKUMÁLASTJÓRI
ÞJÓRSÁ URRIÐAFOSS
Samband rennslis og aurburðar 1962-65
Fnr. 7488

RIVER ICE PRODUCTION

by Sigmundur Freysteinsson
Thoroddsen & Partners, Consulting Engineers

Abstract

Production of sludge ice in rivers may seriously hamper power generation, or even stop it in severe cases. NEA has since 1963 investigated the various factors involved in the ice production of rivers. In 1964 the work was headed by the Norwegian ice experts dr. O. Devik and chief eng. E. Kanavin and in the same time a cooperation with Thoroddsen & Partners was initiated. Further ice research is now carried on by the joint research staff of Thoroddsen and NEA.

Water temperature in rivers and canals.

Neglecting dissipation of turbulent energy and radiation effects the differential equation for the water temperature in a turbulent stream is:

$$k \text{ div grad } \overline{T} - \gamma c \, \overline{V} \cdot \text{grad } \overline{T} - \gamma c \text{ div } \overline{(T'V')} = \gamma c \frac{\partial \overline{T}}{\partial t}$$

where k is the conductivity of the water, T the water temperature, γ specific weight and c the specific heat of water, V the velocity vector and t time. The bars denote time averages and T' and V' are the fluctuations.

For the simplest case: unform flow in a wide rectangular channel, it can be shown that with the usual assumptions and empirical relations of engineering hydraulics the water temperature is a function of several dimensionless groups

$$T = f \; (æ, I, D/Z_o, F, P_t, P, R, N_s, N_b).$$

Here æ is von Kármán's constant, I the slope of the energy line, D the depth, Z_o the equivalent sand roughness of the bottom, F the Froude number, P_t the turbulent Prandtl number, P the Prandtl number, R the Reynolds number and N_s (Nusselt number) is a dimensionless form of the boundary conditions at the surface:

$$N_s = -\frac{S}{k}\,\frac{D}{T_o}$$

where S is the rate of heat loss from the surface per unit area, and T_o is a reference temperature. N_b is a corresponding expression for the boundary conditions at the bottom.

In swift rivers or canals where $\partial T/\partial t$ and $\partial T/\partial x$ can be assumed constant over the depth the water temperature equation simplifies to:

$$\frac{\partial T}{\partial t} + V_m \frac{\partial T}{\partial x} = -\frac{S}{\gamma c D}$$

where V_m is the mean velocity. The heat exchange with the bottom is neglected here, as it is usually small compared to that at the surface. For constant S the general solution to this equation is

$$\Phi(x-vt,\ T + \frac{St}{\gamma c D}) = 0$$

where Φ is an arbitrary function. The solution can also be written

$$T\,(x,t) = \phi(x-vt) - \frac{St}{\gamma c D}$$

which for the characteristics $x-vt = a =$ constant gives

$a < 0;\ T(x,t) = T(0,-a/v) - St/\gamma c D$

$a > 0;\ T(x,t) = T\,(a,o) - St/\gamma c D$

In the stationary case ($\partial T/\partial t = 0$):

$$T(x) = T(o) - \frac{Sx}{\gamma c D V_m}$$

The simplified equations have been used:

1) To calculate the ice-free reaches of proposed canals from reservoirs for given values of S and reservoir temperature.

2) To evaluate the heat loss, S, from measurements of the water temperature in rivers.

Heat loss from rivers

The discharge of frazil ice during frost periods is a problem of considerable importance at certain sites in the Thjorsá Basin. As direct measurements or estimates on frazil ice discharge are very scarce, the ice prodcution has been computed from meteorological and hydrological observations. The different formulas available in the literature for the rate of heat loss from a water surface give widely divergent results and therefore it was necessary to investigate the subject.

The set of formulas for the rate of heat loss per unit area presently used in the Thjorsá Basin is the following:

$$s_1 = \{13{,}18 \cdot 10^{-9} \cdot T_a^4 (0{,}46 - 0{,}06 \sqrt{e_a}) - G_o(1-\alpha)\}(1 - 0{,}012N^2) + 13{,}18 \cdot 10^{-9} (T_w^4 - T_a^4);$$

$$s_2 = 1{,}9 \cdot V_6^{\,0{,}845} (e_w - e_a);$$

$$s_3 = 1{,}2 \cdot V_6^{\,0{,}845} (T_w - T_a).$$

s_1: heat loss by radiation, Mcal km^{-2} s^{-1}

s_2: " " " evaporation, "

s_3: " " " convection, "

T_a: air temperature, degrees Kelvin

T_w: water " , "

e_a: wapour pressure of the air, mb

e_w: " over water (saturation vapour pressure of air at water temperature), mb

G_o: global radiation with clear sky, Mcal km^{-2} s^{-1}

α : albedo of the water surface

N: cloud cover, 0-8

V_6: wind velocity at 6 m height, m s^{-1}

The empirical constants in the formula for radiation are taken from the literature but empirical constants and the exponent on the wind velocity in the formulas for evaporation and convection,were determined by measurement of the heat loss, from the water surface of specially constructed calorimeters. As it is possible that formulas derived from measurements in calorimetres will give to high a rate of heat loss several methods have been considered to verify the formulas, f.i. determination of the heat loss from measurements of water temperature. The only method that has been carried out sucessfully so far is a comparison of calculated ice production to the increase in volume of an ice jan. In March 1965 an increase of 19 $\pm$ 1 million cubic metres of the Burfell ice jam in Thjórsá River was determined with photogrammetric methods. According to the heat loss calculations,13 million tons of ice reached the jam, which agrees with the increase in volume,if the mean water equivalent of the jan is 0,65-0,72, which is rather high, but not unreasonable.