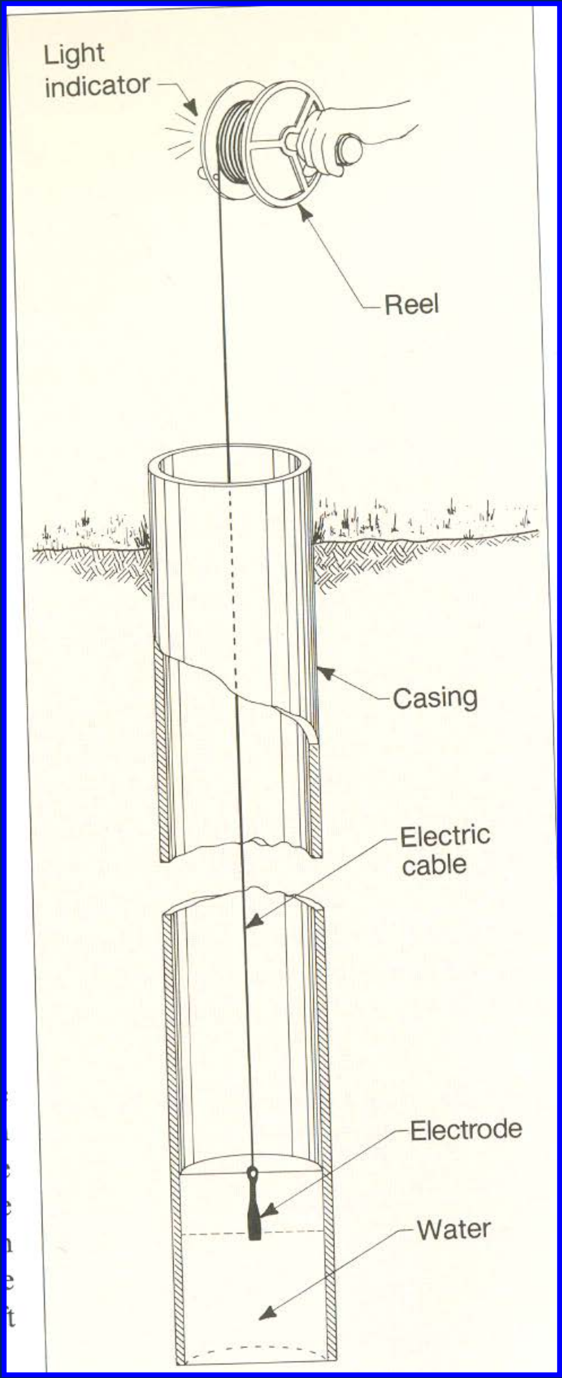
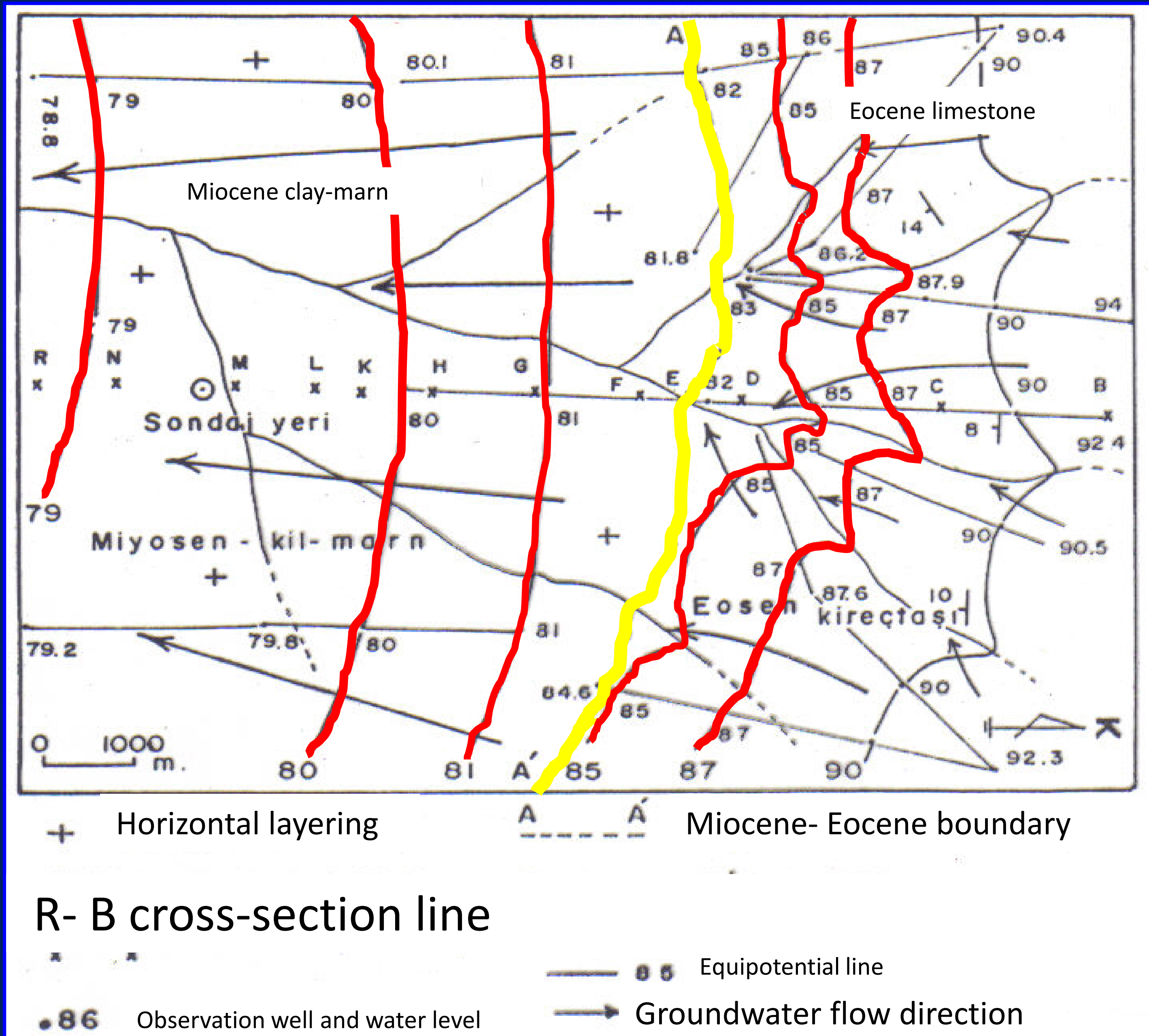
Water Level Meters



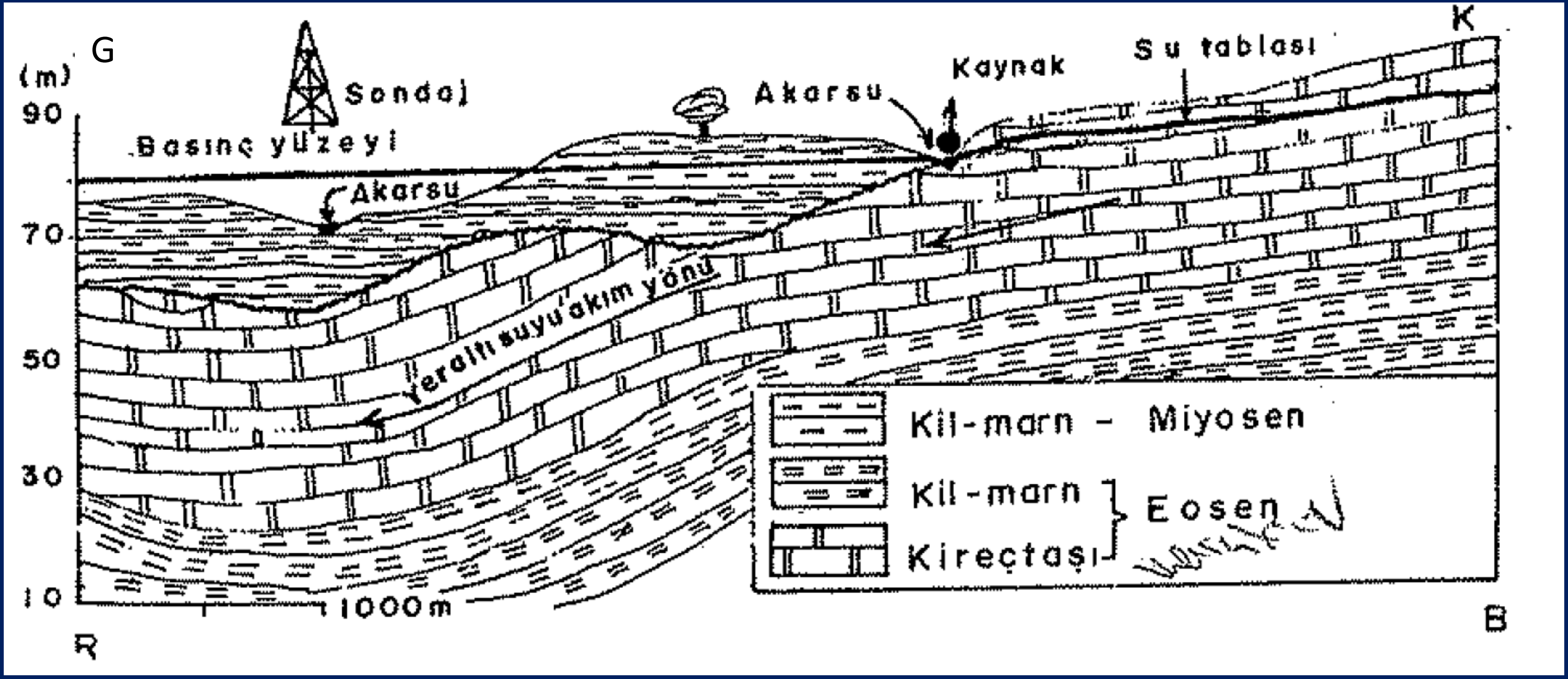


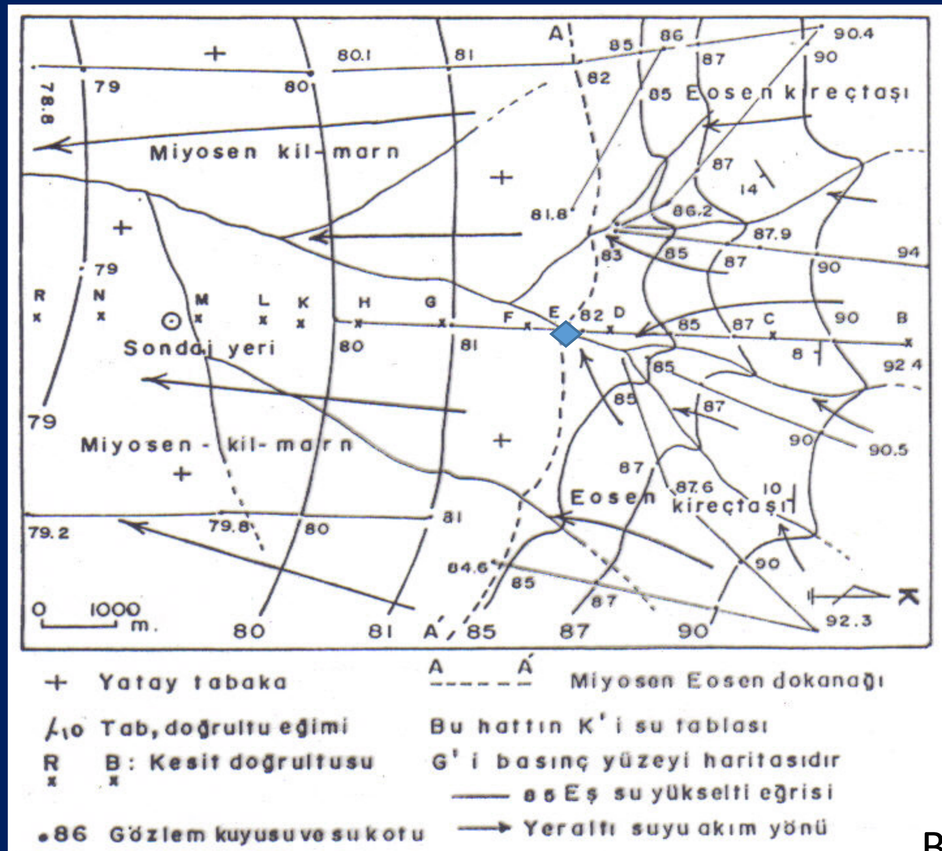
Unconfined aquifer?
Confined aquifer?

Water table and piezometric level maps

Hydrogeological Cross-section

Hydrogeological cross-sections are different from regular geologic cross-sections because they contain hydrologic and hydrodynamic structures like springs, water table or piezometric surface levels, groundwater flow directions.

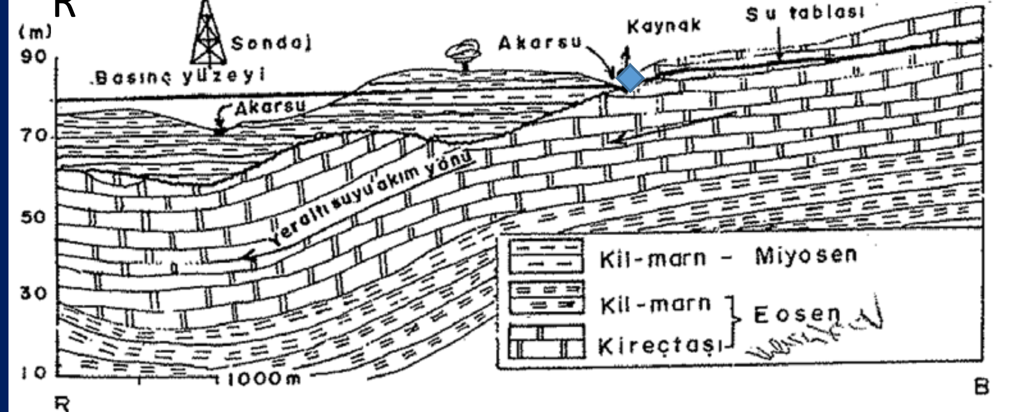




Eosen kireçtaşı
Miyosen kil-marn
Sondaj yeri
Miyosen - kil-marn
Eosen kireçtaşı
R N M L K H G F E D C B
0 1000 m.
Yatay tabaka
Tab, doğrultu eğimi
R B: Kesit doğrultusu
Gözlem kuyusu ve su kotu
A A' Miyosen Eosen dokanağı
Bu hattın K'i su tablası
G'i basınç yüzeyi haritasıdır
86 Eş su yükselti eğrisi
Yeraltı suyu akım yönü

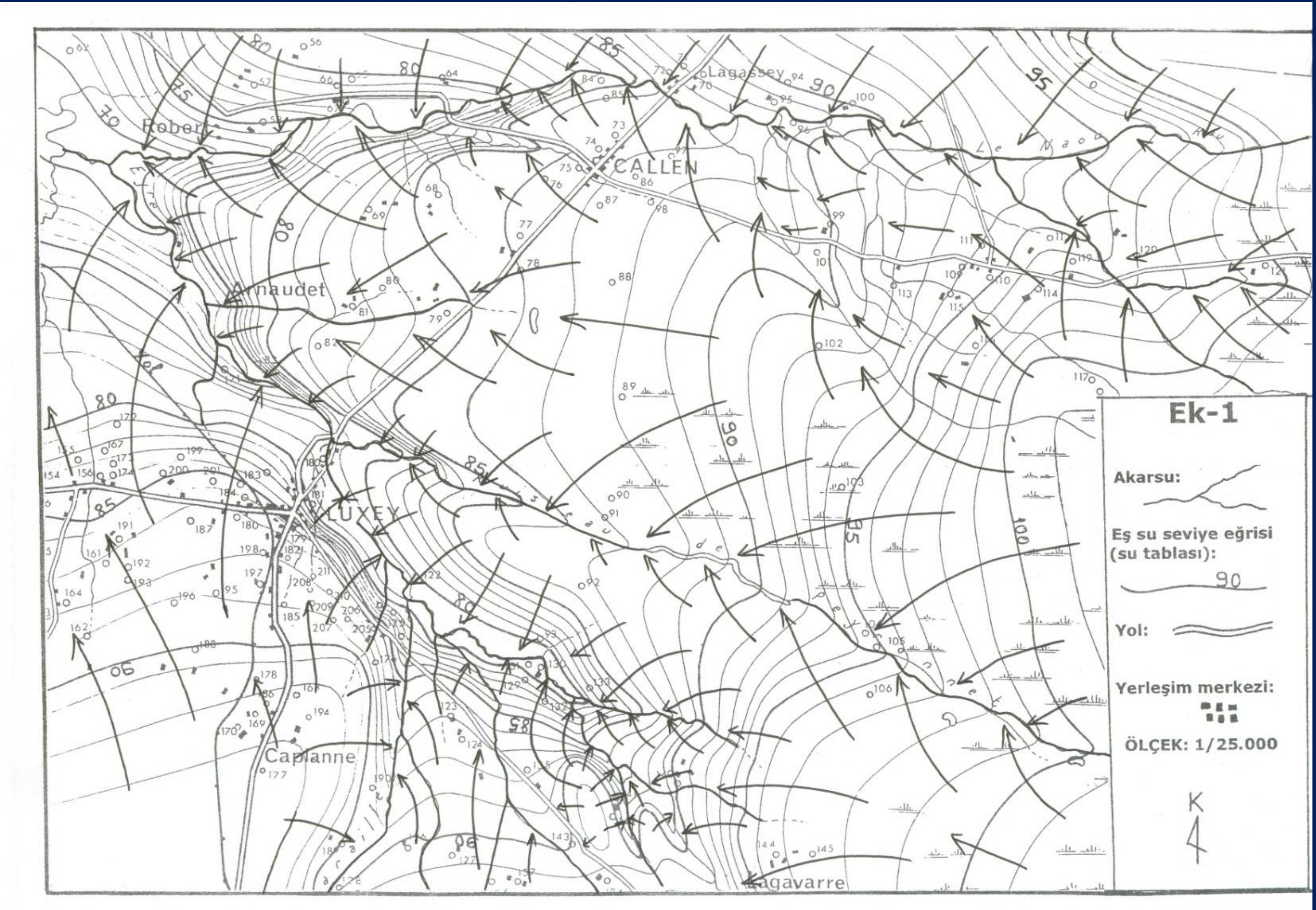
B

R

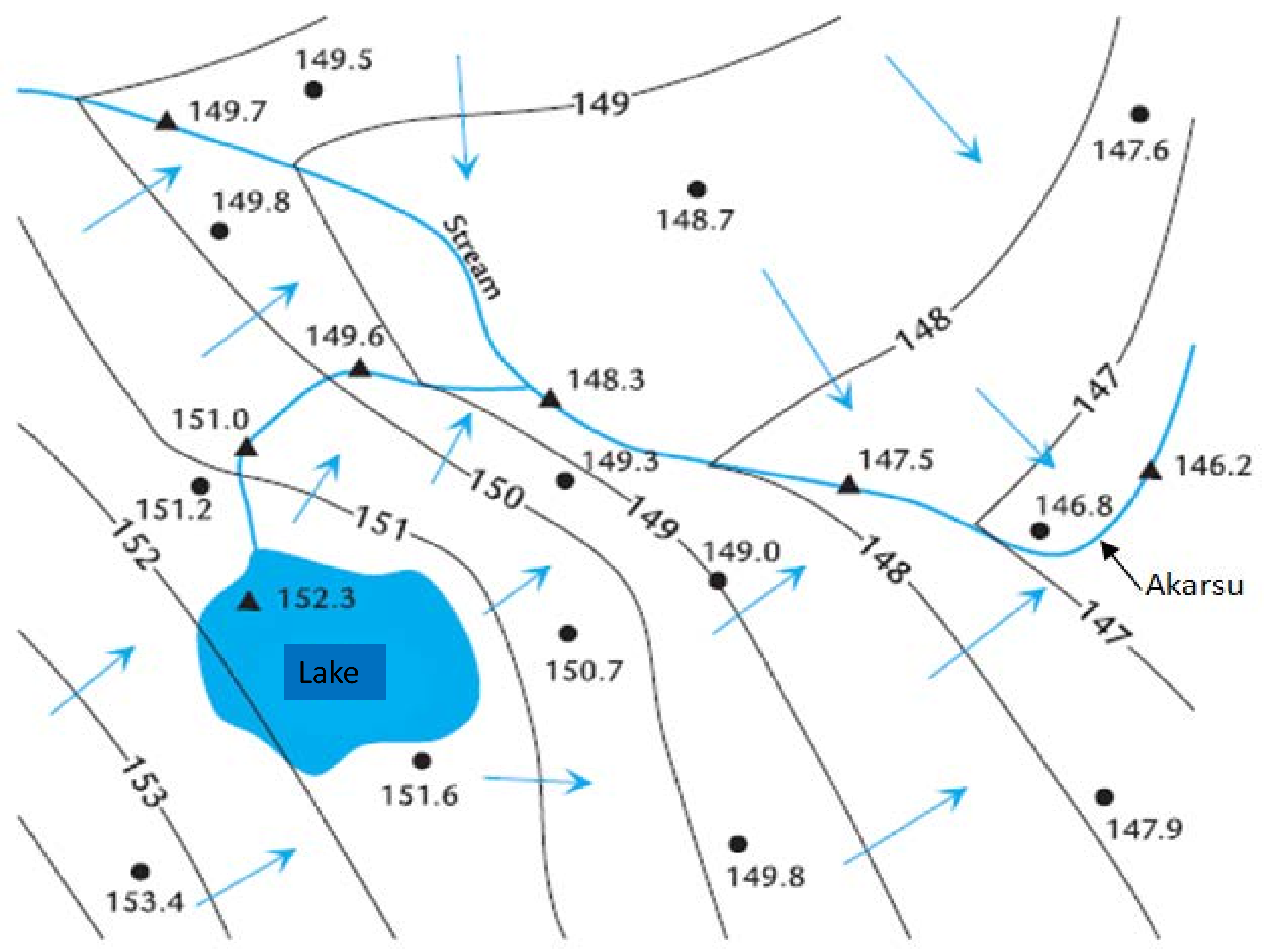


(m)
90
70
50
30
10
Sondaj
Basınç yüzeyi
Akarsu
Akarsu
Kaynak
Su tablası
Yeraltı suyu akım yönü
1000m
R
B
Kil-marn - Miyosen
Kil-marn
Kireçtaşı
Eosen

Water table contour map



Water table contour map



Streamflow and groundwater

When a stream channel is in direct contact with an unconfined aquifer, the stream may recharge the groundwater or receive discharge from the groundwater, depending on the relative levels.
A gaining stream is one receiving groundwater discharge
A losing stream is one recharging groundwater

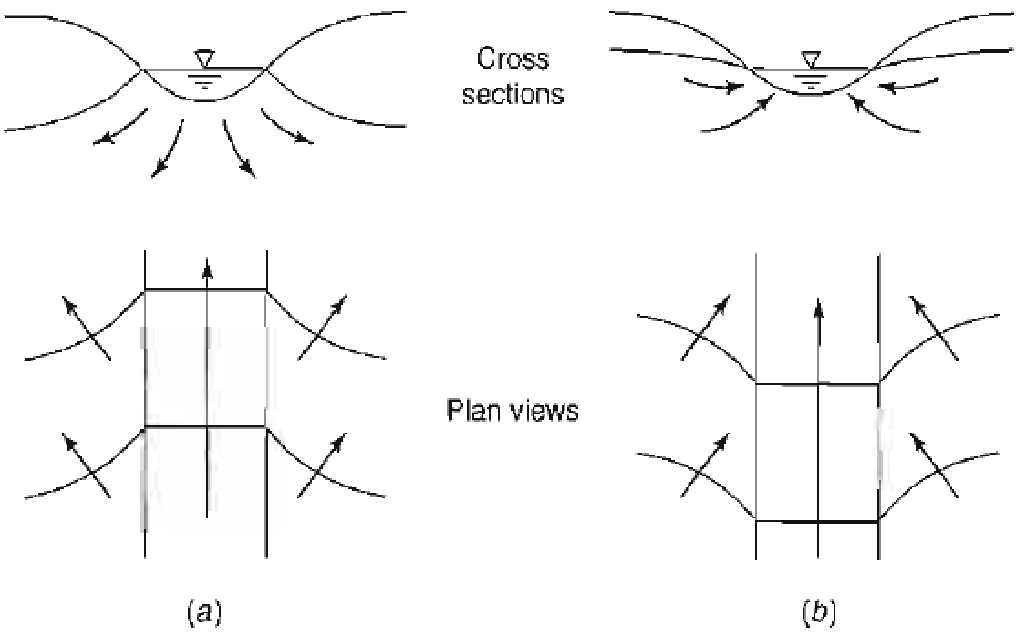


Figure 6.2.2. Water table contours and groundwater flow directions in relation to stream stages. (*a*) Losing stream (*b*) Gaining stream

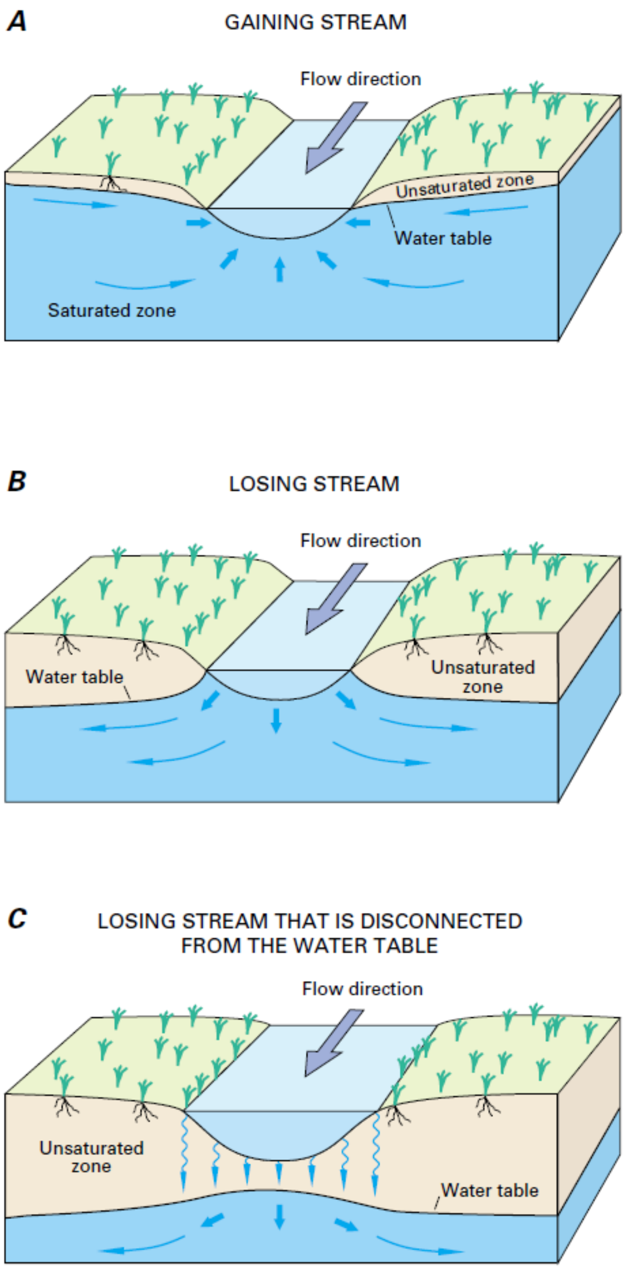


Figure 12. Interaction of streams and ground water. (Modified from Winter and others, 1998.)

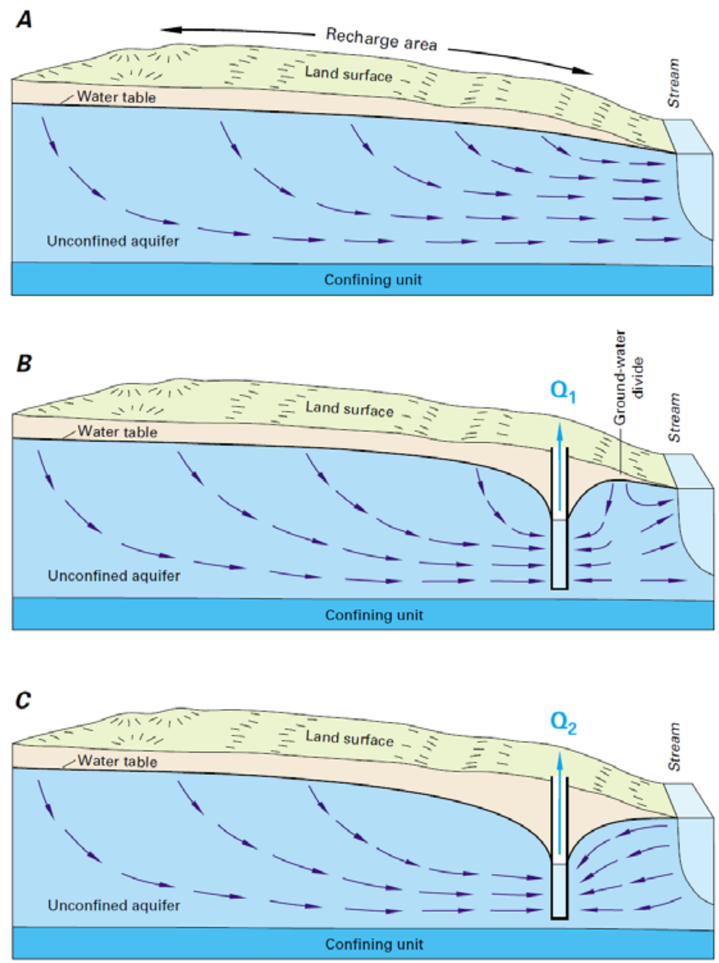
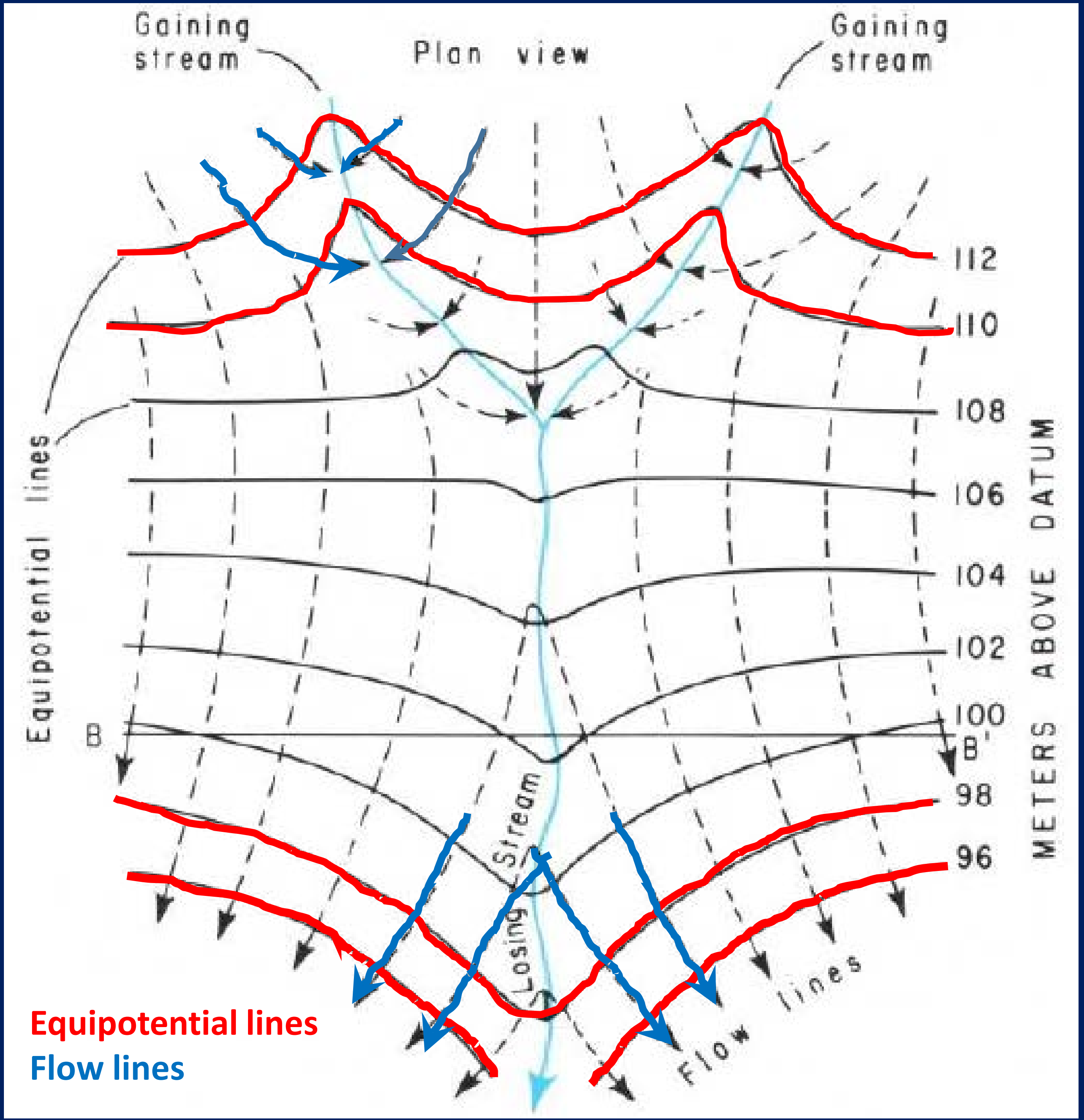
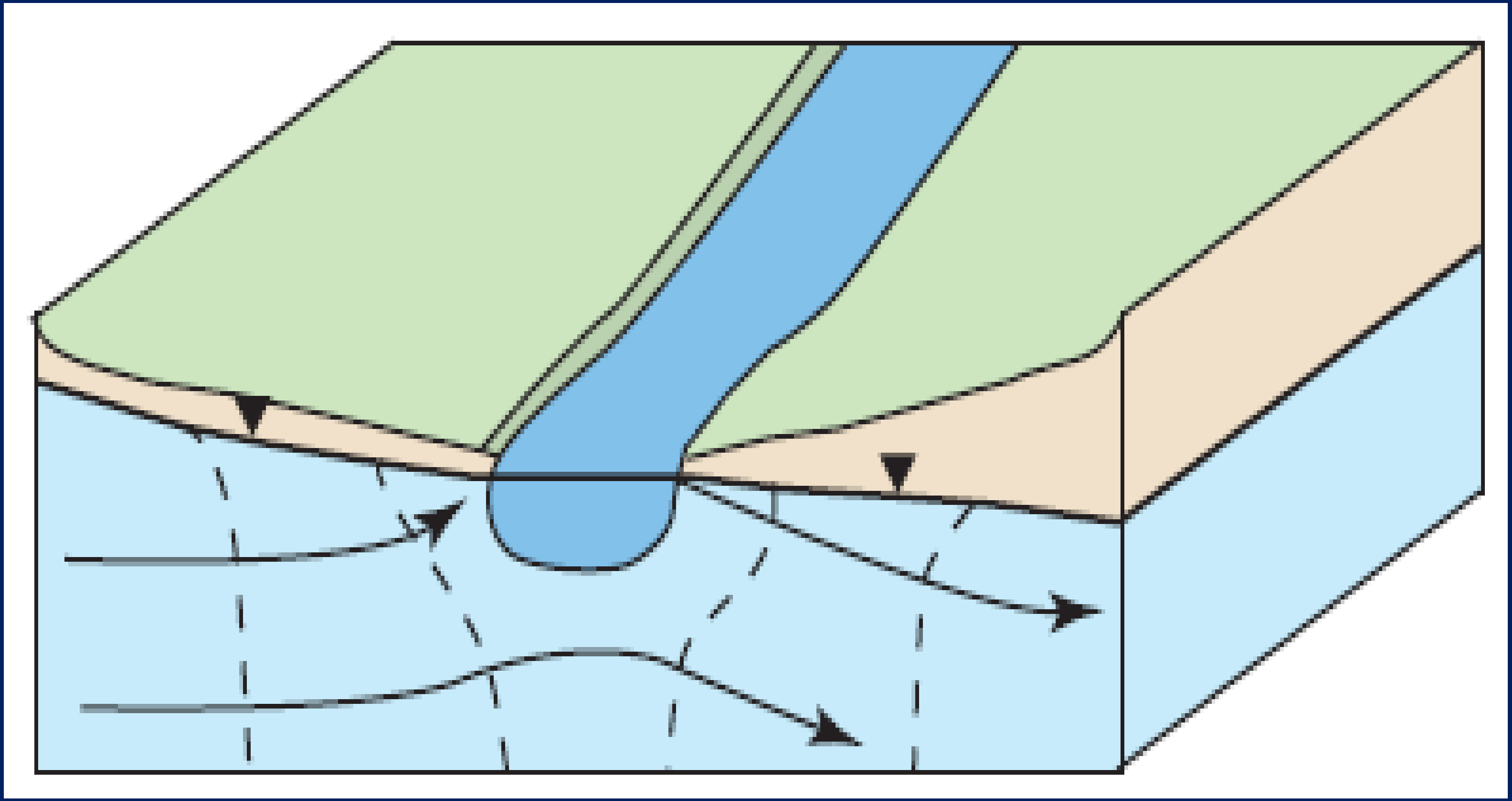


Figure 13. Effects of pumping from a hypothetical ground-water system that discharges to a stream. (Modified from Heath, 1983.)

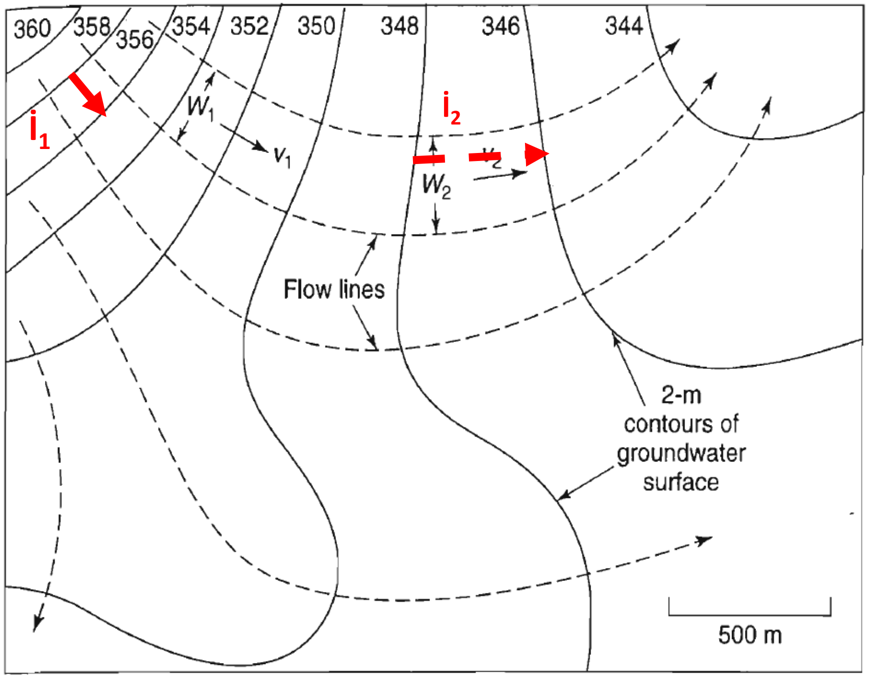


Gaining stream
Plan view
Gaining stream
112
110
108
106
104
102
100
98
96
METERS ABOVE DATUM
Equipotential lines
B
B'
Losing Stream
Flow lines
Equipotential lines
Flow lines

Gaining stream on one side, losing stream on the other side.



If contours are located apart, hydraulic gradient is low, hydraulic conductivity is high



i_1=

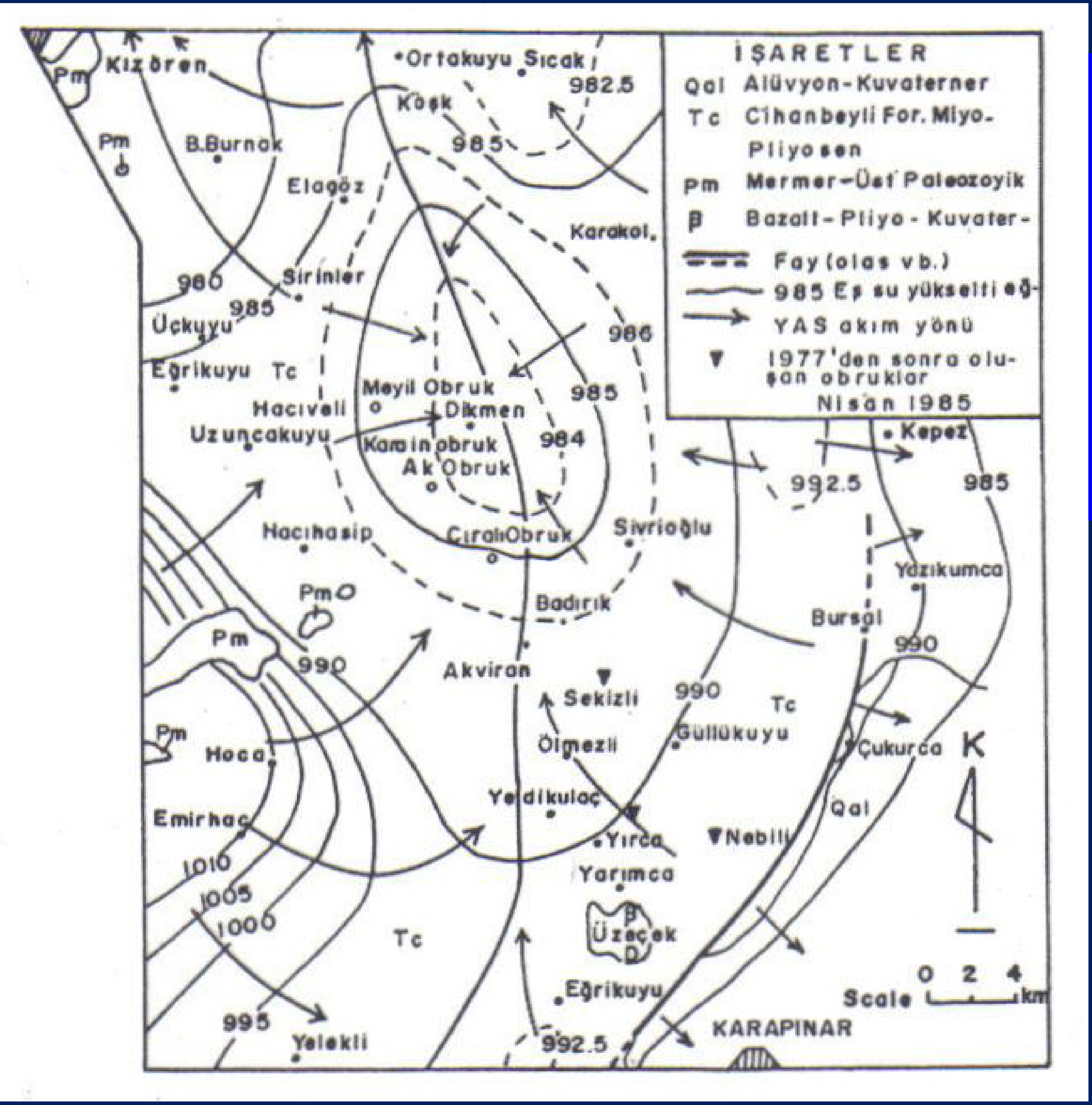
i_2=

Characteristics	Change	Hydraulic gradient (i)	Distance between equipotential lines
Hydraulic conductivity (K)	Increase		
	Decrease		
Discharge (Q)	Increase		
	Decrease		
Cross-sectional area (A)	Broadens		
	Narrows		

Figure 3.6.5. Contour map of a groundwater surface showing flow lines.

Characteristics	Change	Hydraulic gradient (i)	Distance between equipotential lines
Hydraulic conductivity (K)	Increase	Decreases	Increases
	Decrease	Increases	Decreases
Discharge (Q)	Increase	Increases	Decreases
	Decrease	Decreases	Increases
Cross-sectional area (A)	Broadens	Decreases	Increases
	Narrows	Increases	Decreases

$$\frac{K_1}{K_2} = \frac{i_2}{i_1}$$



Çıralı swallowhole,
Konya

DARCY'S LAW IN THREE DIMENSIONS

In 3-D flow, hydraulic head is a function of all three-space coordinates, i.e., h=h(x, y, z). In three dimensions, specific discharge is a vector with components q_x, q_y and q_z. For an isotropic medium,

$$q_x = -K\frac{\partial h}{\partial x}, \qquad q_y = -K\frac{\partial h}{\partial y}, \qquad q_z = -K\frac{\partial h}{\partial z}$$

In vector notation: $\vec{q} = -K\nabla h$ where $\nabla h = \text{gradient vector of the head} = \left(\frac{\partial h}{\partial x}\vec{i} + \frac{\partial h}{\partial y}\vec{j} + \frac{\partial h}{\partial z}\vec{k}\right)$

$\vec{i}, \vec{j}, \vec{k}$ are unit vectors in x, y and z coordinate directions, respectively.

$$\begin{bmatrix} q_x \\ q_y \\ q_z \end{bmatrix} = -\begin{bmatrix} K_{xx} & K_{xy} & K_{xz} \\ K_{yx} & K_{yy} & K_{yz} \\ K_{zx} & K_{zy} & K_{zz} \end{bmatrix}\begin{bmatrix} \partial h/\partial x \\ \partial h/\partial y \\ \partial h/\partial z \end{bmatrix}$$

Hydraulic conductivity tensor

3-D linear parabolic partial differential equation, the solution of which yields h(x, y, z, t) in a heterogeneous, anisotropic, confined aquifer.

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right)=S_s \frac{\partial h}{\partial t}$$

For homogeneous but anisotropic aquifer

$$K_x\left(\frac{\partial^2 h}{\partial x^2}\right)+K_y\left(\frac{\partial^2 h}{\partial y^2}\right)+K_z\left(\frac{\partial^2 h}{\partial z^2}\right)=S_s \frac{\partial h}{\partial t}$$

$$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}+\frac{\partial^2 h}{\partial z^2}=\frac{S_s}{K}\frac{\partial h}{\partial t}$$

For horizontal flow (x, y)

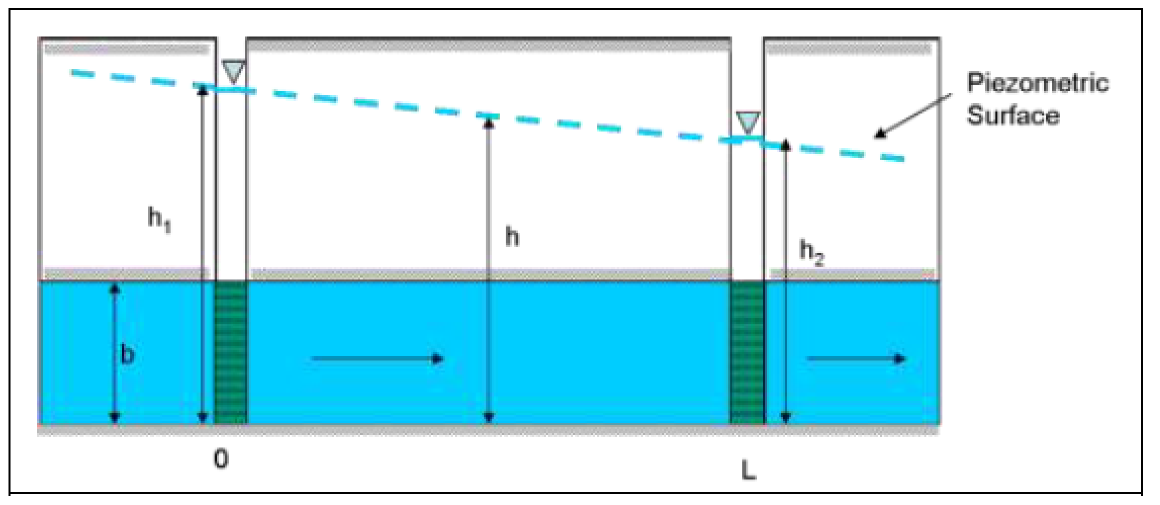
$$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}=\frac{S_s b}{K b}\frac{\partial h}{\partial t}=\frac{S}{T}\frac{\partial h}{\partial t}$$

Laplace's Equation- solution of which gives h= h(x,y,z) in an isotropic and homogenous confined aquifer

$$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}+\frac{\partial^2 h}{\partial z^2}=0$$

For steady-state flow homogeneous and isotropic aquifer

1-D Steady state flow in a confined aquifer (Homogeneous and Isotropic)



$$\frac{d^2 h}{dx^2}=0$$

Integrate twice wrt x

$$\frac{dh}{dx}=c_1 \quad \rightarrow \quad h=c_1 x+c_2$$

$$\text{when } x=0, \quad h=h_1 \quad \rightarrow \quad c_2=h_1$$

$$\text{when } x=L, \quad h=h_2 \quad \rightarrow \quad h_2=c_1L+c_2 \quad \rightarrow \quad h_2=c_1L+h_1 \quad \rightarrow \quad c_1=\frac{h_2-h_1}{L}$$

$$\frac{dh}{dx} = c_1 \quad \rightarrow \quad h = c_1 x + c_2$$

$$\text{when } x = 0, \quad h = h_1 \quad \rightarrow \quad c_2 = h_1$$

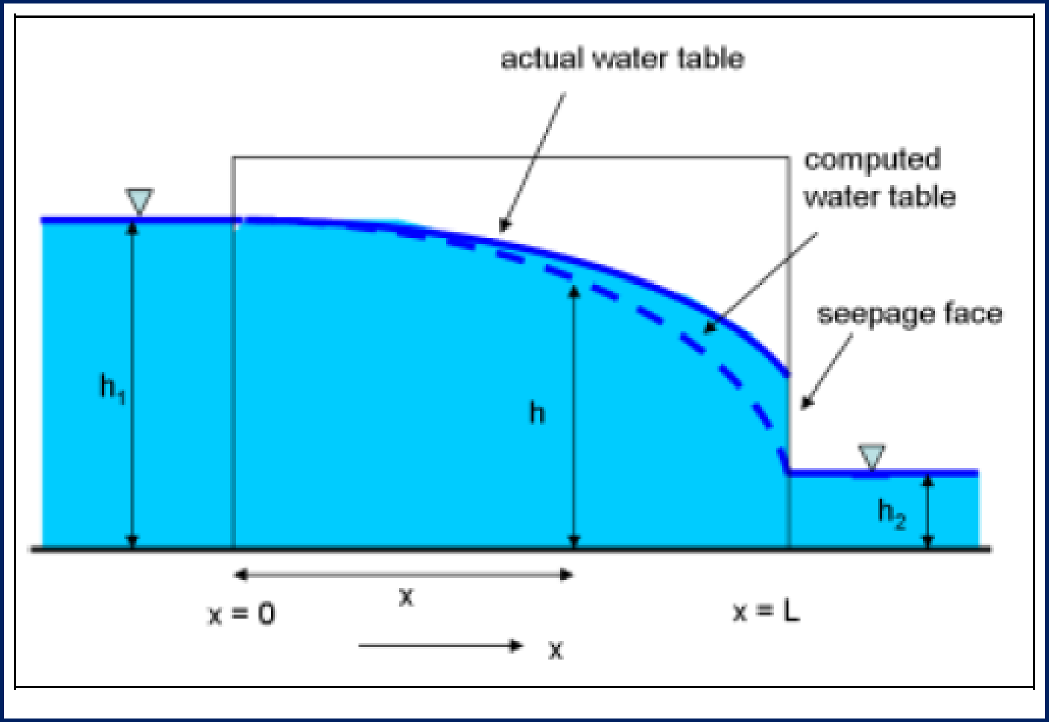
$$\text{when } x = L, \quad h = h_2 \quad \rightarrow \quad h_2 = c_1 L + c_2 \quad \rightarrow \quad h_2 = c_1 L + h_1 \quad \rightarrow \quad c_1 = \frac{h_2 - h_1}{L}$$

$$\boxed{h = \left(\frac{h_2 - h_1}{L}\right) x + h_1}$$

Head decreases linearly in a confined aquifer system

Discharge per unit width, q, is

$$q = -K.b\frac{dh}{dx} \quad \text{or} \quad q = -K.b\frac{h_2 - h_1}{L}$$



In a unconfined aquifer, gradient of the water table is not constant; it increases in the flow direction.

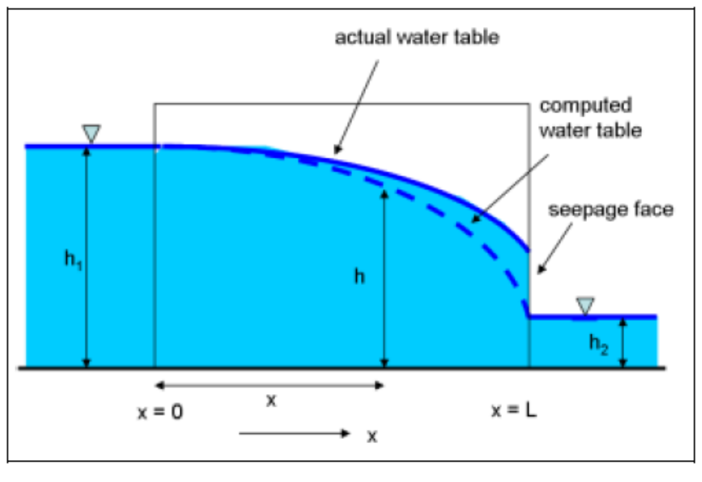
To derive an equation of flow in an unconfined aquifer Dupuit assumptions should be accepted.

1) Hydraulic gradient is equal to the slope of the water table
2) For small water table slopes, the flow lines are horizontal and the equipotential lines are vertical (i.e. Horizontal flow).

From Darcy's law, discharge per unit width

$$q = -K.h\frac{dh}{dx}$$

h: saturated thickness (i.e. height of the water table)

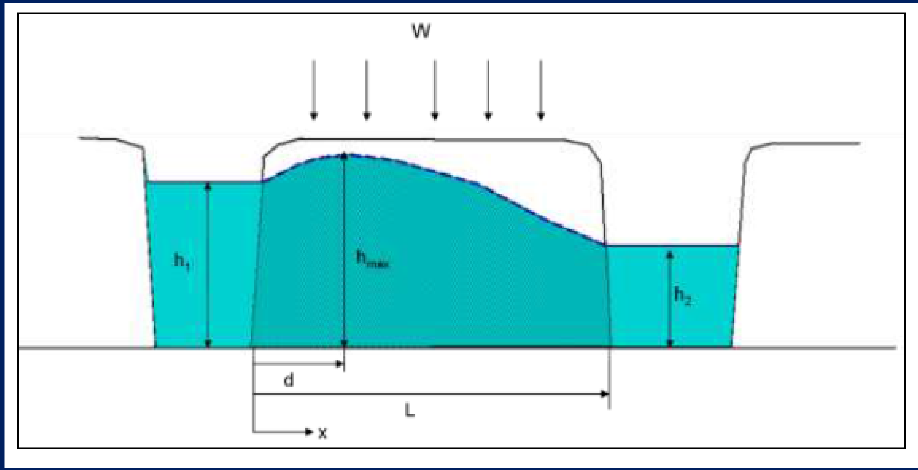


$$q\int_{x=0}^{x=L} dx = -K\int_{h=h_1}^{h=h_2} h.dh \rightarrow$$

$$qx\,|_0^L = -K\frac{h^2}{2}\,|_{h_1}^{h_2}$$

$$qL = -\frac{K}{2}\left(h_2^2 - h_1^2\right)$$

$$q = -\frac{K}{2L}\left(h_2^2 - h_1^2\right)$$



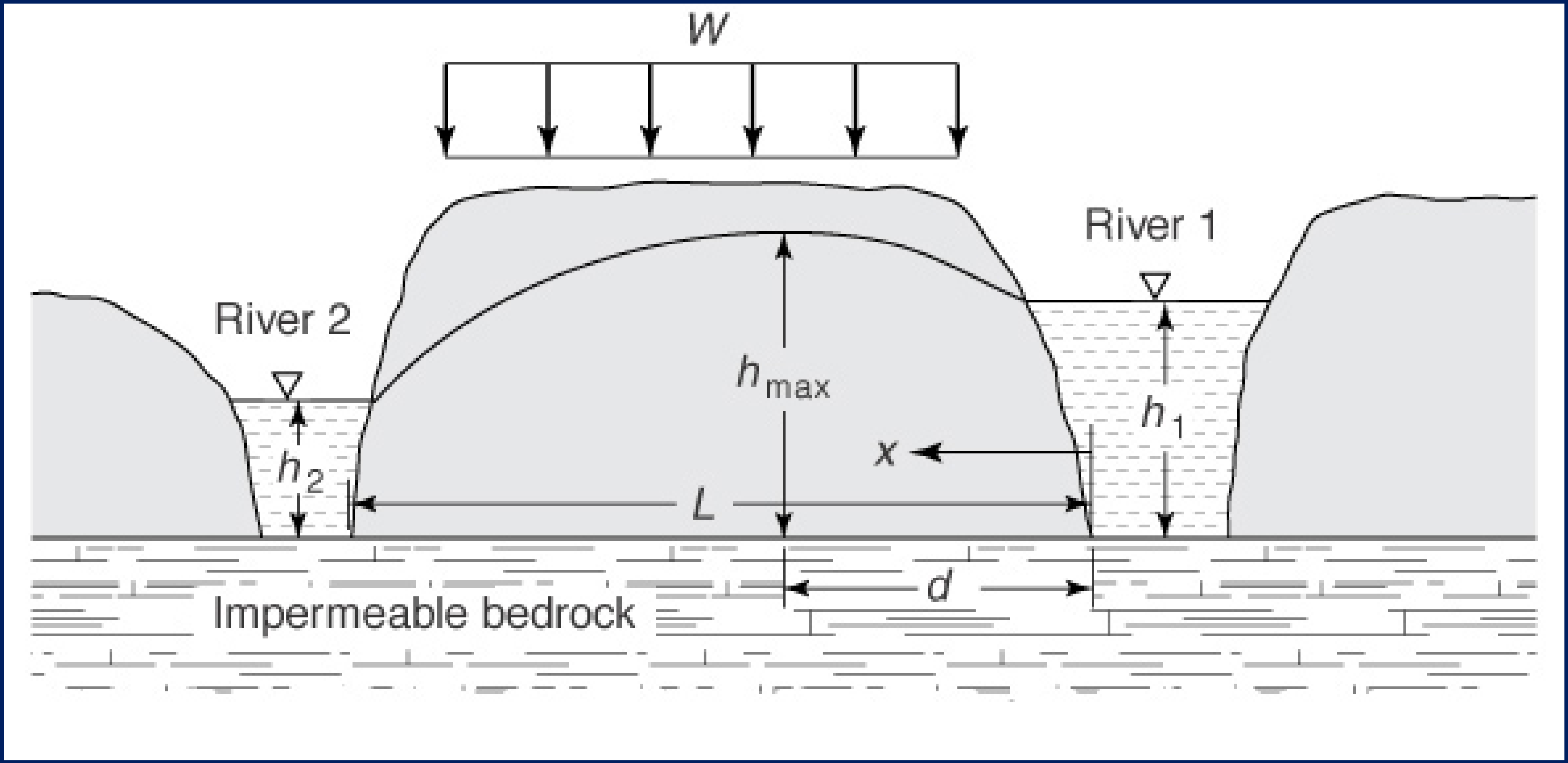
$$q_x = K\frac{(h_1^2 - h_2^2)}{2L} - W(\frac{L}{2} - x)$$

discharge per unit width at any section x distance from the origin.

At the water table divide q_x=0, x=d

Distance to water divide from the origin

$$d = \frac{L}{2} - \frac{K}{W}\frac{(h_1^2 - h_2^2)}{2L}$$



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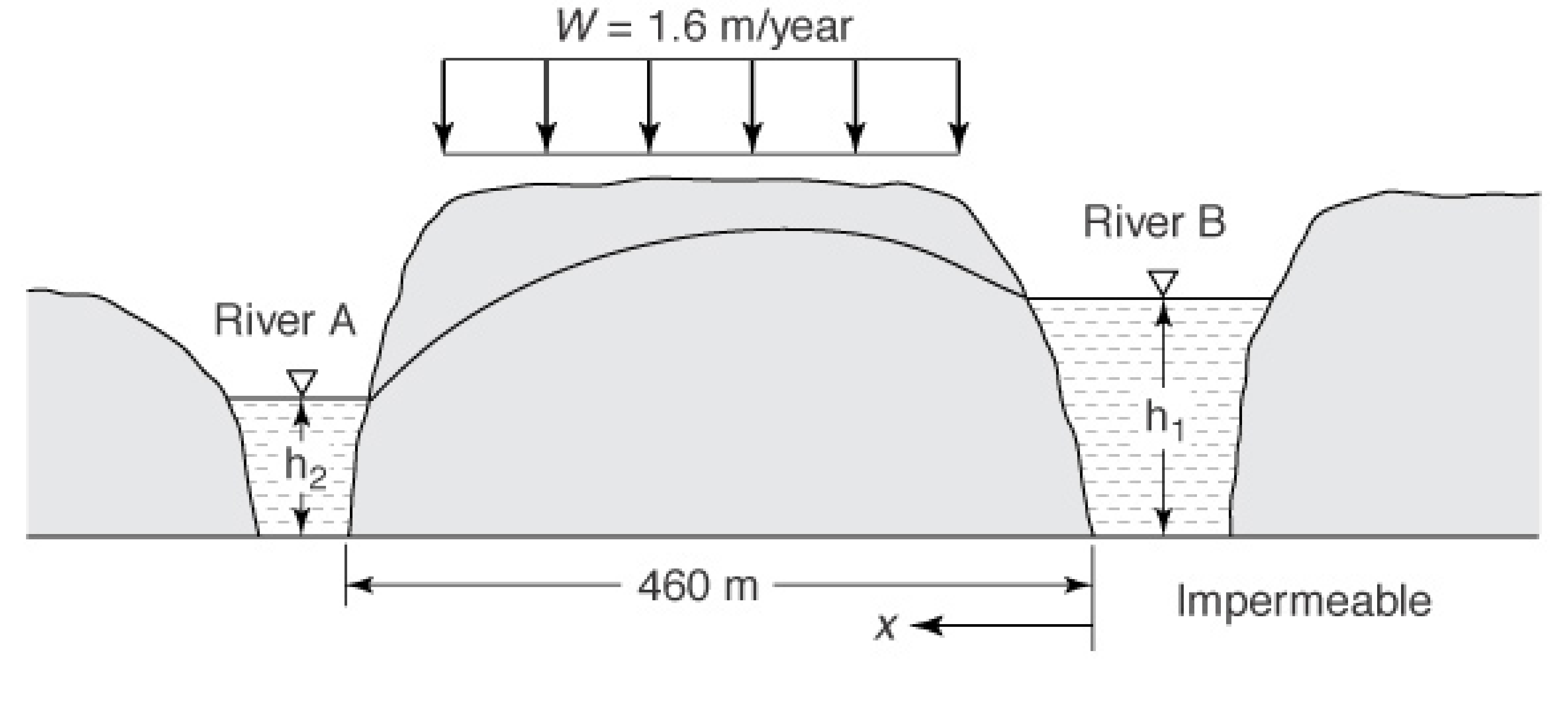

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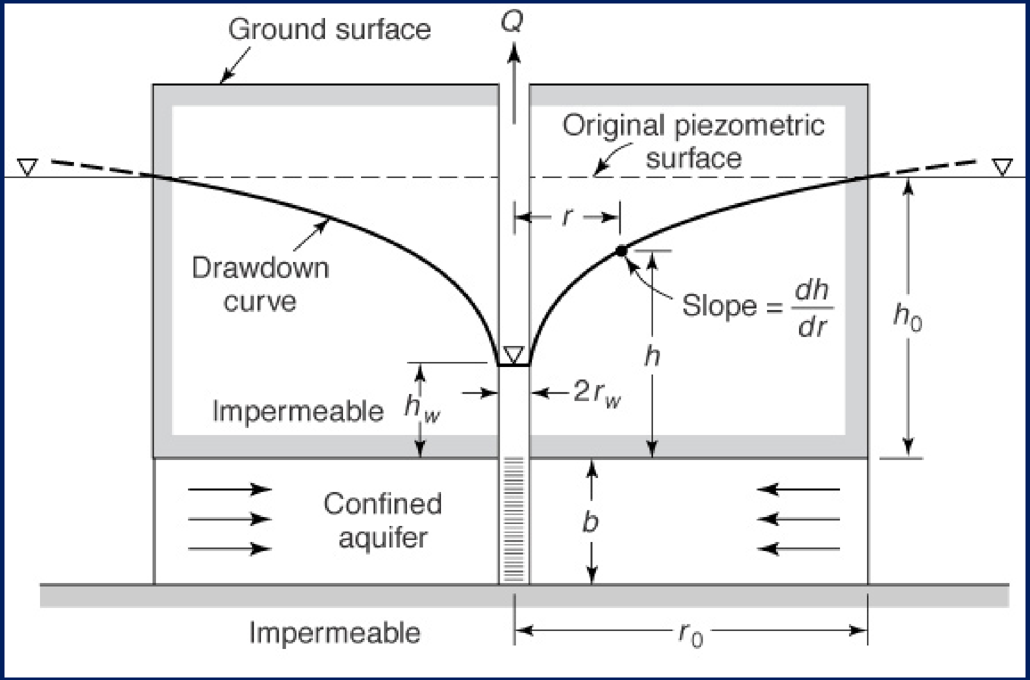
Well Hydraulics in steady state condition

Primary goal of groundwater resources evaluation should be the prediction of hydraulic-head drawdowns in the aquifer under a proposed pumping scheme. The theoretical response of idealized aquifers to pumping can be examined by different methods. Hydraulic characteristics of aquifers (T, K, S) may be determined by means of pumping tests. They involve removal or the addition of water to a well and observation of reaction of the aquifer to change.

Steady-state or equilibrium methods: Yields values of transmissivity and related hydraulic conductivity
Non-steady state or nonequilibrium methods: Yields values of transmissivity, hydraulic conductivity, and storage coefficient.

Steady-state Radial Flow to a Well in a Confined Aquifer



Q= A*v

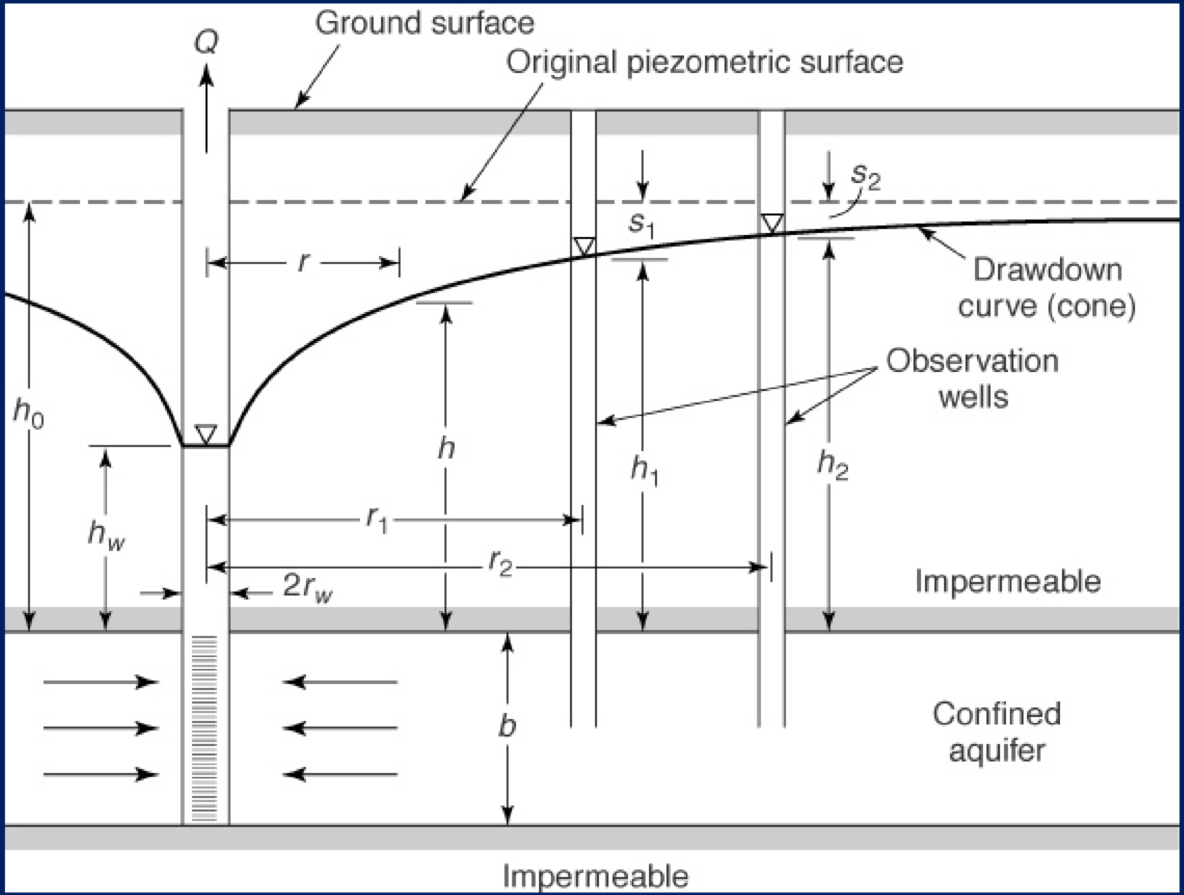
$$=-2\pi rbK\frac{dh}{dl}$$

Curved surface area of the cylinder

Rearrange, integrate for the boundary conditions at the well $h=h_w$, $r=r_w$
$h=h_0$, $r=r_0$

$$Q = 2\pi Kb\frac{h - h_w}{\ln(\frac{r}{r_w})}$$

Steady-state Radial Flow to a Well in a Confined Aquifer



$$Q = 2\pi K b \frac{h - h_w}{\ln(\frac{r}{r_w})}$$

T= Kb

$$T = \frac{Q}{2\pi(h_2 - h_1)} ln \frac{r_2}{r_1}$$

Drawdown change must be negligible.

The observation wells should be located close enough to the pumping well

We assume that;

1. The well is pumped at a constant rate and fully penetrates the aquifer.
2. The aquifer is homogeneous and isotropic, is of uniform thickness, and is of infinite areal extent; that the well penetrates the entire aquifer; and that initially the piezometric surface is nearly horizontal.
3. Water is initially removed from the storage upon a decline of head
4. The flow in the aquifer obeys Darcy's law

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If there are 4 observation wells, you may choose pairs of wells and apply the Thiem equation for each pair. Including pumping well, PW, you will have 10 pairs of calculations of Transmissivity. An Average value of T can then be calculated.

$$T = \frac{Q}{2\pi(s_1 - s_2)} \ln(r_2/r_1)$$

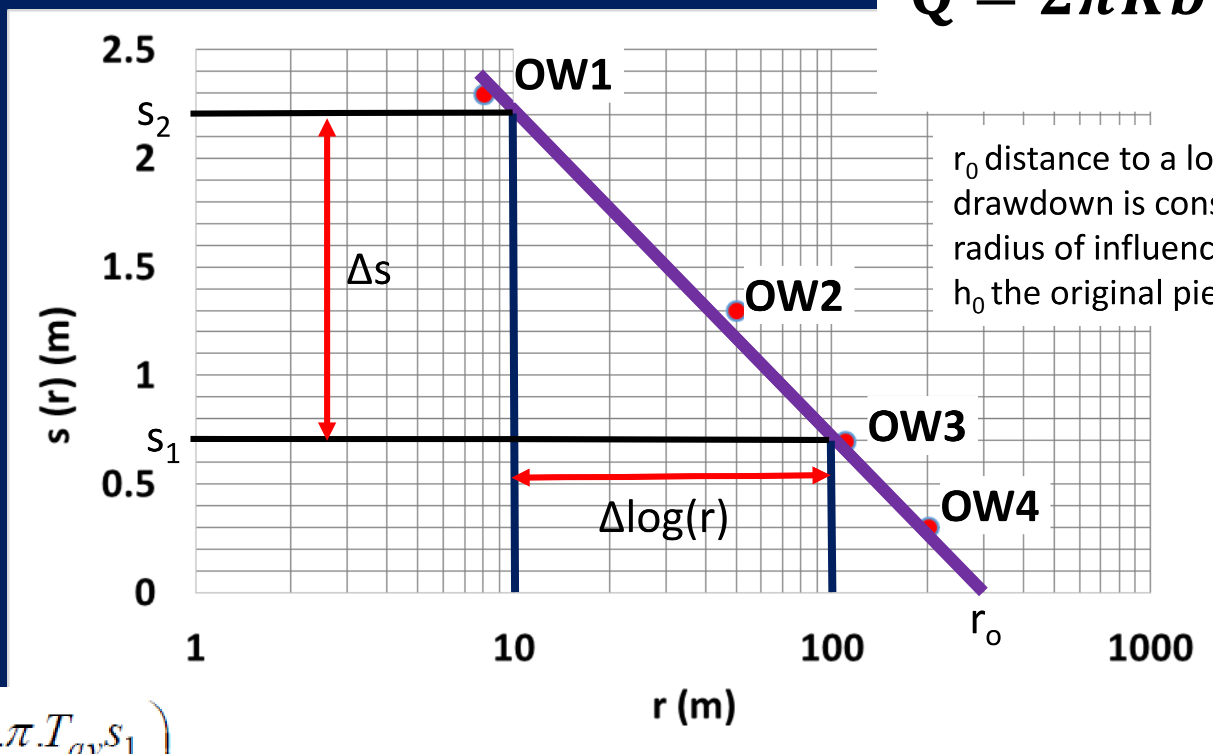
	OW1	OW2	OW3	OW4
PW	x	x	x	x
OW1		x	x	x
OW2			x	x
OW3				x

When there is more than one well, a plot of s(r) versus r on a semi-log paper should yield a straight line from the slope of which we obtain T.

$$T = \frac{Q}{2\pi(s_1 - s_2)} \ln\left({}^{r_2}/_{r_1}\right) = 2.3 \frac{Q}{2\pi(s_1 - s_2)} log_{10}\left({}^{r_2}/_{r_1}\right)$$

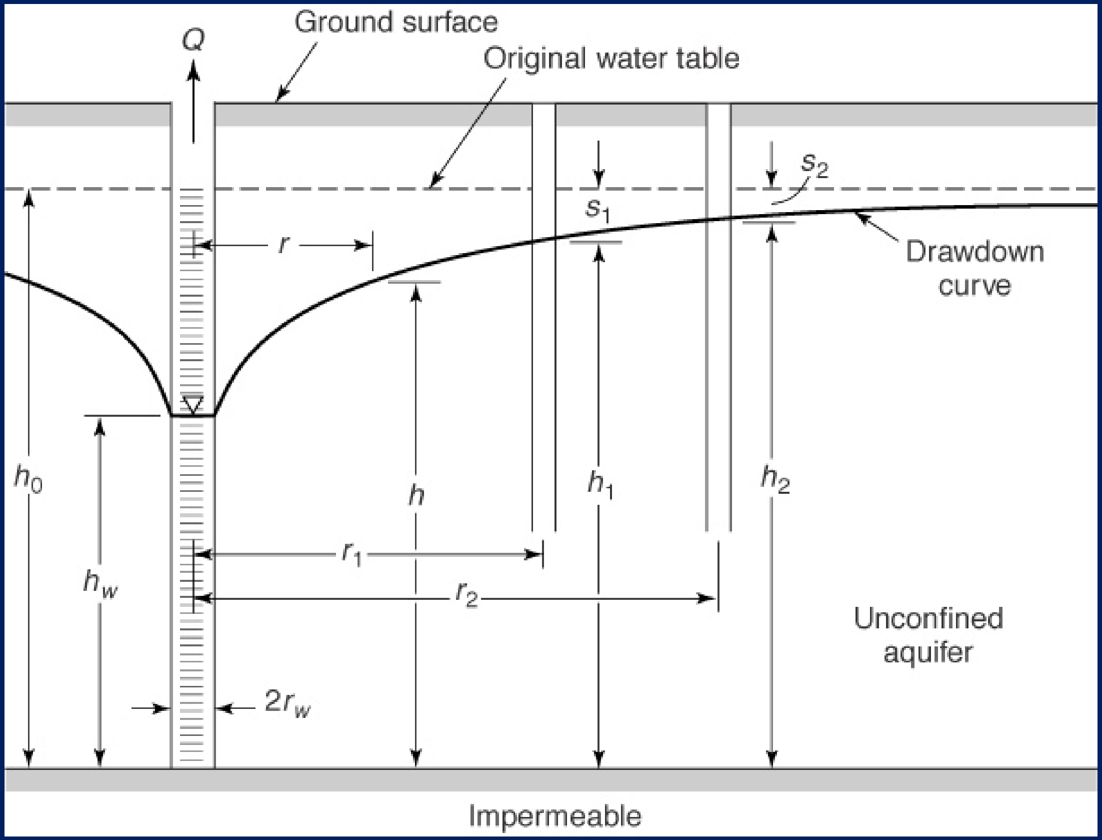
$$\mathrm{Q} = 2\pi K b \frac{2\pi(h_0 - h_w)}{\ln\left({}^{r_0}/_{r_w}\right)}$$

r_0 distance to a location at which drawdown is considered negligible = radius of influence of the pumping well
h_0 the original piezometric height



$$r_0 = r_1 . \exp\left(\frac{2.\pi.T_{av}.s_1}{Q}\right)$$

Steady-state Radial Flow to a Well in an unconfined Aquifer



$$Q=-2\pi rKh\frac{dh}{dr}$$

Integrate between the limits $h=h_w$ at $r=r_w$ $h=h_o$ at $r=r_o$

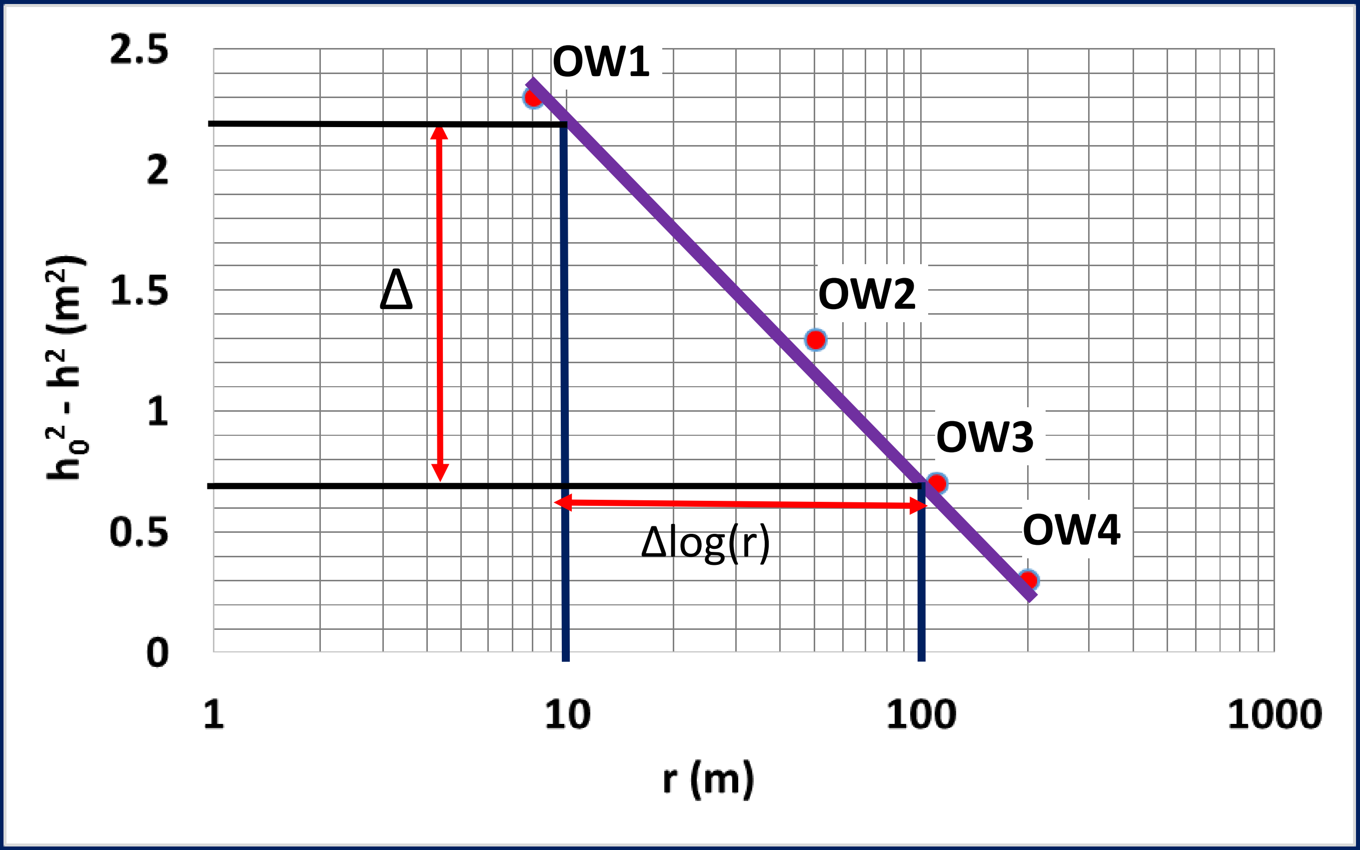
$$Q=\pi K\frac{{h_0}^2-{h_w}^2}{\ln(\frac{r_0}{r_w})}$$

$$Q=\pi K\frac{{h_2}^2-{h_1}^2}{\ln(\frac{r_2}{r_1})}$$

Hydraulic conductivity can be estimated from Thiem Equation as:

$$K=\frac{Q}{\pi({h_2}^2-{h_1}^2)}\ln(\frac{r_2}{r_1})$$

$$K = \frac{2.3Q}{\pi \Delta s} \qquad r_0 = r.\exp\left(\frac{\pi K a v}{Q}\left(h_0^2 - h_{(r)}^2\right)\right)$$



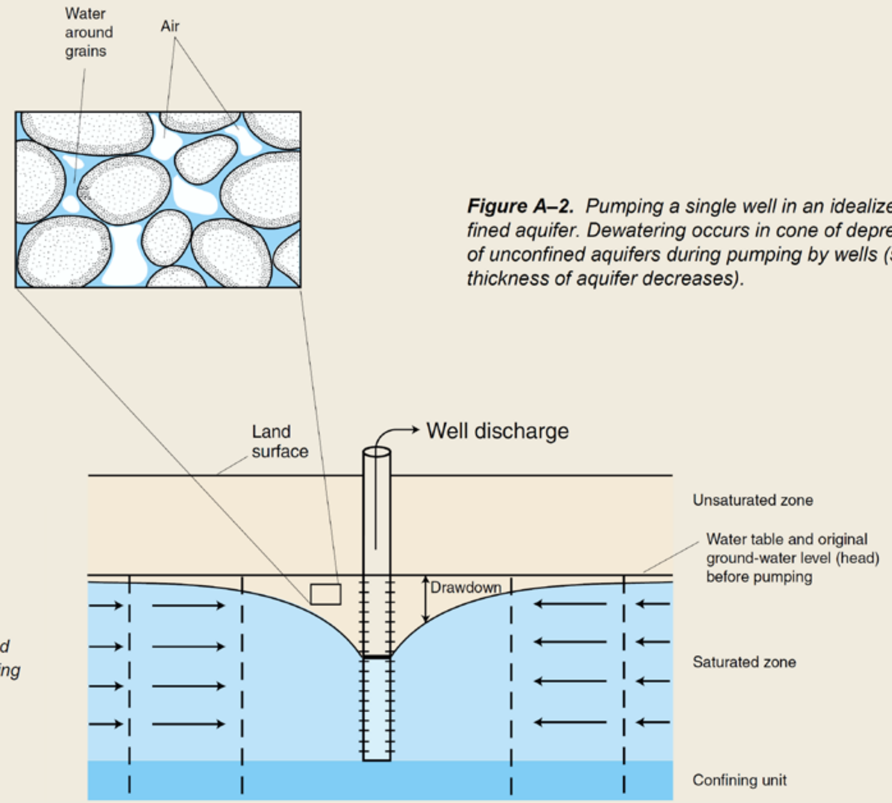


Figure A–2. *Pumping a single well in an idealized unconfined aquifer. Dewatering occurs in cone of depression of unconfined aquifers during pumping by wells (saturated thickness of aquifer decreases).*

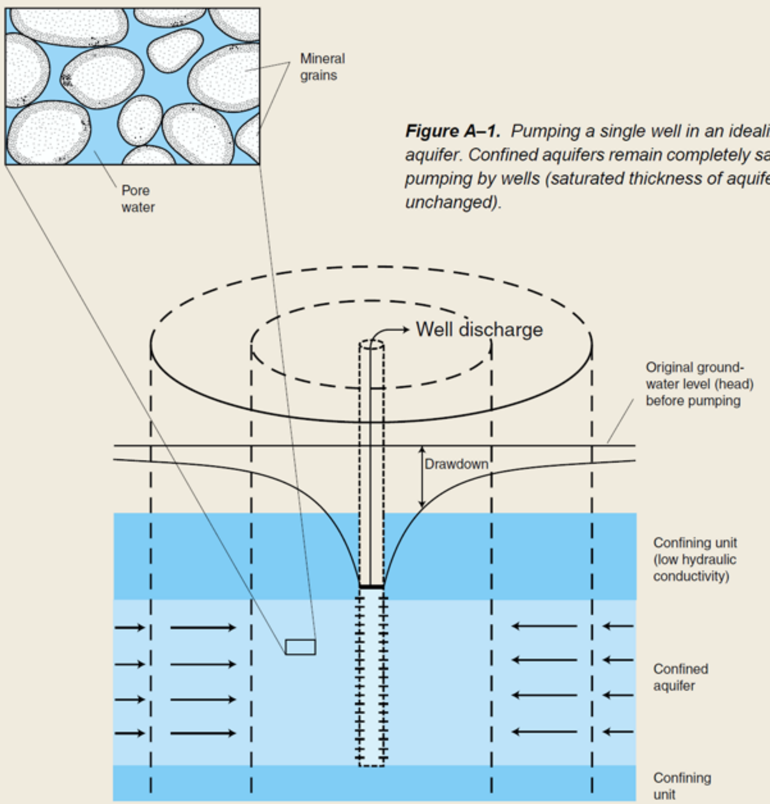


Figure A–1. *Pumping a single well in an idealized confined aquifer. Confined aquifers remain completely saturated during pumping by wells (saturated thickness of aquifer remains unchanged).*

Unsteady State (Transient) Radial flow to a Well in a Nonleaky Confined Aquifer

Assumptions:

1) Darcy's law is applicable
2) Water is discharged instantenously from storage upon a decline of head
3) Aquifer is homogeneous, isotropic and has a constant thickness
4) Horizontal aquifer of infinite areal extent
5) Wells fully penetrate the aquifer and flow is horizontal

Theis (1935) method

$$s = h_0 - h = \frac{Q}{4\pi T}\int_u^\infty \frac{e^{-u}du}{u} = \frac{Q}{4\pi T}W(u)$$

$$W(u) = -0.5772 - lnu + u - \frac{u^2}{2x2!} + \frac{u^3}{3x3!} - \frac{u^4}{4x4!} + \cdots \ldots$$

W(u) well function

$$u = \frac{r^2 S}{4Tt}$$

Q= constant pumping rate

h= hydraulic head at time t since pumping began

h_0= hydraulic head before the start of pumping

s= drawdown at point r at time t

r= radial distance from the pumping well to the observation well

T= Transmissivity

S= storativity or storage coefficient

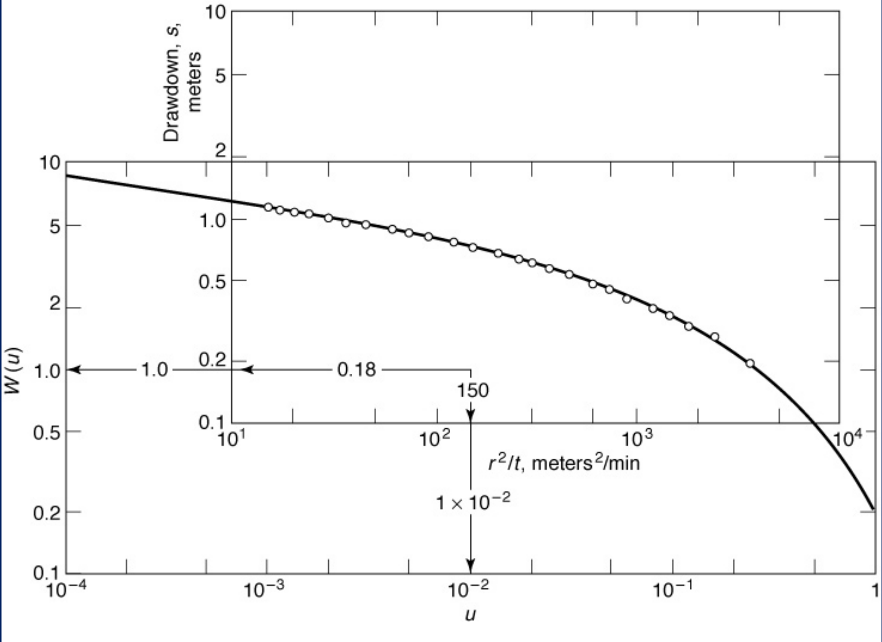
Theis suggested an approximate solution for S and T based on a graphic method of superposition.

A plot on logarithmic paper of W(u) versus u, known as a type curve is prepared.

Values of drawdowns are plotted against values of r^2/t on logarithmic paper of the same size and scale as for the type curve.

The observed time-drawdown data are superimposed on the type curve, keeping the coordinate axes of the two curves parallel, and adjusted until a position is found by trial whereby most of the plotted points of the observed data fall on a segment of the type curve.

Any convenient point is then selected, and the coordinates of this match point are recorded.



$$s = \frac{Q}{4\pi T} W(u) \qquad u = \frac{r^2 S}{4Tt}$$

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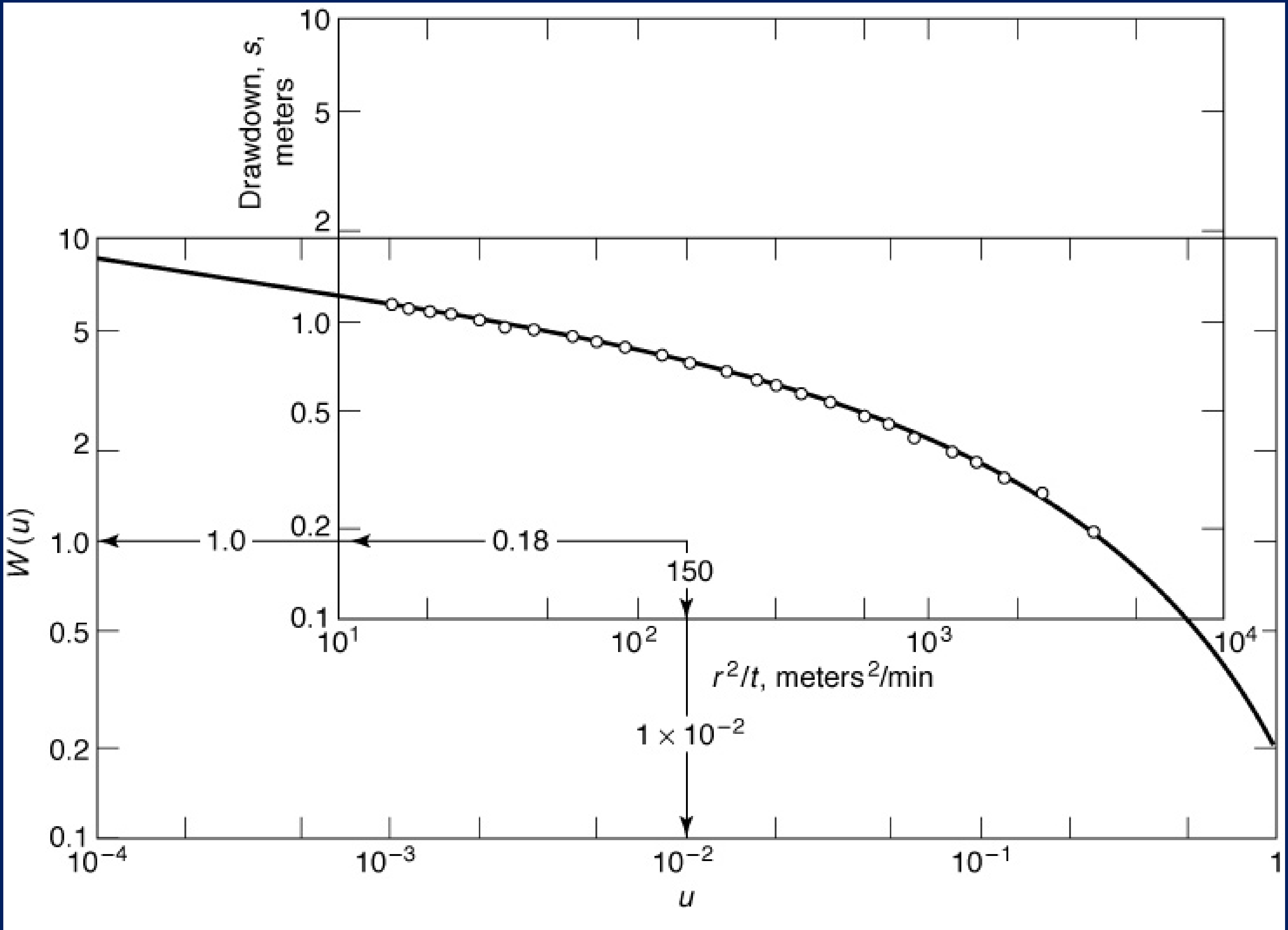
$$s = \frac{Q}{4\pi T} W(u) \qquad u = \frac{r^2 S}{4Tt}$$

Table 4.4.1 Values of $W(u)$ for Values of u

u	1.0	2.0	3.0	4.0	5.0	6.0	7.0	8.0	9.0
$\times 1$	0.219	0.049	0.013	0.0038	0.0011	0.00036	0.00012	0.000038	0.000012
$\times 10^{-1}$	1.82	1.22	0.91	0.70	0.56	0.45	0.37	0.31	0.26
$\times 10^{-2}$	4.04	3.35	2.96	2.68	2.47	2.30	2.15	2.03	1.92
$\times 10^{-3}$	6.33	5.64	5.23	4.95	4.73	4.54	4.39	4.26	4.14
$\times 10^{-4}$	8.63	7.94	7.53	7.25	7.02	6.84	6.69	6.55	6.44
$\times 10^{-5}$	10.94	10.24	9.84	9.55	9.33	9.14	8.99	8.86	8.74
$\times 10^{-6}$	13.24	12.55	12.14	11.85	11.63	11.45	11.29	11.16	11.04
$\times 10^{-7}$	15.54	14.85	14.44	14.15	13.93	13.75	13.60	13.46	13.34
$\times 10^{-8}$	17.84	17.15	16.74	16.46	16.23	16.05	15.90	15.76	15.65
$\times 10^{-9}$	20.15	19.45	19.05	18.76	18.54	18.35	18.20	18.07	17.95
$\times 10^{-10}$	22.45	21.76	21.35	21.06	20.84	20.66	20.50	20.37	20.25
$\times 10^{-11}$	24.75	24.06	23.65	23.36	23.14	22.96	22.81	22.67	22.55
$\times 10^{-12}$	27.05	26.36	25.96	25.67	25.44	25.26	25.11	24.97	24.86
$\times 10^{-13}$	29.36	28.66	28.26	27.97	27.75	27.56	27.41	27.28	27.16
$\times 10^{-14}$	31.66	30.97	30.56	30.27	30.05	29.87	29.71	29.58	29.46
$\times 10^{-15}$	33.96	33.27	32.86	32.58	32.35	32.17	32.02	31.88	31.76



Drawdown, s, meters
10
5
2
1.0
0.5
0.2
0.1
10^1
10^2
10^3
10^4
r^2/t, meters^2/min
W(u)
10
5
2
1.0
0.5
0.2
0.1
1.0
0.18
150
1×10^{-2}
10^{-4}
10^{-3}
10^{-2}
10^{-1}
1
u

Advantages of unsteady state equation:

Only one observation well is required.

A shorter period of pumping is generally necessary.

No assumption of steady-state flow conditions is required.

A value for S can be determined.

Cooper- Jacob Method of Solution

For small values of r and large values of t, u is small.

$$u = \frac{r^2 S}{4Tt}$$

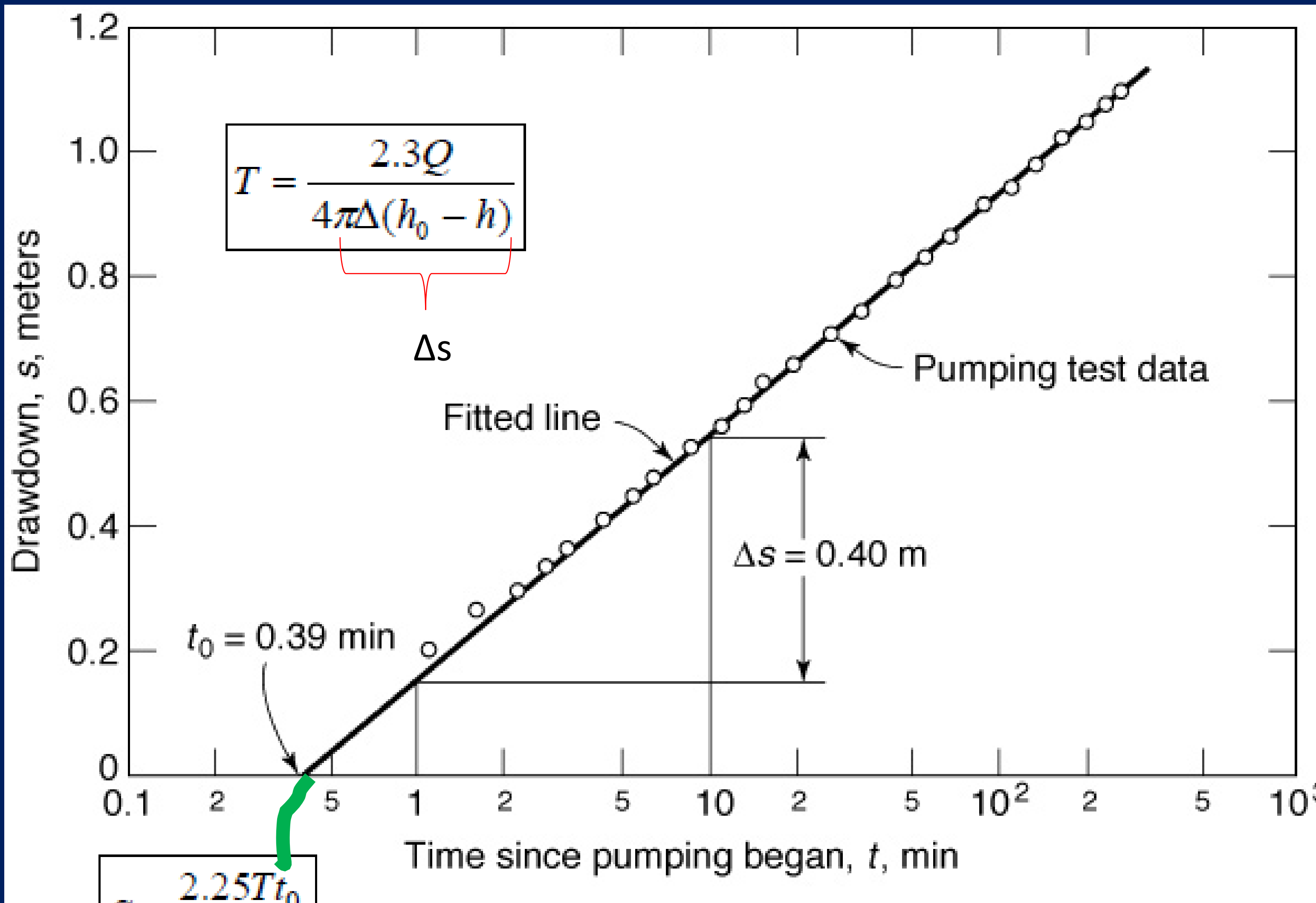
$$W(u) = -0.5772 - lnu + u \underbrace{- \frac{u^2}{2x2!} + \frac{u^3}{3x3!} - \frac{u^4}{4x4!} + \cdots \ldots}_{\text{negligible}}$$

$$s = \ h_0 - h = \frac{Q}{4\pi T} W(u) = \frac{2.3Q}{4\pi T} (log_{10} \frac{2.25Tt}{r^2 S})$$

For this equation to be valid

$u<0.01$

Cooper- Jacob Method of Solution
Time-Drawdown Data in a single observation well (r= constant)



$$S = \frac{2.25 T t_0}{r^2}$$

F04_04_

Time Variations of Groundwater Levels

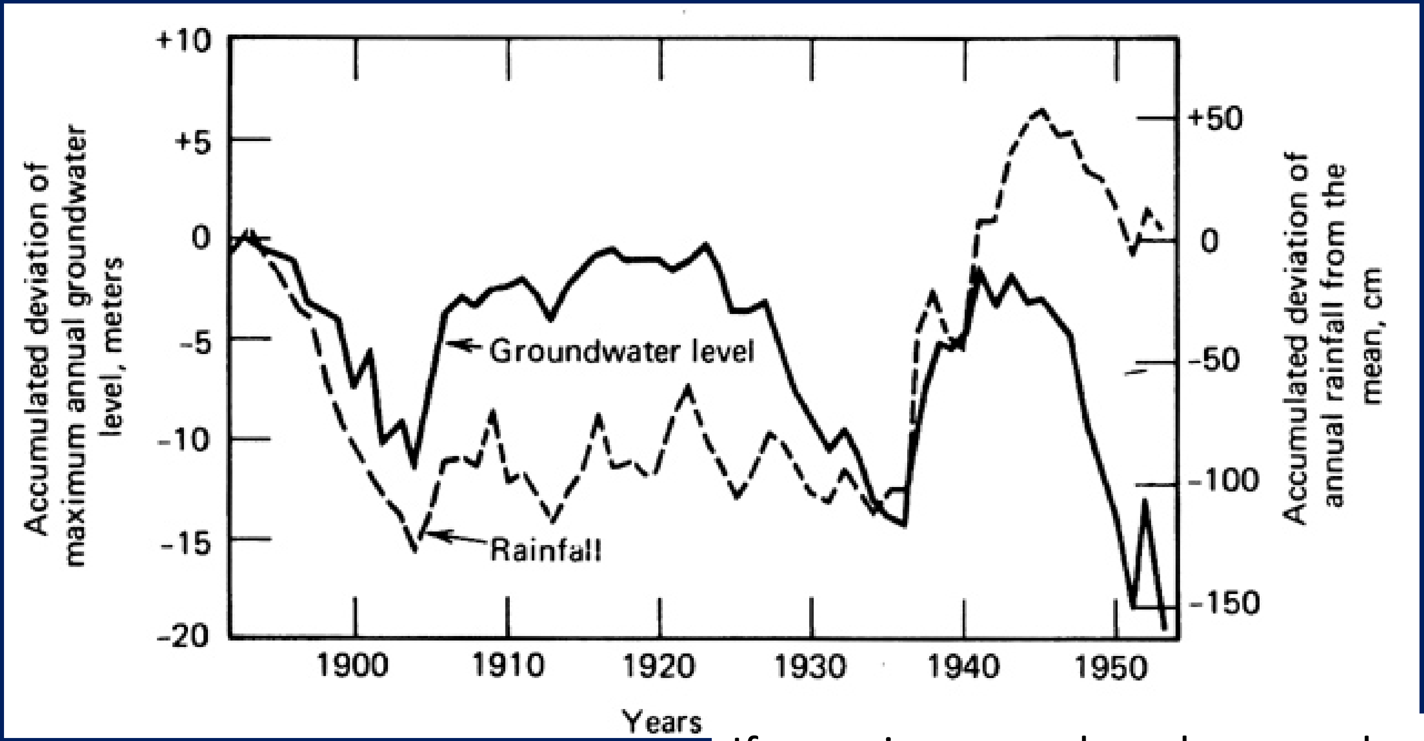
Secular variations (extend over periods of years-alternating series of wet and dry years)

Seasonal variations (influences such as rainfall and irrigation pumping)

A groundwater level, whether it be the water table of an unconfined aquifer or the piezometric surface of a confined aquifer, indicates the elevation of atmospheric pressure of the aquifer.

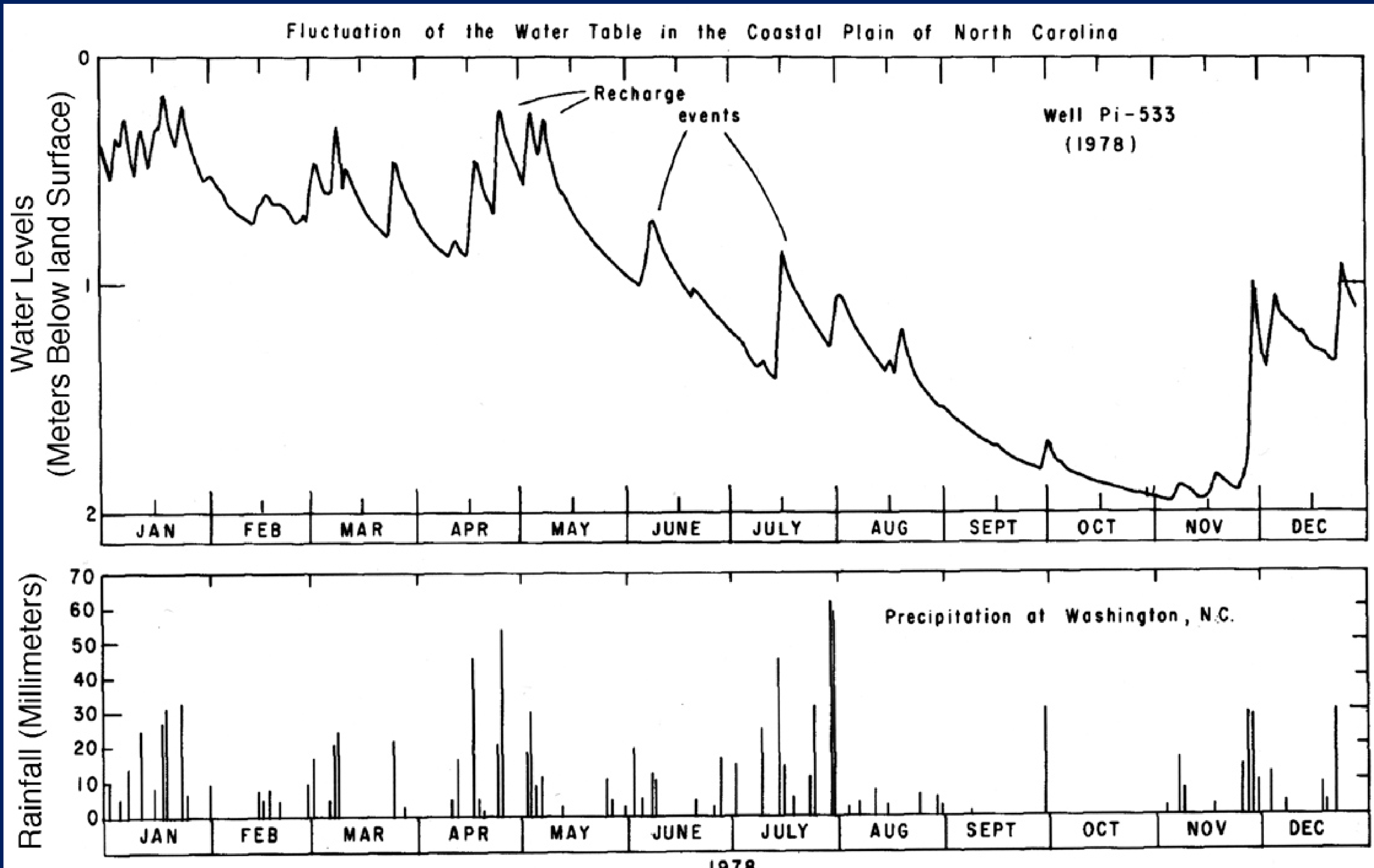
Long-period fluctuations of levels

Alternating series of wet and dry years, in which the rainfall is above or below the mean, will produce long-perid fluctuations of levels. Rainfall is not an accurate indicator of groundwater level changes. Recharge is the governing factor (assuming annual withdrawals are constant). It depends on rainfall intensity and distribution and amount of surface runoff.



If pumping exceeds recharge, a downward trend of groundwater levels may continue for many years.

During summer months, evapotranspiration is high, precipitation is low.



10 ft= 3 m

250 ft-170 ft
Difference 80 ft ~24 m

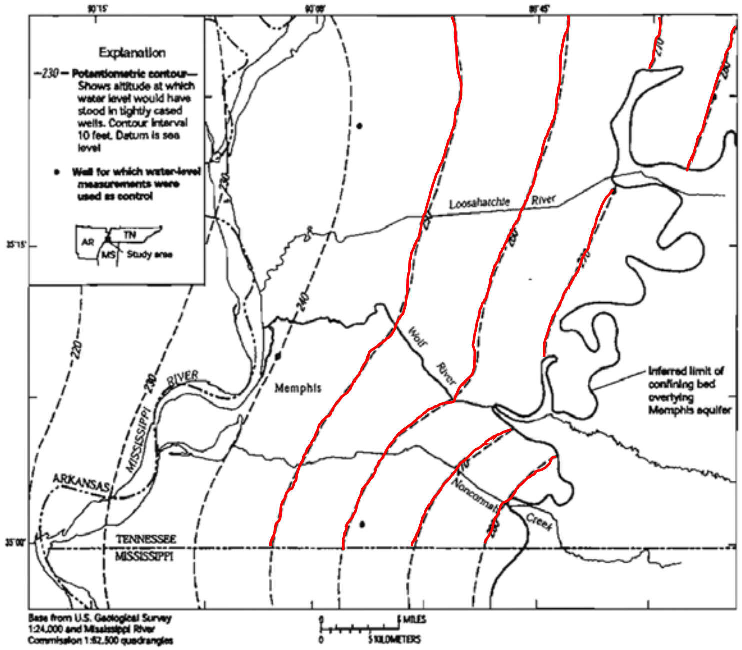


Figure 6.1.4. Inferred potentiometric surface of the Memphis aquifer prior to groundwater development. The observation wells shown were selected for their early records away from initial pumping centers (modified from Criner and Parks, 1976, as presented in Taylor and Alley[80]).

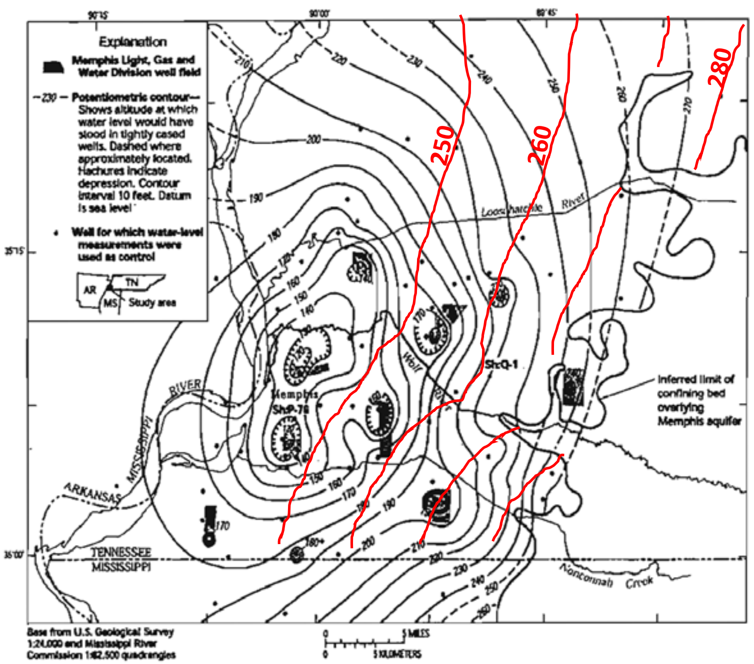
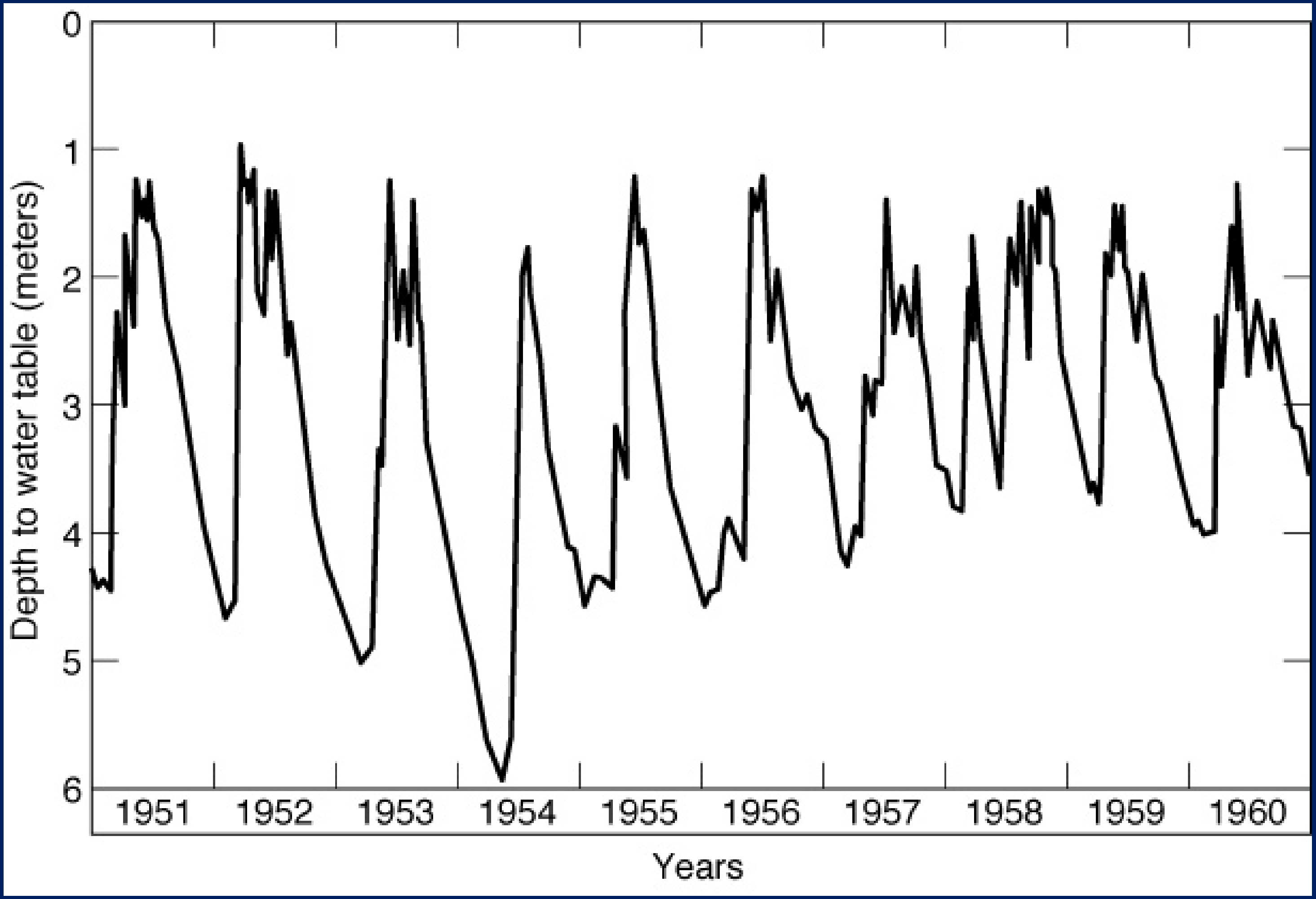


Figure 6.1.5. Potentiometric surface of the Memphis aquifer in 1995 showing cones of depression and location of observation wells Sh:P-76 and Sh:Q-1 (modified from Kingsbury, 1996, as presented in Taylor and Alley[80]).

Seasonal fluctuations of the water table in a glacial till aquifer in Ohia. Well depth is 9 m. – frozen ground in winter. Highest levels in late spring and lowest in winter.



Fluctuations due to meteorological phenomena:

1) Atmospheric pressure- In confined aquifers, when atmospheric pressure increases, water level decreases (the relationship is inverse). No such effect in unconfined aquifers

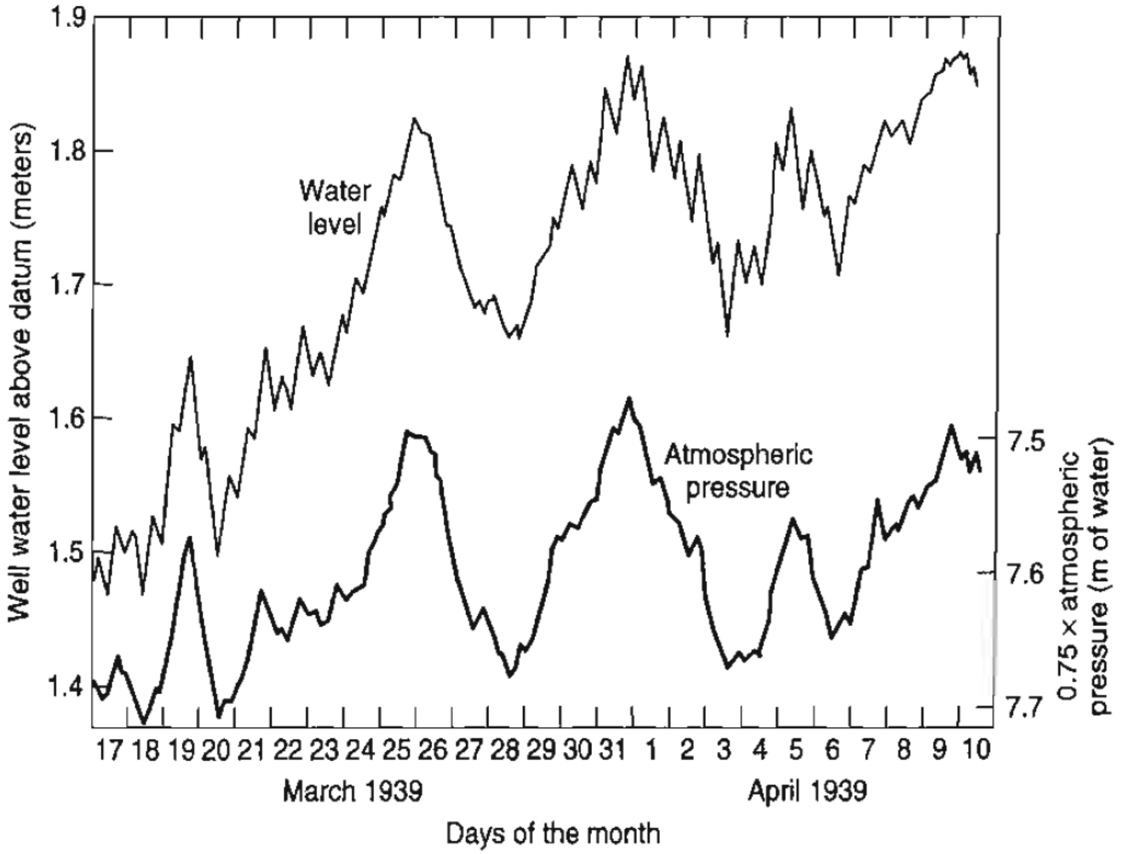


Figure 6.4.1. Response of water level in a well penetrating a confined aquifer to atmospheric pressure changes, showing a barometric efficiency of 75 percent (after Robinson[71]).

When atmospheric pressure changes are expressed in terms of a column of water, the ratio of water level change to pressure change expresses the barometric efficiency of an aquifer.

$$B = \frac{\gamma \Delta h}{\Delta p_a}$$

γ= specific weight of water (9810 N/m^3)
Δh= change in the piezometric level
Δp_a = change in the atmospheric pressure.
1 kPa= 1 kN/m^2
20%-70%

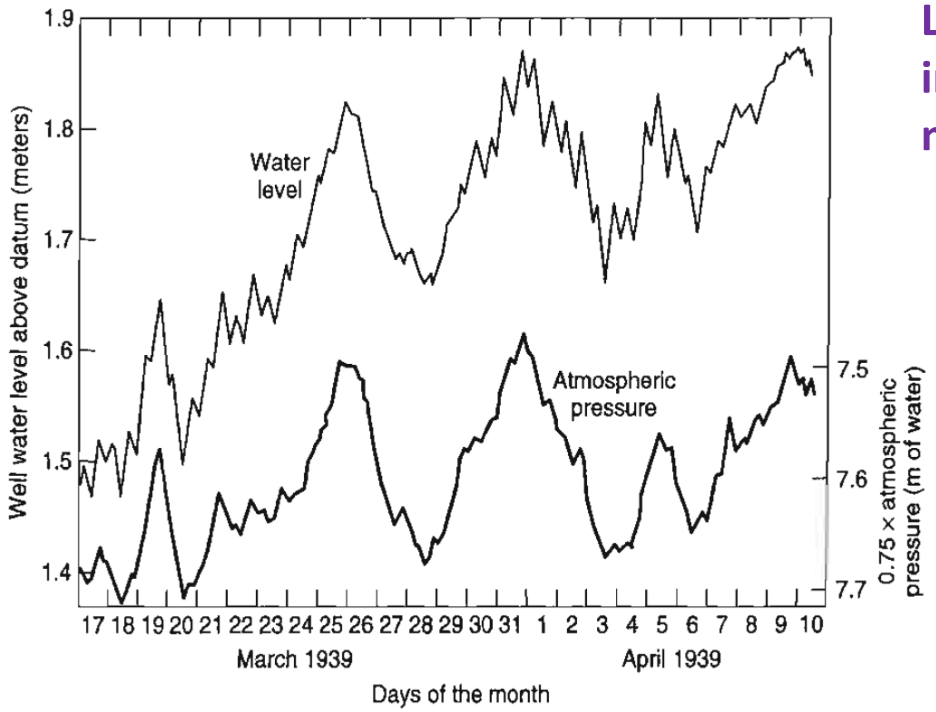


Figure 6.4.1. Response of water level in a well penetrating a confined aquifer to atmospheric pressure changes, showing a barometric efficiency of 75 percent (after Robinson[71]).

Lower curve shows atmospheric pressure inverted, expressed in meters of water, and multiplied by 0.75.

Under normal atmospheric condition of 15°C, the atmospheric pressure is equal to the pressure generated by 760 mm of mercury or 10.33 m of water.

$P = \rho * g * h$

ρ= 1000 kg/m3, g= 9.81 m/s2

P= 1000*9.81*10.33

P= 101325 Pa, P= 1.01325 bar= 1 atm

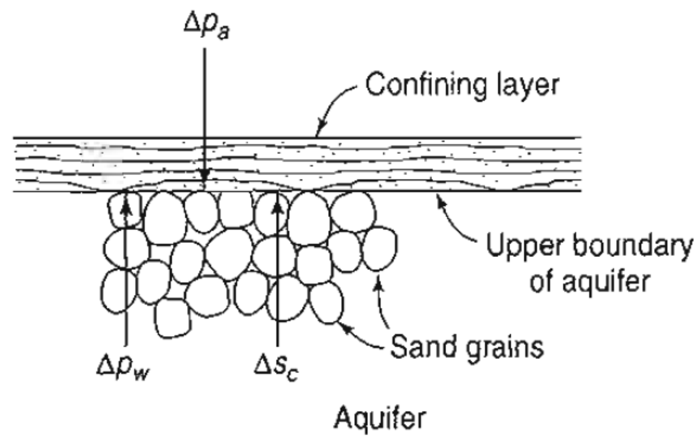


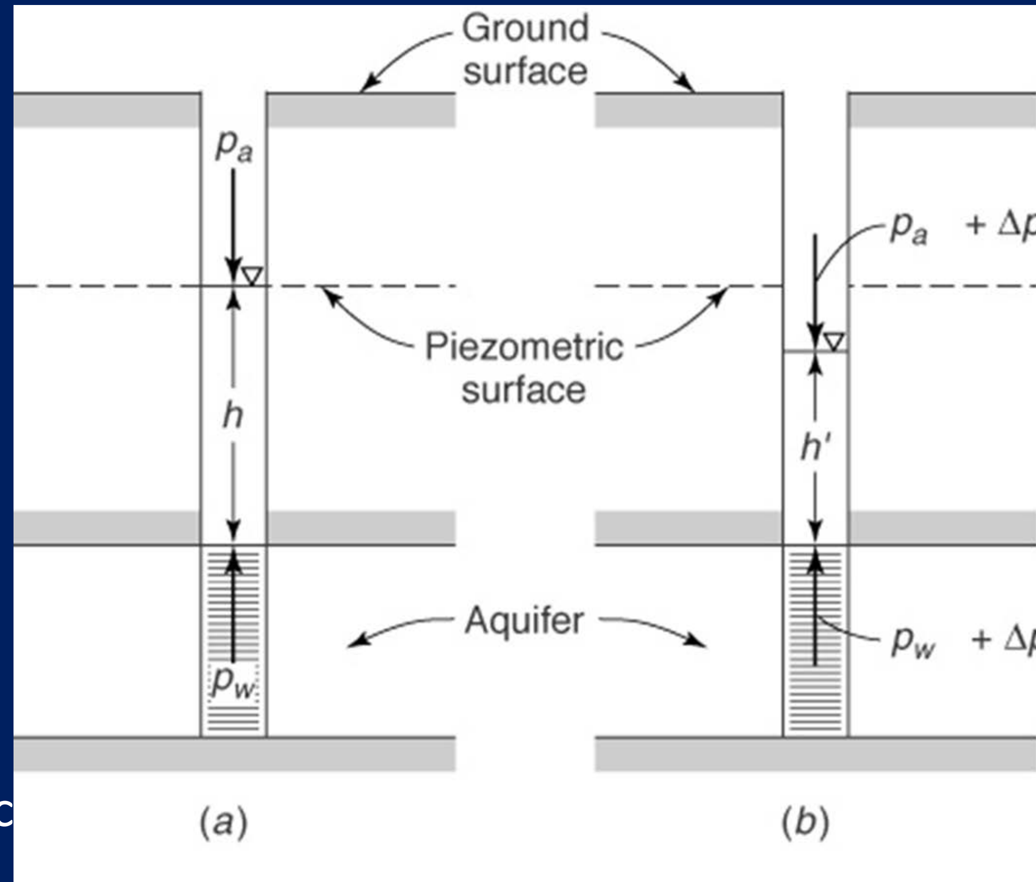
Figure 6.4.2. Idealized distribution of forces at the upper boundary of a confined aquifer resulting from a change in atmospheric pressure.

Δp_w= resulting change in hydrostatic pressure at the top of a confined aquifer

Δp_a = change in the atmospheric pressure

ΔS_c= increased compressive stress on the aquifer

$$p_w = p_a + \gamma h$$
$$p_w + \Delta p_w = p_a + \Delta p_a + \gamma h'$$

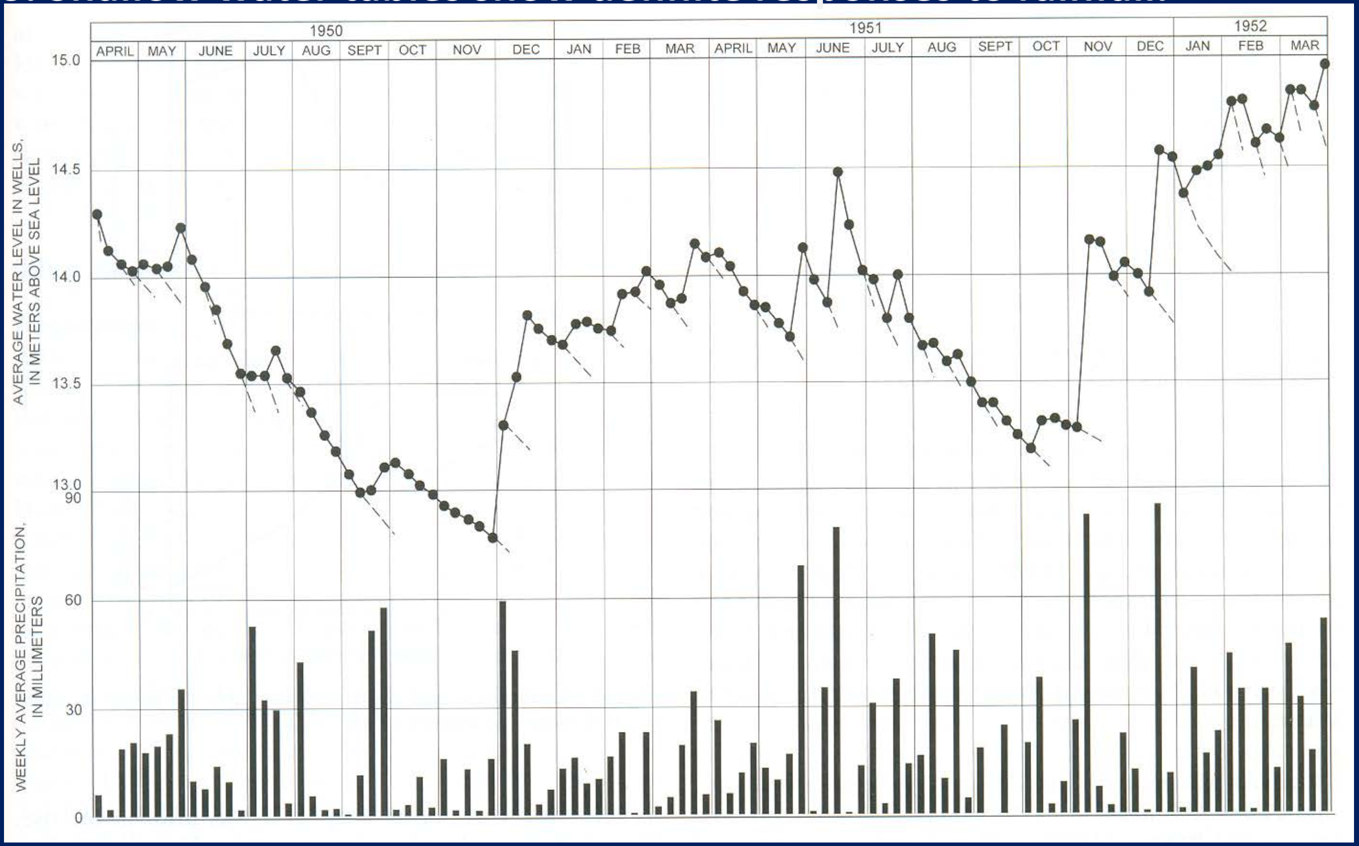


$$\Delta p_w = \Delta p_a + \gamma(h' - h)$$
$$\Delta p_w < \Delta p_a$$
$$h' < h$$

The water level in a well falls with an increase in atmospheric pressure.

2) Rainfall

Rainfall is not an accurate indicatior of groundwater recharge because of surface and subsurface losses as well as travel time for vertical percolation. The travel time may vary from a few minures to months or years. In arid and semi-arid regions- recharge from rainfall might be zero. Shallow water tables show definite responses to rainfall.



Variation in average water table level and weekly average preciptation (Beaverdam Creek Basin, Maryland) (Todd, 1980)

The rise due to response to recharge from rainfall: $\Delta h = P/S_y$ P= portion of rainfall percolating to the water table

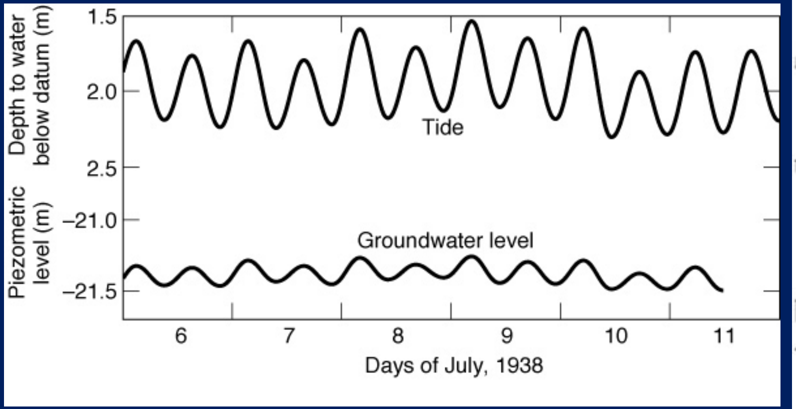
Fluctuations due to Tides

In coastal aquifers in contact with the ocean, sinusoidal fluctuations of groundwater levels occur in response to tides. If the sea level varies with a simple harmonic motion, a train of sinusoidal waves is propagated inland from the submarine outcrop of the aquifer. With distance, inland amplitudes of the waves decrease.

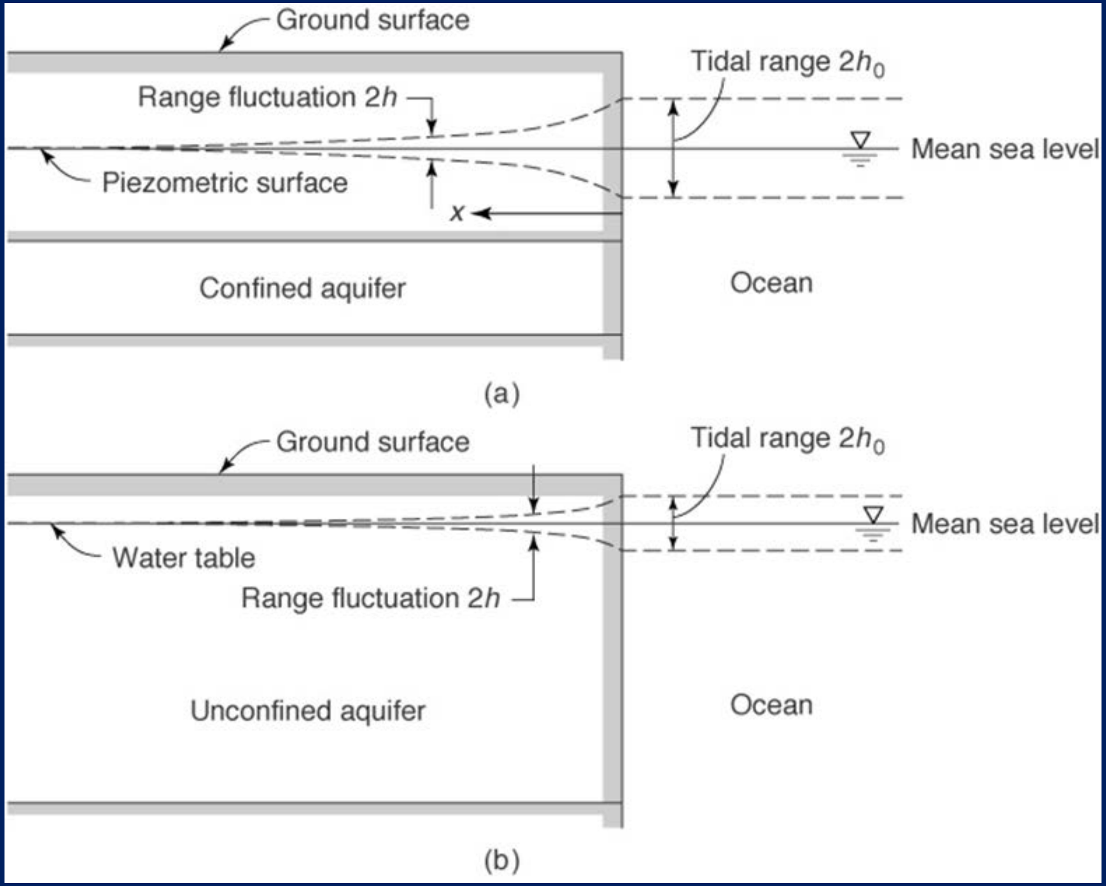
https://www.denizcilikbilgileri.com/gel-git-olayi-nedir/

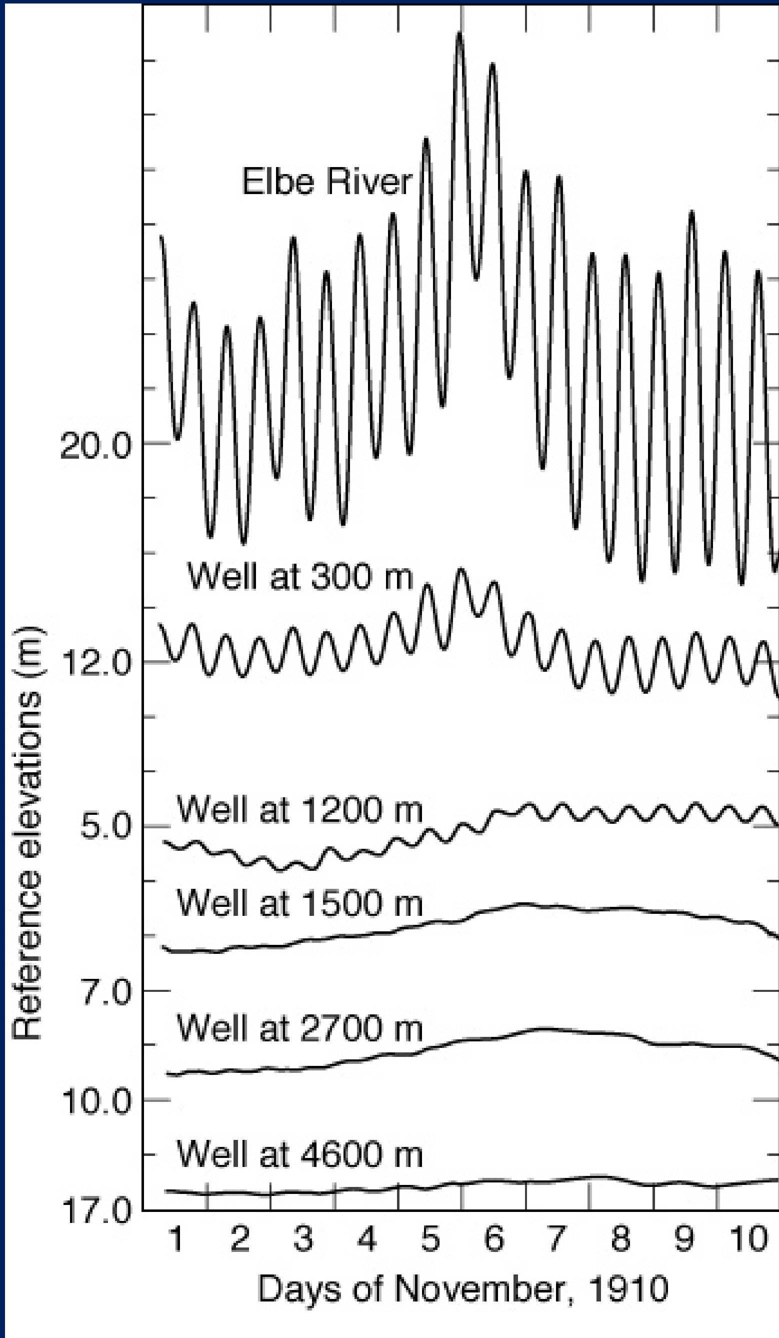
$$h = h_0 * e^{-x\sqrt{\pi n e / T t o}}$$

h= half range of the rise or fall of the piezometric surface with reference to the mean level
X is the distance inland from the outcrop- m
h_o= amplitude, or half-range, of the tide
t_o= tidal period



Tidal fluctuations and induced piezometric surface fluctuations observed in a well 30 m from shore at Mattawoman Creek, Maryland





Fluctuations of the Elbe River and water table levels in wells at various distances from the river

Todd and Mays, 2005, Groundwater Hydrology

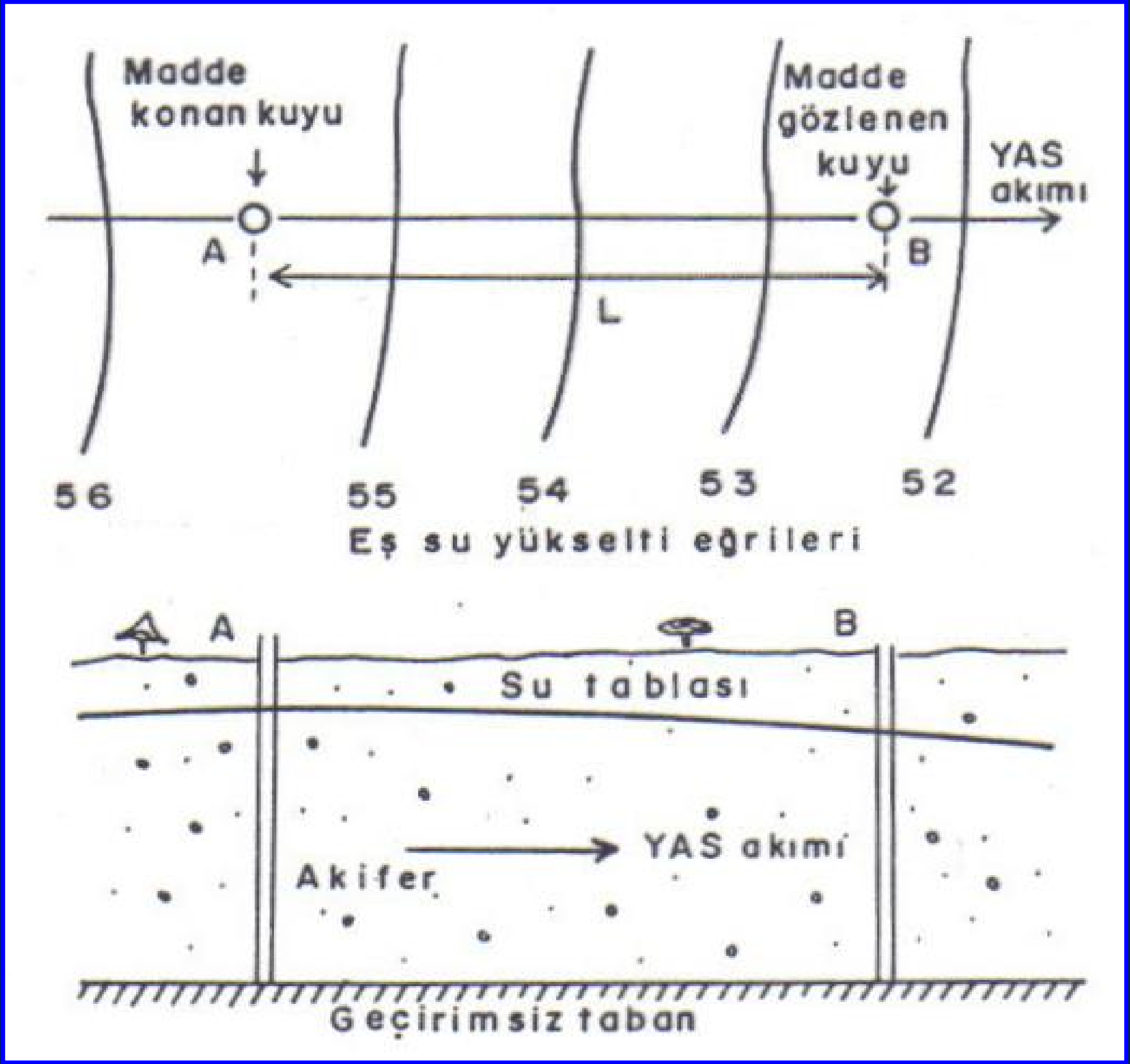
Groundwater Tracers

How to determine groundwater flow velocity and flow direction by using tracers?

An ideal tracer should

a) be susceptible to quantitative determination in minute concentrations,
b) Be absent or nearly so from the natural water
c) Not react chemically with the natural water or be absorbed by the porous media
d) Be safe in terms of human health
e) Be in expensive and readily available.

Water-soluble dyes (sodium fluorescein) detected by colorimetry, soluble chloride and sulfate salts and sugars detected chemically and strong electrolytes, which can be detected by electrical conductivity.

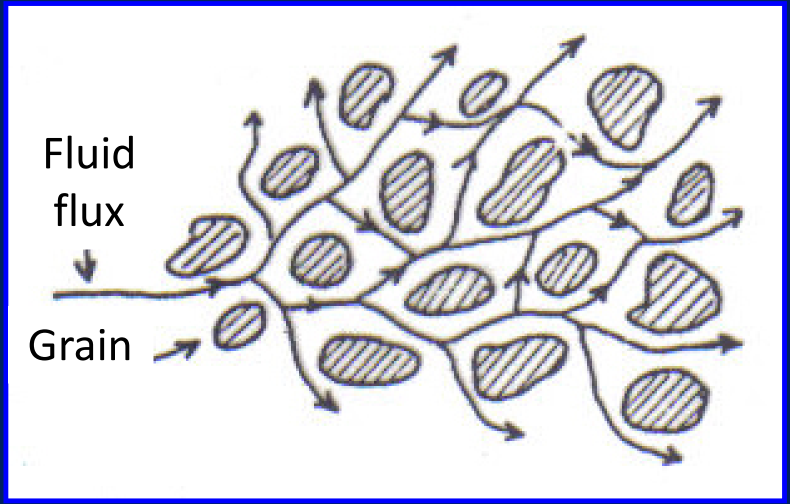

Madde
konan kuyu
Madde
gözlenen
kuyu
YAS
akımı
A
B
L
56
55
54
53
52
Eş su yükselti eğrileri
A
B
Su tablası
YAS akımı
Akifer
Geçirimsiz taban

Groundwater flow direction, groundwater flow velocity, recharge area determination.


UZLAR
N SERFEL

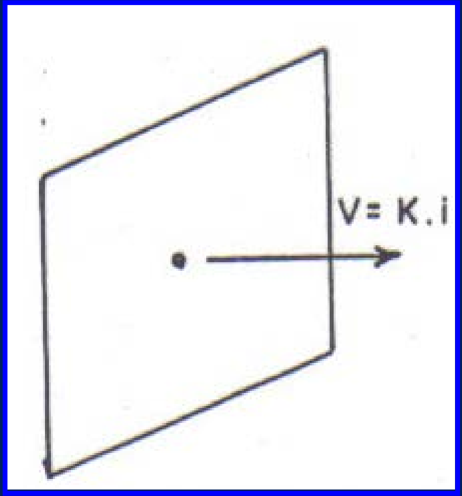

YAVUZLAR
SERİM SERFEL

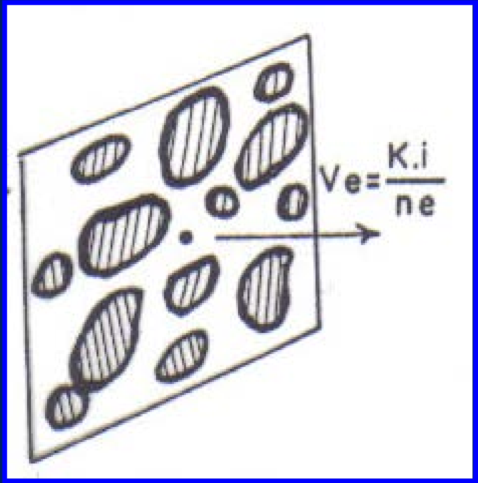
Groundwater Flow Velocity



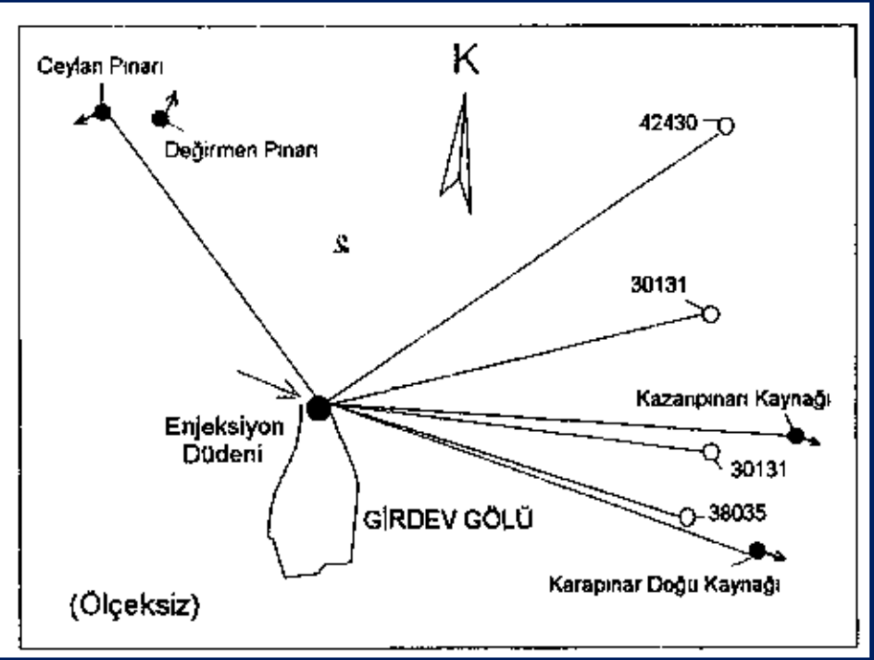
Darcy (Apparent) velocity

Average Linear velocity, seepage velocity
Average interstitial velocity





Calculate the average apparent velocity and average linear velocity if the effective porosity is 25%



Observation point	Total time when the dye reaches the observation point (hour)	The distance from the injection point (km)	Apparent Groundwater velocity (m/day)
30131	309	11	
30132	405	12	
38035	325	8.5	
42430	517	17.5	
Karapınar Doğu Spring	365	9.5	
Kazanpınar Spring	437	14.5	

Çelik, M., Ünsal, N., 1996, Girdev Gölü- Kazanpınarı kaynağı arasındaki (Antalya) allokton birimlerde yeraltı usuyu dolaşımının incelenmesi, Türkiye Jeoloji Bülteni, 69-74

Saline Water Intrusion in Aquifers

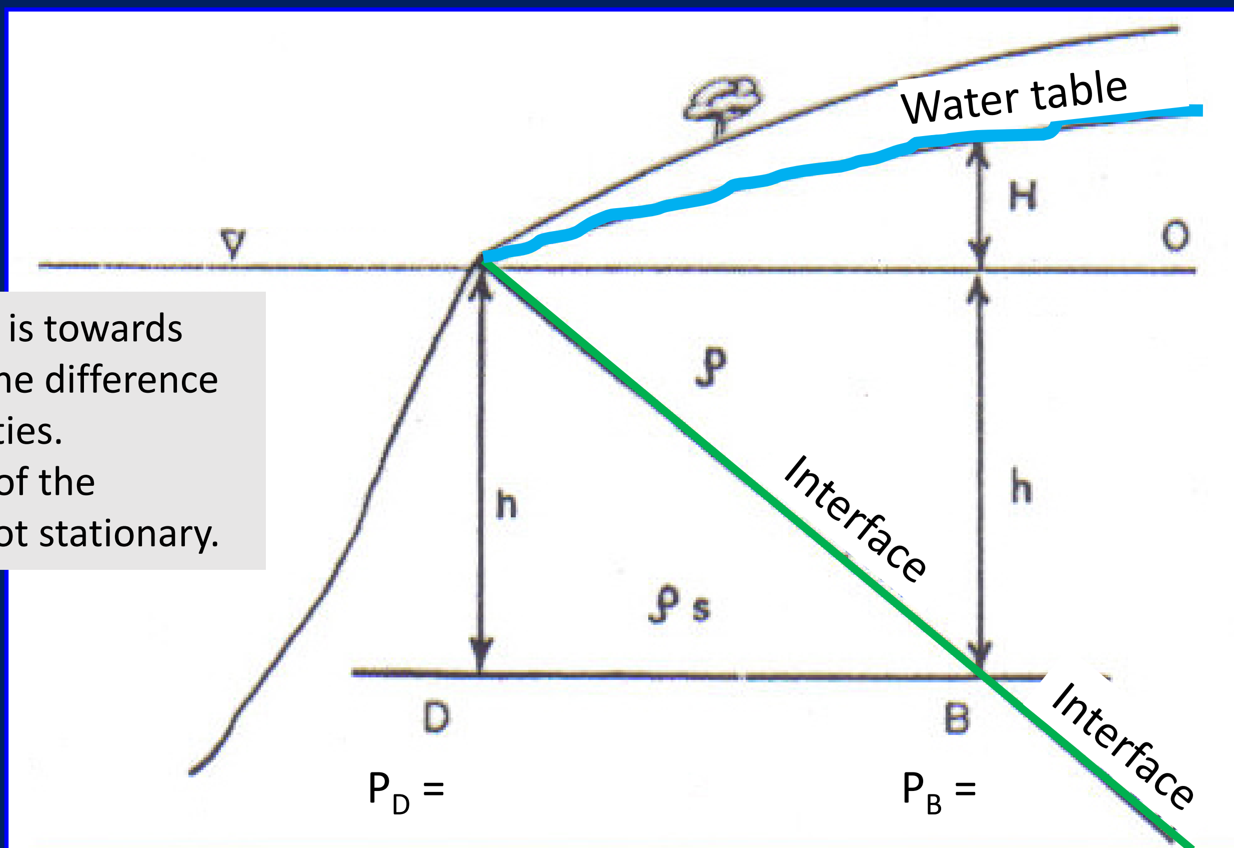
Intrusion of saline water occurs where saline water displaces or mixes with freshwater in an aquifer. The phenomenon can occur

a) in deep aquifers with the upward advance of saline waters of geologic origin,
b) in shallow aquifers from surface waste discharges,
c) and in coastal aquifers from an invasion of seawater.

Ghyben-Herzberg relation between fresh and saline waters

In 1889 Badon Ghyben worked in the Holland coast and in 1901 Herzberg worked in the Baltic Sea. They worked independently, and found out that salt water occured underground, not at sea level but at a depth below sea level of about 40 times the height of the fresh water above sea level. This distribution was attributed to a hydrostatic equilibrium existing between the two fluids of different densities.

The interface is towards land due to the difference in fluid densities.
The location of the interface is not stationary.



Ghyben-Herzberg model

Ghyben-Herzberg relation between fresh and saline waters

Hydrostatic pressures in both sides equal each other

$$\rho_s * g * z = \rho * g * (z + h_f)$$

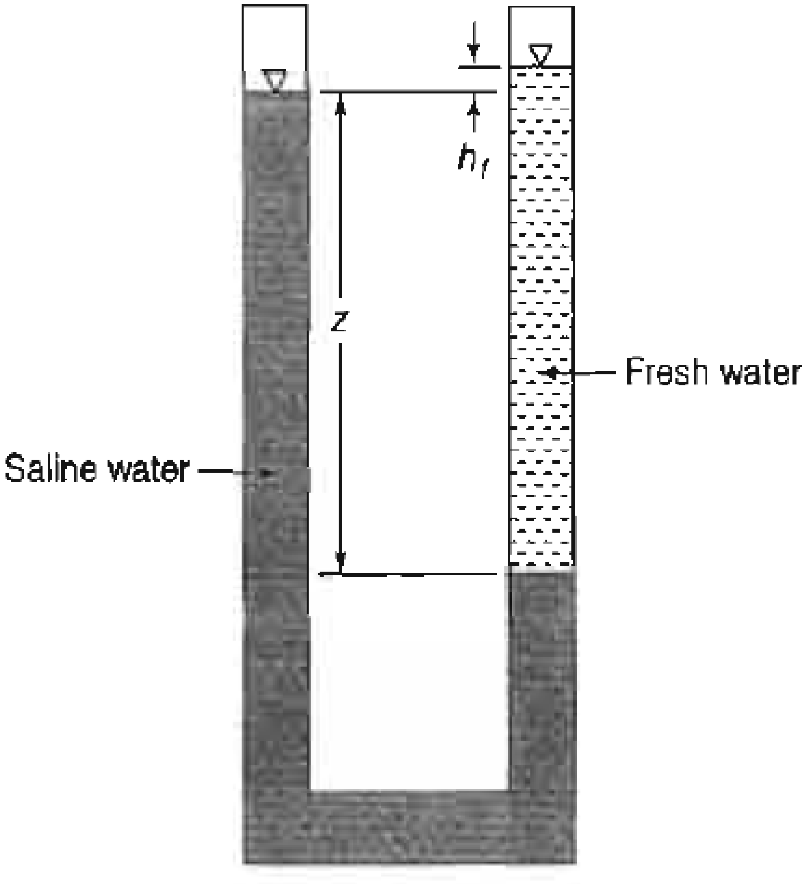
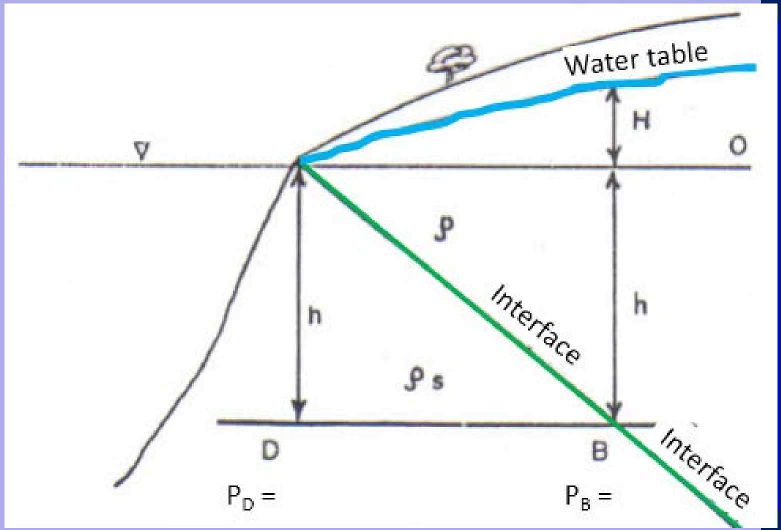


Figure 14.2.1. Hydrostatic balance between fresh water and saline water illustrated by a U-tube.

Hydrostatic pressure at point D $P_D = \rho s * g * h$

Hydrostatic pressure at point B $P_B = \rho * g * h + \rho * g * H$

If there is no flow between fresh and saline water the pressures in BD should equal each other.



$$\rho s * g * h = \rho * g * h + \rho * g * H$$

$$\rho s * h = \rho * h + \rho * H$$

$$h(\rho s - \rho) = \rho * h$$

$$h = \frac{\rho}{\rho s - \rho} * H$$

If density of sea water is 1.025 $\mathrm{h} = 40\ \mathrm{H}$

Fresh water- salt water equilibrium hs=0

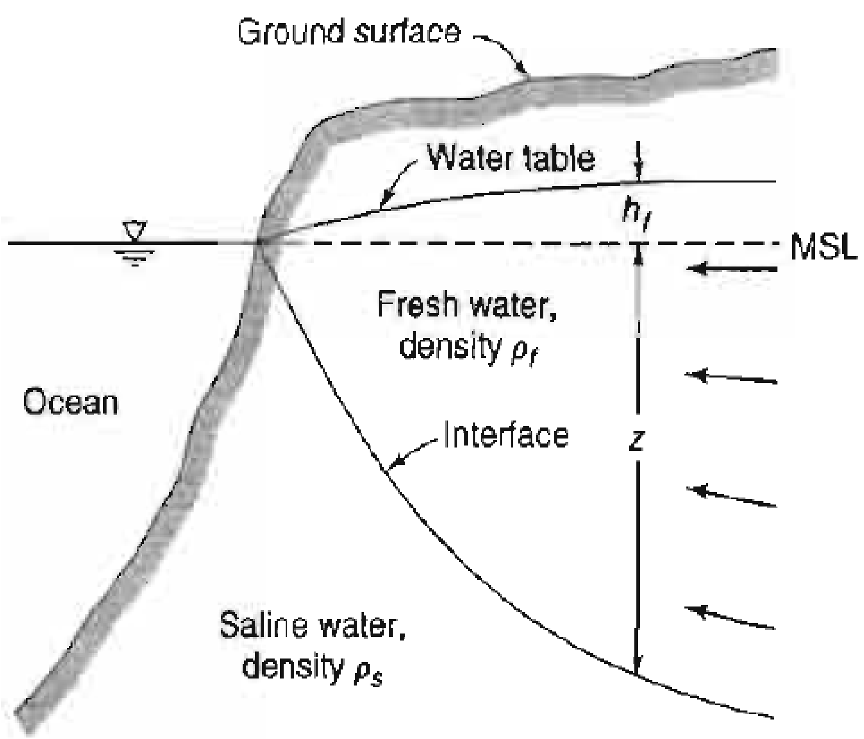


Figure 14.2.2. Idealized sketch of occurrence of fresh and saline ground-water in an unconfined coastal aquifer.

There is a hydrodynamic rather than a hydrostatic balance because fresh water is flowing toward the sea. Without flow, a horizontal interface would develop with fresh water everywhere floating above saline water. Where the flow is nearly horizontal, G-H relation gives satisfactory results. Near the shoreline, vertical flow components become significant.

G-H relation

Fresh-salt water equilibrium requires that the water table, or piezometric surface

1) Lie above sea level
2) Slope downward toward the ocean

OR seawater will advance directly inland.

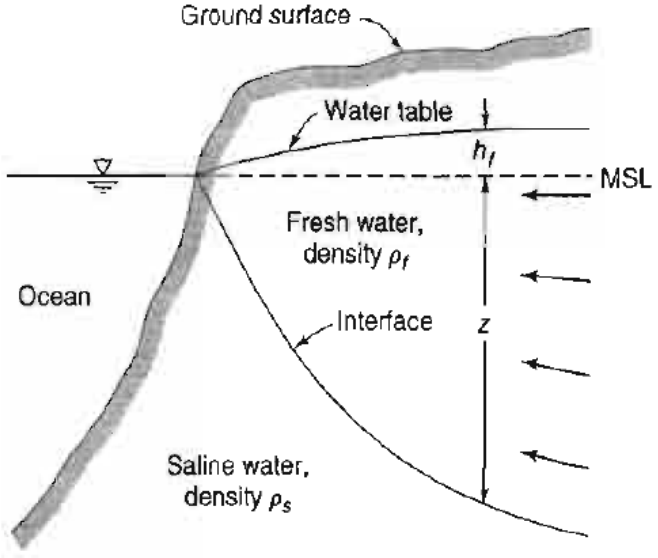
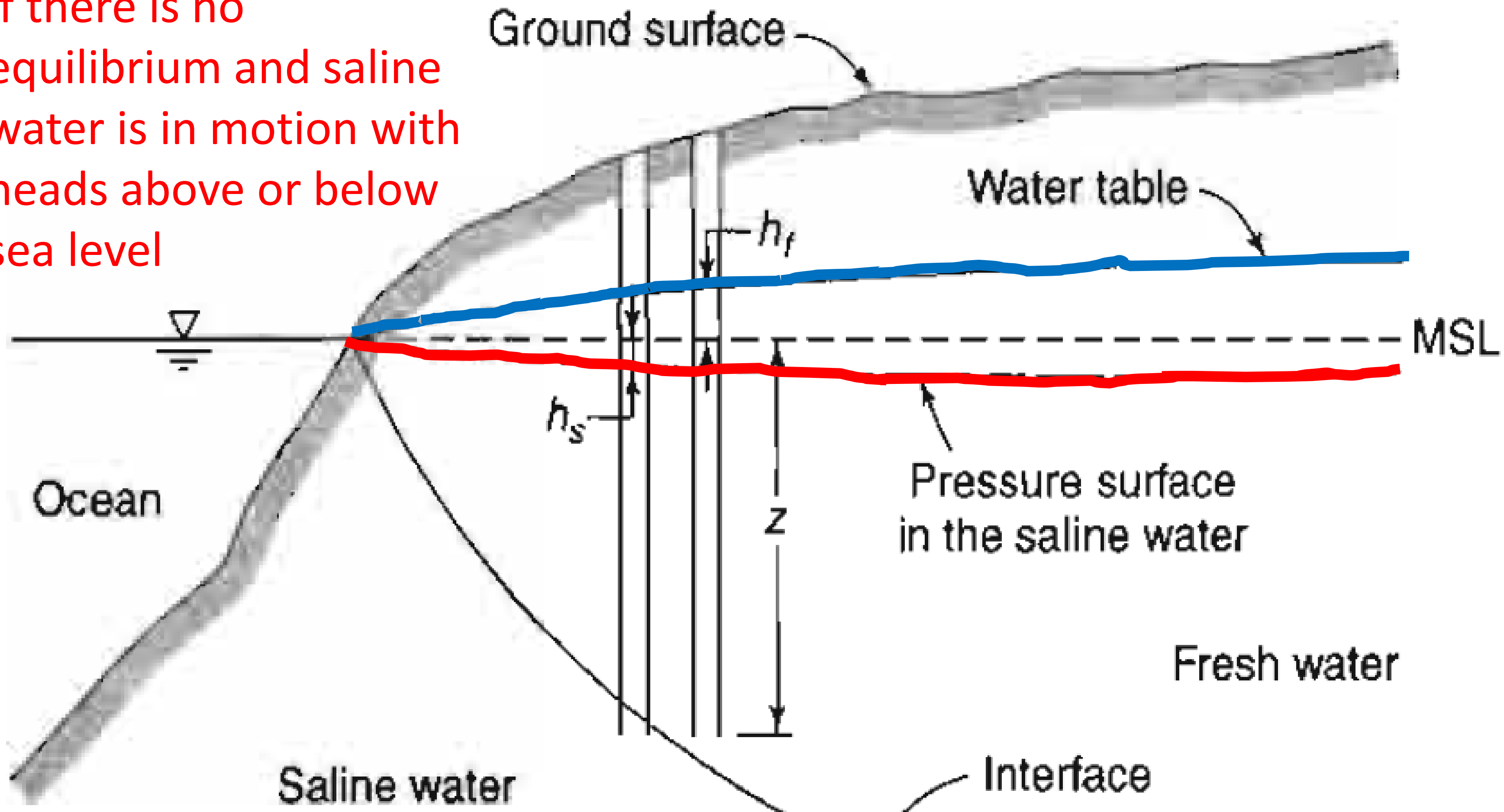


Figure 14.2.2. Idealized sketch of occurrence of fresh and saline groundwater in an unconfined coastal aquifer.

If there is no equilibrium and saline water is in motion with heads above or below sea level

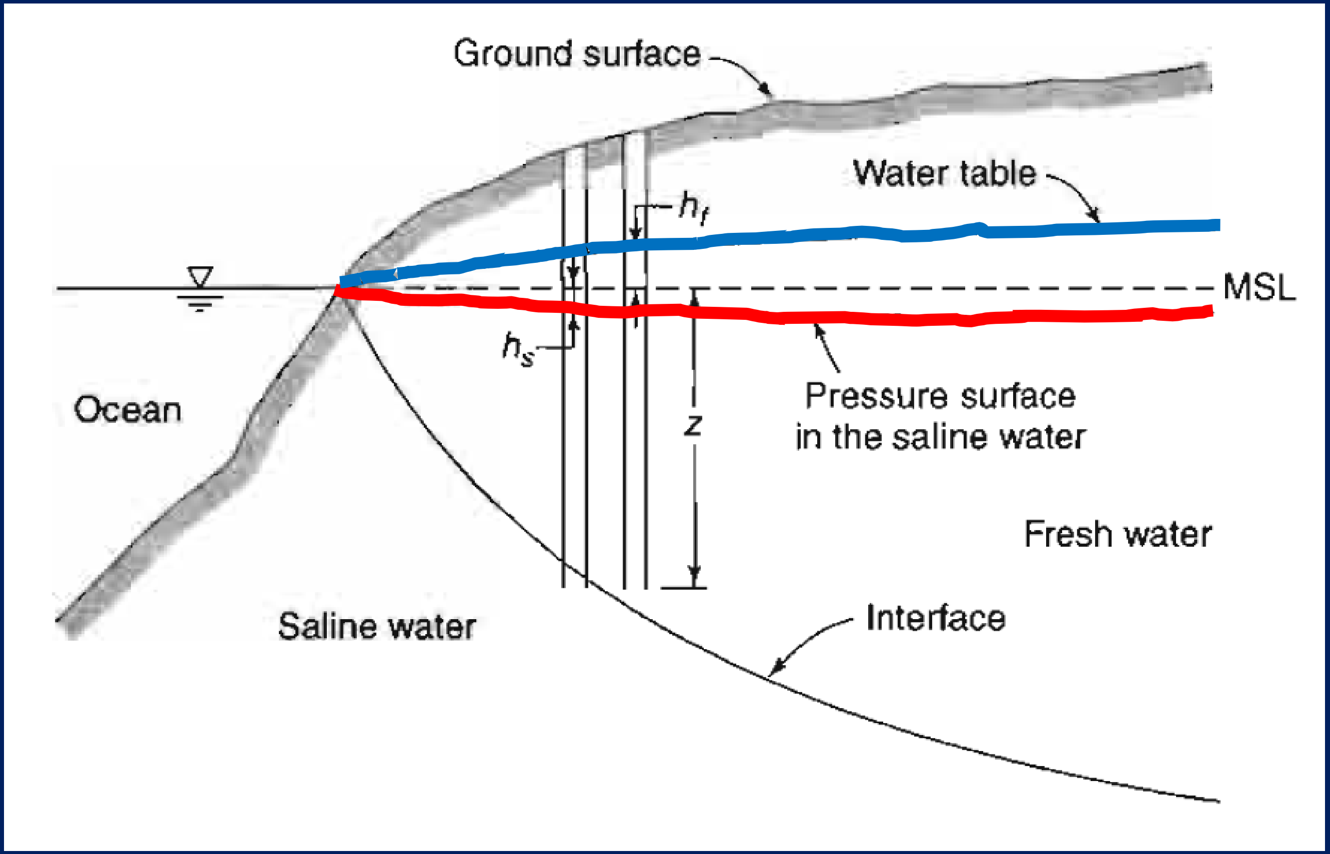


$$z = \frac{\rho_f}{\rho_s - \rho_f} h_f - \frac{\rho_f}{\rho_s - \rho_f} h_s$$

h_f= altitude of the water level in a well filled with fresh water of density ρ_f

h_s= altitude of the water level in a well filled with saline water of density ρ_s and also terminated at depth z

When hs=0, there is equilibrium



Nonequilibrium conditions

$$z = \frac{\rho_f}{\rho_s - \rho_f} h_f - \frac{\rho_f}{\rho_s - \rho_f} h_s$$

The water table 30 m from the shoreline is located 0.65 m above sea level in an unconfined coastal aquifer. Determine the depth to the saltwater interface at this location using the Gyhben-Herzberg relation.

No sharp boundary- landward movement of saline water body

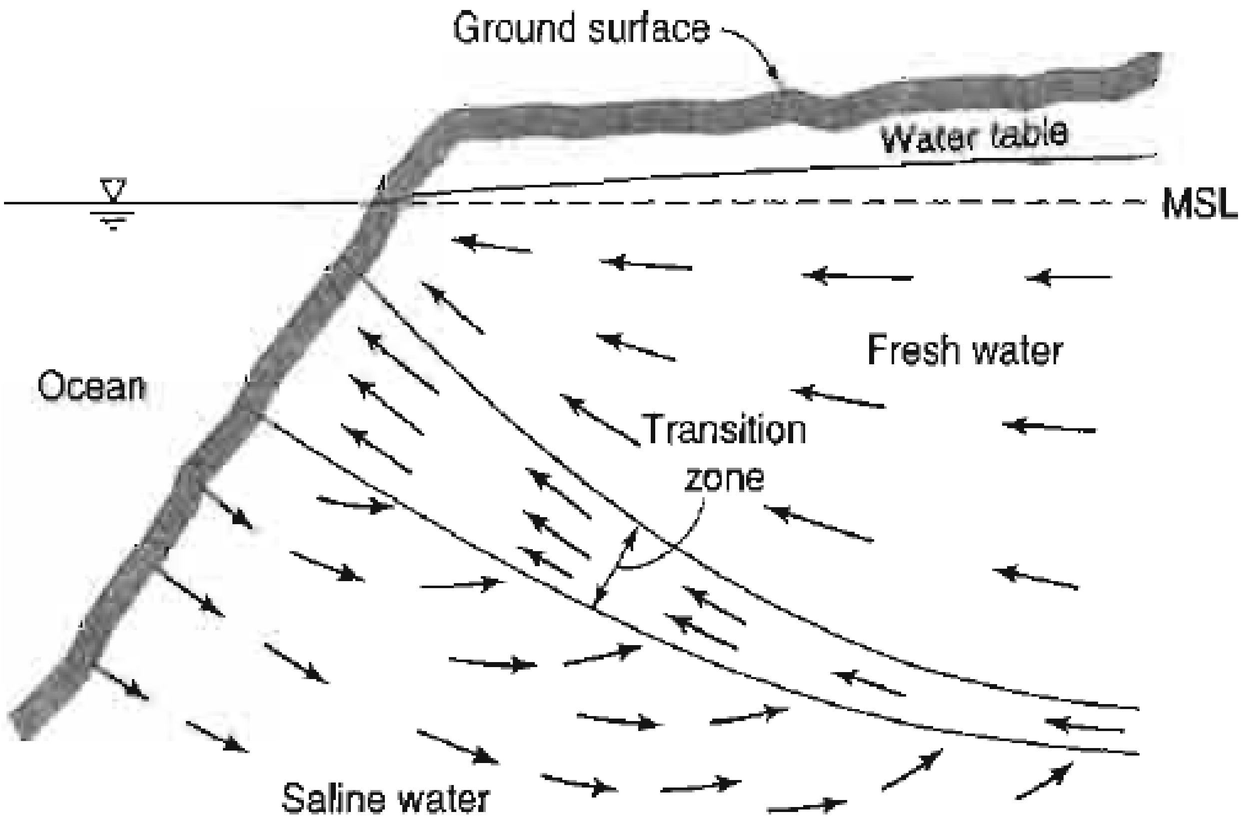


Figure 14.4.2. Vertical cross section showing flow patterns of fresh and saline water in an unconfined coastal aquifer (after Todd[61]; reprinted from the *Journal American Water Works Association*, vol. 66, by permission of the Association, copyright © 1974 by American Water Works Association, 6666 West Quincy Avenue, Denver, CO, 80235).

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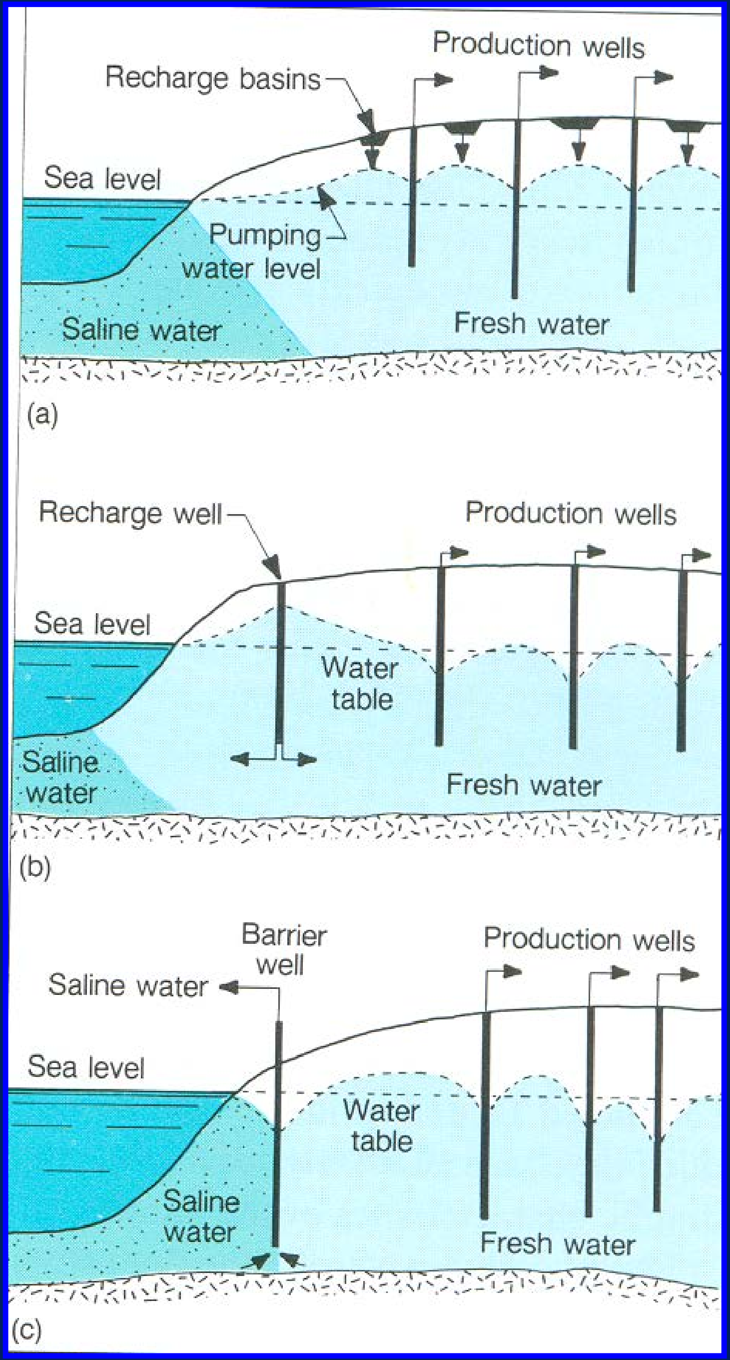
*As an extreme case, concentrated pumping in the Honolulu–Pearl Harbor area of Hawaii has created localized transition zones more than 300 m thick; these occupy essentially the entire vertical extent of the aquifer.[67]



How to prevent saline water intrusion?

SPRINGS

A spring is a concentrated discharge of groundwater appearing at the surface as a current of flowing water. (Seepage areas- indicate a slower movement of groundwater to the ground surface. Well established in the engineering literature in connection with groundwater movement from and to surface bodies, particularly where associated with structures such as dams, canals, etc.

A spring in Kazan Basin (Ankara)

A spring in Kütahya Köprüören Basin

Spring areas may pond.

Springs occur in many forms and have been classified as to cause, rock structure, discharge, temperature and variability.

Springs resulting from nongravitational forces

Volcanic springs

Fissure springs- resulting from fractures extending to great depths in the earth's crsut.

Springs resulting from gravitational forces

Gravity springs result from water flowing under hydrostatic pressure.

1) Depression Springs
2) Contact Springs
3) Artesian Springs
4) Impervious Rock Springs
5) Tubular or Fracture Springs

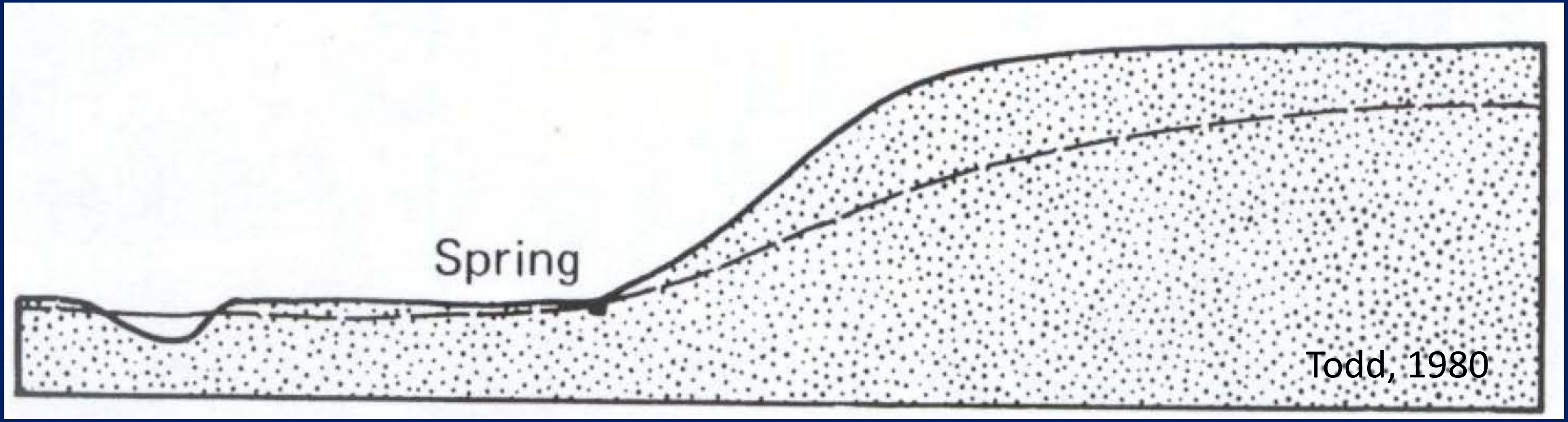
Springs occur in many forms and have been classified as to cause, rock structure, discharge, temperature and variability.

Springs resulting from nongravitational forces

Springs resulting from gravitational forces

Gravity springs result from water flowing under hydrostatic pressure.

1) Depression Springs- formed where the ground surface intersects the water table.



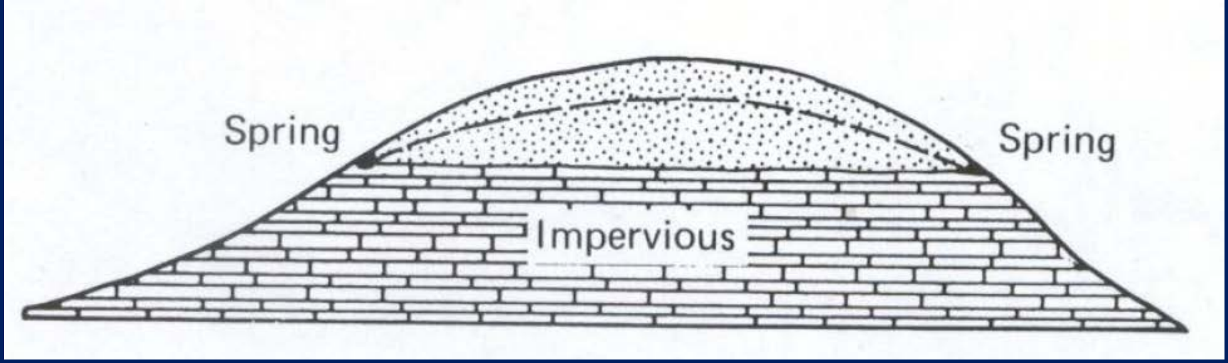
Springs occur in many forms and have been classified as to cause, rock structure, discharge, temperature and variability.

Springs resulting from nongravitational forces

Springs resulting from gravitational forces

Gravity springs result from water flowing under hydrostatic pressure.

1) Depression Springs- formed where the ground surface intersects the water table.
2) Contact Springs- created by a permeable water-bearing formation overlying a less permeable formation that intersects the ground surface.



Todd, 1980

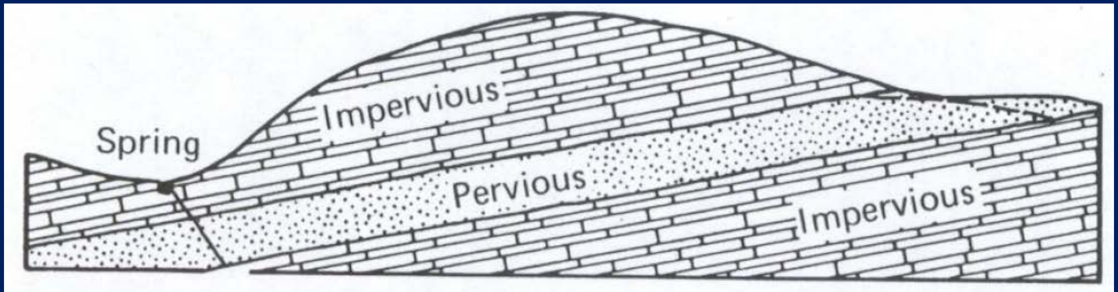
Springs occur in many forms and have been classified as to cause, rock structure, discharge, temperature and variability.

Springs resulting from nongravitational forces

Springs resulting from gravitational forces

Gravity springs result from water flowing under hydrostatic pressure.

1) Depression Springs- formed where the ground surface intersects the water table.
2) Contact Springs- created by a permeable water-bearing formation overlying a less permeable formation that intersects the ground surface.
3) Artesian Springs- Resulting from releases of water under pressure from confined aquifers either at an outcrop of the aquifer or through an opening in the confining bed.



Todd, 1980

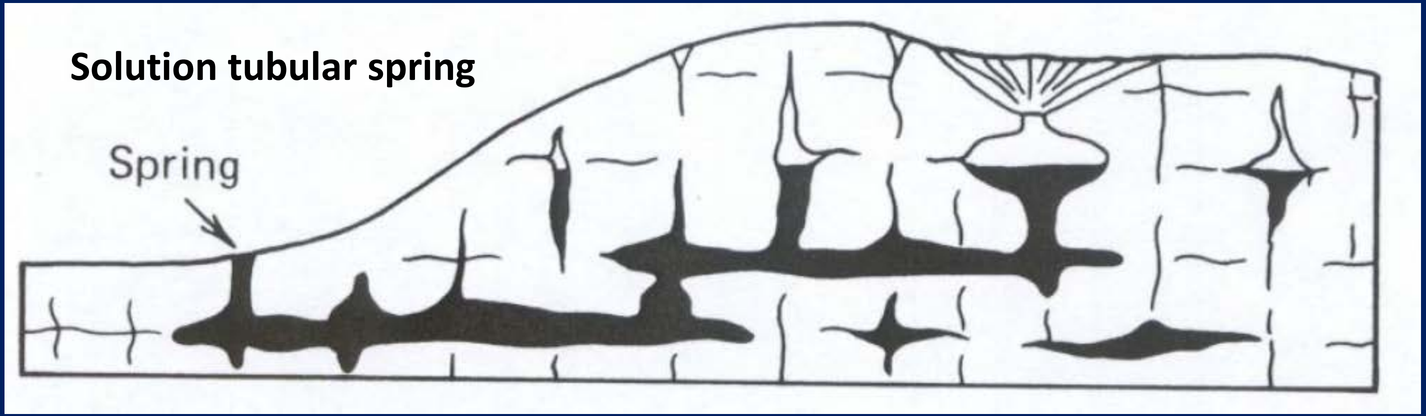
Springs occur in many forms and have been classified as to cause, rock structure, discharge, temperature and variability.

Springs resulting from nongravitational forces

Springs resulting from gravitational forces

Gravity springs result from water flowing under hydrostatic pressure.

1) Depression Springs
2) Contact Springs
3) Artesian Springs
4) Impervious Rock Springs- Occuring in tubular channels or fractures of impervious rocks.
5) Tubular or Fracture Springs- Issuing from rounded channels, such as lava tubes or solution channels, fractures in impermeable rock connecting with groundwater.



Todd, 1980

Springs, especially those in arid regions, are renowned as hotspots of biological and cultural diversity, and the presence of endangered or unique species and ethnological and historical resources often greatly influences their management (Springer and Stevens, 2009).

Ecological and cultural variables relevant to springs classification include:

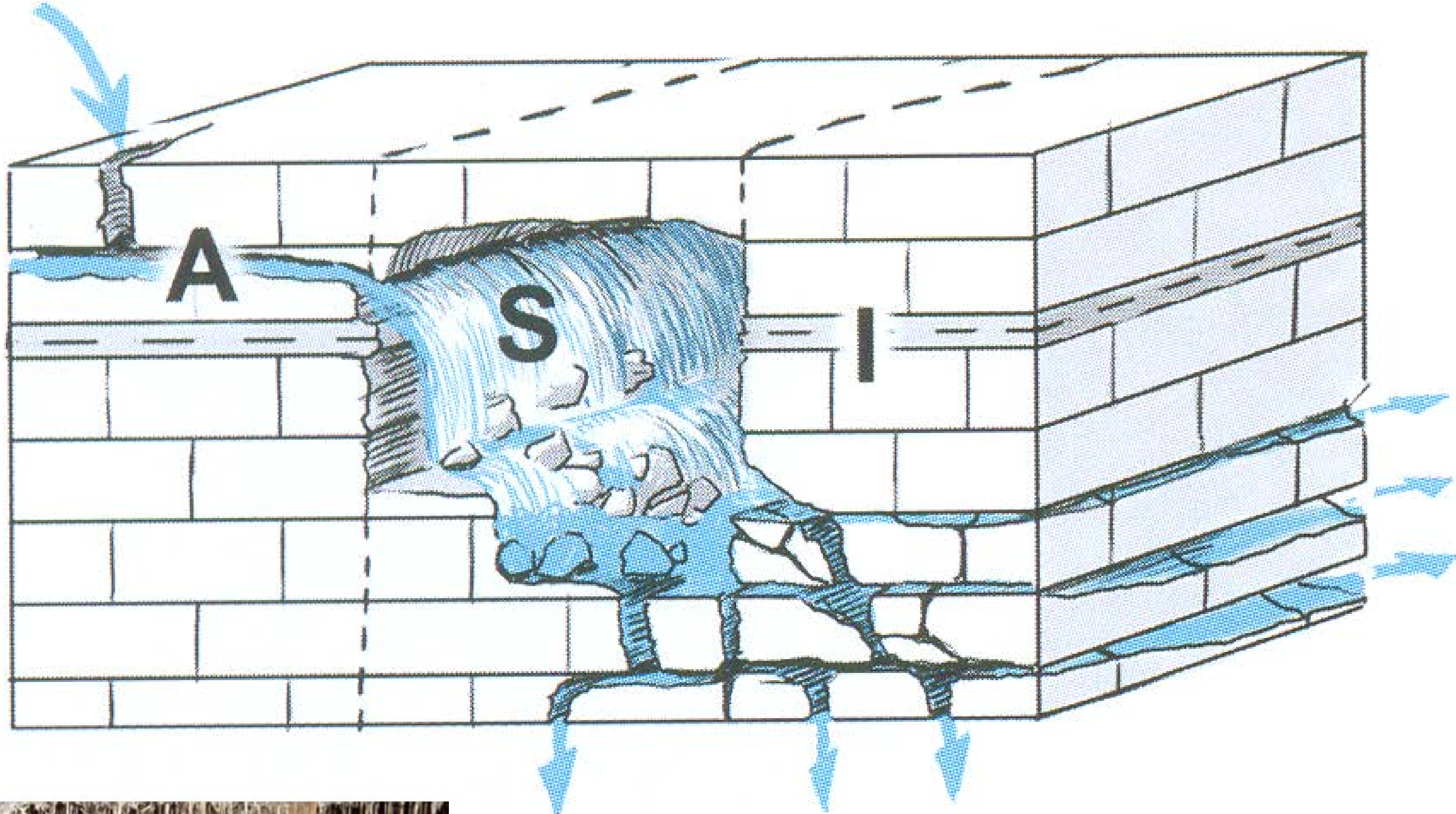
a) size, spatial isolation
b) Microhabitat distribution
c) Paleontological resources
d) The presence of rare or endemic biota
e) Archeological or traditional cultural resources

Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93

The «sphere» into which the aquifer is discharged as described by different scientists. 12 different spheres of discharge of springs defined by Springer et al. (2008)

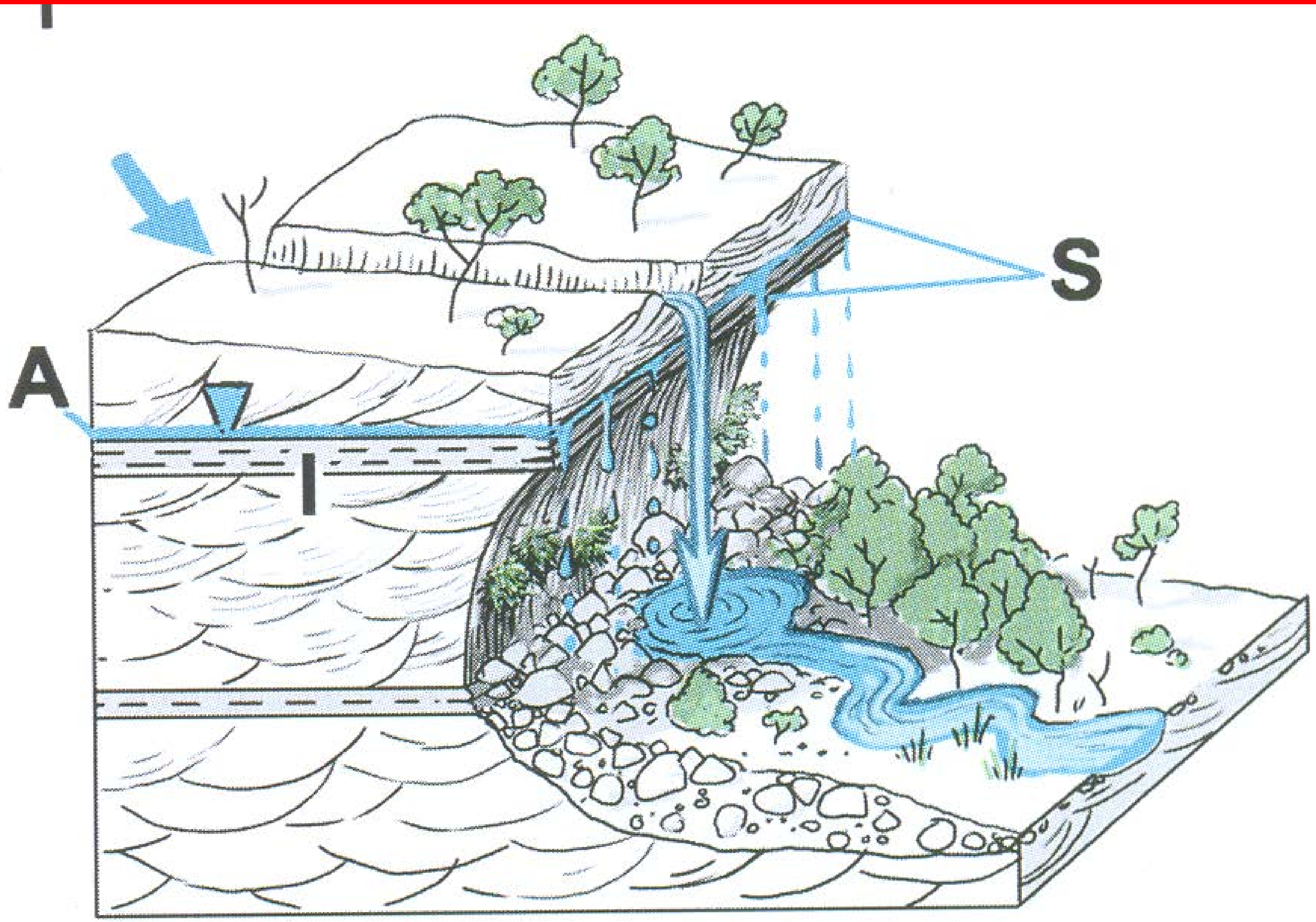
Table 1 Sphere of discharge and types of springs (modified from Springer et al. 2008) with examples of known springs and references of descriptions of sphere of discharge

Spring type	Emergence setting and hydrogeology	Example	Reference
Cave	Emergence in a cave in mature to extreme karst with sufficiently large conduits	Kartchner Caverns, AZ	Springer et al. (2008)
Exposure springs	Cave, rock shelter fractures, or sinkholes where unconfined aquifer is exposed near the land surface	Devils Hole, Ash Meadows, NV	Springer et al. (2008)
Fountain	Artesian fountain with pressurized CO_2 in a confined aquifer	Crystal Geyser, UT	Springer et al. (2008)
Geyser	Explosive flow of hot water from confined aquifer	Riverside Geyser, WY	Springer et al. (2008)
Gushet	Discrete source flow gushes from a cliff wall of a perched, unconfined aquifer	Thunder River, Grand Canyon, AZ	Springer et al. (2008)
Hanging garden	Dripping flow emerges usually horizontally along a geologic contact along a cliff wall of a perched, unconfined aquifer	Poison Ivy Spring, Arches NP, UT	Woodbury (1933); Welsh (1989); Spence (2008)
Helocrene	Emerges from low gradient wetlands; often indistinct or multiple sources seeping from shallow, unconfined aquifers	Soap Holes, Elk Island NP, AB, Canada	Modified from Meinzer (1923); Hynes (1970); Grand Canyon Wildlands Council (2002)
Hillslope	Emerges from confined or unconfined aquifers on a hillslope (30–60° slope); often indistinct or multiple sources	Ram Creek Hot Spring, BC, Canada	Springer et al. (2008)
Hypocrene	A buried spring where flow does not reach the surface, typically due to very low discharge and high evaporation or transpiration	Mile 70L Spring, Grand Canyon, AZ	Springer et al. (2008)
Limnocrene	Emergence of confined or unconfined aquifers in pool(s)	Grassi Lakes, AB, Canada	Modified from Meinzer (1923); Hynes (1970)
(Carbonate) mound-form	Emerges from a mineralized mound, frequently at magmatic or fault systems	Montezuma Well, AZ Dalhousie Springs, Australia	Springer et al. (2008); Zeidler and Ponder (1989)
Rheocrene	Flowing spring, emerges into one or more stream channels	Pheasant Branch, WI, US	Modified from Meinzer (1923); Hynes (1970)



Cave spring- common in karst terrain. Cave type springs are most likely to occur in the «mature» to «extreme» karst conditions where the conduits are sufficiently large enough to allow for emergence.

Hanging gardens are complex, multi-habitat springs emerging along geologic contacts and seep, drip or pour onto underlying walls.

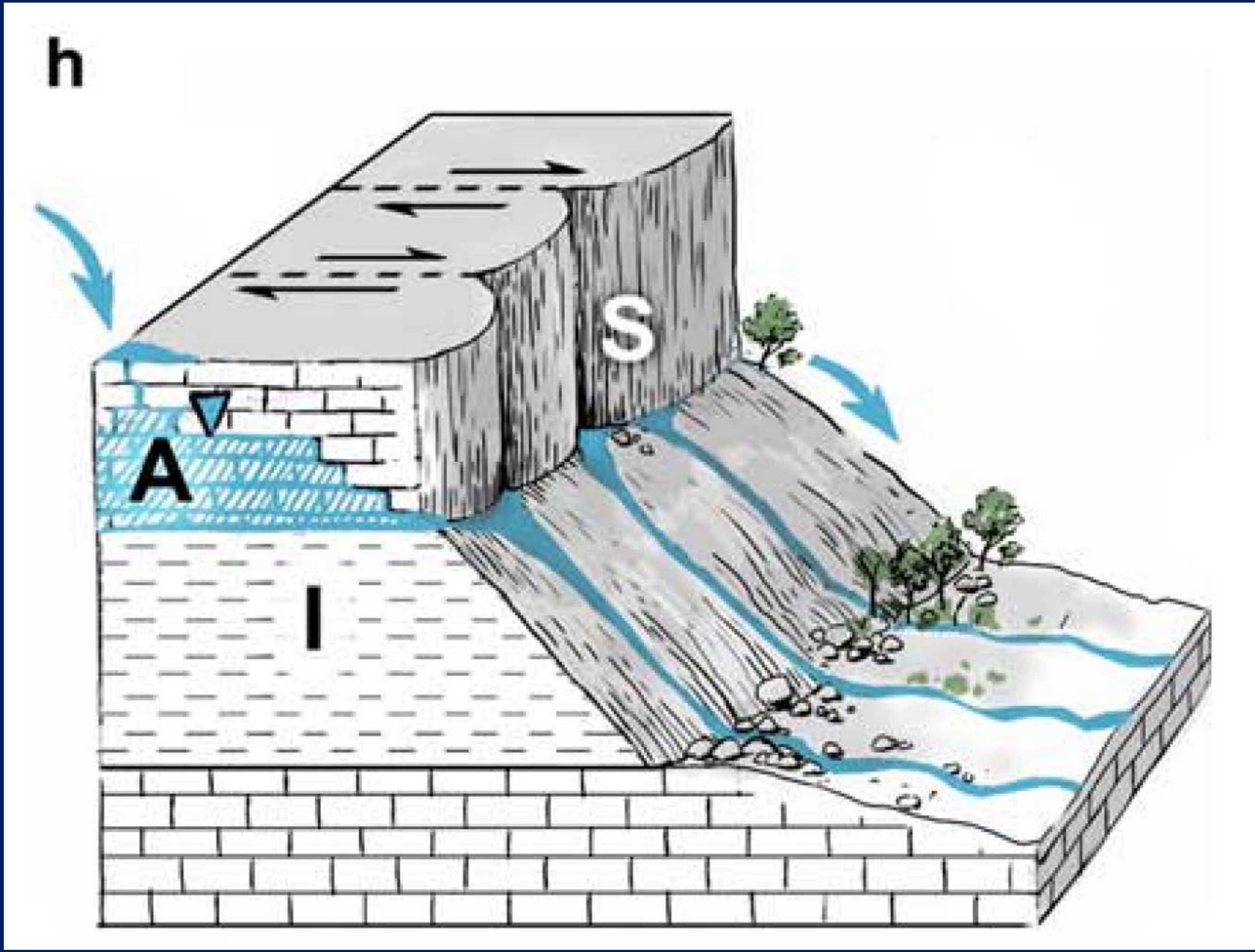


Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93

Poison Ivy Spring, Arches National Park, Utah, USA

Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93

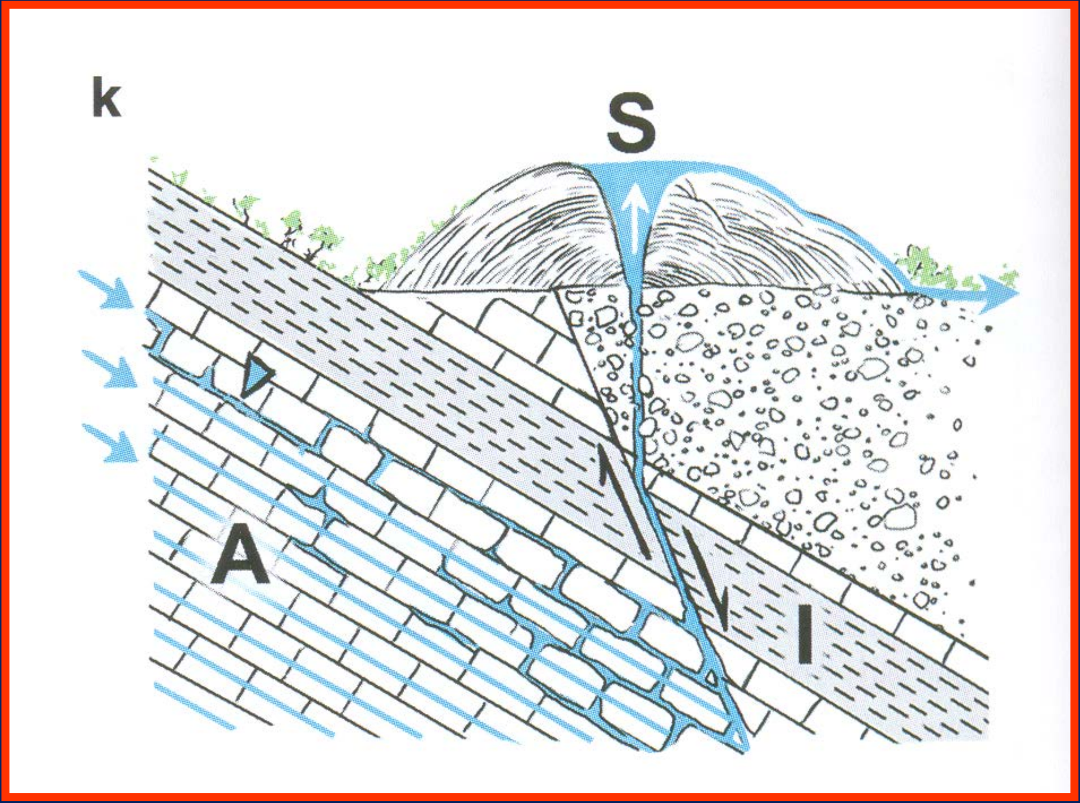
Hillslope springs emerge from confined or unconfined aquifers on non-vertical hillslopes at 30-60° slopes, and usually have indistinct or multiple sources.



Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93

Groundwater flowing out of Minnie Miller Springs, Idaho, USA.

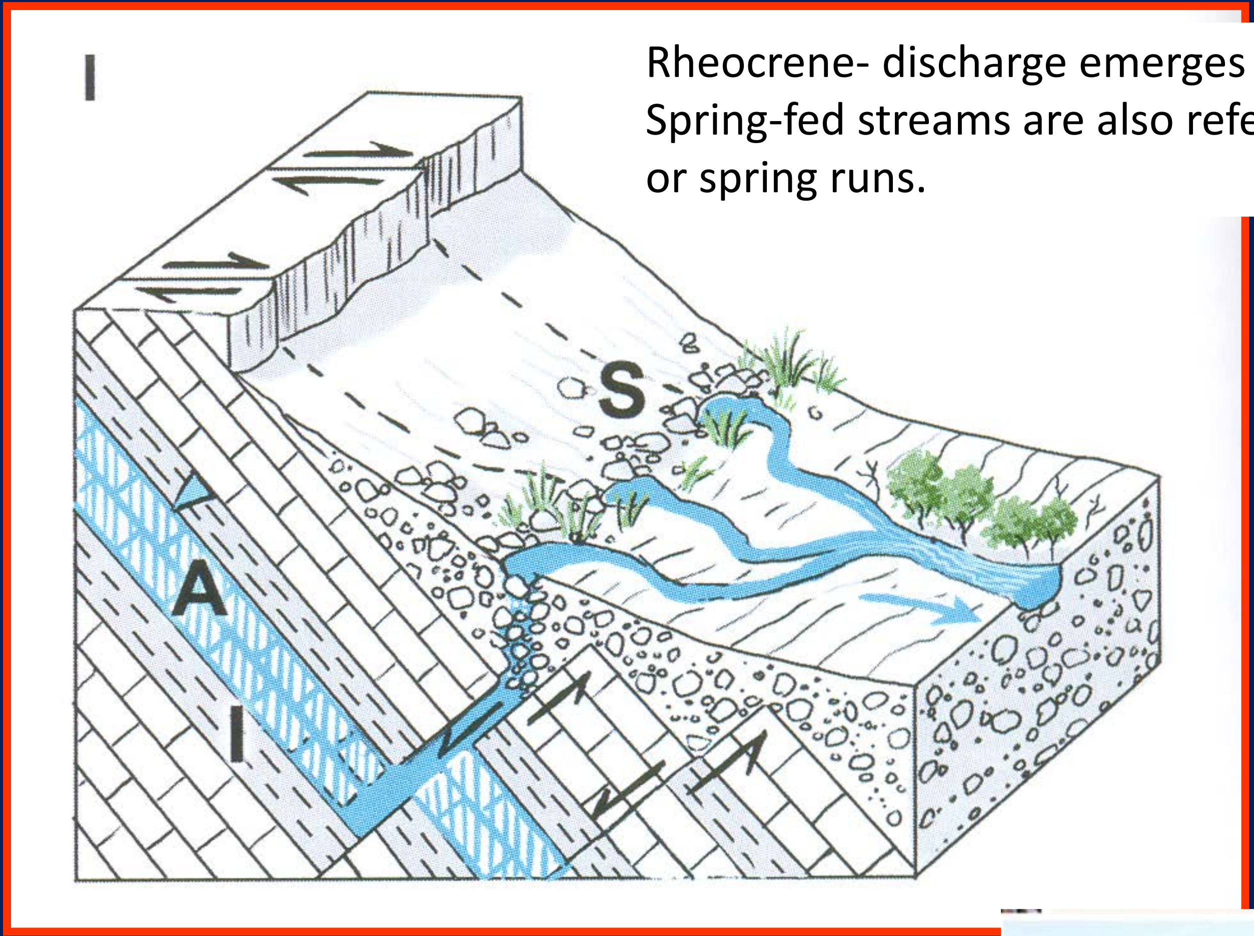
Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93



Mound-form springs emerge from (usually carbonate) precipitate mounds or peat mounds. Travertine-forming mound springs are often located along active magmatic or fault systems and therefore may be hot.

Mound form spring

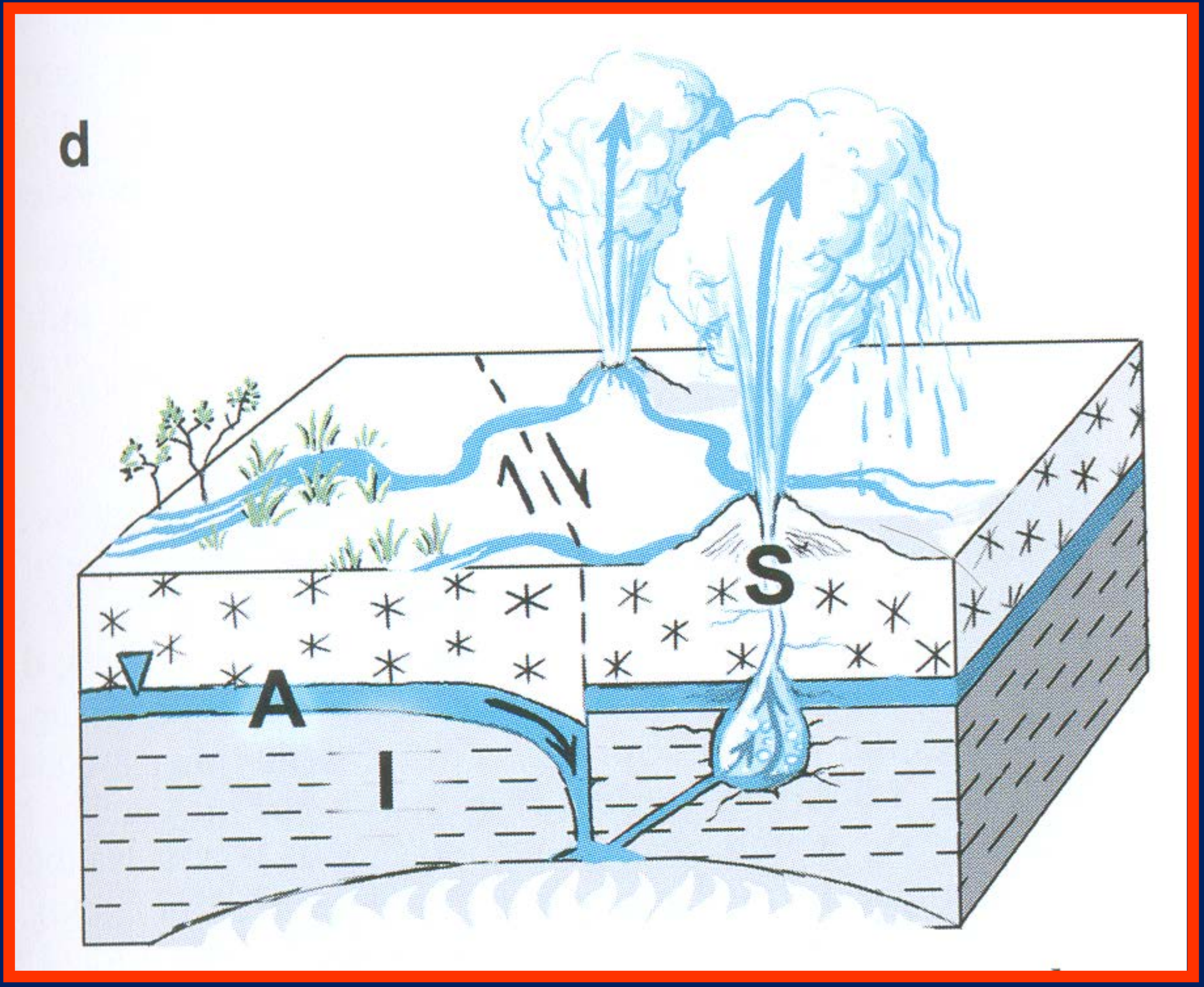
Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93



Rheocrene- discharge emerges as flowing streams. Spring-fed streams are also referred to as springbrooks or spring runs.

Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93

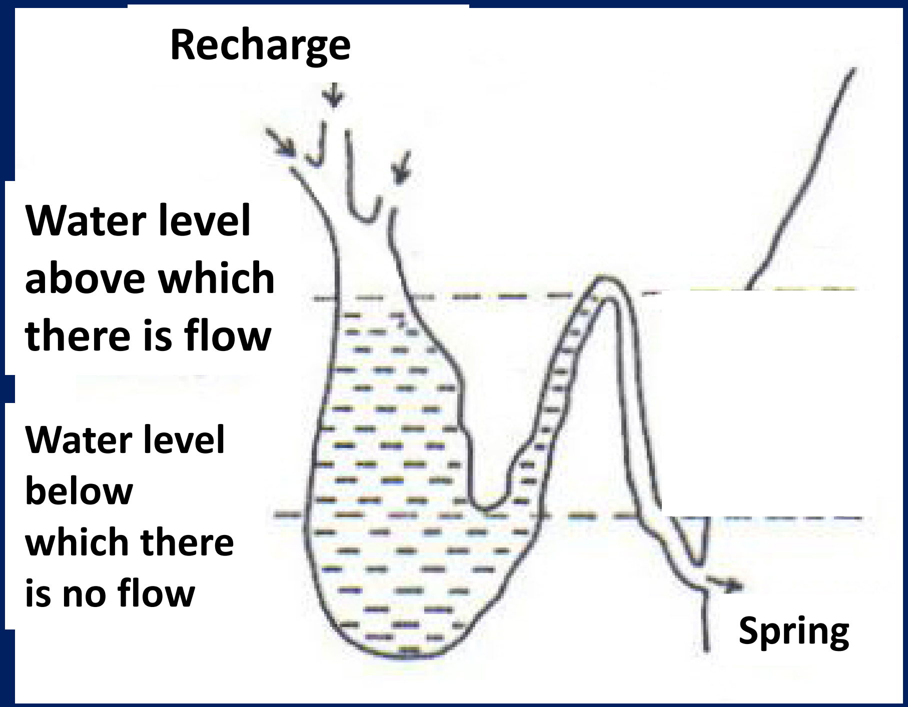
Geyser spring- Geysers are globally rare, geothermal springs that emerge explosively and usally erratically. A geyser is a hot spring characterized by intermittent discharge of water ejected turbulently and accomplished by a vapor phase.

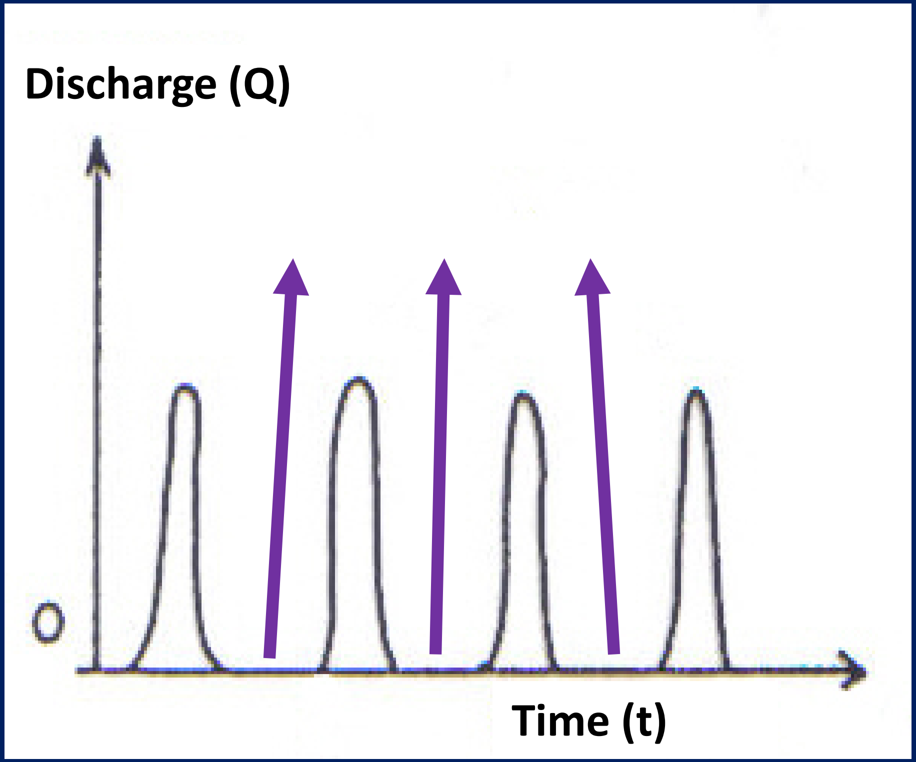


Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93

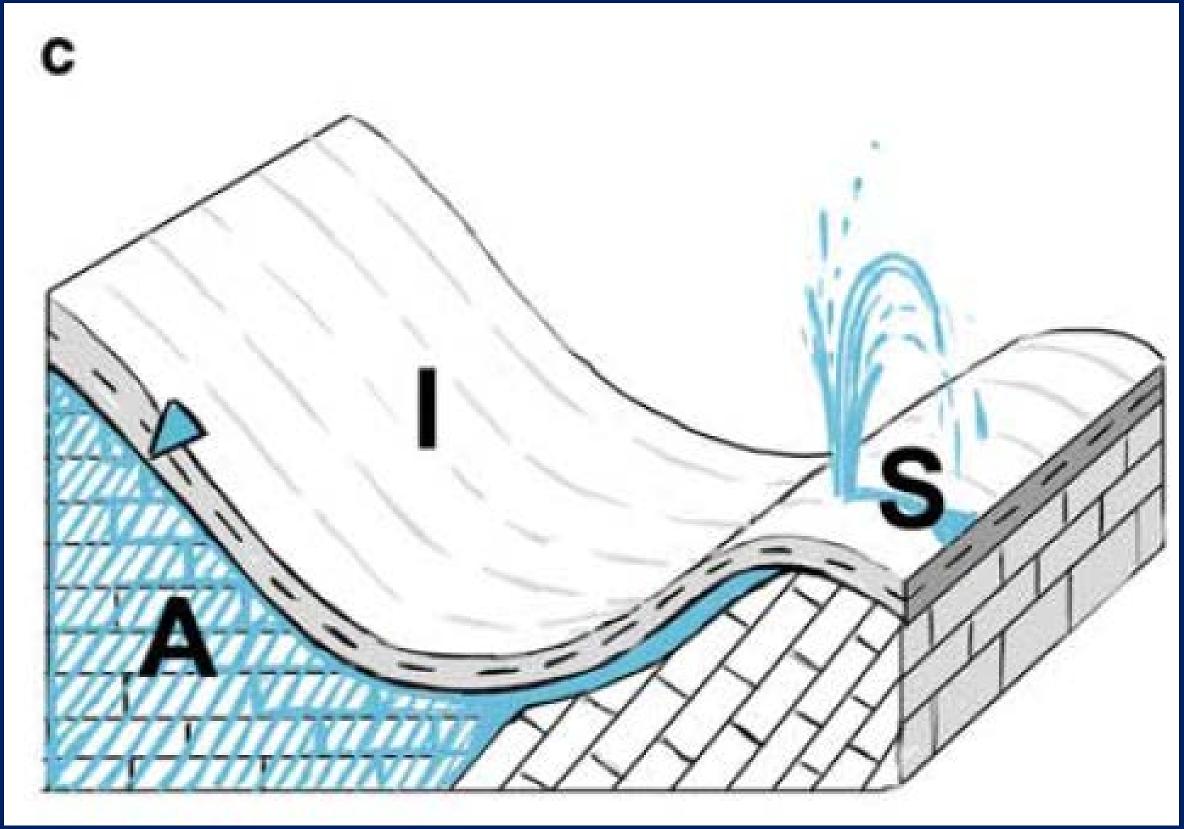
Geyser spring

Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93


Recharge
Water level above which there is flow
Water level below which there is no flow
Spring

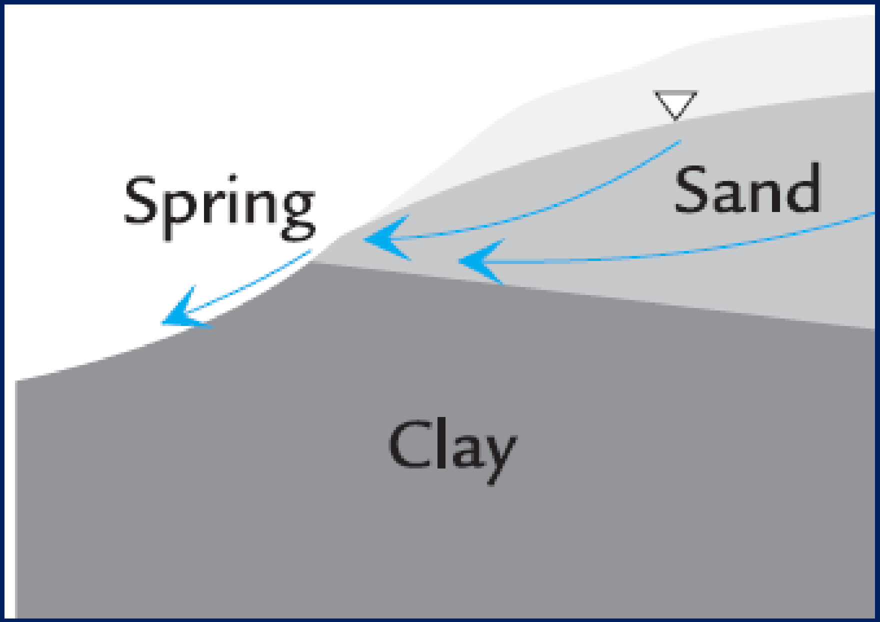

Discharge (Q)
0
Time (t)

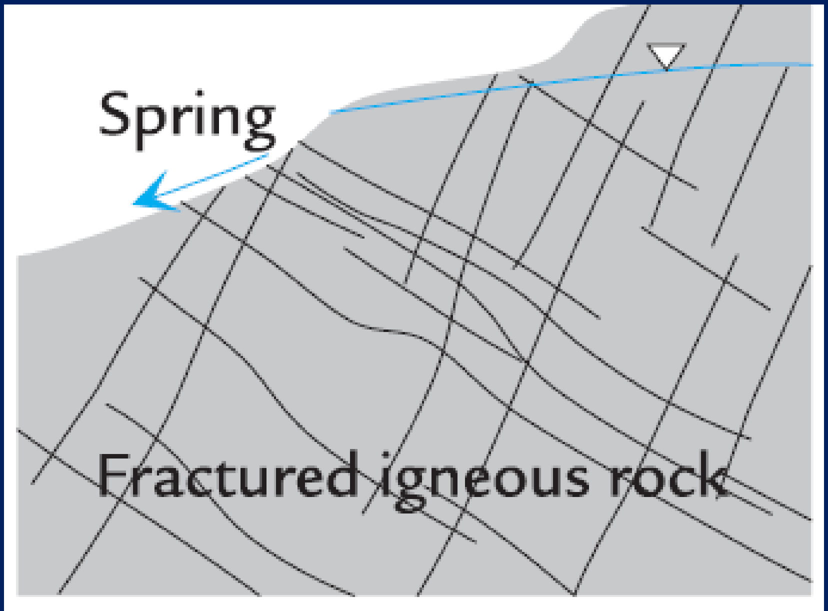
Geyser

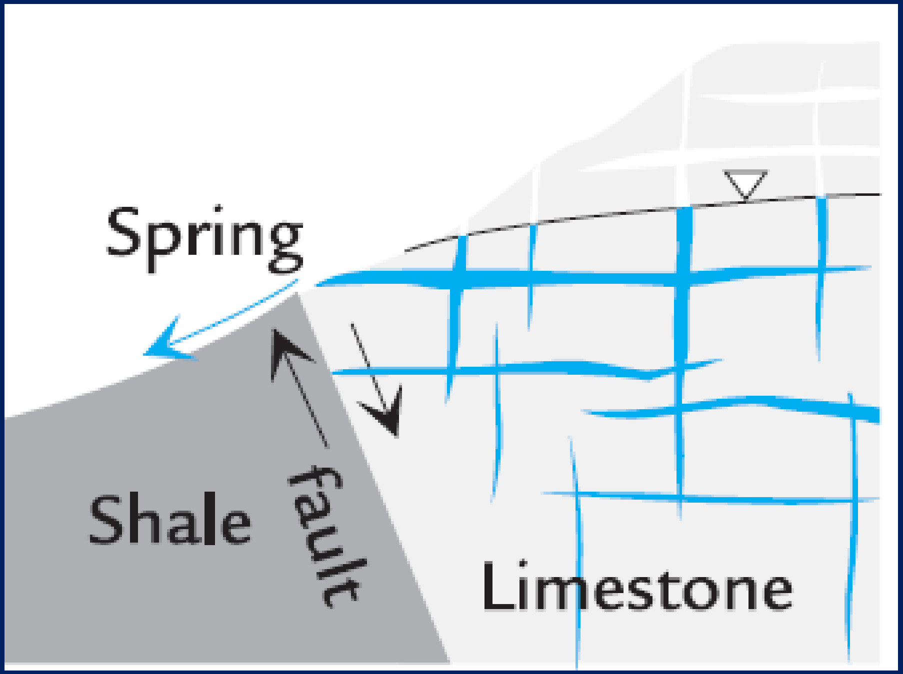


Fountain springs are cool-water artesian springs that are forced above the land surface by stratigraphic head-driven pressure or CO_2. Discharge at fountain springs is not driven by thermal processes, such as geysers, but still require a confined aquifer with water pressurized by CO_2, not heat.

Springer, A. E., Stevens, L. E. 2009. Spheres of discharge of springs. Hydrogeology Journal 17, 83-93





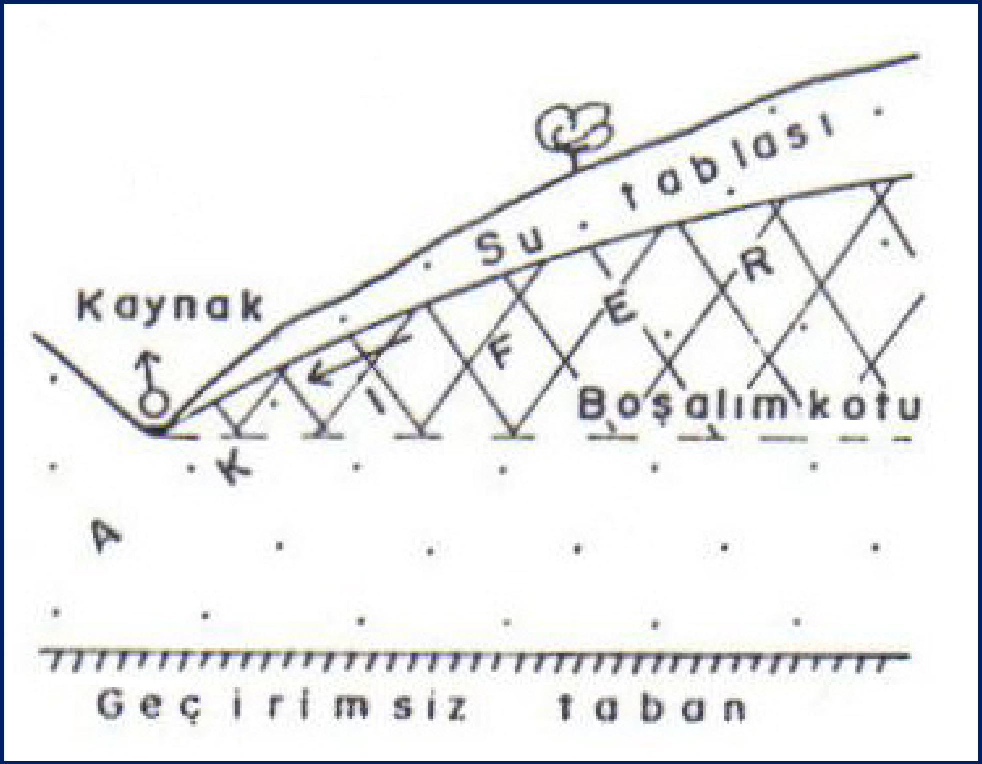


Springs generally form where fractures or the aquifer base cut topography.

Spring formation in different geological environments

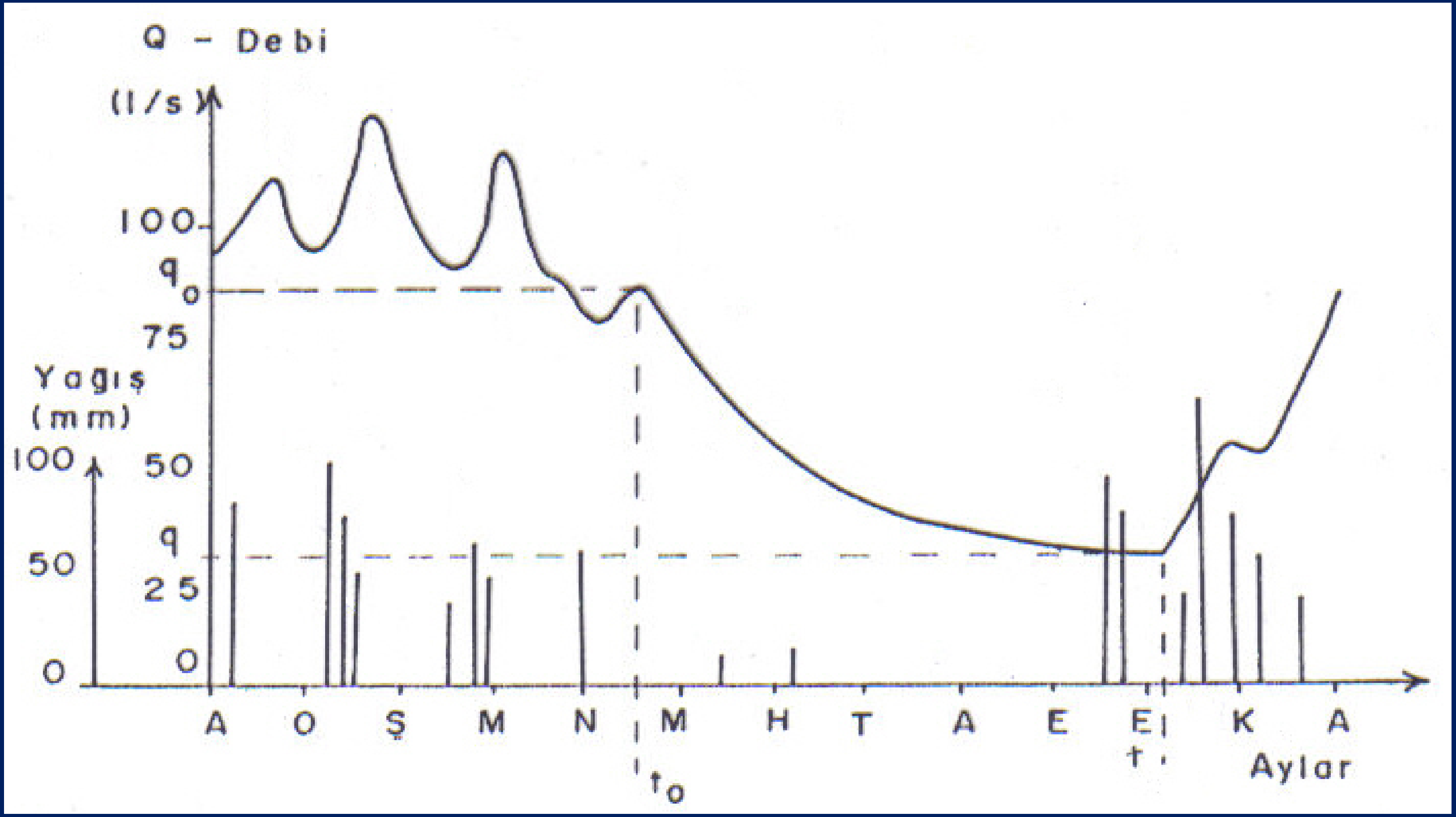
SPRING DISCHARGE HYDROGRAPH
Responses of spring discharge to precipitation
RECESSION CURVES

Akiferin dış etkilerle beslenemediği dönemde kaynaklar akiferin boşalım kotu üzerindeki rezervini boşaltırlar. Bu devredeki rejime AKİFERİN GERÇEK REJİMİ DENİR.



Kaynak boşalım kotu

Kaynağın beslenme havzasında yağışın hiç olmadığı veya akifere etki etmeyecek kadar azaldığı dönemde kaynak debisinde meydana gelen değişimler yeraltısuyu boşalım eğrileri ile gösterilir.

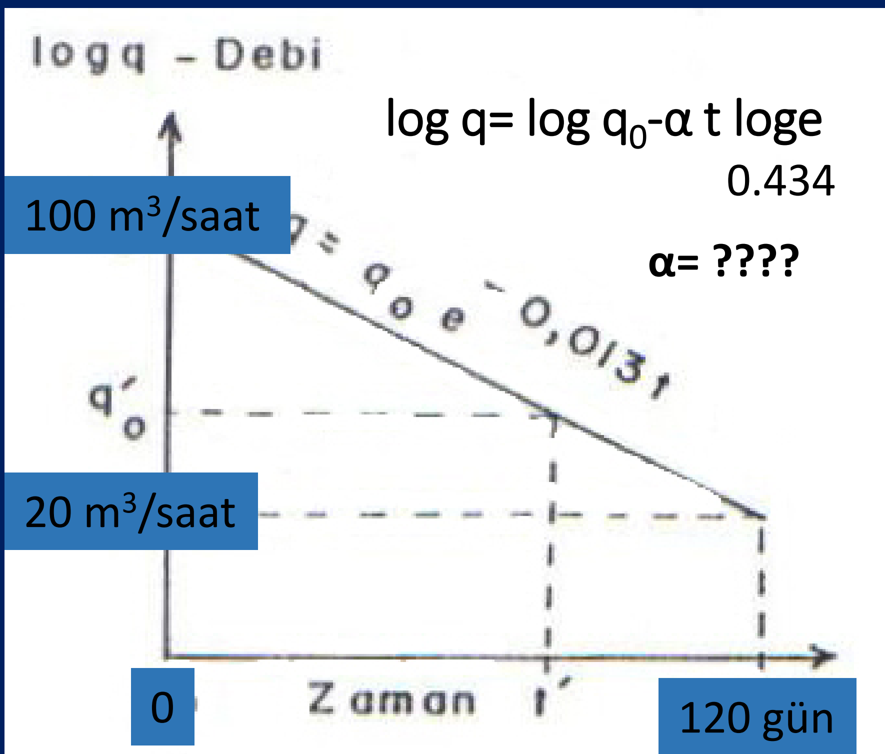


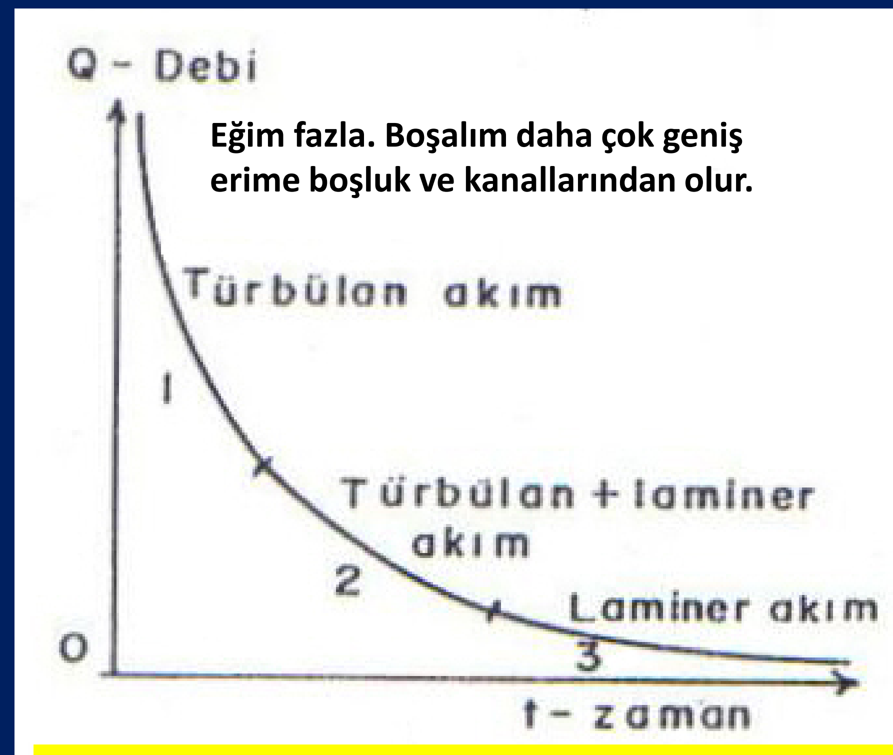
Kaynağın debi değişim grafiği

Kaynakların gerçek rejimdeki boşalım grafiğinin temel formülü Maillet (1905) tarafından verilmiştir.

$$q = q_0 e^{-\alpha(t-t_0)}$$

q= t zamanındaki debi m3/s
q_0 = t_0 zamanındaki debi yani gerçek rejimin başlangıç debisi m^3/s
α = Boşalım katsayısı 1/gün
e: 2.718

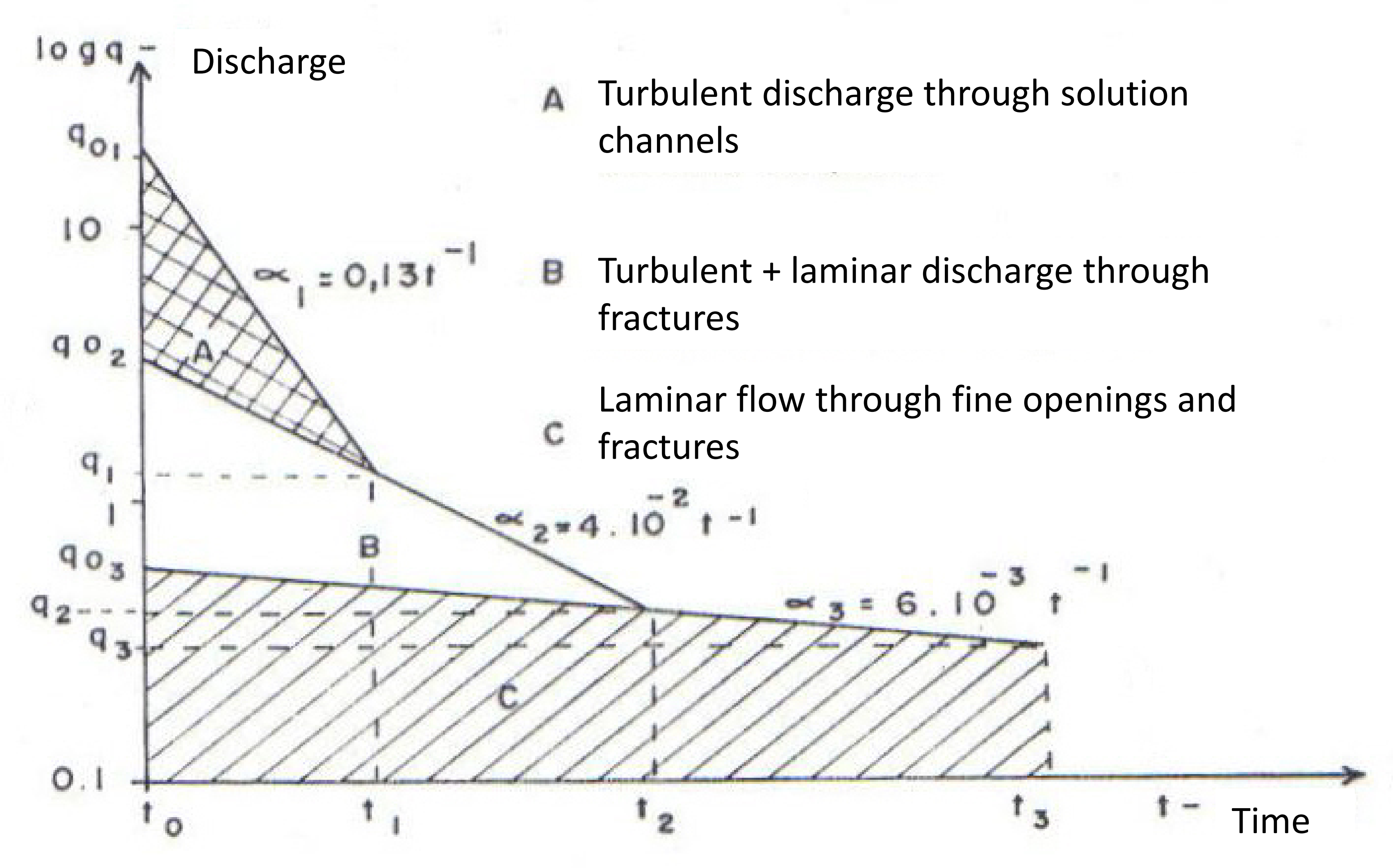




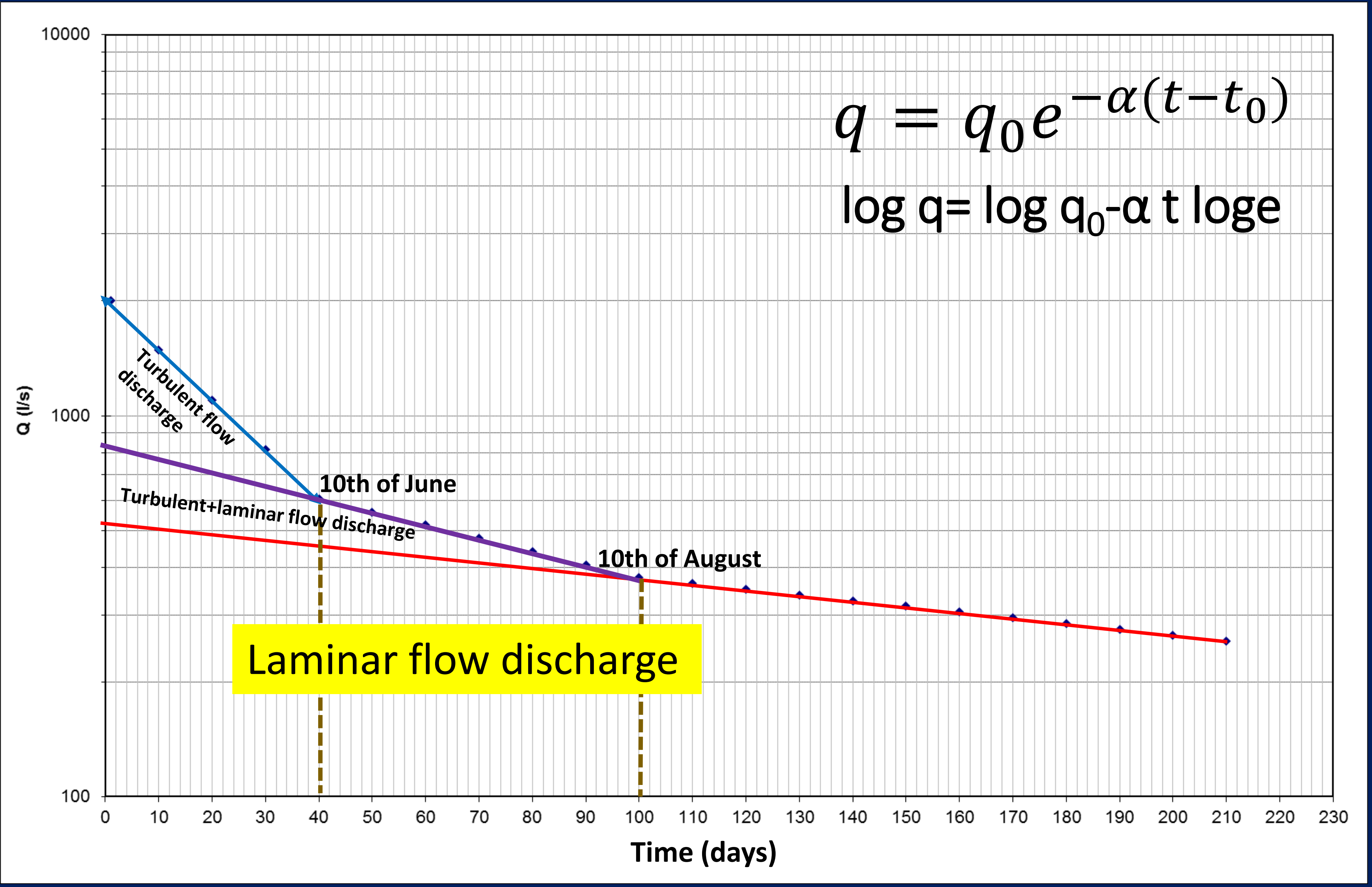
Yağışsız dönem uzadıkça kaynak debisi giderek azalır, sonuçta kaynak kurur.

log q= f(t) boşalım grafiği

Q=f(t) grafiği ve akım türleri



Components of a semi-log hydrograph of a spring



$$q = q_0 e^{-\alpha(t-t_0)}$$
log q= log q_0-α t loge
10000
1000
100
Q (l/s)
Turbulent flow discharge
10th of June
Turbulent+laminar flow discharge
10th of August
Laminar flow discharge
0
10
20
30
40
50
60
70
80
90
100
110
120
130
140
150
160
170
180
190
200
210
220
230
Time (days)

The storage capacity of a karstic spring:

$$Vo = \int_0^{\infty} q dt$$

....

....

$$V_0 = \frac{q_0 86400}{\alpha} \quad \text{1/day}$$

If q_0 m³/s then Vo m³

Çizelge 1. Karstik kaynağın zamana bağlı boşalım (debi) değerleri

Zaman (t, gün)	Q,l/s
1 Mayıs	2000
10 Mayıs	1485
20 Mayıs	1100
30 Mayıs	815
10 Haziran	605
20 Haziran	559
30 Haziran	516
10 Temmuz	477
20 Temmuz	440
30 Temmuz	406
10 Ağustos	375
20 Ağustos	362
30 Ağustos	350
10 Eylül	338
20 Eylül	327
30 Eylül	316
10 Ekim	305
20 Ekim	295
30 Ekim	285
10 Kasım	275
20 Kasım	265
30 Kasım	256

Toros kuşağında yer alan Mesozoyik yaşlı karstik kireçtaşları ile Eosen yaşlı geçirimsiz filiş birimlerinin dokanağından boşalan bir kaynakta zamana bağlı olarak debiler ölçülmüştür (Çizelge 1). Mesozoyik kireçtaşları Eosen fliş üzerine bindirmeli olarak gelmektedir.

Sorular:

a) Kaynağın boşalım katsayılarını bularak, boşalım doğru denklemlerini oluşturunuz.

b) Kaynağın boşalım kotu üzerinde sadece **laminer** akımla boşaltabileceği su miktarını hesaplayınız

c) Kaynağın boşalım kotu üzerinde sadece **laminer** akımla 20 Eylül'e kadar boşaltacağı su miktarını hesaplayınız

d) Kaynağın 20 Eylül-30 Ekim arasında türbülan akımla boşalttığı su miktarını hesaplayınız

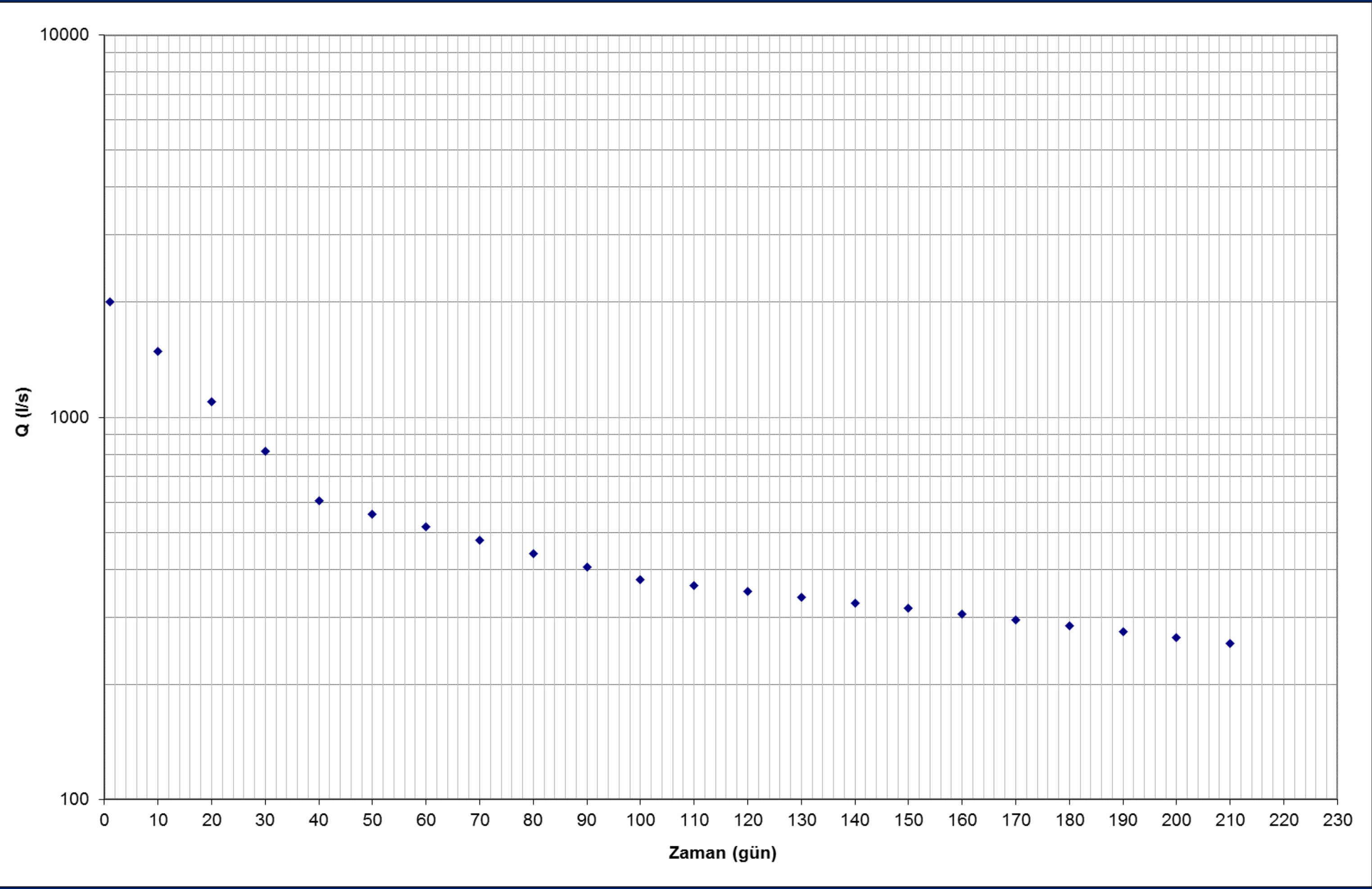
e) Kaynağın **sadece türbülan** akım ile boşaltabileceği su miktarını hesaplayınız.

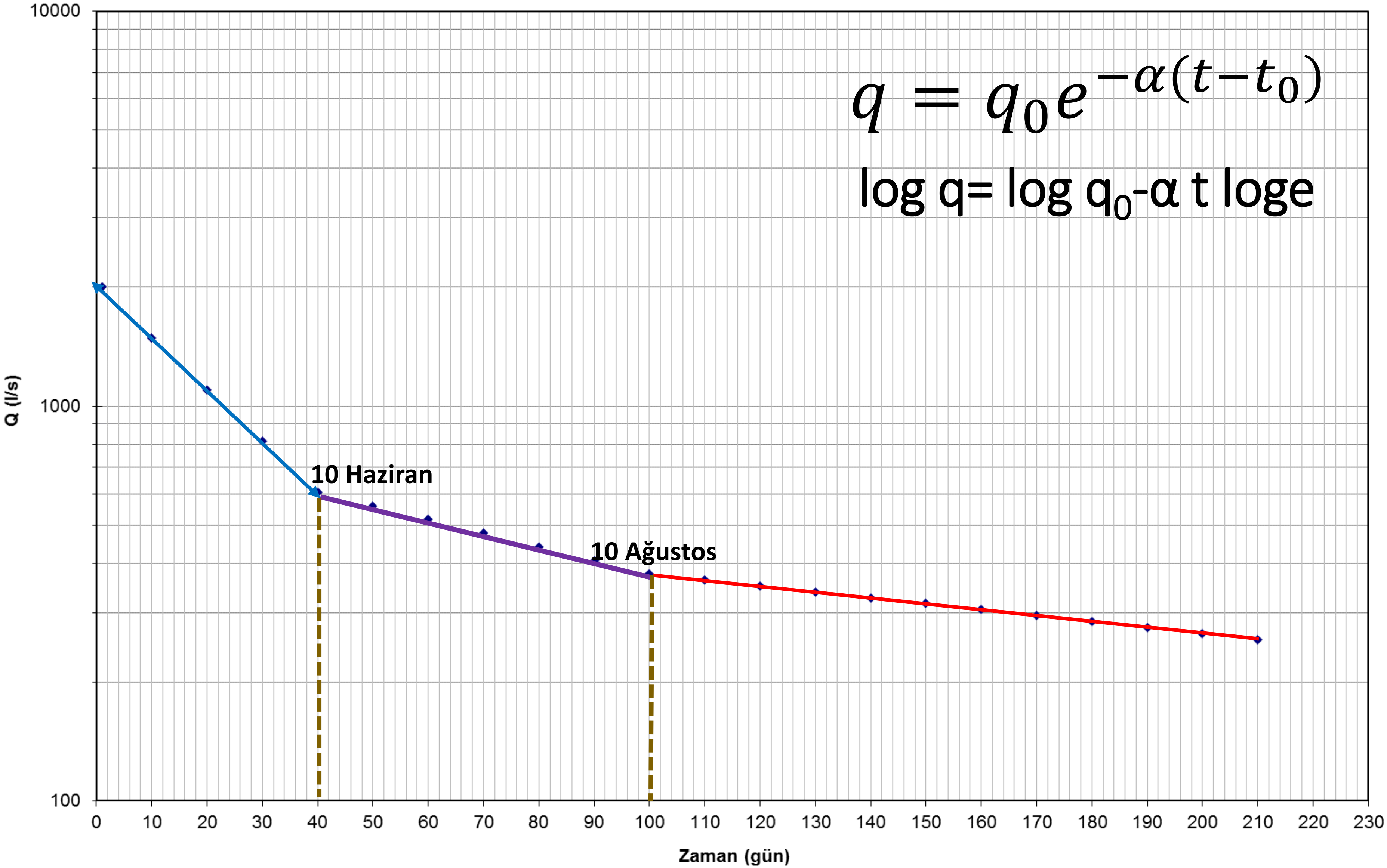
f) Kaynağın boşalım kotu üzerinde boşaltabileceği su miktarını hesaplayınız.

g) Kaynağın türbülan+ laminer akım ile boşaltabileceği su miktarını hesaplayınız.

h) Kaynağın 20 Ekim-20 Kasım arasında boşaltacağı su miktarını hesaplayınız.

Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210





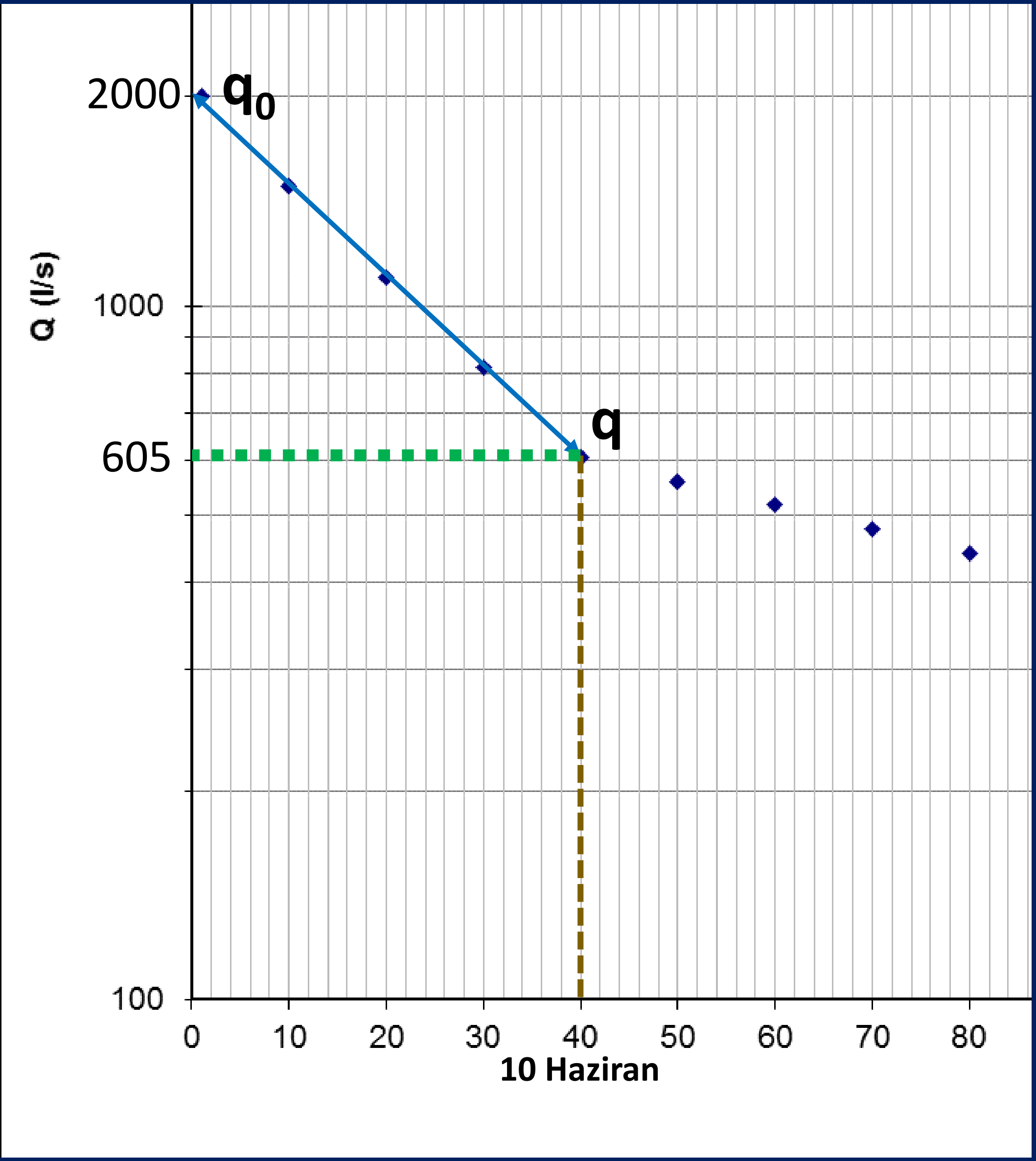
$$q = q_0 e^{-\alpha(t-t_0)}$$

$$\log q = \log q_0 - \alpha\, t \log e$$

$$q = q_0 e^{-\alpha(t-t_0)}$$

$\log q = \log q_0 - \alpha\, t \log e$

$\log 605 = \log 2000 - \alpha * 40 * 0.434$



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
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20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
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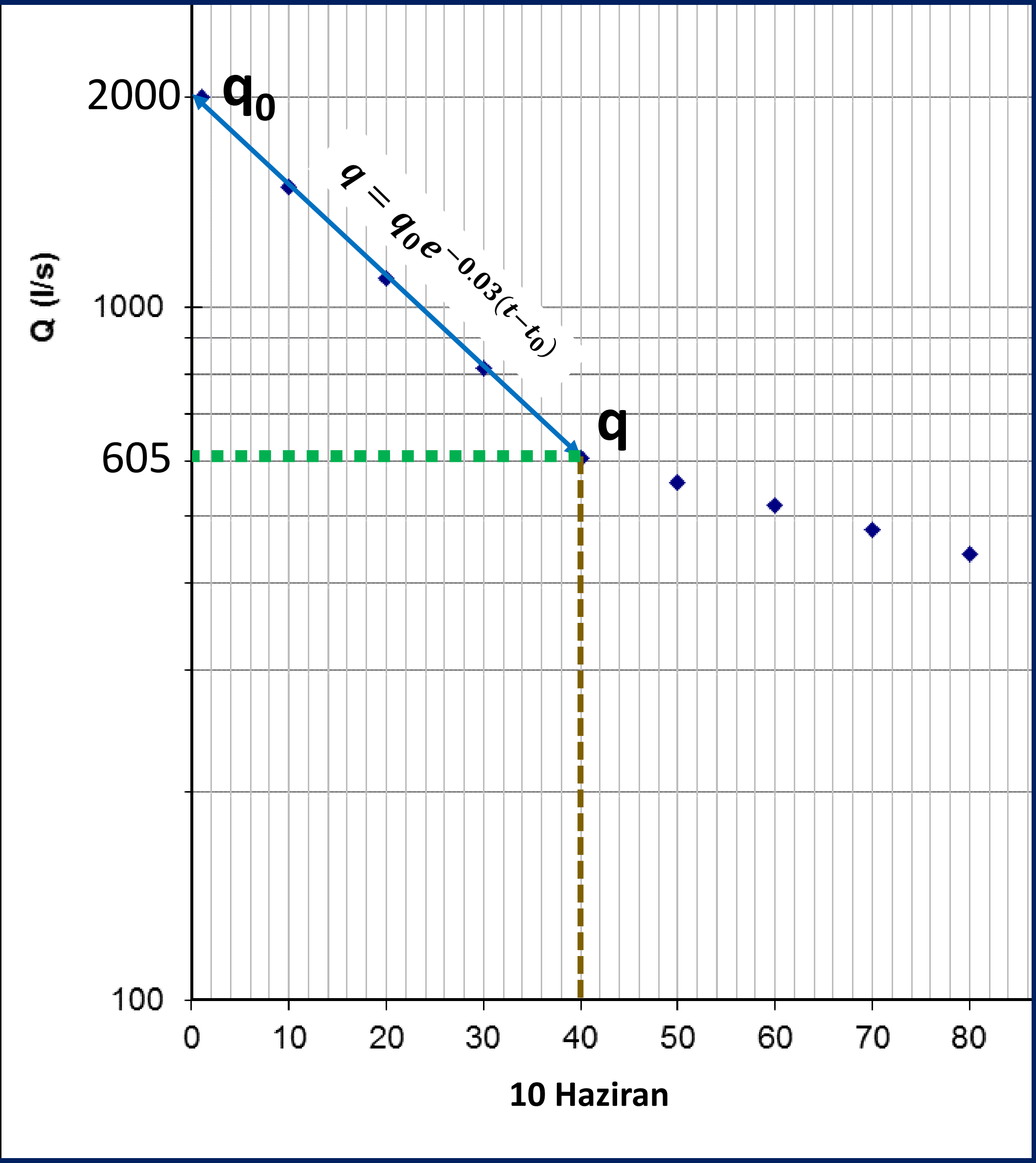
30 Kasım	256	210



$\log q = \log q_0 - \alpha\, t \log e$

$\log 605 = \log 2000 - \alpha * 40 * 0.434$

Dikkat 40 günde debide 1395 lt/s azalma gerçekleşmiş.

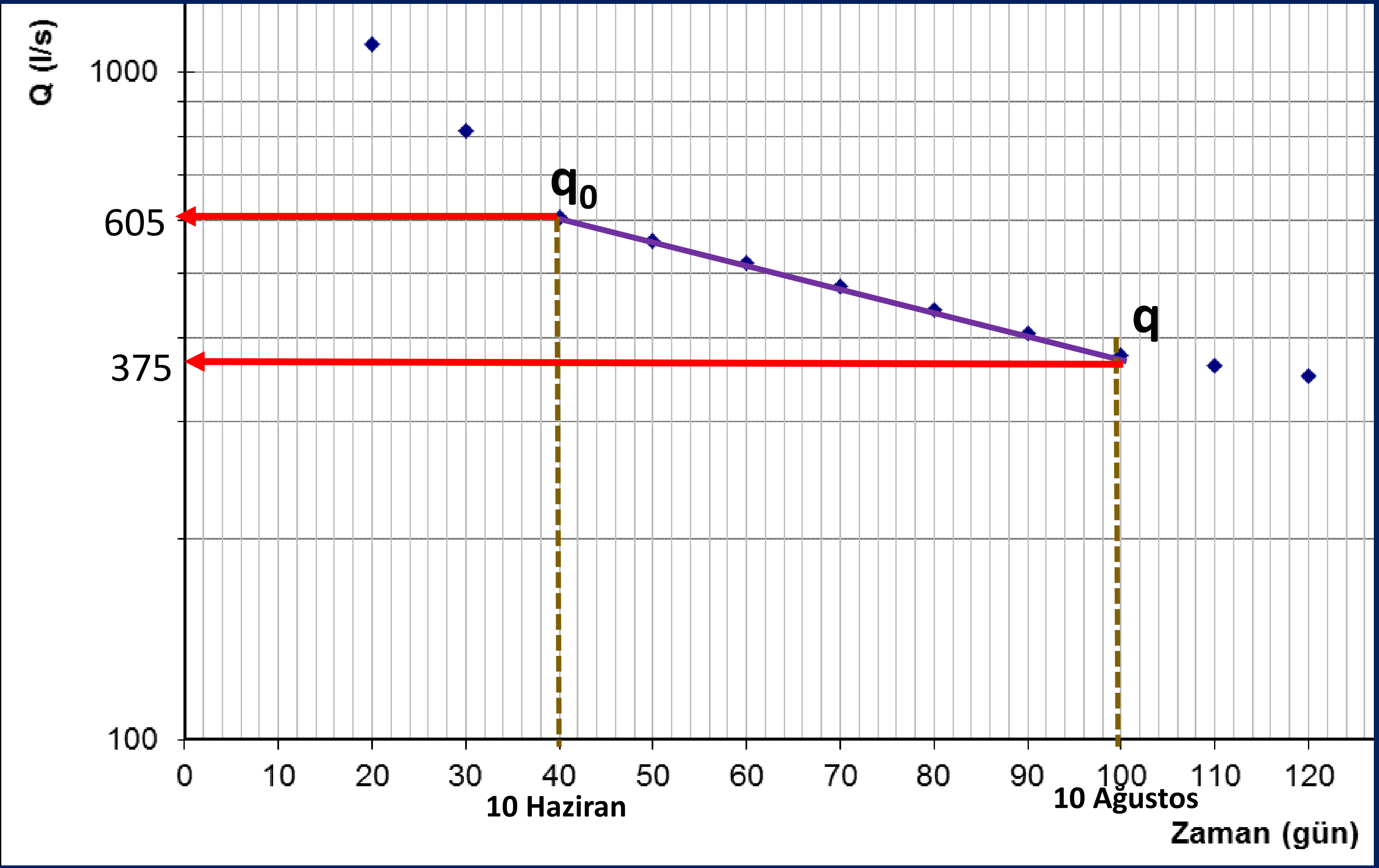
Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

$$q = q_0 e^{-\alpha(t-t_0)}$$

log q= log q_0-α t loge

log375=log605-α*60*0.434

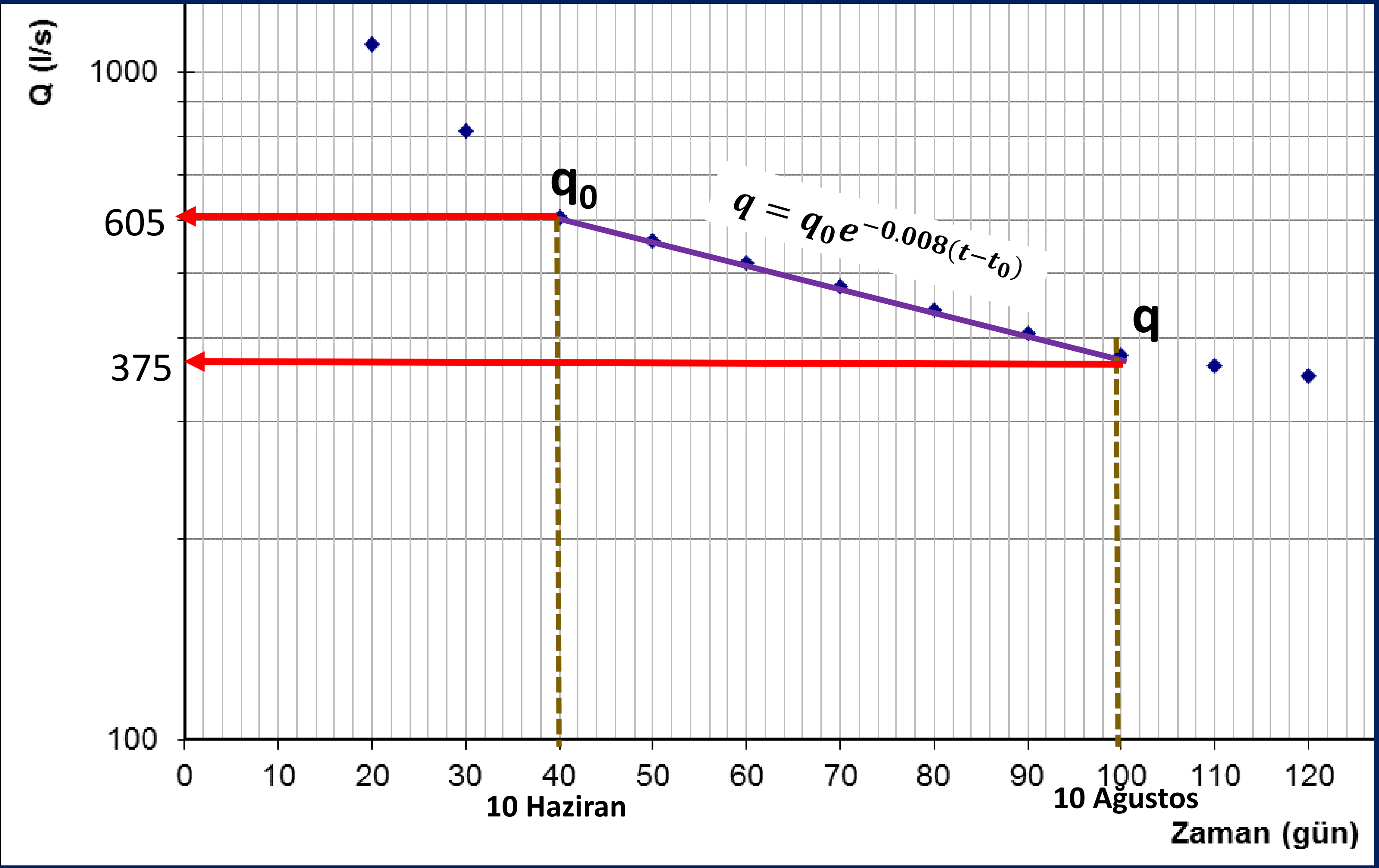
Dikkat 60 günde debide 1395 lt/s azalma gerçekleşmiş.



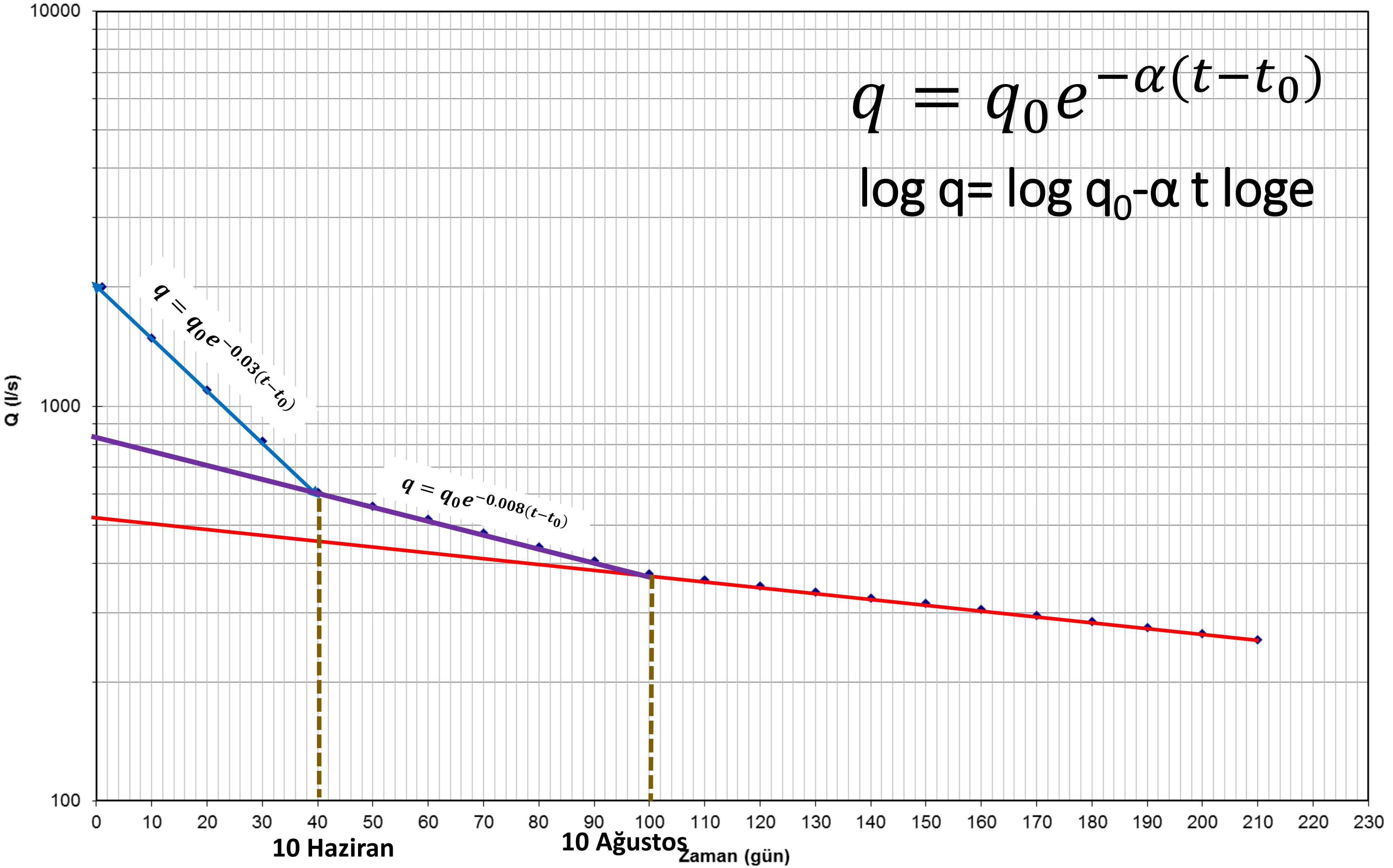
Dikkat 60 günde debide 1395 lt/s azalma gerçekleşmiş.

$$\log q = \log q_0 - \alpha\, t \log e$$

$$\log 375 = \log 605 - \alpha * 60 * 0.434$$

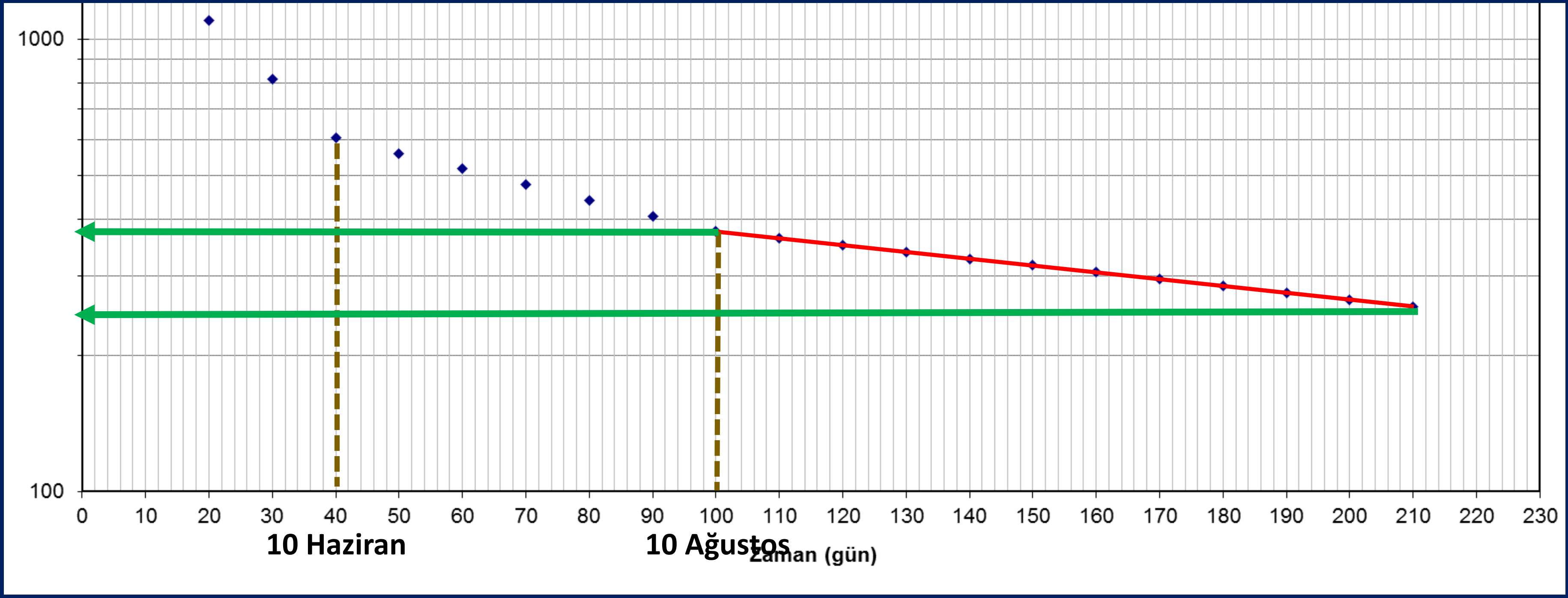


Her zaman $\alpha_1 > \alpha_2 > \alpha_3$



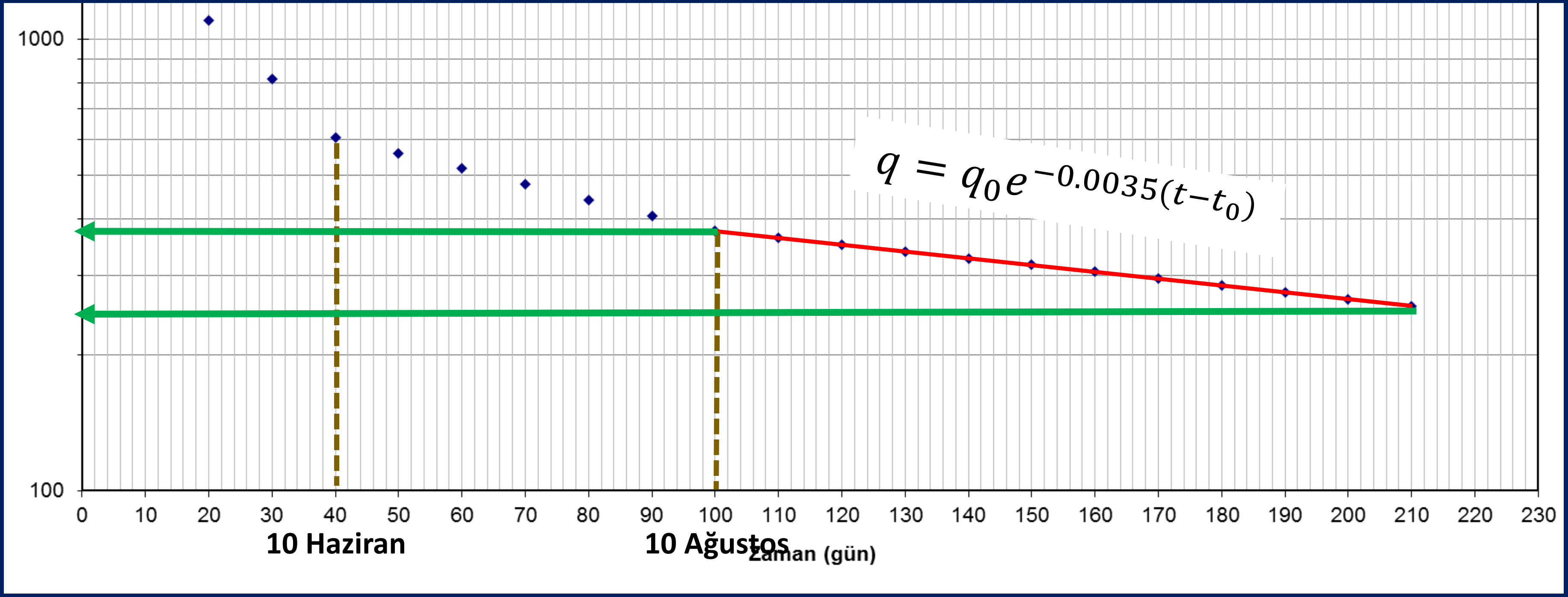
$$\log 255 = \log 375 - \alpha * 110 * 0.434$$

Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

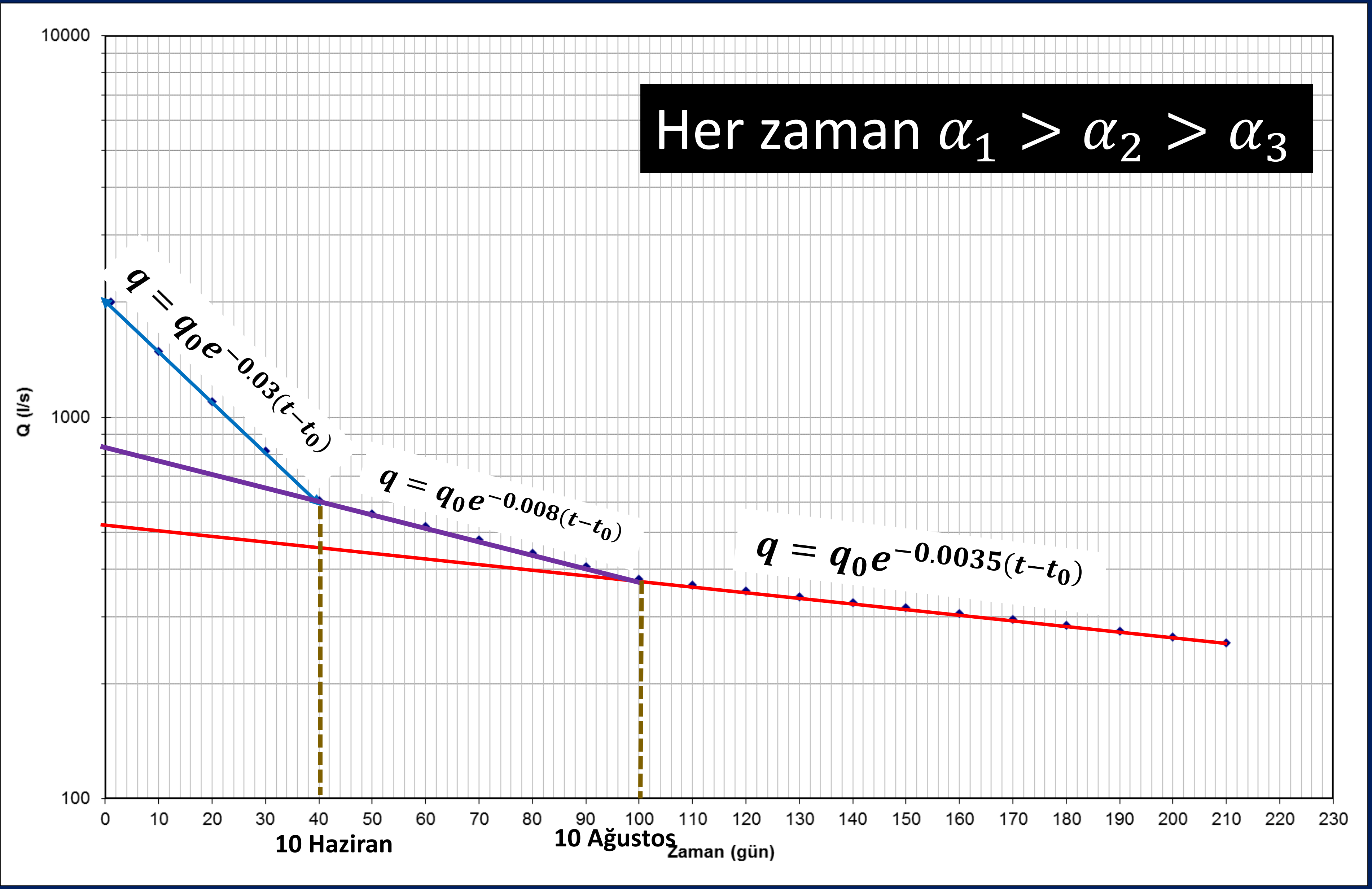


log255=log375-α*110*0.434

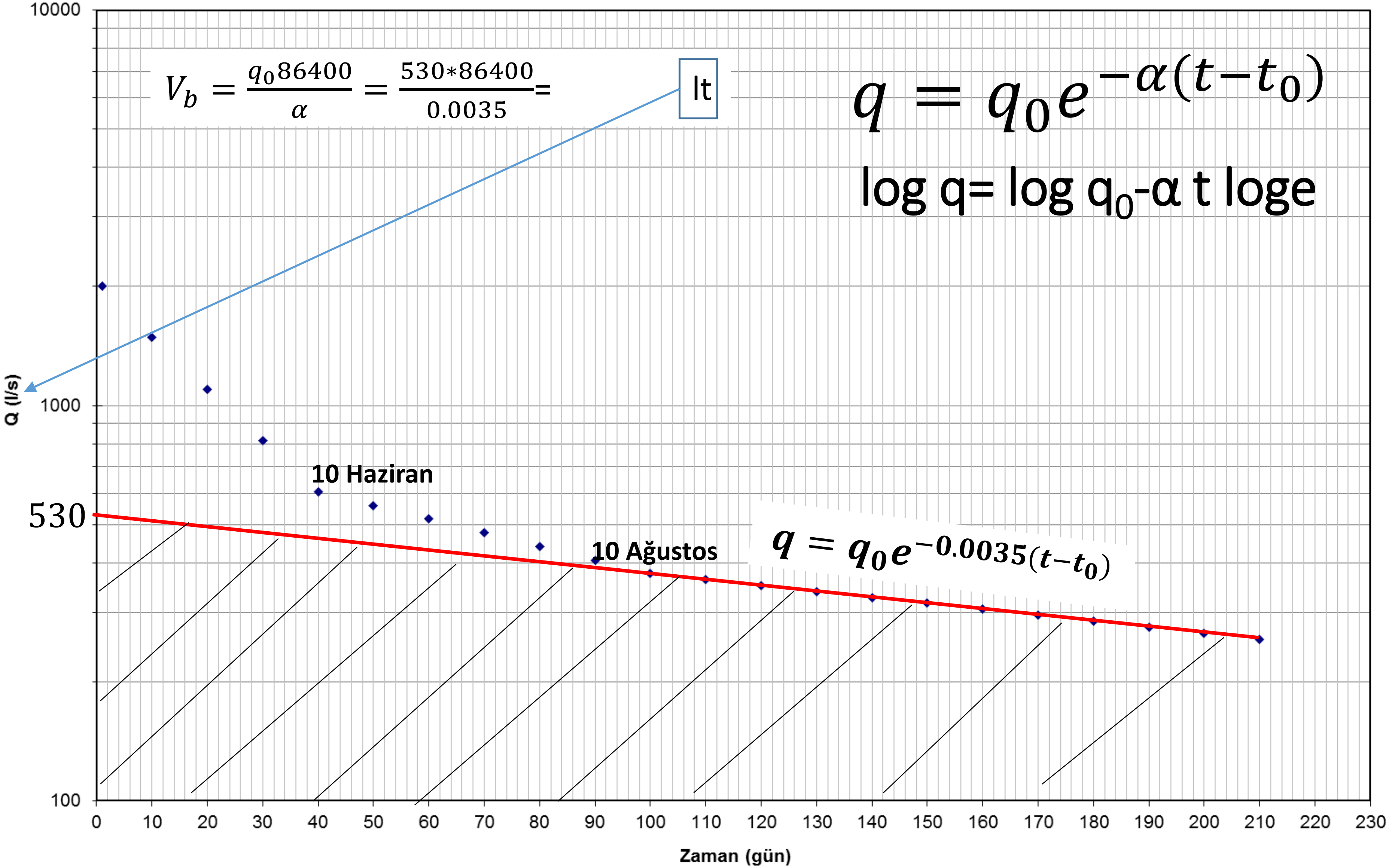
Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210



a) Kaynağın boşalım katsayılarını bularak, boşalım doğru denklemlerini oluşturunuz. ✓

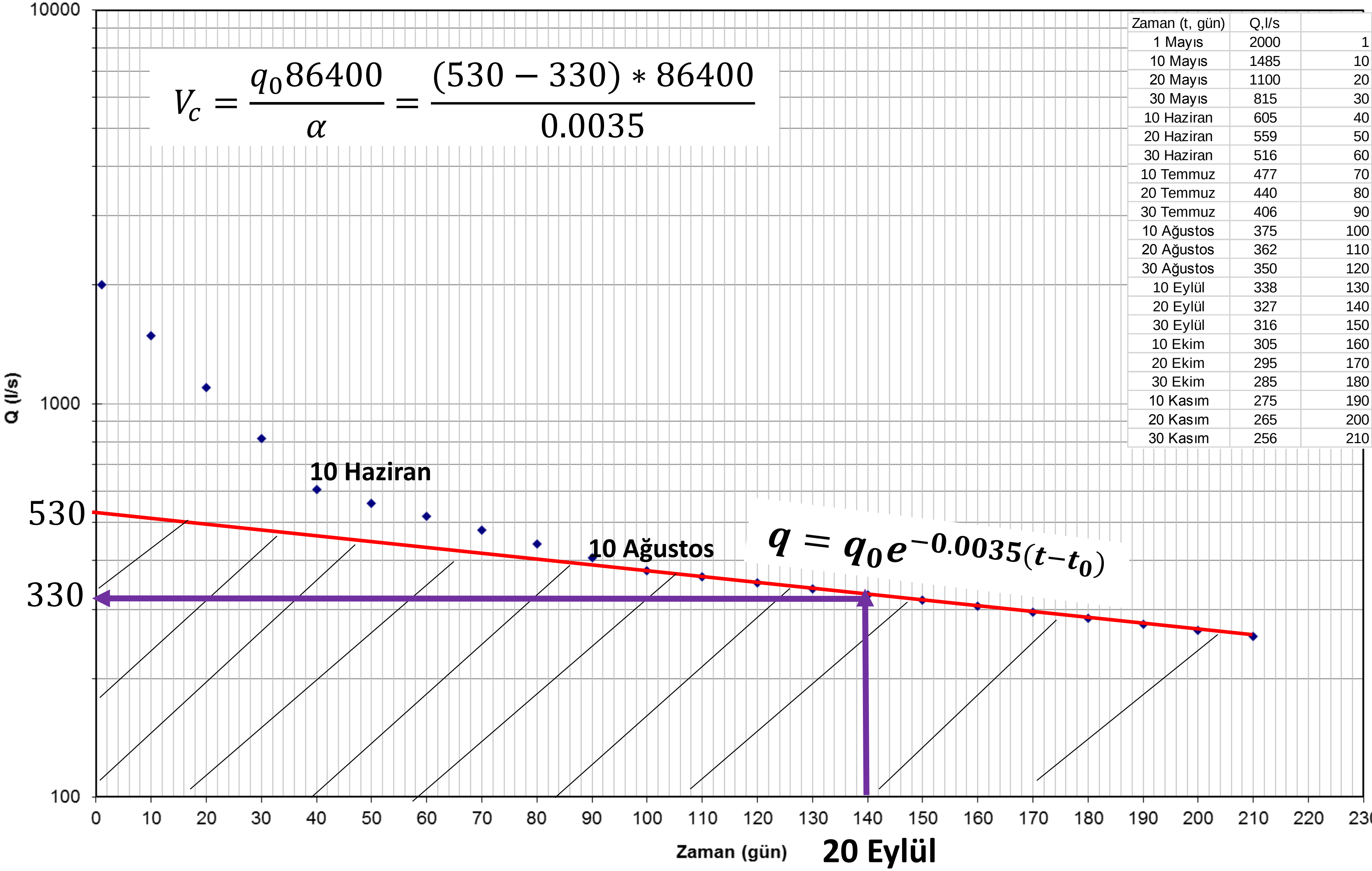


b) Kaynağın boşalım kotu üzerinde sadece **laminer** akımla boşaltabileceği su miktarını hesaplayınız ✓



c) Kaynağın boşalım kotu üzerinde sadece **laminer** akımla 20 Eylül'e kadar boşaltacağı su miktarını hesaplayınız

$$V_c = \frac{q_0 86400}{\alpha} = \frac{(530-330)*86400}{0.0035}$$



$$q = q_0 e^{-0.0035(t-t_0)}$$

Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
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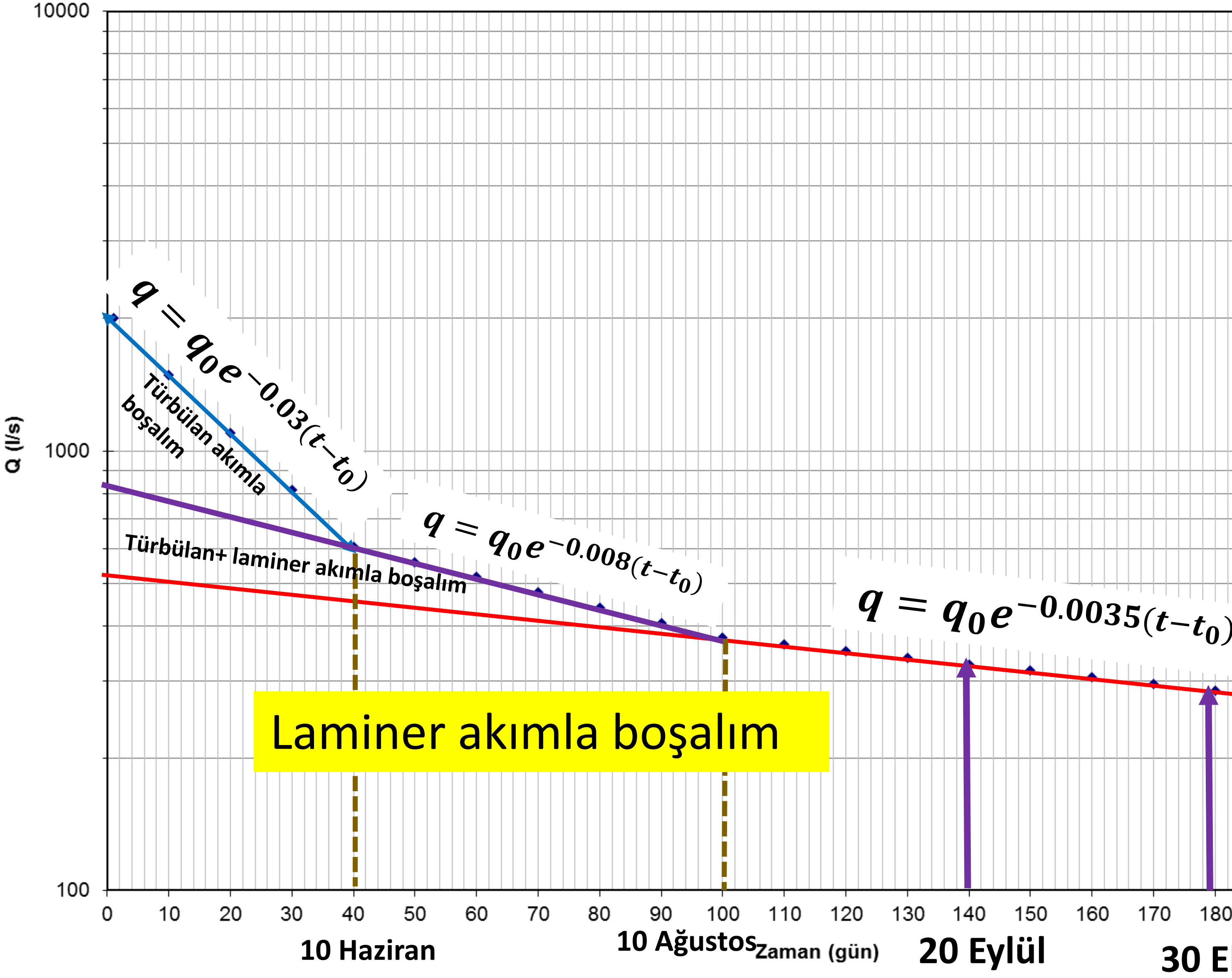
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

d) Kaynağın 20 Eylül-30 Ekim arasında türbülan akımla boşalttığı su miktarını hesaplayınız



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
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30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
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30 Ağustos	350	120
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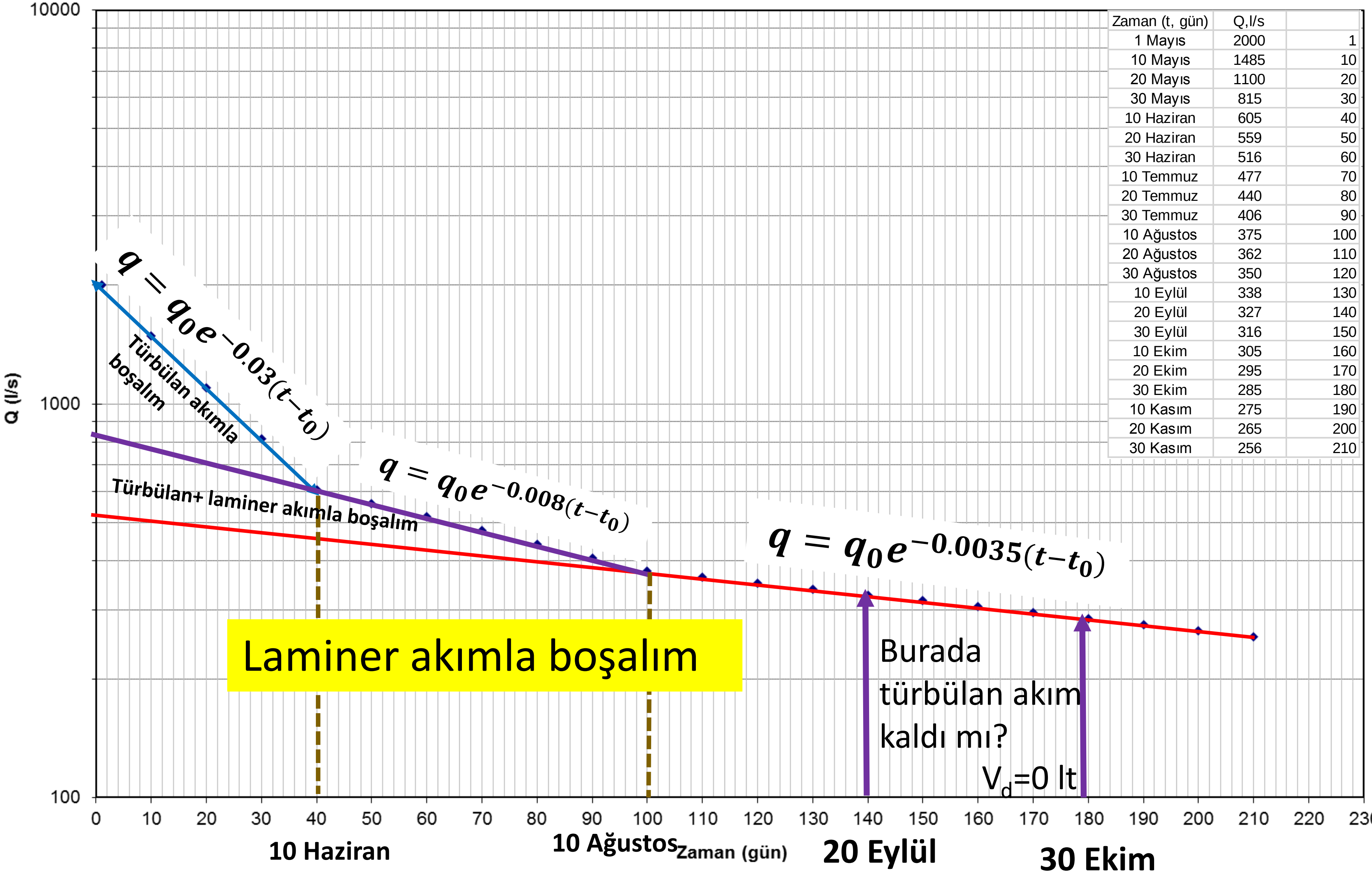
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

d) Kaynağın 20 Eylül-30 Ekim arasında türbülan akımla boşalttığı su miktarını hesaplayınız ✓



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
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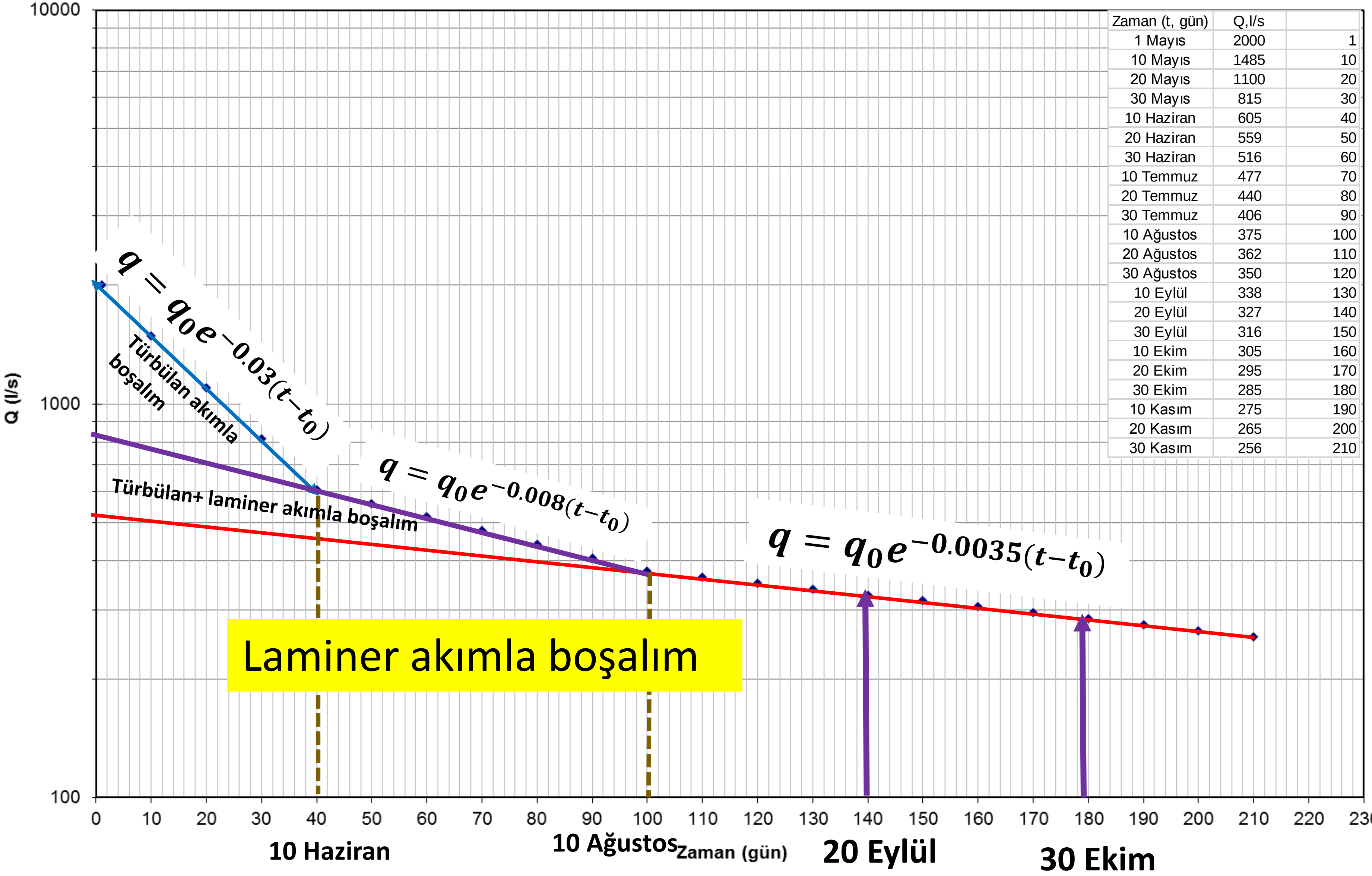
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

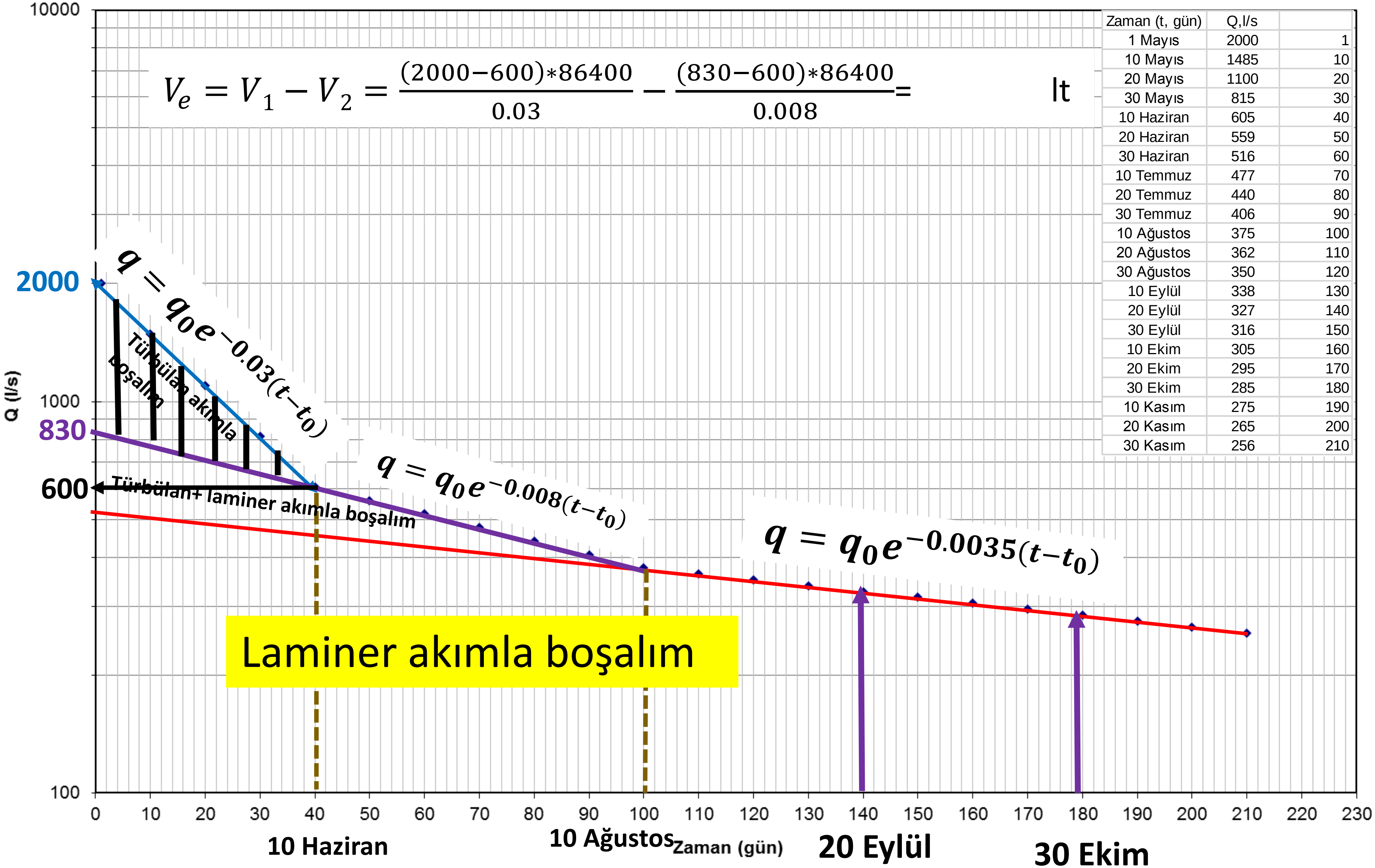
e) Kaynağın **sadece türbülan** akım ile boşaltabileceği su miktarını hesaplayınız.



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

e) Kaynağın **sadece türbülan** akım ile boşaltabileceği su miktarını hesaplayınız.

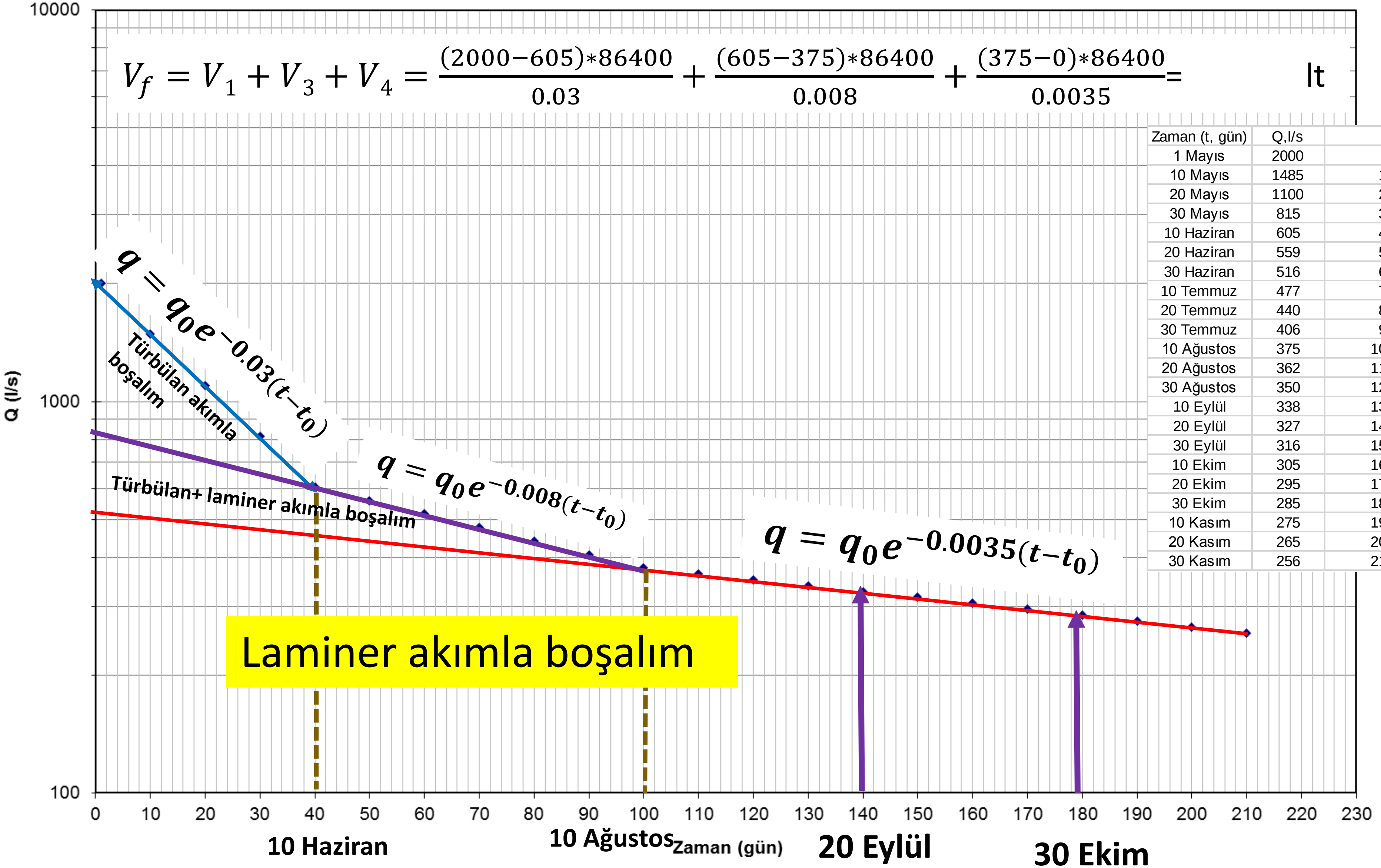
$$V_e = V_1 - V_2 = \frac{(2000-600)*86400}{0.03} - \frac{(830-600)*86400}{0.008} = \quad \text{lt}$$



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

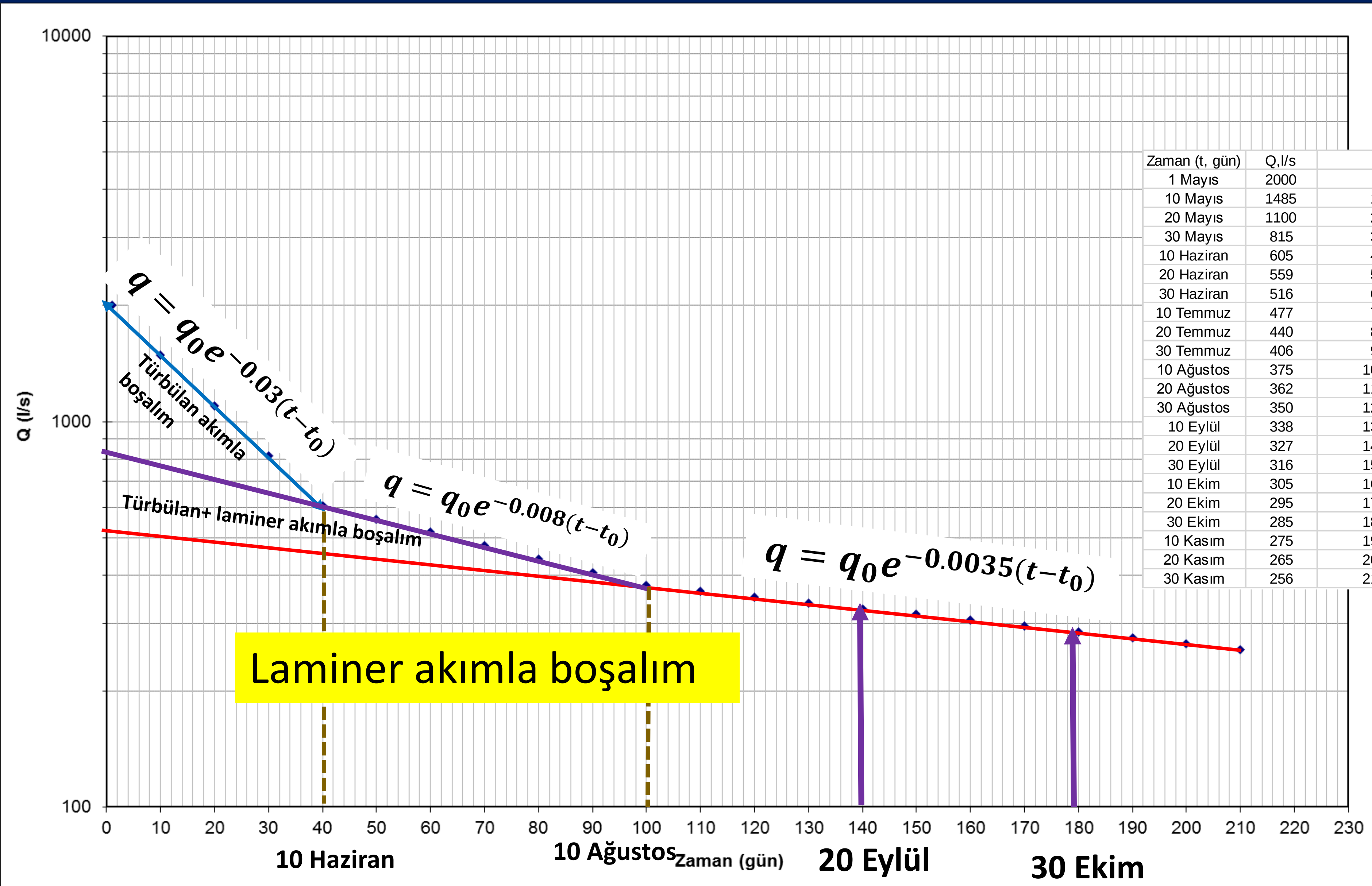
f) Kaynağın boşalım kotu üzerinde boşaltabileceği su miktarını hesaplayınız.

$$V_f = V_1 + V_3 + V_4 = \frac{(2000-605)*86400}{0.03} + \frac{(605-375)*86400}{0.008} + \frac{(375-0)*86400}{0.0035} = \quad \text{lt}$$



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

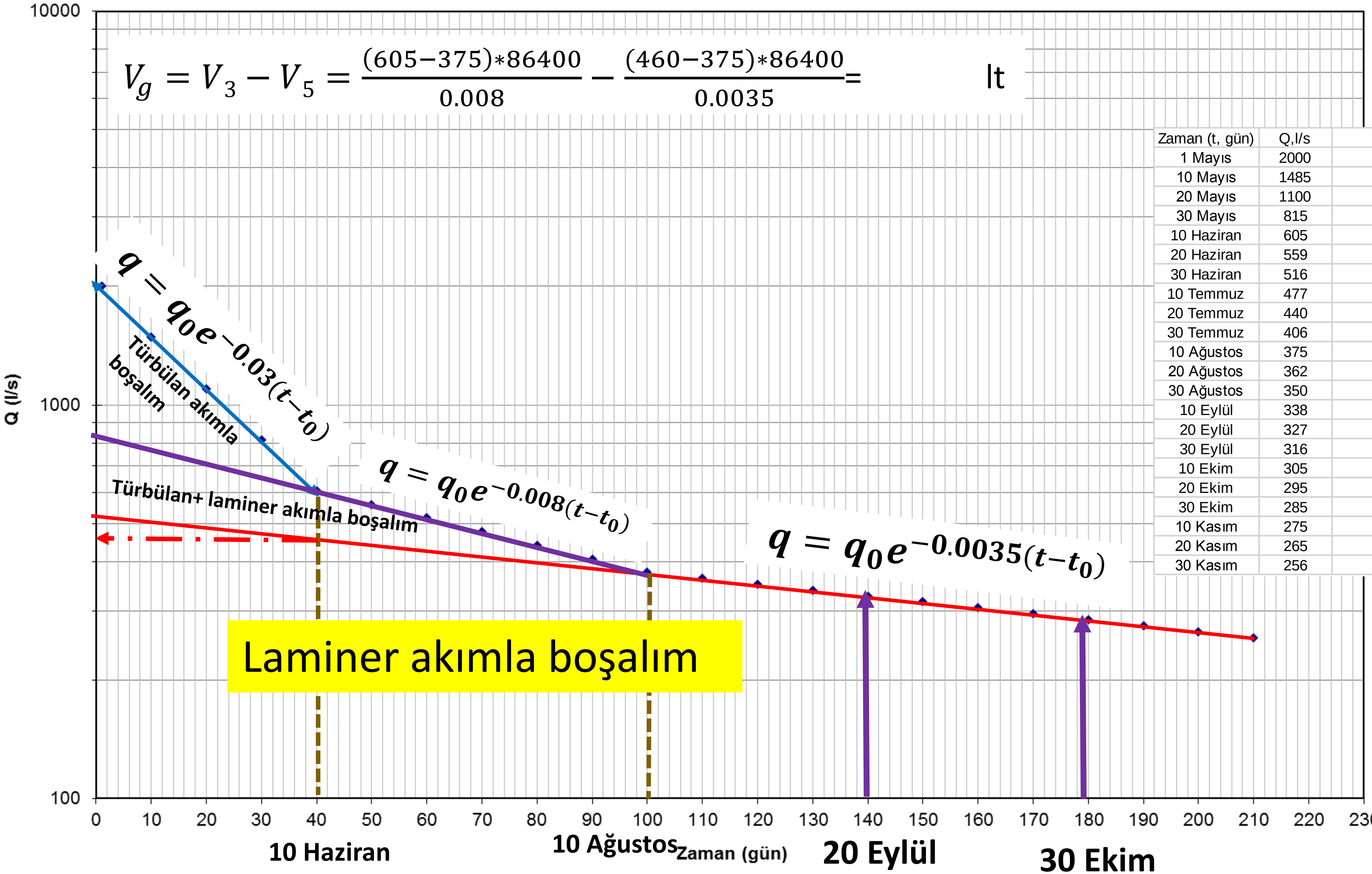
g) Kaynağın türbülan+ laminer akım ile boşaltabileceği su miktarını hesaplayınız.



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

g) Kaynağın türbülan+ laminer akım ile boşaltabileceği su miktarını hesaplayınız.

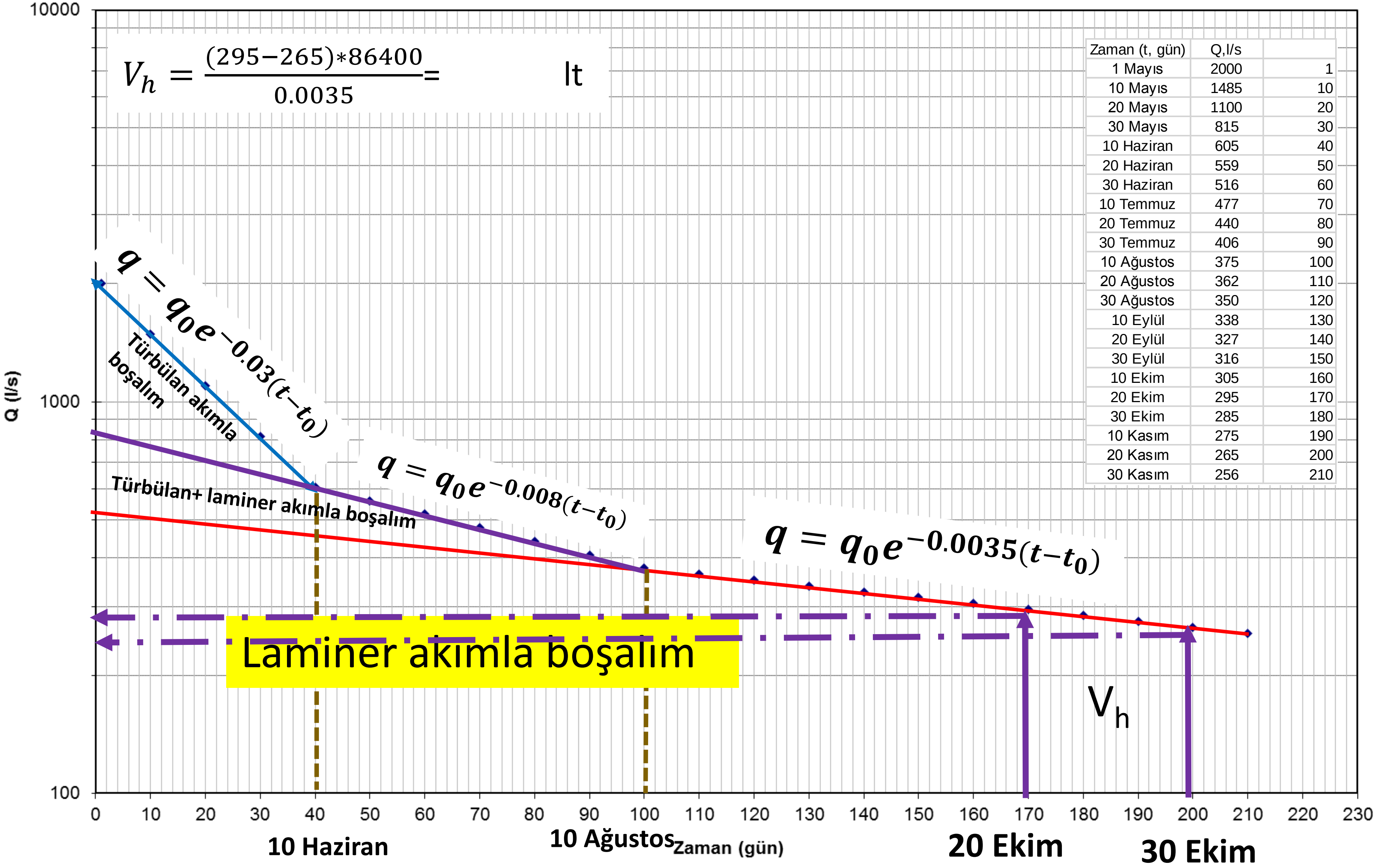
$$V_g = V_3 - V_5 = \frac{(605-375)*86400}{0.008} - \frac{(460-375)*86400}{0.0035} = \quad \text{lt}$$



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

h) Kaynağın 20 Ekim-20 Kasım arasında boşaltacağı su miktarını hesaplayınız.

$$V_h = \frac{(295-265)*86400}{0.0035} = \quad \text{lt}$$



Zaman (t, gün)	Q,l/s	
1 Mayıs	2000	1
10 Mayıs	1485	10
20 Mayıs	1100	20
30 Mayıs	815	30
10 Haziran	605	40
20 Haziran	559	50
30 Haziran	516	60
10 Temmuz	477	70
20 Temmuz	440	80
30 Temmuz	406	90
10 Ağustos	375	100
20 Ağustos	362	110
30 Ağustos	350	120
10 Eylül	338	130
20 Eylül	327	140
30 Eylül	316	150
10 Ekim	305	160
20 Ekim	295	170
30 Ekim	285	180
10 Kasım	275	190
20 Kasım	265	200
30 Kasım	256	210

Soru 1- Toros kuşağında yer alan Mesozoyik yaşlı karstik kireçtaşları ile Eosen yaşlı geçirimsiz fliş birimlerinin dokanağından boşalan bir kaynakta zamana bağlı olarak debiler ölçülmüştür (Çizelge 1). Mesozoyik kireçtaşları Eosen fliş üzerine bindirmeli olarak gelmektedir. Zamana bağlı boşalım değerlerini yarı-logaritmik bir kağıt üzerinde kullanarak Maillet (1905) Yöntemi ile;

a) Kaynağın boşalım katsayılarını bularak, boşalım doğru denklemlerini oluşturunuz.
b) Kaynağın boşalım kotu üzerindeki rezervini veya depolama gücünü hesaplayınız.
c) İnceleme dönemi boyunca kaynağın boşalım kotu üzerinden boşalttığı su miktarını (depolama gücünü) hesaplayınız.
d) 20 Mayıs- 20 Temmuz arası kaynağın laminer akımla boşalttığı su miktarını hesaplayınız.
e) Kaynağın türbülan akımla boşalttığı su miktarını hesaplayınız.
f) Kaynağın 30 Ağustos- 20 Eylül arasında türbülan akımla boşalttığı su miktarını hesaplayınız.
g) Kaynağın 10 Mayıs- 30 Mayıs arasında türbülan akımla boşalttığı su miktarını hesaplayınız.

Çizelge 1. Karstik kaynağın zamana bağlı boşalım (debi) değerleri
Not: Soruları ilgilendiren şekiller çizilecektir..

Zaman (t, gün)	Debi (Q, l/s)	Zaman (t, gün)	Debi (Q, l/s)	Zaman (t, gün)	Debi (Q, l/s)
1 Mayıs	2000	30 Haziran	525	30 Ağustos	380
10 Mayıs	1485	10 Temmuz	500	10 Eylül	367
20 Mayıs	1100	20 Temmuz	475	20 Eylül	350
30 Mayıs	815	30 Temmuz	440	30 Eylül	335
10 Haziran	605	10 Ağustos	412	10 Ekim	330
20 Haziran	560	20 Ağustos	391		