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COLLEGE ON SOIL PHYSICS
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"Heat Transfer and Soil Temperature"

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The knowledge of thermal properties of soils is of importance in engineering works where transport of heat takes place. While designing roads and buildings, laying pipelines and underground cables usefulness of the knowledge of thermal properties is immediately felt. Through the data on heat conduction one recognises why and how the annual temperature variation penetrates into the soil at larger depths than diurnal variation. The subject of soil temperatures tells us what values of maximum and minimum temperature may occur in the dry soil and how those will be modified when the moisture is altered. The germination and the emergence from seed are highly affected by the moisture and temperature of the soil. Besides, insulative nature of dry sand enables one to use it as a solar sand collector and to cut down heat dissipation from alternative heat devices.

1. Soil temperature

The temperature regime of the soil and of the adjoining air layer affects the microclimate of the soil. The temperature of the soil depends on the thermophysical coefficients of the soil and the quantity of heat that enters or leaves the soil. To an open field, the prime source of heat is insolation or solar radiation which on reaching a surface is partly reflected and remaining amount is absorbed in a thin layer. This absorbed amount is used up in heating the soil and the air. In addition some portion is utilised as latent heat for the evaporation

of work. A fraction of the absorbed amount is also emitted by the soil surface as long wavelength radiation.

If we define H_{sh} the heat flux incident to active surface as short wavelength diffuse and direct radiation, r_{sh} be the reflection coefficient, H_{so} the thermal energy transferred into the soil, H_{ai} that into the air, H_{ev} be the heat flux utilised in evaporation and H_{lo} the net out ward flux radiated out by the soil in the form of long wavelength radiation, then the energy balance equation of the active surface is

$$(1-r_{sh})\,H_{sh} = H_{so} + H_{ai} + H_{ev} + H_{lo} \qquad (1)$$

In general, one finds tillage increases the thermal resistance of the upper soil layer in consequence it decreases amount of heat flux into the soil. As such after tillage one would get an increase in the temperature of the surface. Tillage affects H_{ai} and H_{lo} too. The temperature of soil can be calculated from the theory of heat conduction as a function of depth and time, once H_{so} and coefficients of heat transfer of the soil are known.

A bare soil surface is open to direct and diffuse sun rays, that is why it becomes quite warm during the hottest part of the day and in cold seasons bare surface loses its heat to the environment much faster. A vegitative cover protects a large amount of insolation thereby it prevents the underlying soil from becoming as warm as a bare surface. In cold seasons, a vegetative cover behaves as an insulated blanket there by preventing heat loss from the soil. The protected soil is cooler in summer and warmer in winter than the bare one. Fig. 1 gives soil temperature in bare and grass covered soils.

Using some published data (ref. 9) one can obtain useful information about seasonal heat flux H_{so} into and out of a soil. The soils gain between 7 to 10 KJ/cm^2 in the six months from March to August. This amounts to about 4 per cent of the incident value. For the hot months like May and June the net heat flux into the soil is 70 to 90 J/cm^2. On a hot sunny day one might find downward flux of the order of 200 J/cm^2 and upward flux of 80 J/cm^2. Thus, the net flux gain on a sunny day is 10 per cent of incoming radiation. Once again, it should be borne in mind that the values of inflow and out flow of thermal energy are basically governed by local conditions.

The temperature of soil also depends upon its properties. Major factors which differentiate one from the other are:
(i) colour, which controls the amount of solar absorption,
(ii) constituents, compactness and moisture content, which affects specific heat and thermal conductivity. The daily variation in temperature has been found greater with dark soils, and the loss of heat from them during the night is also faster. However, with increasing depth the temperature differences between dark and light coloured soils are smaller. It should be noted that colour influences warmth of soil, but this might not be so important factor as compared to other properties of soil. The dark complexion of soil is usually due to organic material.

The organic materials possess larger specific heat value and low thermal conductivity. The less porous sand warms up quickly because of large conductivity value, peat soils are slow in warming because their pores are filled with non-conductive air. The presence of water causes the large temperature rise because water is much better conductor of heat than air. The frost penetration has been found fastest in sand than in clay and slowest in peat.

2. Heat transport.

The transmission of thermal energy in soil takes place by the process of conduction through the solid grains and the fluid filling pores, by convection and radiation between the walls of a pore. To develop mathematics it is often assumed that the thermal properties of the soil in an aggregate do not depend on space coordinates. All types of heat transfer mechanisms may be reduced to conduction by introducing the equivalent coefficient of thermal conductivity λ of the soil. The real flux density H (which is the quantity of heat flowing per unit time across a unit surface area) is proportional to temperature gradient $\frac{(\theta_2-\theta_1)}{(x_2-x_1)}$ across the given width (x_2-x_1) of the soil sample

$$H = \lambda \frac{(\theta_2-\theta_1)}{(x_2-x_1)} = -\lambda \frac{\partial\theta}{\partial x} \qquad (2)$$

λ, termed as thermal conductivity of the soil is in fact a factor of proportionality in eq. (2) whose value is different for different soils, unit of λ is W $m^{-1}K^{-1}$ or cal/cm/s/°C. The equivalent coefficient of thermal conductivity λ_{eq} is

$$\lambda_{eq} = \lambda + \lambda_{conv} + \lambda_{rad}.$$

with pore diameter less than 5 mm and $\Delta\,\theta = 10^{0}C$ the contribution due to convective transfer is negligibly small. If the pore diameter is less than 0.05 mm then contribution due to radiative heat transfer can also be neglected. One arrives at a conclusion that for capillary radius less than 0.001 cm and with small temperature differences the transport of heat in pores by way of convection and radiation is very very small and the chief mechanism responsible in transporting heat in soils is conduction (ref. 5).

Let $-\partial H/\partial x$ is the rate of change of heat flux density with the depth (x) then in a volume dV(=1x1xdx), heat stored in unit time will be $-(\partial H/\partial x)\,dx$. Also, if one defines c as the volumetric heat capacity i.e. the amount of heat needed to raise the temperature of a unit volume of soil by one degree (cal cm^{-3} $^{o}C^{-1}$) then $C(\partial\theta/\partial t)\,dx$ is the amount of heat stored in the soil. Equating the heat storage found from (a) the difference in heat flux density and (b) the temperature change, one gets

$$-\partial H/\partial x = C\ \partial\theta/\partial t \qquad (3)$$

substituting H from (2) in above one gets the differentiate *l* equation of heat conduction in one dimension as

$$\frac{\partial\theta}{\partial t} = \alpha\,\frac{\partial^2\theta}{\partial x^2} \qquad (4)$$

where, $\alpha = \lambda/C = \lambda/\rho s$ is called the coefficient of thermal diffusivity ($m^2 s^{-1}$) of the soil. Here s is the specific heat (cal g^{-1} $^{o}C^{-1}$) and ρ (g cm^{-3}) is the density of the sample.

Physically, high value of thermal diffusivity of a soil indicates its capability for rapid and considerable changes in temperature. When one considers the heat flux into the soil over a certain period the knowledge of one more coefficient 'b' which is $\sqrt{\lambda C} = \lambda/\sqrt{\alpha}$ is found useful. b is called as thermal effusivity (W m^{-2} $s^{+1/2}$ $^{o}K^{-1}$). This coefficient characterises the soil from view point of its storage qualities.

3. Temperature variations in soil

To obtain an introductory discussion of temperature variations in soil we here take a simple treatment (ref. 2). Let us assume that at all depths the temperature varies as pure harmonic function of time around an average value of temperature.

The surface temperature (i.e. at x = 0) can be written as

$$\theta(o,t) = \theta_{av} + A_{\theta_o}\sin\omega t \qquad (5)$$

Here, θ_{av} is the average temperature in the soil during a period, which is assumed to be the same at all depths. A_{θ_o} is the amplitude of periodic wave at the surface and ω is angular frequency. For diurnal variation $\omega = 2\pi/86400\ s^{-1}$. The temperature at any arbitrary depth x would also be a sine function of time

$$\theta(x,t) = \theta_{av} + A_{\theta_x}\sin\left[\omega t + \phi(x)\right] \qquad (6)$$

The amplitude and phase terms can be determined with the help of (6) and (4) as such temperature at a depth x is

$$\theta(x,t) = \theta_{av} + A_{\theta_o}\exp(-x/D)\sin(\omega t - x/D) \qquad (7)$$

It may be noted on comparing (7) with (5) that at the depth x the amplitude is lessened by a factor exp $(-x/D)$ and there occurs a shift in phase by $(-x/D)$. The constant D is termed as damping depth. This is connected to thermal characteristics of the soil and frequency of temperature variation and is given by

$$D = (2m/\omega)^{1/2} \tag{8}$$

The damping depth in a soil depends on the period of the temperature variation. For annual variation it is $\sqrt{365}$ 19 times larger than diurnal variation. The reader may find an exhaustive description on the subject in refs. 2 and 4. Here we describe only some results of interest.

Using eq. (2) and a similar expression like eq. (6) the heat flux density H (cal $cm^{-2}s^{-1}$) for a sinusoidal variation of temperature is

$$H(x,t) = -\lambda \frac{\partial \theta_{av}}{\partial x} + A_{\theta_0} (\lambda C \omega)^{1/2} \exp(-x/D) \sin(\omega t + \phi - x/D + \pi/4)$$

what we note is flux is also a harmonic function of time with $\pi/4$ phase advanced as compared with temperature variation at a given depth. This phase shift of $\pi/4$ between H and θ corresponds to a time shift of 3 hours for diurnal variation and 1.5 months for annual variation.

4.1. Thermal properties of soils

We have seen that two independent thermal properties of soils enter into the expressions when one deals with quantitative description of heat transfer. These are: thermal conductivity λ

and the heat capacity per unit volume C or specific heat s. Quartz has the lowest value of specific heat and humus the largest, excepting water. As such it is obvious that humus and water will affect heat capacity of the soil greatly. Wet soils have much higher heat capacity than dried. This explains the cold nature of wet soils. However, it may be noted in the range of temperatures occuring in a field the specific heat values vary slightly. In fig. 2 variation of specific heat of dry sand with temperature is shown.

The heat capacity per unit volume of a soil could be calculated by adding the heat capacities of different constituents of soil in a cubic centimeter. If ϕ_s, ϕ_ω and ϕ_a denote the volume functions of solid material, water and air, respectively, then

$$C = \phi_s C_s + \phi_\omega C_\omega + \phi_a C_a \tag{9}$$

For most soil minerals, using values from ref. 2, at 10^oC an average value of $C_s = 0.46$ cal $g^{-1}\,{}^oC^{-1}$ and density 2.7 g cm^{-3} holds. For organic minerals $C_s = 0.6$ and $\rho = 1.3$ are accepted average values. The air contribution in (9) is negligibly small. Thus, if ϕ_{sm} denotes volume fraction of soil minerals, ϕ_o that of organic material and since $C_\omega = 1$ cal $cm^{-3}\,{}^oC^{-1}$ one may calculate heat capacity of the soil through

$$C = 0.46\,\phi_m + 0.6\,\phi_o + \phi_\omega \tag{10}$$

4.2. Thermal conductivity of soils

In the preceeding section we have known that the heat capacity of a soil is weighted arithmatic mean of constituent heat capacities. The thermal conductivity of soil can not be estimated so simply. Thermal conductivity, of the soil is governed by factors like constituent conductivities, volume fractions of constituents, packing, temperature, pressure etc.

Thermal conductivity of soil solids λ_s can be estimated using Johansen's suggestion through

$$\lambda_s = \lambda_q^{\phi_q} \cdot \lambda_o^{(1-\phi_q)} \qquad (11)$$

$\lambda_q = 7.7 \ \ m^{-1} K^{-1}$ is the value of thermal conductivity of quartz and ϕ_q is volume fraction of quartz present in the soil, $\lambda_o = 2.0$ is that of soil minerals other than quartz having volume fraction $(1-\phi_q)$. The thermal conductivity of water $\lambda \ = 0.6$ and thermal conductivity of air $\lambda_a = 0.026$ at 20^oC and 1 atm.

To obtain workable expressions for the estimation of λ it is usual to consider that soil is made of two components. Continuous fluid (f) medium of conductivity λ_f with volume fraction ϕ and dispersed solids (s) of conductivity λ_s with volume fraction $(1-\phi)$. This description holds to dry or fully saturated soil. When one assumes that the flux lines are not markedly altered in presence of solid grains one arrives at semi-empirical expressions. Considering the soil as a composite system made of successive layers of solid and fluid two plans, as shown in fig. 3, are obtained (refs. 6 and 7). Mathematically, the parallel distribution is weighted arithmatic mean of constituent conductivities,

$$\lambda_{\parallel} = \phi\, \lambda_f + (1-\phi)\, \lambda_s \qquad (12)$$

The series distribution is harmonic mean

$$\lambda_{\perp} = \left[\frac{\phi}{\lambda_f} + \frac{(1-\phi)}{\lambda_s} \right]^{-1} \qquad (13)$$

We then have a geometric mean as

$$\lambda = \lambda_f^{\phi} \cdot \lambda_s^{(1-\phi)} \qquad (14)$$

A plot for extremum eq. (12) and eq. (13) with intermediate eq. (14) is fig. 4. Cross-hatched area indicates realisable region on soils. An expression similar to Eq. (14) given as below falls in the realisable region (ref. 7).

$$\lambda = \lambda_{\parallel}^{n}\, \lambda_{\perp}^{(1-n)} \qquad (15)$$

where,

$$n = \frac{0.5\ (1-\log\phi)}{\log\left[\left(\frac{\lambda_s}{\lambda_f}\right) \phi\ (1-\phi) \right]}$$

Expressions (12) and (13) are called as extremum bounds on thermal conductivities. On the basis of variational principles Hashin and Shtrikman (ref. 1) obtained bounds which are nearer to the realisable region. The geometric mean (14) may be applied to frozen soils for the estimation of λ,

$$\lambda = \lambda_s^{(1-\phi)}\ \lambda_w^{\theta}\ \lambda_i^{(\phi-\theta)} \qquad (16)$$

where, θ is the volume fraction of the unfrozen water and λ_i the thermal conductivity of ice. Besides statistical approaches λ may be calculated through rigorous mathematics where one considers idealised geometries and these are called as physical models. Originally developed by Maxwell the following equation has been modified by de Vries (ref. 2) that can be applied on soils

$$\lambda = \frac{\phi\lambda_f + F(1-\phi)\,\lambda_s}{[\phi + F\,(1-\phi)]} \qquad (17)$$

where $F = \frac{1}{3} \sum_{a,b,c} \left[1 + (\frac{\lambda_s}{\lambda_f} - 1)\; g \right]^{-1}$

in which $g_a + g_b + g_c = 1$. Assuming $g_a = g_b = 0.125$ one gets good agreement for saturated soils. ut to obtain an agreement for dry soils one should increase the calculated values by 25 per cent. In the literature one finds a large number of expressions for the calculation of λ. An interested reader may find some of those in references given here.

Thermal conductivity of sand shows a large variation on change of interstitial air pressure. Fig. 5 gives variation of thermal conductivity of dune sand with interstitial air pressures. One may notice that the behaviour of thermal conductivity of dry sand with interstitial air pressure suggests its use as alternative thermal insulators.

4.3. Thermal diffusivity and thermal effusivity

The knowledge of thermal diffusivity becomes essential when one deals with problems when regular change of temperature occurs. Estimation of thermal diffusivity of a soil can be made if one has data of thermal conductivity and heat capacity.

For the sand whose $\lambda = 6.11 \times 10^{-4}$ cal cm^{-1}s^{-1} $^oC^{-1}$ and $C = 0.304$ cal cm^{-3} $^oC^{-1}$, $\alpha = \lambda/c = 2.0 \times 10^{-3}$ cm^{2}s^{-1}.

When one considers regular dispersion of spheres (solid phase) in continuous medium (fluid) and accommodates in the physical model third order interaction an expression for thermal diffusivity follows as (ref. 8)

$$\alpha = \alpha'_f \left| 1+5.6(1-\phi)^{2/3} \frac{\alpha'_s-\alpha'_f}{\alpha'_s+2\alpha'_f} + 6.298\;(1-\phi)\;\frac{\alpha'_s-\alpha'_f}{\alpha'_s+2\alpha'_f} \right| \qquad (18)$$

where $\alpha'_f = \frac{\lambda_f}{c}$ and $\alpha'_s = \frac{\lambda_s}{c}$

Similarly, thermal effusivity or coefficient of heat storage which gives the capability of the soil in storing heat may be evaluated by

$$\sqrt{\lambda C} = \frac{\lambda}{\sqrt{\alpha}} = b$$

For the stated values of sand it is 0.014 cal cm^{-2}s$^{-1/2}$ $^oC^{-1}$

5. Measurement of thermal properties of soil

The guarded hot plate method is a standard method to determine thermal conductivity of soils but it is a steady state method which not only consumes much larger time in finding thermal conductivity but to moist samples this method is not applicable. Here, we describe dynamical methods for the measurement of nductivity and diffusivity and a steady state method to determine specific heat of dry soils.

5.1. Needle method for λ

The needle method also called as thermal probe is a compact instrument to determine the thermal conductivity of insulative materials. In a steel needle an electric heater wire is spread throughout the length. A thermocouple is placed in the middle of the heater. The thermocouple and heater leads are brought out through a plastic handle. Through a reference device thermocouple ends are brought to a potentiometer. Heater wire circuitory consists of a power supply (storage battery) and power measuring unit. The thermal probe is placed in the soil and observations are taken after thermal equilibrium is established. The experiment is conducted in 10 minutes. The probe and its electrical connections are shown in fig. 6a (refs. 6 and 7). Thermal conductivity of the soil is calculated using

$$\lambda = \text{constant} \quad I^2 \frac{\ln\left(\frac{t_2}{t_1}\right)}{\Delta V}$$

where I the current supplied to the heater, ΔV is the rise in emf in duration t_1 to t_2. The constant in expression contains resistance of heater wire, length of the probe and temperature emf conversion factor for the thermocouple. A plot for emf generated on probe is given in fig. 6(b).

5.2. Hot wire method for a

The hot wire method for the determination of thermal diffusivity of the soil is based on similar theory as that of the probe (Refs. 10 and 8). An experimental arrangement for the hot wire is shown in fig. 7a. As in the probe we have a heater encased in a sheath and at a certain distance r (note not adjacent to the heater as in case of the probe) from the heater

lies a thermocouple. This assembly is then inserted in soil sample under investigation. Heating of a sample is illustrated in fig. 7b. From the plot $\partial\theta/\partial t$ vs t the value of t_{max} is noted. Then the thermal diffusivity of the sample may be evaluated through $a = r^2/4t_{max}$.

5.3. Continuous fall electric method for s

In this method for the determination of specific heat, the dry soil is allowed to fall through a double walled glass tube. A steady electric current flows through centrally stretched heater wire. A thermocouple is employed to measure the difference in temperatures of incoming and outgoing sand. To vary rate of fall different diametered nozzles employed. Specific heat s (cal/g/°C) of the sample can be calculated from

$$s = (i_1^2 - i_2^2)\ XR \times 10^7/J\ (m_1 \Delta\theta_1 - m_2 \Delta\theta_2)$$

Here i_1 and i_2 are the values of currents when rates of soil fall are m_1 and m_2 (gs^{-1}) and corresponding steady state temperature differences at the ends of glass tube are $\Delta\theta_1$ and $\Delta\theta_2$. Fig. 8 illustrates schematic diagram of the method.

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Figure Caption

Fig. 1a. Soil temperature in bare and grass covered soils (Wollny).

Fig. 1b. Seasonal variations in soil temperature at various depths (Ref. 3).

Fig. 2. Specific heat of dry Ottawa sand vs temperature (Ref. 1).

Fig. 3. Parallel and series distribution of soil constituents.

Fig. 4. A plot for Eq. (12), Eq. (13) and Eq. (14). Cross-hatched area is realisable region on soils.

Fig. 5. Thermal conductivity of dry dune sand with interstitial air pressure (Ref. 8).

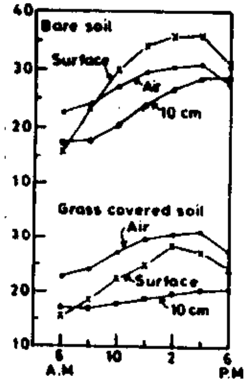
Fig. 6a. Thermal probe for measuring thermal conductivity of soils.

Fig. 6b. Thermo emf generated on a probe.

Fig. 7a. A hot wire arrangement for measuring thermal diffusivity of soil.

Fig. 7b. Heating curve of a sand sample in hot wire experiment.

Fig. 8. Schematic diagram of continuous fall method for the determination of specific heat of dry soil.



g. 1a. Soil temperature in bare and grass covered soils [Wollny]. (ref.3)

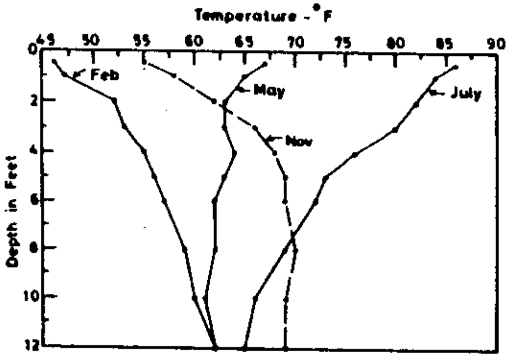
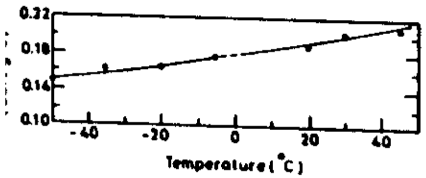


Fig. 1b. Seasonal variations in soil temperature at various depths. (ref.3)



). 2. Specific heat of dry Ottawa sand vs. temperature. (ref.1)

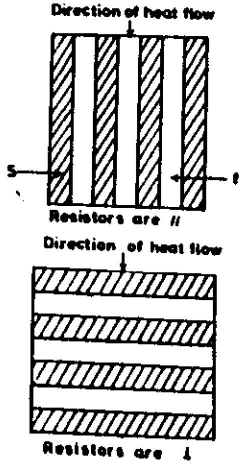


Fig. 3. Parallel and series distribution of soil constituents.

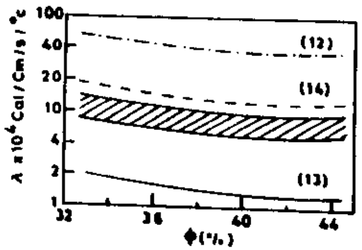


Fig. 4. A plot for Eq (12), Eq (13) and Eq (14). Cross hatched area is realisable region on soils. (ref.8)

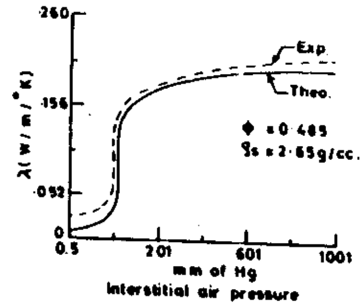


Fig. 5. Thermal conductivity of dry dune sand with interstitial air pressure (ref.8)

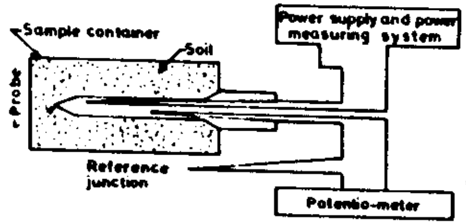


Fig. 6a. Thermal probe for measuring thermal conductivity of soil. (ref.8)

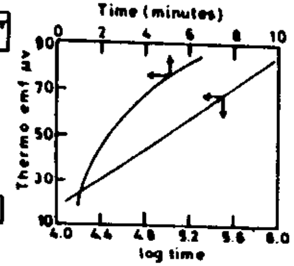


Fig. 6b. Thermo emf generated on a probe. (ref.8)

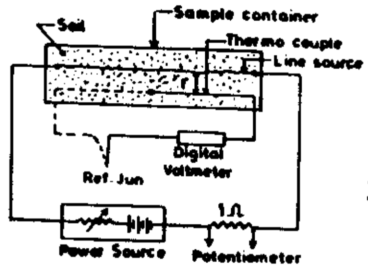


Fig. 7a. A hot wire arrangement for measuring thermal diffusivity of soil. (ref.8)

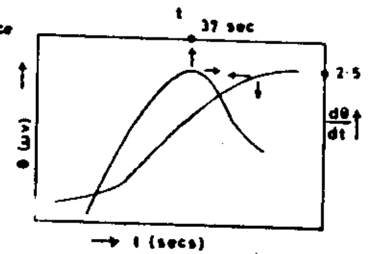


Fig. 7b. Heating curve of a sand sample in hot wire experiment. (ref.8)

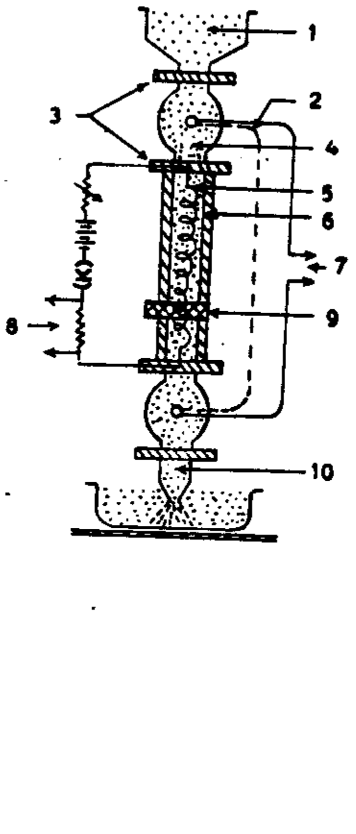


Fig.8. Schematic diagram of continuous fall method for the determination of specific heat of dry soil - (1) Reservoir (2) Thermocouple (3) cork (4) sample (5) heater (6) evacuated and silvered double walled glass tube (7) to potentiometer for the measurement of i (9) glass tube coupled to a vibrator and (10) nozzle. (ref.8)