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Units and their appropriate use in soil-plant-atmosphere processes

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ABSTRACT

The study of processes in the soil-plant-atmosphere continuum either through theoretical considerations or field measurements, involves use of units which are to be expressed in several forms or converted from one system to the other. Though SI (Systeme Internationale) units are the internationally recommended units for use in the modern scientific literature, we still come across in our day to day practice the old systems of units which continue to be popular both in theory and practice. Appropriate use and conversion to the standard system of units thus assumes significance. The units of parameters commonly encountered in the field studies of crops with reference to soil, plant and atmosphere are discussed.

Key words : SPAC, Units, Expressions, Energy transfer

Radiant energy

Stefan-Boltzmann equation

The Stefan-Boltzmann law states that the rate of emission of energy from a blackbody is proportional to the fourth power of the absolute temperature of the blackbody.

$$E = \sigma (T + 273)^4$$

where σ is a proportionality constant known as the Stefan-Boltzmann constant, T is the surface temperature in degrees Centigrade, and E is the radiant emittance of the source in watts per square meter. The value of σ is 5.673×10^{-8} W m^{-2} K^{-4}.

Most objects are not perfect blackbody radiators but have a radiant emittance that is a fraction of blackbody radiant emittance. All such objects are known as gray bodies, and their radiant emittance is given by

$$E = \epsilon \sigma (T + 273)^4$$

where ϵ is the emissivity of the object surface and has a value between 0 and 1.0. Emissivity may be a function of both view angle and wavelength. Emissivity values quoted in the literature are often integrals of the spectral emissivity over some wavelength, typically the 3-5 μm or 8-14 μm range (Table 1). The wavelength intervals are not always clearly identified. Emissivities of vegetative materials are generally greater than 0.95, except when dehydration occurs, in which case the value falls to 0.90. Generally the emissivity values are less in the 3-5.5 μm region but all are greater than 0.85. The emissivity values of soils and rocks are relatively smaller than those for vegetative materials.

Table 1: Integrated emissivities of some representative vegetation materials and soils and rocks in two wavelength regions.

Feature	Wavelength integration over 3.0 - 5.5 µm	8.0 - 14.0 µm
Vegetation		
Green mountain laurel	0.90	0.92
Young willow leaf (dry, top)	0.94	0.96
Holly leaf (dry, top)	0.90	0.90
Holly leaf (dry, bottom)	0.86	0.94
Pressed dormant maple leaf (dry, top)	0.87	0.92
Green leaf winter colour-oak leaf(dry, top)	0.90	0.92
Green coniferous twigs (jack pine)	0.96	0.97
Grass - meadow fescue (dry)	0.82	0.88
Bark - northern red oak	0.90	0.96
Bark - northern American jack pine	0.88	0.97
Bark - Colorado spruce	0.87	0.94
Corn		0.94
Indian fis-cactus		0.94
Prickly pear cactus		0.96
Cotton (upland)		0.96
Tobacco		0.97
Blind-pear cactus		0.98
Fermont cotton wood		0.98
Philodendron		0.99
Sugarcane		0.99
Soils and Rocks		
Granite, rough		0.89
Dunite, rough		0.89
Basalt, rough		0.94
Sand, large grains		0.91
Sand, large grains, wetted		0.93
Sand (Monterey), small grains		0.92
Hainamanu silt loam - Hawaii	0.84	0.94
Barnes fine silt loam - South dakota	0.78	0.93
Gooah fine silt loam - Oregon	0.80	0.98
Vereiniging - Africa	0.82	0.94
Maury silt loam - Tennessee	0.74	0.95
Dublin clay loam - California	0.88	0.97
Pullman loam - New Mexico	0.78	0.93

(Table Contd...)

Grady silt loam - Georgia	0.85	0.94
Colt's neck loam - New Jersey	0.90	0.94
Mesita negra - lower test site	0.75	0.92

Polished materials

	Temperature integration			
	253 K	273 K	293 K	313 K
Quartz (agate)	0.694	0.682	0.672	0.664
Granite	0.787	0.783	0.780	0.777
Feldspar	0.826	0.822	0.819	0.811
Lavrotite	0.813	0.812	0.812	0.811
Obsidian	0.832	0.830	0.828	0.826
Basalt	0.904	0.905	0.906	0.907
Dunite	0.851	0.857	0.861	0.865
Marble	0.941	0.942	0.942	0.943

Solar constant

The amount of sunlight incident upon the earth's surface at any point in space and time is a function of several variables. Assuming that the Sun is a blackbody, the Sun emits a total power of about 3.84×10^{26} W. The radiant emittance of the solar surface is $\sigma\,(T + 273)^4$, or 6.244×10^7 W m^{-2}. The total flux of radiation passing out into space through the surface of the sun is $4\pi R^2\sigma\,(T+273)^4$, where R is the radius of the sun (6.965×10^8 m). The sun radiates an amount of energy equivalent to that which a blackbody at a temperature of 5760 K would radiate. The earth is in the orbit around the sun at a mean distance d of 1.497×10^{11} m. Assuming that earth goes round the Sun in a circular orbit, with the Sun at the centre of the orbit, the total radiation emitted by the Sun would be incident on the inner surface of a sphere, described by earth's orbit, of radius equal to the mean distance of the earth from the Sun. If S_o is the amount of incident energy received per unit surface of earth per unit time, then the total amount of incident energy received is $4\pi d^2 S_o$. Then,

$$4\pi d^2 S_o = 4\pi R^2 \sigma\,(T+273)^4$$

$$S_o = \frac{4\pi R^2 \sigma\,(T+273)^4}{4\pi d^2} = \frac{R^2 \sigma\,(T+273)^4}{d^2}$$

By substituting the values of R, d and σ,

$$S_o = 1360 \text{ W m}^{-2} = 1360 \text{ Jm}^{-2}\text{s}^{-1}$$

(Since 1 watt = 1 joule second^{-1}).

By converting 1 J = 0.239 calorie, and converting second into minute and meter into centimeter, we get $S_o = 1.95$ cal cm^{-2} min^{-1} ≈ 2 langleys/min (1 langley = 1 cal cm^{-2}). The value obtained is based on the assumption that the earth's orbit is circular, but as the orbit, in fact, is elliptical, the value of solar constant (S_o) varies with time of the year. In addition, latitude and longitude, as well as time of day, determine the amount of sunlight received at any time of the year. Turbidity or cloudiness of the atmosphere determines the amount of solar radiation that penetrates to the earth's surface. Table 2 shows the difference in the amount of radiation received at the sea level and above the atmosphere.

Table 2 : Spectral irradiance ($Wm^{-2}\mu^{-1}$) of direct sunlight normal incidence [from Moon 1940, J. Franklin Inst., 230:583 (1940)]

Wavelength, λ (nm)	Above atmosphere (Solar constant = 1322 Wm^{-2})	At sea level (Solar constant = 739.8 Wm^{-2})
300	450	0.081
305	540	1.91
310	616	11
315	676	30
320	726	54
325	762	75
330	796	101
335	826	130
340	856	151
345	886	170
350	916	188
360	976	233
370	1046	279
380	1121	336
390	1202	397
400	1304	470
410	1728	672
420	1766	733
430	1788	787
440	1939	911
450	2036	1006
460	2096	1080
470	2119	1138
480	2127	1183
490	2103	1210
500	2061	1215
510	2000	1206
520	1954	1199
530	1912	1188
540	1894	1198
550	1878	1190
560	1861	1182
570	1841	1178

(Table Contd...)

580	1819	1168
590	1795	1161
600	1762	1167
610	1727	1168
620	1690	1165
630	1653	1176
640	1616	1175
650	1579	1173
660	1543	1166
670	1508	1160
680	1473	1149
690	1439	978
700	1405	1108
710	1371	1070
720	1337	832
730	1304	965
740	1270	1041
750	1236	867
760	1205	566
770	1175	968

(*At the sea level, air mass is taken as 2 which is the ratio of length of path of rays within the atmosphere to the observation point on earth, to the corresponding length of path for a hypothetical position of sun at zenith.)

Wien's Displacement Law

The law states that the product of the wavelength (λ) for the maximum intensity multiplied by the absolute temperature is a constant and is given by

$$\lambda_{max} T = 289.7 \times 10^{-5} \text{ mK.}$$

From this, it is clear that if T = 289.7 K, $\lambda_{max} = 10^{-5}$ m (10μm), if T = 579.4 K, $\lambda_{max} = 0.55 \times 10^{-5}$ m (5μm), and if T = 5794K, which is approximately the apparent black-body temperature of the sun, $\lambda_{max} = 0.05 \times 10^{-5}$ m (0.5μm). One micrometer (μm) is equal to 10^{-4} m. If the temperature of a body is known, it is possible to calculate the wavelength of maximum intensity radiation that the body is likely to emit. All bodies emit radiation over a wide range of wavelengths but the Wien's displacement law gives the wavelength of that radiation whose intensity is maximum.

Photosynthetically Active Radiation (PAR)

Photosynthetically active radiation (PAR) is defined as radiation in the 400 to 700 nm (nano meter) waveband. To measure light intensity in these wavelengths, a unit known as Einstein has been defined in the past in plant science. However, now the use of the mole is recommended since it is a SI unit. With either of the units, the quantity of photons in a mole is equal to the quantity of photons in an Einstein.

1 mole = 1 Einstein

= 6.023×10^{23} photons /quanta.

Table 3 : Energies contained in an Einstein/mole for different wavelengths of radiation

Wavelength λ (nm)	Spectral range	Energy per mole (MJ)
100	UV (ultraviolet)	1.195
200	UV	0.598
300	UV	0.398
400	Visible - violet	0.300
500	Visible - blue/green	0.273
590	Visible - yellow	0.208
650	Visible - Orange	0.184
750	Visible - red	0.159
800	IR (Infrared)	0.149
900	IR	0.133
1000	IR	0.120

According to Planck's law, a quantum of radiation contains energy E, proportional to the frequency, ν,

$E \alpha \nu$ or $E = h\nu = hc/\lambda$

where h is Planck's constant = 6.62×10^{-34} Js (Joule sec), c is the velocity of light = 3×10^{8} ms^{-1} and λ is the wavelength. Substituting these values, we get

$E = 6.62 \times 10^{-34} \times 3 \times 10^{8}\ Jm/\lambda$
$= 19.86 \times 10^{-26}\ Jm/\lambda$.

As per definition, one Einstein or one mole is equal to the energy contained in 6.023 x 10^{23} quanta. Therefore,

1 mole = 1 Einstein
$= 6.023 \times 10^{23} \times 19.86 \times 10^{-26}\ Jm/\lambda$.

This relationship tells us that the energy contained in one mole varies with the wavelength. It is more at shorter wavelengths and less in longer wavelengths (Table 3). One single value cannot be used for all wavelengths.

In other words, for a given amount of energy, less number of photons/quanta carry this energy in shorter wavelengths than in longer wavelengths. Thus the total number of quanta in a given wavelength range depend upon the spectral distribution of the source of radiation. For example, the spectral energy distribution of a fluorescent lamp may have more number of quanta corresponding to wavelengths of cool day light. For an infrared lamp, more number of photons contain energy corresponding to longer wavelengths. For a gas disharge lamp, it will be a characteristic of the gas, that is, for a sodium vapour lamp, maximum number of photons contain energy corresponding to a wavelength around 550 - 600 nm.

Therefore, to express the radiation received from different sources in terms of Einstein or mole, it is necessary to know the number of quanta that carry energies at different wavelengths and then sum up the total quanta, because one mole or Einstein is defined only in terms of number of quanta as

6.023 x 10^{23} quanta. The number of ~~quanta~~ is measured with the help of a quantum sensor which gives readings in terms of micromole per sec per sqare meter (µmole s^{-1} m^{-2} = 6.023 x 10^{17} photons s^{-1} m^{-2}).

On the other hand, radiometric system of units are simple and more acceptable as they are expressed in basic SI units. The radiation received between 400 - 700 nm wavelength range can be expressed in radiometric units as Joules per sec ~~per square~~ meter (J s^{-1} m^{-2}) or watts per square meter (Wm^{-2}).

Conversion of quantum sensor output in µmole s^{-1} m^{-2} (400 -700 nm) to radiometric units in Wm^{-2} (400 -700 nm) is complicated and the conversion factor will be different for each light source as the spectral distribution of each source is different.

The radiometric quantity (W_T) desired is simply obtained from the spectral energy distribution curve of the light source as

$$W_T = \int_{400}^{700} W_\lambda \, d_\lambda \quad ....(1)$$

For obtaining the PAR quantity, the number of quanta received at different wavelengths should be known. At a given wavelength λ, the number of photons per sec is given by

$$\text{Photons } s^{-1} = \frac{W_\lambda}{hc/\lambda}$$

$$= \frac{\text{energy received per sec at wavelength } \lambda}{\text{energy of one photon at wavelength } \lambda}$$

where,

h = 6.62 x 10^{-34} Js (Planck's constant), c = 3 x 10^{8} m s^{-1} (velocity of light) and λ is in nm. The total number of photons in the 400-700 nm range is then obtained by

$$\text{Total no. of photons} = \int_{400}^{700} (W_\lambda \, d_\lambda)/(hc/\lambda)$$

If R is the reading of quantum sensor in µmole s^{-1} m^{-2}, then

$$R\ (\mu\text{mole } s^{-1}\ m^{-2}) = \int_{400}^{700} (W_\lambda \, d_\lambda)/(hc/\lambda)$$

or, because 1 µmole s^{-1} m^{-2} = 6.023 x 10^{17} photons s^{-1} m^{-2}, then

$$6.023 \times 10^{17} R = \int_{400}^{700} (W_\lambda \, d_\lambda)/(hc/\lambda) \quad .....(2)$$

Dividing the radiometric quantity (1) by PAR quantity (2),

$$\frac{W_T}{6.023 \times 10^{17} R} = \frac{\int_{400}^{700} (W_\lambda \, d_\lambda)}{\int_{400}^{700} (W_\lambda \, d_\lambda)/(hc/\lambda)}$$

$$\text{or, } W_T = 6.023 \times 10^{17} R\, hc \; \frac{\int_{400}^{700} (W_\lambda \, d_\lambda)}{\int_{400}^{700} \lambda\, W_\lambda \, d_\lambda}$$

Substituting the values of h and c, and the summation done over a specific interval of Δλ,

$$W_T = 119.8\ R \ \frac{\sum_{400}^{700} W_\lambda \Delta\lambda}{\sum_{400}^{700} \lambda W_\lambda \Delta\lambda}$$

For evaluating the integrals, divide the spectral distribution curve of light source into small intervals of equal wavelength spacing $\Delta\lambda$. Determine the centre of the interval as the average of two limits, say, for the range 400 - 410 nm, λ is 405 nm and $\Delta\lambda = 10$ nm. Find the products of $W_\lambda \Delta\lambda$ and $\lambda W_\lambda \Delta\lambda$ and make the summation over the range 400 - 700 nm. Substitute the value of R (reading obtained in quantum sensor for PAR measurement) and the values of integrals, to get the corresponding value of radiometric units in W m^{-2}.

For example, assume a flat spectral distribution curve of the source over the 400 - 700 nm range. As the curve is flat, the number of intervals is 1, $\Delta\lambda = 700-400 = 300$ nm and average $\lambda = (700+400)/2 = 550$ nm. Then

$$W_T = 119.8\ \ R = \frac{\sum_{400}^{700} W_\lambda 300}{\sum_{400}^{700} 550\ W_\lambda 300}$$

$$= 119.8\, R \ \frac{1}{550} = 0.22\ R\ W\ m^{-2}$$

If the PAR reading is R in μmole $s^{-1}\ m^{-2}$ for the given flat distribution curve of light source, the corresponding value for radiometric reading will be 0.22 R Wm^{-2}. If $R = 4.6\ \mu$ mole $s^{-1}m^{-2}$, then $W_T = 1\ Wm^{-2}$.

Similarly PAR units can be converted to photometric units. Photometry refers to the measurement of visible radiation (light) with a sensor having a spectral response curve equal to the average human eye. This response curve covers the waveband of 380 - 770 nm. The human eye responds differently to light of different colours and has maximum sensitivity to yellow and green. Due to the differences in the human eye responses to light CIE (Commission Internationale de l'Eclairage) has given a standard curve with luminosity coefficient of 1 at 550 nm.

In relating photometric unit of illuminance (Lux = lm m^{-2} = luminous flux (lm) per square meter), we have

$$\text{Lux} = 683 \ \int_{400}^{700} Y_\lambda W_\lambda d\lambda$$

where Y_λ is the luminosity coefficient of standard CIE curve with $Y_\lambda = 1$ at 550 nm and W_λ is the spectral irradiance (W m^{-2} nm^{-1}). Following similar arguement as advanced earlier,

$$\text{Lux} = 683 \times 6.023 \times 10^{17}\ R\ hc \ \frac{\int_{400}^{700} Y_\lambda W_\lambda d\lambda}{\int_{400}^{700} \lambda W_\lambda d\lambda}$$

$$= 8.17 \times 10^4\ R \ \frac{\sum_{400}^{700} Y_\lambda W_\lambda \Delta\lambda}{\sum_{400}^{700} \lambda W_\lambda \Delta\lambda}$$

Table 4 : Conversion factors for different light sources

To Convert	Day light	Metal halide	Sodium HP	Mercury	White fluorescent	Incandescent
			Multiply by			
W m^{-2} to µmole s^{-1} m^{-2}	4.6	4.6	5.0	4.7	4.6	5.0
klux to µmole s^{-1} m^{-2}	18	14	14	14	12	20
klux to W m^{-2}	4.0	3.1	2.8	3.0	2.7	4.0

(*luminous flux 1 lm (lumen) at 555 nm = 0.00147 W = 4.08 x 10^{15} quanta s^{-1})

Table 5 : Radiation terms in different systems of units

Term	Radiometric	Photometric	Photon (PAR)
Flux = energy flow per unit time	Joule/Sec =Watt	lumen (lm)	mole/sec Einstein/sec
Intensity = Flux/ Solid angle(sr)	W/sr	lm/sr = 1 cd (candela)	mole/sec.sr
Flux per unit area area of receiving surface	W/m^2 (Irradiance)	lm/m^2 = lux (Illuminance)	mole/sec.m^2
Flux per unit area area of emitting surface	W/m^2 (radiant emittance)	lm/m^2 (Luminous emittance)	mole/sec.m^2
Radiant emittance per unit solid angle	W/m^2.sr (Radiance)	lm/m^2.sr (nit) (Luminance)	mole/sec.m^2.sr

[Full Sunlight = 100klux = 500 W m^{-2}

Full Moonlight = 0.68 lux = 2.3 mW m^{-2} = 0.063 foot-candles
(at a lunar altitude of 60° with clear sky)

sr = steradian, is a unit of solid angle defined as the ratio of the area which subtends the solid angle at the centre of the sphere to the square of the radius of the sphere.]

In evaluating the integral in the numerator, W_λ is to be multiplied at each interval by the luminosity coefficient Y_λ which is different at different wavelengths (at 400 nm, $Y_\lambda = 0.0004$, at 410 nm, $Y_\lambda = 0.0012$, at 420 nm, $Y_\lambda = 0.004$, ... and at 700 nm, $Y_\lambda = 0.0041$).

If the spectral distribution curve is divided into 30 intervals between 400 - 700 nm, of $\Delta\lambda = 10$ nm each, with Y_λ values ranging from 0.0004 to 1 (given by CIE), we obtain

$$\text{Lux} = 8.17 \times 10^4\, R\, \frac{10.682}{17050} = 51.2\, R$$

where R is in µmole s^{-1} m^{-2}. Or

1000 lux = 1 klux = 19.5 µmole s^{-1} m^{-2}

Based on these calculations, conversion factors for various light sources can be worked out (Table 4). In the three different systems of units discussed earlier, various terms are encountered which are compared in Table 5.

Foot-candle is an obsolete unit used earlier to measure the degree of illuminance (or brightness). It is defined as the illumination of a white screen held vertically at a horizontal distance of 1 foot from a standard candle. The illuminance varies as the inverse square of the distance. Thus the same illuminance would be obtained from a 16 cp (candle power: means that the intensity of the light given by the lamp is 16 times as great as that given by a standard candle) lamp at a distance of 4 feet.

Thermal conductance

The amount of heat q transferred through a material is proportional to the temperature difference between the hot side T_h and the cold side T_c and the cross-sectional area of the material A, inversely proportional to the thickness of the material d, and proportional to the time duration such that

$$q = \frac{A(T_h - T_c)}{d}\, t$$

where k is the proportionality constant known as thermal conductivity of the material. Hence

$$k = \frac{qd}{A(T_h - T_c)t}$$

Therefore, the units for thermal conductivity are watts per meter per degree centigrade. Values of conductivity vary from about 421 Wm^{-1} °C for silver to 0.013 Wm^{-1} °C for cotton wool.

Thermal conductance K is equal to the conductivity k divided by the thickness of the material d as follows:

$$K = \frac{k}{d} = \frac{q}{A(T_h - T_c)t}$$

The conductivity of any material is a very specific quantity, whereas the conductance varies with the thickness. Thermal conductance is essentially the heat flow rate per unit area per unit temperature difference, or watts per square meter per degree centigrade.

The insulation I is the reciprocal of the thermal conductance. Hence

$$I = \frac{d}{k} = \frac{A(T_h - T_c)t}{q}$$

The units for insulation are square meters per degree centigrade per watt. Insulation is also referred to as specific resistance. As with conductance, the term insulation is not strictly an intrinsic property of the material but includes the extrinsic characteristics which involve thickness of the material.

Convection

The rate of heat transfer by convection between, say, leaf and the air or water around it is proportional to the temperature difference between the leaf and the air. This proportionality constant is called the convection coefficient h_c.

The rate of heat transfer by convection is

$$C = h_c [T_s - T_a]$$

where T_s is surface temperature and T_a is air temperature. The convection coefficient is a complex function of wind speed or rate of fluid flow, properties of the fluid, and characteristics of the surface over which the flow occurs.

Energy budget of a leaf

A plant leaf absorbs a certain fraction of the incident radiation and partitions this energy into three outgoing streams: radiation, convective heat exchange with the air, and evaporation of water, or transpiration. The radiation emitted by the leaf is proportional to the fourth power of the absolute temperature of the leaf according to the Stefan-Boltzmann equation. The emissivity of a leaf is approximately 1.0, or, to be more exact, is usually about 0.96 and may be as low as 0.92. The rate at which a broad leaf (considered as a flat plate) exchanges heat with the air is proportional to the square root of the wind speed V and the temperature difference between the leaf and the air $(T_l - T_a)$ and is inversely proportional to the square root of the width D of a leaf in the direction of the air flow (approximately the characteristic dimension of the leaf). If the leaf absorbs an amount of radiation Q_a, expressed in watts per square meter, and dissipates this energy by radiation, convection and transpiration, one can write, for steady-state conditions,

$$Q_a = \epsilon\sigma (T_l + 273)^4 + k_1 \frac{V^{1/2}}{D^{1/2}} (T_l - T_a) + \lambda E$$

where $\lambda = 5.67 \times 10^{-8}$ W m^{-2} °K^{-4},

$k_1 = 9.14$ J m^{-2} s$^{-1/2}$ °C^{-1},

V = wind speed (m s^{-1})

D = width of leaf in the direction of flow (m)

E = rate of transpiration (kg m^{-2})

and λ = the latent heat of vapourization, a quantity which is a function of the temperature of the liquid water. The value of λ is approximately 2.430×10^6 J kg^{-1} at 30°C and 2.501×10^6 J kg^{-1} at 0°C.

The transpiration rate is determined not only by the amount of energy available, but also by the vapour gradient between the leaf mesophyll and the free air beyond the boundary layer adhering to the leaf surface. Water vapour must be vapourized at the mesophyll cell walls in the substomatal cavity and then pass through the stomatoes and across the boundary layer of air. As with a fluid passing through any tube or pipe, there is internal leaf resistance r_l expressed in seconds per meter, to the fluid flow. The simplest way to see the relative importance of each mechanism affecting the transfer of energy between a leaf and its environment is to take them one at a time, adding them up as one goes along.

Latent heat

The quantity of heat necessary to produce a change of state from solid to liquid is called the latent heat of fusion, and that quantity necessary to produce a change from liquid to vapour is the latent heat of vapourization. The latent heat of fusion is about 3.33 x 10^5 J kg^{-1} at 0°C. The latent heat of vapourization of water is

2.260 x 10^6 Jkg^{-1} at 100°C
2.356 x 10^6 Jkg^{-1} at 60°C
2.406 x 10^6 Jkg^{-1} at 40°C
2.454 x 10^6 Jkg^{-1} at 20°C
2.501 x 10^6 Jkg^{-1} at 0°C.

The latent heat of water, L, is therefore a function of temperature which can be expressed as

$$L = [(2.490) - (0.00213)(°C)] \times 10^6 \text{ J kg}^{-1}$$

$$= [(594.9) - (0.51)(°C)] \text{ Cal g}^{-1}.$$

The value often used for biological purposes is 2.427 x 10^6 J kg^{-1} at 30°C since many organisms exist at about this temperature much of the time. The error involved with using 2.427 x 10^6 J kg^{-1} is only 2.5 per cent at 40°C and 2.0 per cent at 10°C. Evaporation or evapotranspiration (mm day^{-1}) is often expressed as the amount of energy received per day, by converting the amount of energy received per day into grams of water evaporated/transpired per day using latent heat of vapourization. If the energy received on a surface is given in W hr (= 3600 J), then the energy in Joules is converted into calories by using the mechanical equivalent of heat, i.e., 1 cal = 4.186 J. This is divided by the heat of vapourization to get the amount of water vapourized in g or kg, or in cm^3 or m^3, if divided by the density of water. Once the amount of water vapourized in m^3 is known, the evaporation depth equivalent of water is calculated by dividing the volume of water with the area of surface on which the radiation ~~is~~ received. For example, if the energy received on a surface of 1 m^2 is 1 W hr, then 3600/4.186 = 860 cal of energy is received on 10^6 mm^2 of the surface. Since 2.454 x 10^6 J of energy is required to evaporate 1 kg (or 585 cal for 1 g) of water at 20°C, the amount of water that evaporates is 3600/2.454 x 10^6 = 1.47 x 10^{-3} kg (or 860/585 = 1.47 g). This amount of water has a volume of (1.47 x 10^{-3} kg)/(10^3 kg m^{-3} = density of water)=1.47 x 10^{-6} m^3 [or 1.47 g/(1g cm^{-3} = density of water) = 1.47 cm^3]. The depth equivalent of this water evaporated from a surface of 1 m^2 = 10^4 cm^2 is then equal to 1.47 x 10^{-6} m^3/1 m^2 = 1.47 x 10^{-6} m^3 (or 1.47 cm^3/10^4 cm^2 = 1.47 x 10^{-4} cm). Both these values, either calculated in SI units (J, kg, m) or in conventional units (cal, g, cm) are equivalent to 1.47 x 10^{-3} mm of evaporation/evapotranspiration. Therefore the evaporation depth equivalent, d, of a quantity, Q, of heat flux density is given by

$$d = \frac{Q}{AL\rho_w}$$

where Q is expressed as W m^{-2}, or mW cm^{-2}, or cal cm^{-2} min^{-1}, or in Langley min^{-1}, A is the area on which the radiant energy is incident (m^2, cm^2, or in^2), L is the latent heat of vapourization of water at temperature T, and the ρ_w is the density of water in kg m^{-3} or g cm^{-3}. Since L and ρ_w change relatively slowly with temperature, we often use the value at 20°C (= 293 K), to approximate within the range of 0 to 40°C within ± 2.6 per cent. At 20°C,

Table 6 : Density of pure water vapour at saturation over water (g cm^{-3})

Temperature (°C)	Relative humidity 0.2	0.4	0.6	0.8	1.0
0	0.969	1.839	2.908	3.678	4.847
5	1.359	2.719	4.078	5.438	6.797
10	1.880	3.760	5.639	7.519	9.399
15	2.566	5.132	7.698	10.264	12.83
20	3.460	6.920	10.380	13.840	17.30
25	4.610	9.220	13.830	18.440	23.05
30	6.076	12.152	18.228	24.304	30.38
35	7.926	15.852	23.778	31.704	39.63
40	10.238	20.476	30.714	40.922	51.19
45	13.100	26.200	39.300	52.400	65.50
50	16.612	33.224	39.836	66.448	83.06

1 W hr m^{-2} ≈ 1.47 x 10^{-3} mm evaporation = 5.78 x 10^{-5} inches

1 langley ≈ 1.71 x 10^{-2} mm evaporation = 6.73 x 10^{-4} inches

1 langley min^{-1} ≈ 1.03 mm evaporation hr^{-1} = 4.05 x 10^{-2} inches hr^{-1}

1 W m^{-2} ≈ 1.47 x 10^{-3} mm evaporation hr^{-1} = 5.78 x 10^{-5} inches hr^{-1}

Absolute humidity

Absolute humidity (ρ_w) is the density of water vapour and depends on temperature (T) and pressure (p_w) of water vapour, according to equation:

$$\rho_w = \frac{p_w}{RT} M_w$$

where M_w is the molecular weight of water (= 18). The density of water vapour is also referred to as the absolute humidity (Table 6).

Dry air also obeys the ideal gas law to a first approximation. If ρ_a is the density of dry air at a pressure p_a and temperature T, then

$$\rho_a = \frac{P_a}{RT} M_a$$

where M_a is the molecular weight of dry air (=28.97)

It is desirable to show how density of moist air, ρ, relates to the density of dry air ρ_a. The total pressure p of moist air is the sum of the pressure of dry air and water vapour.

$$p = p_a + p_w$$

Also, the density of moist air, ρ, is the sum of the density of dry air and water vapour.

$$\rho = \rho_a + \rho_w$$

$$= \frac{P_a}{RT} M_a + \frac{P_w}{RT} M_w$$

$$= \frac{(p-p_w)}{RT} M_a + \frac{p_w}{RT} M_w$$

$$= \frac{pM_a}{RT} - \frac{p_w M_a}{RT} + \frac{p_w M_w}{RT}$$

$$= \frac{pM_a}{RT} \left[1 - \left(\frac{M_a - M_w}{M_a}\right) \frac{p_w}{p} \right]$$

$$\text{Density of moist air } (\rho_w) = \frac{pM_a}{RT} \left[1 - 0.379 \frac{p_w}{p} \right]$$

where $[(M_a - M_w)/M_a] = 0.379$. Therefore density of moist air is always less than the density of dry air at the same temperature and pressure. This, of course, is the reason why moist air coming off vegetation, ponds, lakes or the ground surface will rise.

Relative humidity

Relative humidity, h, is the ratio of the actual vapour pressure of the moist air, p, to the saturation vapour pressure, p_s. Hence

$$h = \frac{p}{p_s}$$

Saturation of moist air at a temperature T and total pressure p occurs if the liquid and vapour coexist when the vapour is above a plane surface of pure liquid water. At constant temperature, the ratio of the vapour pressure is also the ratio of the vapour densities.

$$h = \frac{\rho_w}{\rho_s}$$

Specific humidity

Specific humidity, q, is the ratio of the density of water vapour to the density of moist air. Hence

$$q = \frac{\rho_w}{\rho} = \frac{(p_w/RT)\, M_w}{(pM_a/RT)\, [1 - 0.379\, (p_w/p)]}$$

$$= \frac{p_w M_w}{pM_a\, [1 - 0.379\, (p_w/p)]}$$

$$= 0.622 \frac{p_w}{p - 0.379\, p_w}$$

where $M_w/M_a = 0.622$. The specific humidity, q, is usually given in g kg^{-1}. For cold relatively dry air, the values of q range from 1 to 3, and for warm, relatively moist air from 12 to 25.

Humidity mixing ratio

Humidity mixing ratio, w, is the ratio of the density of water vapour to the density of dry air.

$$w = \frac{\rho_w}{\rho_a} = \frac{p_w M_w}{p_a M_a} = 0.622 \frac{p_w}{p - p_w}$$

Usually, p is about two orders of magnitude greater than p_w, or $p - p_w \approx p$. Then

$$w = 0.622 \frac{p_w}{p} = q$$

Thus mixing ratio and specific humidity can be used interchangeably. It also leads to the result that relative humidity is the ratio of the mixing ratio of the moist air to the mixing ratio of the air at saturation w_s.

Water potential

Transport of water through the plant from soil to atmosphere is a simple process to conceive, but difficult to describe in an exact way because there are a number of factors, both biological and physical, that influence the process in one way or the other. The basic driving force for both liquid and vapour movement is the difference in water potential (ø) of soil, plant and atmosphere, which can be calculated by (Van den Honert ,1948)

$$\text{ø} = \frac{RT}{M} \ln \frac{p}{p_s}$$

where R is the gas constant per mole, T is the absolute temperature, M is the molar mass of water and p/p_s is the relative humidity of air in moist soil, and likewise the relative humidity of air in the atmosphere. Thus, from the relative humidity of the atmosphere, the water potential of air at large can be calculated using the above equation. For example, the water potential of air at a relative humidity of 47 per cent at 20°C is

$$\text{ø} = \frac{8.3144 \text{ J/°mole} \times 293° \times \ln (0.47)}{18 \text{ g/mole}}$$

$$= -102182 \text{ J/kg} = -10^5 \text{ J/kg}$$

$$= -10^5 \text{ Nm/kg} = \frac{10^5 \times 10^7}{10^3} \text{ erg/g}$$

$$= -10^9 \text{ dyne-cm g}^{-1}$$

The potential, when expressed on volume basis, is obtained in pressure units. To change the mass (g) to volume, the mass must be divided by the density. Then, as the mass term occurs in the denominator

$$\text{ø} = - \frac{10^9 \text{ dyne-cm}}{\text{g. cm}^3 \text{ g}^{-1}} = -10^9 \text{ dyne cm}^{-2}$$

$$= -10^3 \text{ bars}$$

since 10^6 dyne/cm^2 = 1 bar. As the negative pressure is called the suction,

$$\text{ø} = -10^3 \text{ bars} = 10^3 \text{ bars of equivalent suction.}$$

The potential actually is not in pressure units and therefore it is denoted as equivalent suction. Air water potential forms the demanding component of the soil-plant-atmosphere continuum (SPAC), while considering water flow through evaporation or transpiration. Potentials of the order 0 to -15 bars are encountered in soils, while in plants, the order of potential is about -50 bars and that in atmosphere about -1000 bars.

In defining the potential, as the potential energy per unit quantity of water, the quantity of water can be expressed as mass, volume or weight. Potential per unit mass of water is expressed as J kg^{-1}. Potential per unit volume is expressed in joules per cubic meter (J m^{-3}). Since this has the dimensions of pressure, in the equivalent pressure units, this becomes newton per square meter (Nm^{-2}) or Pascal(Pa). Potential per unit weight is expressed as length (head) in meters. A potential per unit weight of h meters is equivalent to gh J kg^{-1} on a mass basis or ρgh Pa on a volume basis.

Pressure

Pressure is measured by a mercury barometer which is a long glass tube that has been filled with mercury and then inverted in a reservoir of mercury. The space above the mercury column contains only mercury vapour, whose pressure, at room temperature, is so small that it may be neglected. Then the mercury in the reservoir is forced into the inverted tube due to atmospheric pressure until the height of mercury column is 76 cm. The height of mercury column depends on ρ and g as well as on atmosphereic pressure. The density, ρ, varies with temperature, g with latitude and elevation above sea level. In calculating the atmospheric pressure (p_a), g can be taken as 9.80 m s^{-2} and ρ as 13.6 x 10^3 kg m^{-3} for most purposes, then

$$p_a = \rho gh = 13.6 \times 10^3 \text{ kg m}^{-3} \times 9.8 \text{ ms}^{-2} \times 0.76 \text{ m}$$

$$= 101,300 \text{ N m}^{-2} = 1.013 \times 10^5 \text{ Pa}$$

As 0.76 m = 30 in = 2.5 ft, and g = 850 lb.ft^{-3},

$$p_a = 2120 \text{ lb.ft}^{-2} = 14.7 \text{ lb in}^{-2}$$

a pressure of 1.013 x 10^5 P_a is called one atmosphere (1 atm). A pressure of exactly 10^5 N m^{-2} is called one bar, and a pressure of one one-thousandth as great is one millibar.

1 bar = 1020 cm of water ≈ 1000 cm of water
= 401.5 inch of water
= 75.01 cm of mercury
= 0.9869 atm ≈ 1 atm
= 100 cbar ≈ 1000 mbar
= 100 J/kg
= 14.5 lb/in^2
= 10^6 erg/g
= 10^6 dyne/cm^2.

The SI unit of pressure is Pascal (Pa). Atmospheric pressures are of the order of 1000 mbars and are now stated in terms of this unit by the weather services. A number of units are used to express suction, tension, stress or potential. They are: atmosphere, bar, centibar, millibar, dyne per square centimetre, erg per gram, joule per kilogram, pound per square inch, centimetre of water, centimetre of mercury, inch of water. All these units are convertible from one to the other. The Table 7 gives the conversion of pressure units from one form to the other.

Expression of soil water

Water content traditionally has been expressed as the ratio of mass of water present in a sample to the mass of a sample after it has been dried to constant weight, or as the volume of water present in a unit volume of sample. Thus water content is a dimensionless ratio of two masses or two volumes and if the ratio is multiplied by 100, these values become percentages. Therefore, it becomes necessary to indicate the percentages, whether they are on mass or on volume basis (Hanks and Ashcroft, 1980).

$$\theta_m = \frac{\text{mass of water}}{\text{mass of dry soil}}$$

$$\theta_v = \frac{\text{volume of water}}{\text{bulk volume of soil}}$$

For example, taking a soil sample of 10 cm x 10 cm x 10 cm having a total wet mass of 1800 g, of which 300 g is water, and assuming the density of water as 1 g cm^{-3}, we have:

Table 7 : Conversion of pressure from one unit to the other

		Pressure units convert to										
c		mbar	bar	torr	Pa(Nm^{-2})	atm	lbf in^{-2}	kgf cm^{-2}	in Hg	mm Hg	in H$_2$O	mm H$_2$O
o	1 mbar =	1	1x10^{-3}	0.75	10^{2}	9.869x10^{-4}	1.45x10^{-2}	1.02x10^{-3}	2.953x10^{-2}	0.75	0.402	10.197
n	1 bar =	10^{3}	1	7.5x10^{2}	1x10^{5}	0.987	14.5	1.02	29.53	7.5x10^{2}	4.015x10^{2}	1.02x10^{4}
v	1 torr =	1.333	1.333x10^{-3}	1	1.333x10^{2}	1.316x10^{-3}	1.934x10^{-2}	1.36x10^{-3}	3.937x10^{-2}	1	0.535	13.69
c	1 P$_a$(Nm^{-2}) =	0.01	1x10^{-5}	7.5x10^{-3}	1	9.87x10^{-6}	1.45x10^{-4}	1.02x10^{-5}	2.953x10^{-4}	7.5x10^{-3}	4.015x10^{-3}	0.102
r	1 atm =	1.013x10^{3}	1.013	7.6x10^{2}	1.013x10^{5}	1	14.7	1.033	29.92	7.6x10^{2}	4.068x10^{2}	1.033x10^{4}
t	1 lbfin^{-2} =	68.95	6.895x10^{-2}	51.71	6.895x10^{3}	6.805x10^{-2}	1	7.03x10^{-2}	2.036	51.71	27.68	7.03x10^{2}
	1 kgfcm^{-2} =	9.807x10^{2}	0.981	7.356x10^{2}	9.807x10^{4}	0.968	14.22	1	28.96	7.356x10^{2}	3.937x10^{2}	10^{4}
f	1 in Hg =	33.86	3.386x10^{-2}	25.4	3.386x10^{3}	3.342x10^{-2}	0.491	3.453x10^{-2}	1	25.4	13.6	3.45x10^{2}
r	1 mm Hg =	1.333	1.333x10^{-3}	1	1.333x10^{2}	1.316x10^{-3}	1.934x10^{-2}	1.36x10^{-3}	3.937x10^{-2}	1	0.535	13.59
o	1 in H$_2$O =	2.491	2.491x10^{-3}	1.868	2.491x10^{2}	2.458x10^{-3}	3.613x10^{-2}	2.54x10^{-3}	7.356x10^{-2}	1.868	1	25.4
m	1 mm H$_2$O =	9.807x10^{-2}	9.807x10^{-5}	7.354x10^{-2}	9.807	9.677x10^{-5}	1.42x10^{-3}	10^{-4}	2.896x10^{-3}	7.354x10^{-2}	3.394x10^{-2}	1

Note : also 1 dyne cm^{-2} (barye) = 0.1 Pa(Nm^{-2}) = 10^{-3} mbar

$$\theta_m = \frac{\text{mass of water}}{\text{mass of dry soil}} = \frac{300}{1800\text{-}300}$$

$$= \frac{300}{1500} = 0.2 \text{ (g/g; kg/kg)} \quad \text{(20\% gravimetric)}$$

$$\theta_v = \frac{\text{volume of water}}{\text{bulk volume of dry soil}}$$

$$= \frac{\text{mass of water/density of water}}{\text{bulk volume of dry soil}}$$

$$= \frac{300/1}{10 \times 10 \times 10} = \frac{300}{1000} = 0.3 \text{ (cm}^3\text{/cm}^3\text{; m}^3\text{/m}^3\text{)} \quad \text{(or 30\% volumetric)}$$

If water content is desired on a volume basis, the volume of the sample must be known, or the bulk density of the soil (mass of soil per unit volume) must be known. Mass based water contents can be converted to volume figures by using the relationship,

$$\theta_v = \theta_m (\rho_b / \rho_w)$$

where, ρ_b is the bulk density of the soil, and ρ_w is the density of water (usually taken as unity in g cm^{-3} or Mg m^{-3}). In the above example, the bulk density is 1.5 g cm^{-3}, which is obtained by dividing dry mass of sample of soil by its volume. It can be seen that the volumetric (volume basis) moisture content is obtained by multiplying the gravimetric (mass basis) moisture content with the value of bulk density ($0.2 \times 1.5 = 0.3$). Occasionally, soil water is expressed on wet basis (θ_{wm}, θ_{wv}) or on dry basis (θ_{dm}, θ_{dv}). Conversion from one base to the other is accomplished by the following equations (Klute, 1986):

<u>Gravimetric</u> <u>Volumetric</u>

$$\theta_{wm} = \theta_{dm}/(1+\theta_{dm}) \text{ and } \theta_{wv} = \theta_{dv}/(1+\theta_{dv})$$

$$\theta_{dm} = \theta_{wm}/(1-\theta_{wm}) \text{ and } \theta_{dv} = \theta_{wv}/(1-\theta_{wv})$$

When water is applied to soil by irrigation, the quantity applied is reported as the depth of water, if it were accumulated as a layer. In case of rainfall also, the water received is expressed as cm or mm. When the volume of water received is divided by the area of the receiving surface, the rainfall is obtained in cm or mm. Common expressions for water in soil are acre-foot or acre-inch, which indicate volume of water, and if divided by the area (acre), water in depth units (feet, inches) is obtained. In soils, the volume of water that is stored in a layer (e.g., root zone) is obtained in depth units cm or m by dividing the figure of volumetric moisture content in the layer by the area of soil surface.

$$\text{Depth of water} = \frac{\text{volume of water}}{\text{bulk volume of soil}} \times \frac{\text{bulk volume of soil}}{\text{area of soil surface}}$$

The volumetric moisture content in any soil layer should be multiplied by the volume of that soil layer to get the volume of water, which then should be divided by the area of soil surface of the soil layer under consideration (acre, etc.). Conversely, if the depth of water is to be converted into volumetric moisture content, it must be multiplied by the area of soil surface and then divided by the bulk volume of soil. These operations are necessary to maintain the dimensional require

Standard Notations

Prefix					Prefix
E	EXA	10^{18}	10^{-18}	ATTO	a
P	PETA	10^{15}	10^{-15}	FEMTO	f
T	TERA	10^{12}	10^{-12}	PICO	p
G	GIGA	10^{9}	10^{-9}	NANO	n
M	MEGA	10^{6}	10^{-6}	MICRO	μ
k	KILO	10^{3}	10^{-3}	MILLI	m
h	HECTO	10^{2}	10^{-2}	CENTI	c
da	DEKA	10^{1}	10^{-1}	DECI	d

ments. Otherwise, the depth of water is obtained simply by multiplying volumetric moisture content by the depth of soil layer. For example, if a soil 80 cm deep has a volume water content, θ_v, of 0.12, the quantity of water that must be added to bring the volume water content to 0.30 is the difference between the initial (0.12 x 80 cm = 9.6 cm) and final (0.30 x 80 cm = 24.0 cm) water contents, i.e., 14.4 cm. Similarly, 10 cm rain will wet the soil to a depth, which can be obtained by dividing the rainfall (cm) by the difference between the final and initial volumetric moisture contents. These calculations assume that the bulk density is uniform throughout the soil profile, which is rarely the case. If the bulk density varies from layer to layer, calculations have to be done for each of the layers and the combined value (total profile moisture) must be worked out from there. For example, if a soil has a field capacity volume water content of 0.36, from the values given for different layers the depth to which a given amount of rain (say 7 cm) penetrates is calculated as

Depth of layer (cm)	Initial mass water content	Bulk density (g cm^{-3})
0 - 15	0.08	1.50
15 - 30	0.15	1.54
30 - 45	0.16	1.55
45 - 60	0.17	1.52

The difference in volumetric moisture content between the field capacity and the initial moisture for the first layer is 0.36 - 1.5 x 0.08 = 0.24. So the amount of water to wet a layer of top 15 cm is 15 x 0.24 = 3.6 cm. Similarly for the next layer, it is 1.94 cm. The remaining water is 7 cm - (3.60 + 1.94) cm, i.e., 1.46 cm is not enough to wet completely the next 15 cm layer.

The depth to which it will penetrate is obtained by dividing the remaining rain water by the difference in the volumetric field capacity and the initial volumetric water content of that layer, i.e., 0.36 - 1.55 x 0.16 = 0.112. Then the depth to which it will penetrate in the third layer is 1.46 ÷ 0.112 = 13.0 cm. The total depth of penetration is then equal to 15 cm + 15 cm + 13.0 cm = 43.0 cm. Similar calculations can be made for the depth to which a given amount of irrigation can penetrate.