

Hydrology

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1 Flows

1.1 Types of flow

If a liquid flow is in contact with the atmosphere, the flow is said to have a free surface, as in the case of an open irrigation channel, for example. If the flow takes place in a pipe, occupying the whole section and at a pressure higher than atmospheric, it is termed a pressurised flow. This is the case in water-distribution systems. If the liquid flows through a porous material, it is regarded as having flow in a porous environment.

If the characteristics of the flow at each point are independent of time, it is known as a permanent regime. Otherwise, the regime is called variable or transient.

When every particle of the liquid describes a well-defined trajectory, and has a velocity only in the direction of flow (parallel flow lines), the motion is laminar. When some particles of the liquid are in agitated motion due to velocities transverse to the flow, the motion is called turbulent.

1.1.1 PRESSURISED FLOW

Turbulence is basically caused by viscosity, defined by Reynolds number (R_e) as follows:

$$R_e = \frac{V.D}{\nu}$$

where V is the flow velocity (m/s), D the pipe diameter (m), and ν the kinematic viscosity of the water (m²/s).

According to Carlier:

- when $R_e < 2\,000$ the regime is laminar,
- when $R_e > 2\,000$ the regime is turbulent.

The transition between the two regimes is not very clear, and there is often what is termed a ‘transition zone’. In practice, and for water-supply systems, the regime is always turbulent.

1.1.2 FLOW IN A POROUS ENVIRONMENT

To define the flow in a porous environment the Reynolds number is used as follows:

$$R_e = \frac{V.d}{\nu}$$

where R_e is the Reynolds number (non-dimensional), V the filtration speed (Darcy) in m/s, d the characteristic grain diameter, either d_{10} (De Marsily) or d_{50} (Lencastre) in m, and ν the kinematic viscosity of water, in m^2/s .

According to De Marsily, flow in a porous environment is considered laminar when $R_e < 10$. Therefore, when $R_e > 100$, the flow is clearly turbulent (and Darcy’s law is no longer valid). In practice, flow only becomes turbulent in a karstic environment, or in the immediate surroundings of a pumped well. De Marsily has proposed an empirical formula to estimate the transition from laminar to turbulent flow using the hydraulic gradient. It defines the gradient limits as $i = 1/15 (K)^{1/2}$.

To calculate R_e , the values of kinematic viscosity of the water depending on temperature are:

$T(^{\circ}C)$	5	10	15	20	30
$\nu \cdot 10^6 (m^2/s)$	1.52	1.31	1.14	1.0006	0.8

1.2 Types of energy

Energy is defined as the product of force and displacement. In hydraulics, energy is usually related to the unit of weight, and designated in a simplified manner by load or height; in fact it has the dimension of a length (energy $[m^2.kg.s^{-2}]$ / weight $[m.kg.s^{-2}] = [m]$).

A water particle travelling at velocity V , subject to a pressure p , and located at a height z above a horizontal plane of reference, possesses the following energy (Lencastre):

- potential energy: $E_z = z$
- pressure energy: $E_p = p / \omega$
- kinetic energy: $E_k = V^2/2g$

where p is the relative pressure (relative to atmospheric), ω the specific gravity of water, z the altitude of the water particle in relation to a reference line, V the velocity, and g the acceleration due to gravity.

Thus, the total energy, or total head, is:

$$E = z + \frac{p}{\omega} + \frac{V^2}{2g}$$

1.2.1 PRESSURISED OR FREE-SURFACE FLOW

This energy is usually represented graphically.

1.2.2 POROUS ENVIRONMENT

In a porous environment, flows usually travel very slowly. It is therefore considered that the dynamic head, $E_k = V/2g$, is negligible. For instance, a velocity of 1 cm/s, which would be quite a fast underground flow, gives a dynamic head of $E_k = 1/1\,962 = 5$ microns.

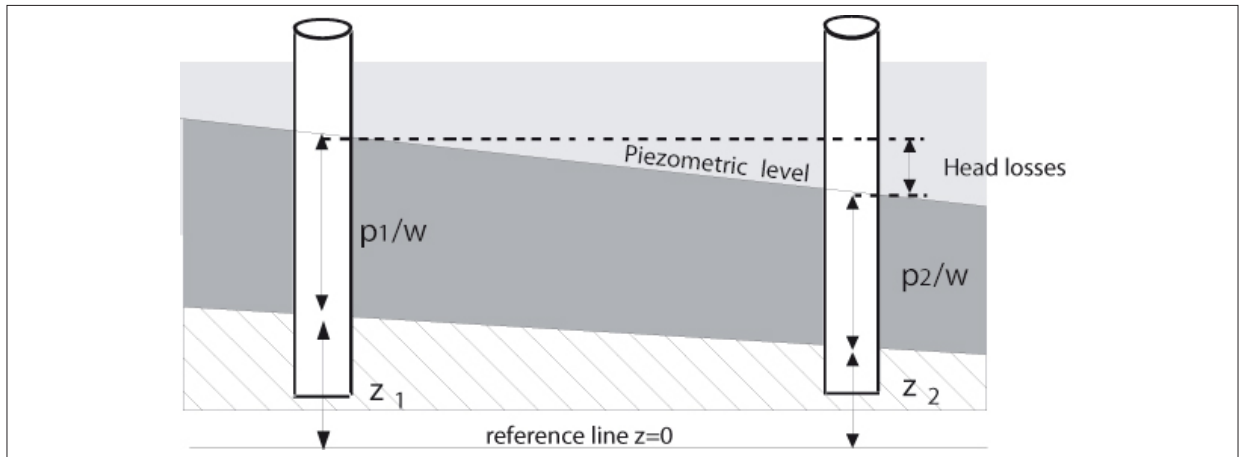


Figure 1: Flow in a porous environment.

Therefore, the total head becomes $E = z + p/\omega$, and is called the static or piezometric head (Figure 1).

1.3 Bernoulli's theorem

Bernoulli's theorem is a direct result of the principle of conservation of energy in a permanent regime, for an incompressible fluid:

$$E = z + \frac{p}{\omega} + \frac{v^2}{2g} = \text{constant}$$

This principle produces a horizontal energy line for zero flow, and explains that the energy line falls due to head losses when a flow occurs. If the head losses (J) are not neglected, the following can be written:

$$E + J = \text{constant}$$

Head losses correspond to the difference between the static and dynamic energy lines.

1.4 The continuity equation

In a cylindrical pipe, the flow Q is:

$$Q = V S = V \frac{\pi D^2}{4}$$

where Q is the flow (m^3/s), V the velocity (m/s), S the cross-sectional area of the pipe (m^2), and D the interior diameter of the pipe (m).

The continuity equation represents the physical principle of conservation of mass, and can be written in a simplified manner:

$$Q = V_1 S_1 = V_2 S_2$$

In other words, the flow that enters a pipe is equal to the flow that leaves it, whatever the changes in diameter or velocity.

2 Simplified hydrological balance

2.1 Upstream approach

The procedure is as follows:

- calculation of total flow: the total flow (E), which corresponds to the surface flow (run-off), plus underground flow (useful precipitation, P_u), is calculated from rainfall data;
- estimation of inputs ($P_u + I_s$): P_u is estimated by establishing the proportions of the total flow as run-off and infiltration (P_u). The other inputs (I_s) are measured;
- estimation of changes in stock (Δs);
- calculation of recharge from the balance of the various terms.

2.1.1 ESTIMATION OF TOTAL FLOW

The total flow (E) is a fraction of precipitation. Part of the precipitation evaporates or is used by plants, and part becomes surface or underground flow.

2.1.1.1 Rainfall - temperature

It is advisable to collect all available rainfall data and/or to install a network of rain gauges or pluviographs.

The density of the rain-gauge network is chosen according to the objectives, and environmental conditions. Within a climatic zone, rainfall depends on many different factors: altitude, exposition of side of valley, distance from the sea etc. In a restricted and relatively homogeneous area, such as some refugee camps, one or two rain gauges may be enough. At district level, the network must be more extensive to reflect the differences in the environment.

Temperature summaries are made with a minimax thermometer. The average temperature is calculated daily. These summaries are recorded on a measurement sheet prepared for the purpose.

The measurement sequence must be over as long a period as possible, ideally covering several years. However, the rain-gauge network should be installed as quickly as possible, because even though there may be no prospect of installing a more permanent network at the beginning of a programme, some ACF projects for wells and boreholes take place over periods of more than 3 or 4 years.

An automatic recording station for rainfall and temperature data can be installed to facilitate recording (ACF Somaliland, 1998).

A geographic representation is then assigned to each rain gauge. The standard methods are:

- arithmetic mean, the least accurate but easiest method;
- Thiessen's method (balancing by area), usable in fairly flat areas (Box 1);
- the isohyet method (linear variation proportional to distance), more accurate for hilly and mountainous relief.

Total rainfall estimated in this way feeds evapotranspiration and total flow.

2.1.1.2 Evapotranspiration

Evapotranspiration is the fraction of water taken up by direct evaporation, and by the transpiration of plants. A large number of factors influence this parameter: temperature, wind, humidity, plant cover, solar radiation etc.

There are different methods for calculating evapotranspiration, but in ACF programmes only estimations from empirical formulas can be used.

The timescale used for calculation must be as small as possible, so as not to flatten out the variations excessively: periods of ten or thirty days are best, but formulas on an annual basis may be used in order to establish orders of magnitude.

Box 1

Interpretation of point data.

Thiessen's method

This method is generally used for rainfall data. It assumes that the area covered by each rain gauge extends halfway to the next rain gauge. Over this area, rain is uniform and equal to the amount recorded by that particular rain gauge.

The geometry of the various areas is determined geometrically. Neighbouring rain gauges are connected by segments of a line, and then the mid-perpendiculars of these segments are drawn. Where they meet, they form polygons, each of which encloses a rain gauge. Polygons on the edge of a catchment basin follow its contour (Figure 1).

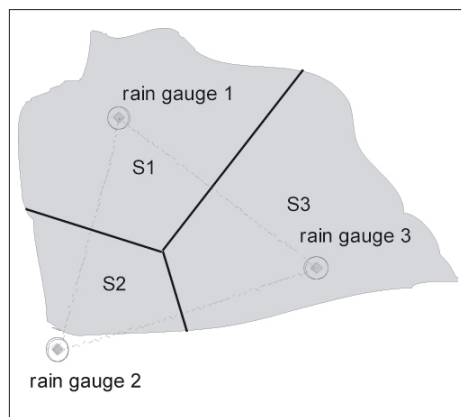


Figure 1: Thiessen's method: weighting by area.

Average rainfall in the basin is:

$$P = \frac{(\sum p_i S_i)}{S}$$

where P is the average rainfall for the whole basin (mm), p_i the rainfall recorded by rain gauge i (mm), S_i the area of the polygon enclosing rain gauge i (m^2), and S the total area of the drainage basin (m^2).

Thiessen's method should be used when there is a high density of rain gauges: then it is a weighting technique which is easy to use. On the other hand, it tends to give a larger weighting to isolated stations when the network is not dense. If these isolated stations are not representative of the catchment basin, a significant error in the total rainfall could be introduced.

Isovalue method

An isovalue curve is formed by all the points that present the same value of a given parameter (level, rainfall, conductivity etc.) for a given date or period of time. Thus, isopiezometric lines are curves of the same piezometric level, contours are curves of equal altitude, isohyets are curves of equal rainfall, isoconductivity curves join points of the same conductivity value etc.

This method assumes that between two neighbouring measurement points, the distribution of the parameter value is linear and directly proportional to the distance (Figure 2).

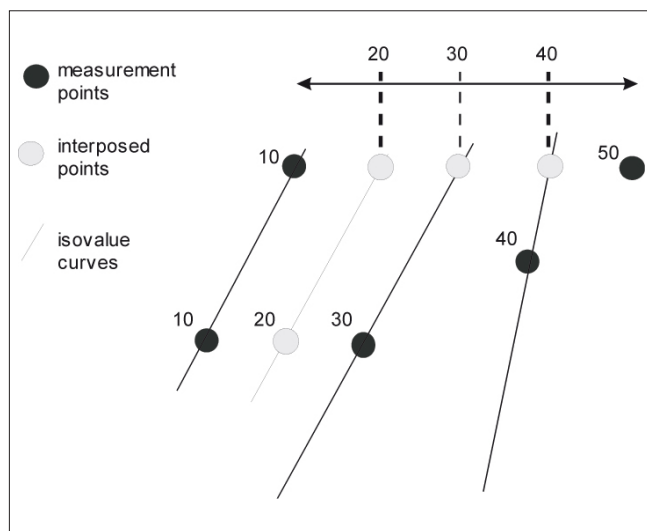


Figure 2: Isovalue method: linear variations proportional to distance.

When drawn by hand, isovalue curves help to take into account important features: surface water in the case of isopiezometric lines, relief or exposition in the case of isohyets etc. This is also possible when curves are drawn using software such as Surfer, but this requires painstaking input of parameters to modify the programme: if this adjustment is not carried out, the results can be erroneous. Therefore, it is generally better to draw curves by hand to integrate all the different features of the environment more quickly.

Measurement of area

It is often necessary to calculate areas from maps or diagrams, e.g. for calculation of the area of a catchment basin, Thiessen polygons, areas defined by curves etc.

The simplest method in the field is to draw or trace the areas to be calculated on squared paper and then count the squares, which is a time-consuming but effective method.

It is also possible to use a planimeter, which is a device that facilitates the measurement of areas, whatever their shape. It consists of an articulated arm, with one of its extremities fixed to a pedestal, and the other one supporting a view-finder and an area meter. The purchase of such a device is only justified if areas must be measured regularly.

Finally, images handled with certain software (Surfer, Mapinfo etc.) can also be processed by, for example, drawing polygons, which are then easy to measure.

In the formulas given in Box 2, a distinction is made between potential evapotranspiration (PET), and real evapotranspiration (RET). The potential evapotranspiration is always a maximum theoretical figure, assuming that the plant is in its optimum condition of transpiration capacity (especially that the plant has access to the required quantity of water to fully transpire).

When rainfall P, infiltration I, and real evapotranspiration (RET) are known, it is possible to calculate the total flow:

$$E = P - RET - I$$

Box 2

Calculation of RET.

Thornthwaite balance method

This method allows work at monthly or ten-days intervals. First, it is necessary to calculate the potential evapotranspiration (PET), which can be regarded as the evaporating power of the atmosphere. Then the real evapotranspiration (RET) is calculated, by comparing the PET and the available water resources for evaporation and transpiration. Even if potential evapotranspiration is high, water must be available to allow it to take place. This water availability is calculated from the 'easily usable reserve' (EUR). This is a theoretical reserve: it does not have a physical reality, even though the EUR could be regarded as the quantity of water stored in the first few metres of soil.

Calculation of the monthly PET using Thornthwaite's formula

$$PET = 1.6 \left(\frac{10t}{TE} \right)^a$$

where PET is the monthly potential evapotranspiration (cm), t the monthly average temperature (°C), TE the annual thermal index ($TE = \sum_{i=1}^{12} i$), i the monthly thermal index [$i = (t/5)^{1.514}$], and $a = 0.016 TE + 0.5$ (Serra's simplified a).

This formula gives good results in maritime temperate climates (e.g. in Normandy, France), but underestimates the PET in hot dry climates by up to 30%.

Calculation of the PET on a monthly or ten-day basis using Turc's formula

$$PET = c \left(\frac{t}{t + 15} \right) (I_g + 50) b$$

where PET is monthly or ten-day potential evapotranspiration (mm), t the monthly or ten-day average temperature (°C), I_g the global solar radiation (cal/cm²/day), I_0 the theoretical maximum radiation (cal/cm²/day), H the astronomical length of the day (h), and h the measured duration of sunshine (h).

- For the monthly PET, c is 0.40 for 30 and 31 day months, and 0.37 for February; for ten-day PET, c is 0.13.
- b is 1 if the average relative humidity of the atmosphere (rh) is greater than 50% (the most frequent case). In desert climates, $rh < 50\%$, and $b = [1 + (50 - rh)/70]$.
- Parameters I_0 , H, and h are tabulated: $I_g = I_0 [0.18 + 0.62 (h/H)]$.

Although it is more difficult to use than Thornthwaite's, this formula gives good results even in hot climates (in very hot climates, it still underestimates the RET, but by less than 10%).

Thornthwaite balance

This method compares the PET calculated by one of the two previous formulas with the rainfall and EUR, over a given period of time. The calculation principle must allow the rain first to feed the PET, then to complete the EUR up to its top level, and finally to feed surface run-off when the PET and the EUR have been satisfied. The following procedure must be followed:

2.1.2 ESTIMATION OF INPUTS

2.1.2.1 Inputs

The main input to the system is generally represented by useful precipitation, which is the fraction of rain that filters into the ground and that will therefore effectively contribute to the recharge of the aquifer. However, there are numerous other sources which can feed an aquifer: infiltration of surface water from rivers, lakes etc., other aquifers, leaks from irrigation channels etc. These sources can be indicated by piezometry, water chemistry (the simplest method being the use of a conductivity meter) and, of course, geology.

– if $P + EUR \geq PET$, $RET = PET$, $EUR = P - RET$, and $E = P - (EUR + RET)$;

– if $P + EUR < PET$, $RET = PET - (P + EUR)$, $EUR = P - RET$, and $E = 0$,

where P is the precipitation, PET the potential evapotranspiration, RET the real evapotranspiration, EUR the easily usable reserve, and E the flow (surface and underground). An example is given in Table I.

Table I: Thornthwaite balance on a monthly basis with initial EUR of 100 mm.

P = precipitation (mm); EUR = easily usable reserve; PET = monthly potential evapotranspiration (mm); RET = real evapotranspiration (mm).

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
P	100	90	80	70	60	50	50	60	70	80	90	100	900
EUR	100	100	100	86	35	0	0	0	0	8	54	100	
PET	34	40	61	84	111	138	156	140	105	72	44	32	
RET	34	40	61	84	111	85	50	60	70	72	44	32	743
E	66	50	19	0	0	0	0	0	0	0	0	22	157

The initial EUR must be chosen on the basis of geological and pedological contexts. The value EUR = 100 mm is generally adopted for contexts similar to those in which Thornthwaite developed his method: humid, moderate climate with cultivated land (eastern United States). For a soil which is not very suitable for storing water, the EUR can be decreased: in a karst context (Vaucluse fountain), Puig chose an EUR of 50 mm. The EUR value can be verified by comparing the actual annual flow value with the values calculated by Turc's and Coutagne's formulas.

Calculation of the annual RET using Coutagne's formula

$$RET = P - \lambda P^2$$

where RET is the real evapotranspiration (m), P the annual average precipitation (m), $\lambda = 1/(0.8 + 0.14 T)$ if $1/8 < P < 1/2$, T is the average annual temperature ($^{\circ}\text{C}$).

If:

– $1/8 > P$, then RET is equal to precipitation, and therefore there is zero flow;

– $P > 1/2$, then RET is independent of precipitation, and is calculated by:

$$RET = 0.2 + 0.035 T$$

Calculation of the annual RET using Turc's formula

$$RET = \frac{P}{[0.9 + (P^2/L^2)]^{1/2}}$$

where RET is the real evapotranspiration (mm), P the annual average precipitation (mm), T the average annual temperature ($^{\circ}\text{C}$), and $L = 300 + 25T + 0.05T^3$.

Table I: Hydrogeological indexes calculated from annual data taken over several years (according to Castagny & BRGM).

	Sedimentary zone (chalk, Hallue basin)	Bedrock zone (schists and granites, Burkina Faso)		
Average precipitation (mm/year)	740	513	807	1076
Average PET (mm/year)	514	310	446	674
Average E (mm/year)	226	203	361	402
Index (%)	92	67	61	66
Average P_u (mm/year)	208	136	221	266
Surface flow average (mm/year)	18	67	140	136

The volume of these sources is calculated by Darcy's law, and by direct observation. However, the constraints involved in the implementation of humanitarian programmes do not always allow enough time to carry out these investigations. Therefore, only simple systems with easily quantifiable inputs can be studied.

2.1.2.2 Useful precipitation

It is possible to establish a ratio between the portion of flow which becomes surface water and the portion which infiltrates. This ratio is based on the distributing power of the soil, calculated from the hydrogeological index (Table I).

It is essential to choose an index which corresponds to the conditions of the survey area. This choice is very tricky, and it greatly influences the results: it is therefore necessary to validate the result of the balance with that obtained using the hydrogeological method.

2.1.3 VARIATIONS IN STOCK

The volume of water stored in the aquifer at any time t is given by the volume of the reservoir, and the useful porosity.

The volume of the reservoir is estimated from its area and depth. Its area is estimated from the geological map (topographic) and/or land surveys. Its power is known from the results of borehole drilling and geophysical studies.

Useful porosity (or the storage coefficient if the aquifer is confined) is calculated from the results of pumping tests.

Without doing this calculation, which is sometimes difficult to carry out in the field, simple piezometric monitoring over several years enables the changes in the stock of underground water (Δs) to be measured. If variations are negligible, this term can be removed from the balance.

2.1.4 ESTIMATION OF RECHARGE

Inputs have been quantified, and changes in stock evaluated. Now it is simple to calculate recharge from the expression for the balance:

$$\text{recharge} = P_u + I_s \pm \Delta s$$

where P_u is the useful precipitation (m^3/year), I_s other sources (m^3/year), and Δs the changes in stock of the aquifer (m^3/year).

The example given in Table II represents a simplified balance, carried out in Corsica. Different values of EUR were tested: the EUR value = 50 mm was taken because the pedology of the site (little soil) allowed the assumption that the water storage potential of the soil accessible to plants was

Table II: Simplified hydrogeological balance carried out in Corsica from rainfall data and monthly average temperatures (1961-1992).

Calculation method for total flow		E (mm/year)		E (l/s/km ²)	
Coutagne annual		97.2		3.1	
Turc annual		60.9		2	
Thornthwaite balance with Turc monthly PET		108.5		3.5	

Catchment basin	Area (km ²)	Calculated total flow (balance)		Outlets		Inputs (P _u = E-ESF) (m ³ /year)	Balance (m ³ /day) (non-exploited renewable resources P _u -Q _p)
		E (l/s/km ²)	E (m ³ /year)	Estimated surface flow (m ³ /year)	Pumping Q _p (m ³ /year)		
1	7.0	3.5	772 632	515 088	73 000	257 544	505
2	4.0	3.5	441 504	220 752	146 000	220 752	204
3	3.0	3.5	331 128	0	73 000	331 128	707
4	4.5	3.5	496 692	248 346	146 000	248 346	280
5	4.9	3.5	540 842	180 280	146 000	360 561	587
TOTAL	23.4	3.5	2 582 798	1 164 466	584 000	1 418 331	2 285

ESF = estimated surface flow.

low. Also, calculations carried out with a EUR of 50 mm gave flow values comparable to those calculated from annual formulas.

2.2 Downstream approach

The procedure is as follows:

- calculate outlets by estimating pumping withdrawals (Q_p), and measure the underground flow (Q_s) which feeds rivers and sources;
- assess change in the stock (s);
- estimate refilling from the computable balance of the various terms.

2.2.1 ESTIMATION OF OUTLETS

2.2.1.1 Withdrawals

Sometimes, with motorised pumping, the pumped volume of water is recorded. If this is not the case, and for manual pumps, it can be calculated from the number of users, average daily use, and pumping flow. A handpump used intensively, for example for 6 h/day, can produce (6 h x 800 l/h), that is, 4.8 m³/day.

2.2.1.2 Springs and rivers

The flow of springs and rivers fed by the aquifer corresponds to outlets from the system. It is therefore necessary to measure this flow, and to take it into account in the balance. However, this is often the balance term most difficult to obtain.

The flow of springs is measured at fixed times (on a monthly or ten-day basis).

Strictly speaking, it is essential to have a sufficiently long flow record to be able to calculate the base river flow, i.e. that generated by the aquifer flow, and not by surface run-off. In ACF programmes, such records are difficult to obtain. It can therefore be taken that the lowest flow level results only from the contribution of the aquifer (i.e. no rain and so no run-off).

2.2.1.3 Other outlets

Aquifer outlets to the sea, flow between different reservoirs etc., are difficult to measure in humanitarian programmes. Therefore, it is not possible to carry out balances on complex systems.

2.2.2 VARIATIONS IN STOCK

Variations are calculated in the same way as for the upstream approach (see Section 2.1).

2.2.3 ESTIMATION OF RECHARGE

The estimation results from the balance of the different terms:

$$\text{Recharge} = Q_s + Q_p \pm \Delta s$$

where Q_s is underground flow (m^3/year), Q_p pumped flow (m^3/year) and Δs the variations in stock (m^3/year).

3 World meteorological data

3.1 January temperatures ($^{\circ}\text{C}$)

Latitude	Longitude																	
	105°W	90°W	75°W	60°W	45°W	30°W	15°W	0°	15°E	30°E	45°E	60°E	75°E	90°E	105°E	120°E	135°E	150°E
30°N																		
20°N	12	16						13	10	12	8	6	16	17	16	8		
10°N	21	24	24					22	21	15	15	16	16	22	24		21	
0°				22	23			24	22	22	22	24		27	20	26	21	
10°S		22	26	23	25			23	23	25					20	26	26	27
20°S		10	24	24	21					23	25	21					30	27
30°S				26	23					23	22	22				27	32	31

3.2 July temperatures ($^{\circ}\text{C}$)

Latitude	Longitude																	
	105°W	90°W	75°W	60°W	45°W	30°W	15°W	0°	15°E	30°E	45°E	60°E	75°E	90°E	105°E	120°E	135°E	150°E
30°N																		
20°N	27	23						30	30	25	22	28	30	31	23	27		
10°N	21	20	21					27	27	27	28	29	27	28	24		27	
0°				21	23			24	22	23	27	27		32	27	24	28	
10°S		14	23	23	26			22	22	16					27	27	23	23
20°S		10	16	21	16					15	13	20					25	14
30°S				0	13					11	11	14				12	13	12

3.3 Annual average precipitation (mm)

Latitude	Longitude																	
	105°W	90°W	75°W	60°W	45°W	30°W	15°W	0°	15°E	30°E	45°E	60°E	75°E	90°E	105°E	120°E	135°E	150°E
30°N	500	1600					50	50	50	100	60	200	1500	2000	2200			
20°N	2000	2500	1200				400	200	200	100	70	400	1000	2500		3000		
10°N			2000	2500			2000	1700	1500	600	100		1000	4000	3000	3500		
0°		1000	2700	2000	700			1000	1500	700				4000	2000	2800	3000	
10°S		300	2200	2000	1000				600	700	1500					400	500	
20°S			600	1500					300	600	1700				300	150	160	
30°S																		

3.4 Solar radiation on a horizontal surface (°C)

$$I_g = I_0 (0.18 + 0.62 h/H)$$

Global solar radiation I_g (also denoted R_g) is given in cal/cm²/day, with 1 cal = 4.1855 J, and 1 cal/cm²/day = 0.011625 kWh/m²/day; h is the measured number of hours of sun per day, H the astronomical length of the day (astronomical day), and I_0 the maximum theoretical radiation (cal/cm²/day), also known as I_{ga} or PSH (peak sun hours)

Values of H , length of the astronomical day (h)

Latitude North	30°	40°	50°	60°
January	10.45	9.71	8.58	6.78
February	11.09	10.64	10.07	9.11
March	12.00	11.96	11.90	11.81
April	12.90	13.26	13.77	14.61
May	13.71	14.39	15.46	17.18
June	14.07	14.96	16.33	18.73
July	13.85	14.68	15.86	17.97
August	13.21	13.72	14.49	15.58
September	12.36	12.46	12.63	12.89
October	11.45	11.15	10.77	10.14
November	10.67	10.00	9.08	7.58
December	10.23	9.39	8.15	6.30

Values of I_{ga} (kWh/m²/day) on a horizontal surface

Latitude North	30°	40°	50°	60°
January	5.9	4.2	2.6	1.0
February	7.3	5.8	4.2	2.5
March	8.9	7.8	6.5	5.0
April	10.2	9.7	8.8	7.9
May	11.0	11.0	10.7	10.2
June	11.3	11.5	11.4	11.3
July	11.1	11.1	10.9	10.6
August	10.4	10.0	9.3	8.5
September	9.2	8.3	7.1	5.7
October	7.6	6.2	4.7	3.0
November	6.1	4.5	2.9	1.3
December	5.5	3.8	2.1	0.6