10  Energy budget at the earth’s surface

10.1  Introduction

The energy exchange at the earth's surface occupies an especially important place in the climate system. About 60% of the total absorption of solar radiation by the atmosphere/earth's surface system takes place at the surface. The absorbed solar radiation is converted into enthalpy and is readily transferred to the atmosphere through turbulence. This process is strongly influenced by atmospheric conditions, such as temperature, humidity and wind speed on the one hand and by the roughness of the surface on the other. Further, heat exchange takes place between the surface and the subsurface. Snow or ice if present may consume heat for the melt. Finally, the surface loses heat through radiative emission. Since Chapter 4 deals exclusively with radiation of the entire atmosphere and the earth’s surface, the present chapter deals with non-radiative components and the total energy budget at the earth's surface. For special cases other energy sources and sinks must be taken into account. In cities with heavy energy consumption, anthropogenic heat is a significant energy source and at high latitudes ranks above solar radiation in winter. Near the crater of a volcano, the heat supply from the subsurface can be a significant source. The solar radiation used for photosynthesis and the heat supplied or consumed by rain usually remains within 1% of global radiation at the site. Averaged over the globe all these additional sources and sinks remain well below 1% of the solar radiation at the top of the atmosphere (TOA) (c.f. Chapter 4).

A half century ago, two pioneers, Mikhail Ivanovich Budyko and Julius London completed the first realistic global energy balance studies of the earth. (56Bud) published an empirical estimate of the global energy balance, and later a global atlas with accompanying explanations (57Lon, 63Bud) evaluated the global energy balance with more physically based methods. Although their approaches were very different, Budyko and London laid the foundation of the energy balance studies for decades to come. Numerically no longer up-to-date, (65Sel, 92Pei) are still outstanding text books as an introduction to the subject of the present chapter. The developments which have taken place during the last 40 years in instrumentation, observational networks and computational capacities have brought the subject to a new higher level and it is justified to compile the present-day knowledge. For presenting the data, priority has been given to instrumentally measured energy fluxes, so that the present chapter can be used to evaluate the quality of computational works which are usually based on models and satellite-outputs.

10.2  Earth's surface and active layer

As solar radiation comes into contact with the surface, a part is reflected and the rest is absorbed and transmitted. Although referred to as the surface, the actual reflection and absorption occur not at an infinitesimally thin surface in a mathematical sense, but in a layer of finite thickness. It is in this layer that the transformation of the absorbed solar radiation into enthalpy takes place. The enthalpy raises the local temperature and evaporates water, which are both diffused into the atmosphere and the subsurface. Finally terrestrial radiation is collectively emitted from the surface of the active layer. This layer where the majority of the energy transformations take place is referred to as an active layer (74Bud). This concept is similar to the boundary layer often used by plant ecologists (81Mon). The thickness of an active layer depends on the nature, structure and condition of the material which constitutes the surface.

The active layer can be very thin, less than 1 mm in a case of smooth surfaces such as asphalt and concrete. On bare soil, more than 90% of solar radiation is absorbed within the surface roughness, that is, 1 cm as an order of magnitude. Solar radiation penetrates the surface more easily when the surface is loose such as vegetation. Although dependent on the thickness of the vegetation cover and the elevation angle of the sun, 80% of solar radiation is usually absorbed within the top 20% of the mean tree height
In the case of a short grass of 10 to 20 cm in length, 70% to 80% of solar radiation is absorbed by blades of the grass, and the rest by the soil surface. In a partially transparent material such as water, ice or snow, absorption of solar radiation can take place through a relatively thick layer. On an average 90% of solar radiation is absorbed in the first 3 m of clean water. In a clean snow cover of average conditions (density = 0.35 kg/m³), 90% of solar radiation is absorbed within the first 15 cm. The exchange of enthalpy and water vapour with the atmosphere, however, takes place in a very thin layer near the surface which is considered as the active layer rather than the entire depth of the medium in which solar radiation is absorbed.

10.3 Processes of energy transformation at the surface

There are three distinctive processes with which the energy is exchanged at the surface. They are radiation, turbulent diffusion and molecular conduction. The divergence of these fluxes is the rate at which the heat (enthalpy) storage changes in the active layer. Firstly, radiation dominates the surface energy processes in most circumstances and regions. The absorption of solar radiation sets the whole system into action. Terrestrial atmospheric radiation is, however, numerically as important as solar radiation. At the end of the chain of the entire energy exchange, the surface emits radiation in the longwave range. Secondly, there is a continuous exchange of energy and substance between the surface and the atmosphere. Because the atmosphere is always turbulent, the exchange of enthalpy (sensible heat flux) and water vapour (latent heat flux) takes place in the form of turbulent diffusion. If the energy exchange on the water surface is considered, the turbulence is also responsible for the exchange between the surface and the interior of the water body. On the land the surface and the subsurface exchange heat mainly through molecular conduction. The actual storage change through the temperature change within the active layer is very small. The only exception is the latent heat of fusion which makes up the majority of the storage change. The above consideration can be formulated in the following manner that is an expression of the law of energy conservation applied for the earth's surface (more strictly for a volume of an active layer). The sign is positive when the flux is directed towards the surface, following the convention in micrometeorology:

\[
S(1-a) + L_{\downarrow} + H + L_{e}E + G = L_{s}M + \sigma T_s^4
\]  

(1)

where

- \(S\): global solar radiation; \(a\): albedo; \(L_{\downarrow}\): terrestrial incoming radiation from the atmosphere; \(H\): sensible heat flux; \(L_{e}E\): latent heat flux of vapourisation; \(G\): subsurface heat flux; \(L_{s}M\): latent heat of fusion or melt; \(\sigma\): Stefan-Boltzmann constant; \(T_s\): surface temperature.

Since Eq. (1) describes the budget of energy for an active layer, strictly speaking the net horizontal advection and the storage term within the active layer should also be considered. This point is considered in detail by Han. The net horizontal advection term is mostly negligible owing to the very thin nature of the active layer and the horizontal homogeneity of temperature and moisture fields. The storage term is likewise very small due to the small heat capacity of the layer. The storage term, however, is not negligible when it takes the form of the latent heat of fusion during the melt or the freeze period, which is expressed as the first term on the right hand side of the equation.

This equation expresses not only the balance of incoming and outgoing energy fluxes, but also a cause and effect relationship between the heat fluxes, storage and the surface temperature. Among the seven terms in Eq. (1), the five terms on the left hand side and the two terms on the right hand side have fundamentally different attributes. The five terms on the left hand side can be determined independently of each other, although they are to some extent mutually related. For example, solar radiation \(S\) is determined independently of the other terms. The solar constant, the sun/earth distance and the scattering and absorbing characteristics of the atmosphere and the earth's surface will determine \(S\). All these factors are quite independent of the other terms in Eq. (1). Another example can be taken from the sensible heat flux \(H\). It is clearly related to solar radiation \(S\) and evaporation \(E\), but also strongly affected by wind speed and the air temperature in the advecting atmosphere both of which are not part of Eq. (1).
Therefore, \( H \) is also partially influenced by external factors. On the contrary, the terms expressing the heat of the melt \( L_M \) and the terrestrial outgoing radiation, \( \sigma T^4 \) on the right hand side are totally passive and determined entirely by the other five terms on the left hand side. Although grouped together on one side, these two terms are self-determinable as they are never simultaneously variables. For example, if the fusion term is a variable, that is, melting or freezing is in progress, the terrestrial outgoing radiation is a constant at the melting point. If the fusion term is a constant and non-zero, that is, melting or freezing is in progress at a constant rate, the terrestrial outgoing radiation must be again a constant at the melting point. If the fusion term is constant and zero, that is, the melt is not in progress, the terrestrial outgoing radiation is a variable. The equation expresses how the heat exchange at the surface determines the surface temperature. It also expresses how the surface restores its new heat balance by adjusting the surface temperature when a perturbation is introduced in the fluxes on the left hand side.

The second term on the right hand side is expressed as if it were a black body emission, although the earth's surface is never black. This is not an attempt to approximate the surface emission with the black assumption. This term represents not only the surface emission but also the reflected portion of the terrestrial incoming radiation. The black body approximation for the terrestrial outgoing radiation as the sum of the emission and reflection usually gives an error well below 1%.

Eq. (1), the equation of the energy balance for the surface layer is written in this manner, thus avoiding conventionally used expressions. The concept of the net radiation is also not used for the above reasons. Net radiation, however, is a technically useful concept, as it represents the total energy exchanged between the electromagnetic wave and enthalpy. Net radiation can be measured on its own by an inexpensive instrument called a net radiometer. When all irradiances are individually measured, the concept of net radiation helps to assess the accuracy of each component by examining how closely the sum of the individual terms comes to the measured net radiation. Therefore, net radiation is also used in this work both in text and tables. The use of net radiation is pragmatic and there is a profound reason to treat terrestrial outgoing as a separate and distinctly different quantity.

### 10.4 Non-radiative energy fluxes

The present section concentrates on the non-radiative components and the total energy budget, as radiative fluxes are already dealt with in Chapter 4. Two important non-radiative fluxes are sensible and latent heat fluxes. As enthalpy (content of the sensible heat flux) and water vapour (content of latent heat flux) are carried from and to the earth's surface by turbulence in the lowest layer of the atmosphere, a brief description is given on this subject. The characteristics of this layer come into existence by friction originally generated by the moving atmosphere over continents and oceans. The friction is an important mechanism to generate turbulence. Another cause for turbulence is the heterogeneity of the density which is mainly the result of the heat exchange between the surface and the atmosphere. This turbulence-dominated layer with significant friction above the surface is called the atmospheric boundary layer (ABL). This layer is usually several hundred meters to several kilometres thick, depending on wind speed, surface roughness and the effect of buoyancy force.

The thinnest ABL is observed in the central regions of Greenland and Antarctica where it can become only 200 to 300 m thick. The thickest boundary layer develops over the roughest surface of the earth, namely major mountain ranges such as the Himalayas and the Alps. Over the Alps the ABL can develop to 4 kilometre thickness above the mean altitude of the mountain ranges. The ABL undergoes substantial changes in a course of the day. The bottom 10% or so of the ABL is usually dominated by small-scale turbulence generated by the surface friction. Within this layer the vertical turbulent fluxes are fairly constant. As we go higher the movement of air tends to be influenced more by density stratification and fluxes usually diminish with increasing height. The lowest layer of the ABL directly influences the exchanges of heat, gases and mechanical characteristics of the atmosphere with the earth's surface. This layer is also best investigated and is the main topic of the following sections.
10.4.1 Turbulent fluxes of sensible and latent heat

The majority of the enthalpy converted from radiation at the surface is transferred into the atmosphere. Since the atmosphere is in a constant state of turbulence, heat and water vapour are transported aloft through turbulence. A brief description of the theory of turbulent transport is presented. This summary is not a reproduction of general theory on turbulence but a description aimed at the application of the theories for determining turbulent fluxes at the earth’s surface. Readers with interest for further details of the subject are recommended to refer to specialised literature (e.g.: 59Pri, 64Lum, 73Hau, 82Bru, 92Gar).

The rate at which the concentration \( c \) is transported over time by turbulence in the vertical direction through a horizontal surface of a unit area is defined as

\[
F = \overline{cw} 
\]

where the prime denotes a deviation of the variable from the mean, and the bar a time-mean. \( w \) is the vertical velocity component. In this section the convention of signs follows that of micrometeorology, with positive taken in the direction towards the surface (i.e., downward positive above the earth’s surface and upward positive below the earth’s surface). The right hand side is the covariance of \( c \) and \( w \). In the case of the enthalpy transport which is referred to as the sensible heat flux, \( c = \rho c_p T \); for the latent heat flux, \( c = \rho L_v q \), where \( \rho \) is the density of air, \( c_p \) is the specific heat of air under constant pressure, \( T \) is the air temperature in K, \( L_v \) is the latent heat of vaporisation of water and \( q \) is the specific humidity. For the momentum flow, \( c = \rho V \) where \( V \) is the wind velocity.

The sensible and latent heat fluxes have the following forms, respectively:

\[
H = -\rho c_p T \overline{w} \\
L_v E = -\rho L_v q \overline{w}
\]

To use these formulations for the flux measurement at a practical height of 1.5 to 2 m, the required frequency of the measurement of \( w, T \) and \( q \) is in the order of 10 Hz. The 1 Hz frequency should be regarded as the minimum. This method is called an eddy correlation method (more appropriately, eddy covariance method), and has become practical only after the development of sonic anemo-thermometers and fast-response hygrometers. This method offers the most direct measurement of turbulent fluxes. The practical problems are not small, however.

Firstly, it is not possible to install the instrument in a perfectly vertical position, so that \( \overline{w} = 0 \) would always be kept. The visual or even geodesic setting of the vertical direction may fail to give \( \overline{w} = 0 \), because the ground surface may have a tilt. Geodetically and aerodynamically horizontal surfaces may not be parallel. To obtain a normal orientation with respect to an aerodynamically horizontal surface some computational adjustment is always necessary. This is done by a co-ordinate transformation between the orientation of the actual installation of the instrument and the designated orientation which gives \( \overline{w} = 0 \). To find out the ideal orientation of the co-ordinate system, one needs \textit{a priori} a 3-D sonic anemo-thermometer, although the appearance of Eq. (2) gives an impression as if a 1-D anemo-thermometer might suffice the purpose.

Secondly, the volume occupied by the instrument already distorts the flow-field. Additionally, a space of stagnating air builds up on the lee- and luff-sides of the probe, which causes systematic under-estimations. Therefore, all instruments of this type must be calibrated in an accurate wind tunnel. This problem is often underestimated, as the instrument is assumed to be absolute.

Thirdly, the condensation and accretion of hydrometeors, such as raindrops, dew, frost and super-cooled cloud droplets on the surfaces of sensors cause serious errors, as the acoustic speed in water and ice is one order of magnitude higher that that in the atmosphere. Because of these difficulties, substantial experience is compulsory. Even with the best available skill, the weather-caused disturbances limit the continuous use of this method for a sustained period. The method becomes truly powerful when it is backed up with more robust methods and instruments which are explained in the following sections.

Fourthly, the eddy correlation method is not applicable if the concentration \( c \) is not sampled quickly at the same frequency as wind \( w \). This problem arises for a number of flux measurements of chemical material, particularly for volatile organic compounds. Eddy accumulation method is used for such cases. This relatively new technique is presently well established, and practical aspects of this method can be found in 00Rin, 02Rin, 02Amm.
Partly because of the technical difficulties in using the eddy correlation method in earlier days, but also because of the desire to understand the turbulent transport in terms of the mean fields, there is about one hundred years of effort to express these turbulent fluxes as functions of the mean values of the involved variables. This development has intensified recently, as the boundary layer in atmospheric models must be equipped with turbulent fluxes which are expressed as functions of mean fields.

The effort to express turbulent covariance as a function of mean values started with shear stress or momentum flux which is friction when realised on a wall, such as the earth's surface. A turbulent exchange coefficient or eddy diffusivity was introduced analogous to the kinematic viscosity of a laminar flow. The shear stress in a laminar flow and turbulence is expressed as below:

\[ \tau = \rho \kappa \frac{du}{dz} \]

Subscripts \( l \) and \( t \) denote laminar and turbulent, respectively; \( \kappa \) and \( K \) are kinematic viscosity and eddy diffusivity, respectively; \( u \) is the horizontal wind speed taken in the main wind direction; and \( z \) is the height. The horizontal bar means again a time mean value. As a desirable length for the time mean computation, the period should be long enough to capture effective turbulent eddies, but should not be so long that the variations other than turbulence such as the diurnal change or synoptic changes might set in.

This type of approximation is called a Boussinesq approximation, but also variously referred to such as a gradient approximation or \( K \)-method. This approximation became useful only when the numerical nature of \( K \) was clarified, after the introduction of the Prandtl's mixing length theory and von Kármán's quantification of the mixing length, namely,

\[ \tau = \rho l \left( \frac{d\bar{u}}{dz} \frac{d\bar{u}'}{dz'} \right), \quad l = k \frac{d^2 \bar{u}}{dz^2} \]

where \( l \) is the mixing length and \( k \) is the von Kármán constant (about 0.4). Upon integration of the above equations with the assumption of flux constancy, one obtains the classic logarithmic wind profile:

\[ \frac{u_2 - u_1}{u_1} = \frac{\sqrt{\tau_1 / \rho}}{k \ln \frac{z_2}{z_1}} \]

where subscripts 1 and 2 refer to lower and higher levels, respectively. \( \sqrt{\tau_1 / \rho} \) has the dimension of speed and contains the information of shear stress, and often called a friction velocity \( u_c \). This relationship is the same as

\[ \tau = \rho k^2 \left( \frac{u_2 - \bar{u}}{u_1} \right)^2 \]

This type of approximation of a flux with the observed mean difference of necessary quantities at two levels is called aerodynamic method (also referred to as the flux/gradient or profile method) and will be used also for other fluxes, such a sensible and latent heat. The integration of the same equation from the level very near the surface \( z = z_0 \) where \( \bar{u} = 0 \) can be assumed is often used but not recommended, as the value of the roughness length \( z_0 \) varies over a wide range for a similar surface. This approximation which is often referred to as the bulk method is, nevertheless, often used in field experiments and atmospheric models as it suffices to measure values only at one level. The accuracy is, however, very poor. The aerodynamic method demands the measurement of wind speed at two levels theoretically and practically at least at four levels.
10.4.1.1 Sensible heat flux

The analogue to the above discussion for the enthalpy transport yields the following equation:

$$H = \rho_c K_h \frac{dT}{dz}. \quad (8)$$

$K_h$ is the eddy diffusivity for heat. Strictly speaking, potential temperature should be used instead of actual temperature $T$. However, in practice for the height range of several meters used often for the measurement, measured air temperature suffices. Upon integration of Eq. (8) very much like the earlier treatment with Eq. (5), one obtains the following equation:

$$H = \rho_c \frac{\left(\overline{T_f} - \overline{T_i}\right)(\overline{u_z} - \overline{u_i})}{\ln z_i / z_f}. \quad (9)$$

This type of approximation is often called the Sverdrup-equation (36Sve) and yields a relatively accurate result under near neutral conditions, that is, when the wind speed is high and the temperature gradient is small. When the wind speed is low and the temperature gradient is strong, the effect of the buoyancy force starts to play a role either to promote or damp the vertical mixing. For such non-neutral conditions the Monin-Obukhov (54Mon) parameterisation can be used. The Monin-Obukhov function is defined for the momentum and sensible heat fluxes in the following manner:

$$\frac{kz}{u^*} \frac{du}{dz} = \phi_m(z/L), \quad \frac{kz}{T^*} \frac{dT}{dz} = \phi_h(z/L), \quad (10)$$

where $T^*$ is a scaling temperature for a turbulent heat flow defined as $T_u = H / \rho c_p$; $\phi_m$ and $\phi_h$ are dimensionless wind-shear and temperature gradient, respectively called the Monin-Obukhov functions for momentum and heat. The Monin-Obukhov function is assumed to be a function of a dimensionless height $z/L$ which was introduced earlier by Obukhov (46Obu). The Obukhov length $L$ is the height above the surface at which the production rate of the turbulent kinetic energy through wind shear will be exactly the same as the consumption rate of the turbulent kinetic energy through the work against the gravity force. Obukhov obviously considered initially the case for a stable condition, but it is easily seen that the same consideration can be extended to an unstable stratification. In this case $L$ becomes the height at which the production rates of turbulent kinetic energy through shear and buoyancy force will be exactly the same and the sign is negative. From the above definition the Obukhov length takes the following form:

$$L = \frac{u^* T}{kg (H / \rho c_p)}. \quad (11)$$

Here, $k$ is the von Kármán constant as before and $g$ is earth’s acceleration of gravity. The integration of Eq. (10) for the scaled height $z/L$ gives the following flux representation:

$$H = \rho_c k^2 \overline{T_f} - \overline{T_i})(\overline{u_z} - \overline{u_i}) \left[ \ln(z_i / z_f) - [\Psi_m(z_i / L) - \Psi_m(z_f / L)] \right] \left[ \ln(z_i / z_f) - [\Psi_m(z_i / L) - \Psi_m(z_f / L)] \right] \quad (12)$$

where $\Psi_h$ and $\Psi_m$ are the integrated Monin-Obukhov functions for sensible heat and momentum, respectively, defined as

$$\Psi(z/L) = \int_{z/L}^{1 - \phi_h(z/L)} \frac{1 - \phi_h(z/L)}{z/L} \, d(z/L) \quad (13)$$
During the last 40 years various forms of the Monin-Obukhov functions have been proposed. Some of them are illustrated in Fig. 10.1. The mathematically corrected description of this subject can be found in Paulson (70Pau). Recent investigations with accurate shapes of \( \Psi \) are presented by Beljaars and Holtslag (91Bel) and Högström (96Hög). This type of description is called a flux/gradient relationship and is widely used to estimate the turbulent fluxes based on the gradient measurement.

Although Eq. (12) gives an impression that the measurements at two levels might suffice the purpose, the temperature and wind speed should be measured at least at four levels. By using this method, the measurement of the wind shear usually gives the largest error so that measurements at least at four levels are necessary to keep the error to a minimum.

The Monin-Obukhov theory described above can not be used directly in the field for the purpose of determining fluxes, as the Obukhov length \( L \) involves at least two (momentum and sensible heat) fluxes which are not yet known. If the buoyancy force due to water vapour is considered, \( L \) is a function of three fluxes. One way to solve this dilemma is to iteratively approach the likely solution by starting at the neutral condition, that is \( \phi_m = \phi_h = 1 \). This method is most frequently used but not recommended. The iterative method requires firstly, the proof of convergence of the solution and secondly, the convergence around a true solution. Usually there is no problem to find a convergence, but the solution in this problem tends to converge around near neutral conditions. The other method is to approximate the Obukhov length ratio \( z/L \) with a gradient Richardson number \( R_i \) which can be evaluated based on the time-mean variables. From the definitions of these stability criteria the following relationships can be derived:

\[
R_f = \frac{\phi_m}{\phi_h} R_i = \phi_m^{-1} \frac{z}{L},
\]

where \( R_f \) is the flux Richardson Number. This relationship enables one to re-write the Monin-Obukhov functions originally written in terms of \( z/L \), with respect to the gradient Richardson number. This can be done if one possesses the shape of the Monin-Obukhov function for momentum \( \phi_m \), and the turbulent Prandtl number \( K_h/K_m \) or \( \phi_h/\phi_m \). For the functions presented in Fig. 10.1, this conversion was possible. For evaluating the gradient Richardson number, the two levels should be chosen from the two lowest atmospheric levels. In other words, the choice of the lower level at the earth's surface should be avoided. Although this one level approximation (bulk method) appears attractive, the result reduces \( R_i \) usually to a falsified neutral state. This error is in fact the largest component of the errors in the surface fluxes produced in most climate models. If one allows this error to happen, there is no point in elaborately attempting to identify the best Monin-Obukhov function or improving the shapes of the Monin-Obukhov functions, as the actual computation is effectively done using the Sverdrup neutral equations.
10.4.1.2 Latent heat flux

For the measurement of latent heat flux in the field, the direct observation with the eddy correlation method described in Section 10.4.1 can also be used. The technical difficulty is two-fold, however. Firstly, one needs a fast-responding hygrometer, comparable to the sonic anemometer. The recently developed Krypton hygrometer most closely fulfils this requirement. Some instruments of this type show a drift of the calibration within several weeks. It is, therefore, indispensable to re-calibrate the Krypton hygrometers against the standard dew-point hygrometer. Secondly, it is not possible to measure the same volume of the atmosphere for the velocity and the humidity. Therefore, one is forced to measure the velocity and humidity over a certain distance, whereby losing a correlation.

Similar to the measurement of the sensible heat flux, the eddy correlation method becomes more powerful when some gradient approximations are used in parallel. The Boussinesq approximation for latent heat can be written in the following manner:

\[ L_E = \rho L_w K_w \frac{\bar{q} - \bar{q}_0}{\bar{u}_2 - \bar{u}_1} \]

(14)

where \( q \) is specific humidity and \( K_w \) is the turbulent exchange coefficient for water vapour. This equation can be integrated into the following Sverdrup equation for latent heat flux:

\[ L_E = \frac{\rho L_w K_w \left( \bar{q}_2 - \bar{q}_0 \right) \left( \bar{u}_2 - \bar{u}_1 \right)}{\left( \ln \frac{z_2}{z_1} \right)^2} \]

(15)

This equation is practically useful for neutral to near neutral conditions. For the range outside the near neutral conditions, the Monin-Obukhov parameterisation in the following form offers a good approximation:

\[ L_E = \frac{\rho L_w K_w \left( \bar{q}_2 - \bar{q}_0 \right) \left( \bar{u}_2 - \bar{u}_1 \right)}{\left[ \ln \left( \frac{z_2}{z_1} \right) - \Psi \left( \frac{z_2}{L} - \Psi \left( \frac{z_1}{L} \right) \right) \right]} \]

where \( \Psi \) is the integrated Monin-Obukhov function for water vapour. The shape of the Monin-Obukhov function for water vapour for the near-neutral and unstable region is similar to that of sensible heat. The shape of this function for the stable range is uncertain for water vapour and can be the main reason for overestimating evaporation in winter in many GCMs.

Especially for the problems related to the plant canopy layer, the authors recommend interested readers following articles: 81Rau, 96Rau and 00Fin.

The measurement of the latent heat flux so far was conceived based on information in the atmosphere. The latent heat flux for land surfaces can, however, be measured also with non-atmospheric methods. One such method is based on the hydrological balance of an isolated hydrological drainage basin, and is suited for long-term observation at permanent stations. Lysimeters are an example of an artificially created hydrological basin, although the surface areas are very small. There are two types of lysimeters. One is a weighing lysimeter illustrated in Fig. 10.2. A weighing lysimeter is designed to measure short-term evapotranspiration by detecting a small loss of the water mass from a soil column contained in a vessel. As the entire apparatus must be weighed, the vessel can not be too heavy, several tons at most. This severely limits the surface area and the depth of the soil column. The main disadvantages of this instrument come mostly from this confinement of the soil column in a geometrically small volume.

Especially difficult is the simulation of the aerodynamic conditions of the surface and the moisture content of the soil in the vessel to the prevailing conditions in the surrounding ground. The surface can be planted with light weight vegetation like short grass. To avoid an unnatural over-saturation, a seepage mechanism is usually made at the bottom of most weighing lysimeters. The drainage of water happens at the bottom of the vessel under the atmospheric pressure instead of the suction force at a comparable depth in the surrounding natural soil. When a dry period persists and the upper soil column is dried, the contraction of the soil column creates a gap between the soil and the wall of the vessel, enlarging an
effectively evaporating surface. Precipitation during the observation adds mass which must be subtracted. The precipitation measurement to accompany a lysimeter must be done with the comparable resolution and accuracy to the lysimeter. One needs a precipitation gauge with a larger than usual receptacle and its orifice must be adjusted to the ground surface. All these details must be taken into account for the observation and the interpretation of the data.

A special kind of weighing lysimeter is a snow lysimeter to measure sublimation or evaporation from the snow cover. This is nothing other than an artificially cut surface layer of a snow cover contained in a transparent receptacle, buried back into the snow cover in such a manner that the surface of the snow in the container is adjusted to that of the surrounding area. As the continuous monitoring of the weight of the container under such circumstances is difficult, the weight is measured on a separate balance by taking the vessel out of the hole at a regular time interval. Depending on the intensity of the sublimation and evaporation, the time interval can be chosen between 6 and 24 hours. The measurement of precipitation for this purpose must be made by a gauge whose orifice is also adjusted to the level of the snow surface. Usually another container of the same geometrical dimensions buried in a snow cover suffices the purpose. This method with modest investment provides relatively accurate snow evaporation for time intervals of synoptic to daily totals. The accuracy depends greatly on skill, however. It is especially important to cut the snow-monolith usually with a knife with minimum disruption of the snow surface. The method usually does not work in drifting and blowing snow.

The other type is a run-off lysimeter. A run-off lysimeter relies basically on the following principle of surface hydrological balance, that the difference between the precipitation and the run-off on a long term basis becomes very close to evapotranspiration. This type of instrument is basically the confinement of a soil column in a vessel which has a run-off mechanism usually near the bottom. This instrument can be made very large both in surface area and depth, as it is not necessary to weigh the mass. One might imagine a swimming pool filled with soil. The surface can be planted with large and heavy plants such as trees. To determine the storage change, there is an additional network of sensors to detect the soil moisture. However, an accurate measurement of the representative change in storage for the entire vessel is difficult, and this method is not suited for short-term measurement. It is safe to assume that this type of instrument is conceived only for the annual total evaporation. All lysimeters are meant to be used in areas where precipitation exceeds evaporation.

A natural hydrological basin located in a geological region with impermeable rocks offers an excellent opportunity to determine evapotranspiration under natural conditions. The basin should be equipped with a reasonably dense network for measuring precipitation, soil moisture and snow cover. At the outlet of the basin, precision discharge measurements must be carried out on a continuous basis.

10.4.1.3 Bowen ratio

The Bowen ratio was first introduced by Bowen (26Bow) to partition two turbulent fluxes, sensible and latent heat. A method to obtain sensible and latent heat fluxes by combining the Bowen ratio and the energy balance equation in Eq. (1) is normally referred to as the Bowen ratio-energy balance method. The
Attractive points of this method are its simple principle, economical instrumentation and relatively high accuracy. This method does not require the use of anemometers, an important advantage, as the most serious error of the aerodynamic method is induced through the measurement of wind shear. The Bowen ratio $\beta$ is defined as the ratio of sensible to latent heat fluxes,

$$\beta = \frac{H}{L_E}$$  \hspace{1cm} (17)

By combining Eq. (17) with Eqs. (8) and (14), the Boussinesq approximations of sensible and latent heat fluxes, and further replacing the differential terms with differences, one obtains,

$$\beta = \frac{c_p(T - T_i)}{L_e(q - q_i)}.$$  \hspace{1cm} (18)

The only assumption made for this expression is the similarity of the eddy diffusivities for heat and water vapour, that is

$$\frac{K_h}{K_w} = 1.$$  

Despite the intricate differences among eddy diffusivities, this assumption is not a bad one for most practical purposes.

The Bowen ratio-energy balance method finds more practical importance, when a profile measurement with a tower is not available. For an earth’s surface with an unlimited water supply for which the saturation vapour pressure at the surface can be assumed, the following equation for the Bowen ratio is derived (84Ohm):

$$\beta = \frac{H}{L_E} = \frac{c_p \Delta T}{L_e \Delta q} = \frac{c_p \bar{P}}{L_e \left( \frac{\Delta e_s}{dT} \right)}$$  

where $P$ is the atmospheric pressure, $\epsilon$ is the ratio of the molecular mass of water to that of dry air, 0.622, $e_s$ is the saturation vapour pressure at temperature $T$. $\Delta$ denotes the difference of variables at two levels. This relationship can be used to estimate the Bowen ratio only with the temperature $T$ at the site when micrometeorological observations are not available. This Bowen ratio is close to that for potential evaporation under minimum advection of dry air.

Upon substituting this relation into the energy balance equation, one obtains,

$$L_e E = -\frac{R + G - L_f M}{1 + \beta}$$  

and,

$$H = -\frac{R + G - L_f M}{1 + 1/\beta}.$$  \hspace{1cm} (19)

Unless extremely accurate instrumentation for the eddy correlation method and experienced personnel are constantly available at the site, the Bowen ratio energy balance method summarised above gives the best results, especially for long-term observations. The method, however, becomes impossible to apply when $\beta$ becomes $-1$. In reality, however, the method becomes often impractical when the Bowen ratio turns negative. This condition is frequently observed just after the sunrise and near the sunset. The loss of the observations during these hours may not be serious, as the absolute magnitudes of the fluxes are very small. The negative Bowen ratio can happen, however, under large fluxes. This phenomenon is observed, for example, under intense foehn when the downward sensible heat flux from warmer air evaporates water from the surface. A similar condition is observed over the Mediterranean when Sirocco from the
Sahara blows over the sea. The Bowen ration is also usually negative over melting snow cover and ice surfaces. The applicable range of the data for the Bowen ratio energy balance method is analytically considered by Ohmura (83Ohm).

The Bowen ratio energy balance method is adapted in various ways. One of the best known adaptations is the Penman/Monteith equation which is the result of an effort to mobilise Eq. (19) with only one-level measurements (48Pen; 95Mon). For further reading the authors recommend 79Mon, 81Gra, 82Bru and 91Sch.

### 10.4.1.4 Spatial distribution of the turbulent heat fluxes

In the present section geographical distributions of sensible and latent heat fluxes will be discussed. The annual mean distributions of these fluxes are presented in Fig. 10.3.

For the global mean, latent heat flux of vapourisation with 85 Wm$^{-2}$ accounts for majority (82%) of net radiation at the surface. Sensible heat flux takes (19 Wm$^{-2}$) only 18% of surface net radiation. They correspond to 25% and 6% of the primary solar radiation at the TOA, respectively. It is noteworthy that the two turbulent heat fluxes being carried by similar eddies, and being propelled by the same energy source, namely the surface net radiation, differ as much as by factor 4. The main reason is found not on the surface but in the middle and higher troposphere where water vapour is removed by cloud and precipitation formation, and the same condensation produces a corresponding amount of sensible heat. This wet convection process is responsible for creating two fundamentally different vertical gradients in the atmosphere, the sharp decrease in water vapour concentration and the increase in enthalpy with increasing altitude.

This situation provides also the basic reason for different geographical distributions of latent and sensible heat fluxes. While latent heat flux is directed upward almost everywhere on the earth's surface, sensible heat flux is often directed downward. Henceforth, in a large area in high latitudes, pole-ward of 70º N and S sensible heat flux is directed to the surface. Sensible heat flux can easily flip downward even in an area with positive net radiation. In lower latitudes when an arid continent is located windward, sensible heat flux turns downward as is witnessed over the Arabian Sea. The lower SST on the equator suffices to depress the magnitude of evaporation but draws sensible heat flux downward in many equatorial regions.

The ocean evaporates 88% of the global evaporation. Further the surface of the ocean within 25º N and S of the equator supplies 50% of global evaporation. The land surface contributes only 12%. Per unit area the latent heat flux on the ocean (110 Wm$^{-2}$) is far larger than over the land where it stands at 36 Wm$^{-2}$. This difference is due to the limit of surface water available on the land. Hence the land surface generates sensible heat more easily than on the oceans: 26 Wm$^{-2}$ on land and 15 Wm$^{-2}$ on oceans. Taking the surface areas into account 40% of the global sensible heat is transferred into the atmosphere from the land, and the remaining 60% from the ocean.

Seasonally analysed, however, the largest sensible heat flux is seen on the ocean in winter and on the desert in summer. In winter monsoon regions where cold air flows from the continents on to the surface of the warm western boundary currents, some of the largest sensible heat fluxes can occur. In the regions off New England and Japan, sensible heat flux in January can reach more than 100 Wm$^{-2}$. The outbursts of the winter monsoon are also dry, and the same regions also see the largest latent heat flux of evaporation in the world (250 Wm$^{-2}$). Both these large fluxes are supported by strong convection in the ocean mixing layer which delivers as much as 400 Wm$^{-2}$ ocean surface heat flux.

Oceans are in general actively evaporating regions in winter. Although often overlooked, the hemispheric total evaporation is larger in winter than in summer in both hemispheres (02Ohm). This situation shows that temperature alone does not determine large-scale evaporation. The strength of the atmospheric circulation is equally influential in determining evaporation.
Fig. 10.3a Annual mean distribution of **sensible heat fluxes**, based on data of the re-analysis project ERA-40. This data set covers the period 1991-1995. Note the downward fluxes of sensible heat over larger ice fields and some sections of the Southern Oceans.
Fig. 10.3b  Annual mean distribution of latent heat fluxes, based on data of the re-analysis project ERA-40. This data set covers the period 1991-1995.
As winter is the active season with evaporation on oceans, summer can be regarded as the active season for generating sensible heat on land. The year's peak of sensible heat flux on land is usually observed in early summer (May and June in NH, October and November in SH). During these months the sensible heat flux in arid regions of both hemispheres reaches 50 to 100 Wm\(^{-2}\), and becomes a heat sink in a magnitude similar to terrestrial net radiation. Annual mean energy fluxes are graphically represented in Fig. 10.3.

### 10.4.2 Subsurface heat flux

Since the subsurface heat flux has fundamentally different mechanisms on land and on a water body, both will be discussed separately. First, the main characteristics of the ground (soil) heat flux will be discussed. Ground heat flux is normally one order of magnitude smaller than other fluxes and plays a limited role in surface energy balance. The main reason for this is the relatively small thermal diffusivity of the ground material in comparison with the eddy diffusivity of the near surface atmosphere. The thermal diffusivity of common soils and rocks is 10\(^{-4}\) to 10\(^{-6}\), when usually observed values for the eddy diffusivity in the first 10 m of the atmosphere is taken as unity. Nevertheless, this heat flux determines the thermal conditions of the near surface layers of the pedosphere and lithosphere.

Ground heat flux turns downward in spring and reaches the year's maximum flow in early summer, usually one month ahead of the summer solstice. The flux declines throughout the summer and turns upward in autumn until the following spring. At the year's maximum it attains about 10% of the net radiation in most climatic zones. The phase and magnitude of the ground heat flux can seriously be affected by the snow cover, the thickness, and the timings of its onset and melt. The annual total ground heat flux observed for more than several years is very close to zero or becomes the same as the geothermal heat flux.

The ground heat flux is normally measured using a combination of the heat flux plate installed in the ground near the surface and the ground temperature profile. The temperature profile should be measured from the surface down to the depth where the annual fluctuation becomes insignificant, that is usually at least 15 m. The main thermal characteristics of various soils are summarised in Table 10.1. Further details are available in the specialised literature on this subject (e.g. 81Far).

#### Table 10.1 Thermal properties of soil constituents and soils (adapted with alterations from 81Far)

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Density ([\text{kg m}^{-3}])</th>
<th>Specific heat ([\text{J kg}^{-1} \text{K}^{-1}])</th>
<th>Heat capacity ([\text{J m}^{-3} \text{K}^{-1}])</th>
<th>Heat conductivity ([\text{W m}^{-1} \text{K}^{-1}])</th>
<th>Thermal diffusivity ([\text{m}^2 \text{s}^{-1}])</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>(2.65 \times 10^3)</td>
<td>(0.732 \times 10^3)</td>
<td>(1.94 \times 10^6)</td>
<td>8.4</td>
<td>(4.33 \times 10^{-4})</td>
</tr>
<tr>
<td>Minerals</td>
<td>(2.8 \times 10^3)</td>
<td>(0.8 \times 10^3)</td>
<td>(2.2 \times 10^6)</td>
<td>6</td>
<td>(2.7 \times 10^{-6})</td>
</tr>
<tr>
<td>Organic matter</td>
<td>(1.3 \times 10^3)</td>
<td>(1.93 \times 10^3)</td>
<td>(2.5 \times 10^6)</td>
<td>0.25</td>
<td>(0.1 \times 10^{-4})</td>
</tr>
<tr>
<td>Water</td>
<td>(1.00 \times 10^3)</td>
<td>(4.1855 \times 10^3)</td>
<td>(4.1855 \times 10^6)</td>
<td>0.6</td>
<td>(0.14 \times 10^{-4})</td>
</tr>
<tr>
<td>Ice</td>
<td>(0.917 \times 10^3)</td>
<td>(2.093 \times 10^3)</td>
<td>(1.919 \times 10^6)</td>
<td>2.2</td>
<td>(1.15 \times 10^{-4})</td>
</tr>
<tr>
<td>Air</td>
<td>1.2</td>
<td>(1.005 \times 10^3)</td>
<td>(1.206 \times 10^6)</td>
<td>0.026</td>
<td>(2.16 \times 10^{-8})</td>
</tr>
</tbody>
</table>

II. Heat conductivity of soils \([\text{W m}^{-1} \text{K}^{-1}]\) (Heat conductivity of soil strongly depends on water content and weakly on temperature. Number in brackets are water contents in volume per cent)

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Conductivity ([\text{W m}^{-1} \text{K}^{-1}])</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay</td>
<td>0.28 (1.8)</td>
</tr>
<tr>
<td>Silt</td>
<td>0.38 (3)</td>
</tr>
<tr>
<td>Sand</td>
<td>0.34 (0.01)</td>
</tr>
</tbody>
</table>
The subsurface heat on the ocean surface is the ocean surface heat flux or net downward heat flux used in oceanography. This is a substantial component in the diurnal and annual energy budget as well as in long-term climate changes. The importance of the ocean in the climate system is mainly due to its large heat storing function. This function is not only due to the large heat capacity of water, but to a great extent to the transparency of the ocean water to solar radiation, and to the oceanic turbulence and convection which promote heat diffusion. The annual amplitude of the ocean surface heat flux reaches as high as 200 W m\(^{-2}\) on the mid-latitude ocean. In the course of a year the ocean flips between a heat source and a sink on the hemispheric scale.

Globally seen there are three belts of heat sinks for the surface energy balance (sources for the ocean), one on the equator and the two others in the higher mid-latitudes around 50° N and S where a substantial amount of heat flows downward. Of these three belts, the region under the equator is a permanent energy sink. In the equatorial ocean, large shortwave incoming radiation, supported by relatively weak terrestrial net radiation (due to cloud amount), and smaller turbulent heat losses (due to higher humidity and lower SST), forms the basis of the permanent energy sink for the atmosphere (source for the ocean). The other two belts in the higher mid-latitudes are sinks on an annual basis, but are in a delicate balance between the sink in summer and the source in winter. Unlike ground heat flux, the annual total ocean surface heat flux assumes large values. The eastern half of the higher mid-latitude ocean surfaces where the temperature is lower is especially important as a sink region. Further in higher latitudes, in the Arctic Ocean (Central Polar Ocean including marginal oceans, such as Norwegian, Barents, Kara, Laptev, East Siberian and Beaufort Sea) and the Antarctic Seas near the continent, however, the ocean surface flux is an important heat source for the atmosphere keeping the polar and sub-polar regions relatively mild.

10.4.3 Latent heat of fusion

Latent heat of fusion (melt) globally averaged, is less than 1% of the extraterrestrial solar radiation. Because this value gives an impression of numerical insignificance, the latent heat of fusion is often not considered in energy balance climatology (56Bud). Regionally viewed, however, the latent heat of melt plays an extremely important role. This component, the first term on the right hand side in Eq. (1) is the main heat sink and averages around 100 W m\(^{-2}\) during the melt season, on glaciers, sea ice and snow cover. The reason for the low air temperature in summer in the Arctic, despite one of the largest global radiation of the hemisphere, should be attributed not only to high albedo but also to the heat consumption by the melt. Latent heat of melt is determined in the field either by measuring the lowering of the surface or the discharge of the melt water. The amount of the melt can be determined by obtaining all available energy sources based on the flux measurements.

Table 10.2 shows the surface energy budget on the snow and ice for various regions and surfaces. In theory, one can explain the amount of the melt by evaluating all energy balance terms on the left hand side in Eq. (1). The relative importance of the energy source terms for the melt varies considerably depending on the site characteristics, such as altitude, albedo, air temperature and humidity. One common feature to all sites is the fact that the longwave incoming radiation is by far the most important energy source for the melt. It was shown further in Chapter 4 that the majority of the terrestrial radiation originates from the atmospheric layers very close to the surface. This is the main reason why air temperature alone can be such an effective parameter for estimating the rate of the melt (01Ohm). Fig.10.4 shows the relationship between the total summer melt-ablation and the mean air temperature.
Table 10.2  Radiation and heat balance for glaciers and snow cover during the melt (Unit in W m⁻²). Values in brackets are percentages of total sources and sinks.

<table>
<thead>
<tr>
<th>Site</th>
<th>Coordinates</th>
<th>Altitude</th>
<th>Global radiation</th>
<th>Albedo</th>
<th>Absorbed global radiation</th>
<th>Longwave incoming radiation</th>
<th>Sensible heat flux</th>
<th>Latent heat flux</th>
<th>Sub-surface heat flux</th>
<th>Heat of melt</th>
<th>Longwave outgoing radiation</th>
<th>Black body temperature of surface</th>
<th>Observed period</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Near equilibrium line</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ward Hunt Ice Shelf</td>
<td>83°12' N, 74°30' W</td>
<td>15</td>
<td>207.0</td>
<td>64.0</td>
<td>23 (20)</td>
<td>181 (80)</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>-66 (16)</td>
<td>-316 (84)</td>
<td>273 (61)</td>
<td>60 hr in 6, 1960</td>
<td>62Lis</td>
</tr>
<tr>
<td>Main Ice, McMurdo Ice Cap</td>
<td>79°59' S, 163°19' W</td>
<td>42</td>
<td>202.0</td>
<td>74.0</td>
<td>299 (85)</td>
<td>4 (1)</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>-13 (29)</td>
<td>-116 (70)</td>
<td>243 (59)</td>
<td>1.5 - 16.0, 1960</td>
<td>74Gay</td>
</tr>
<tr>
<td>ETH Camp, Grevent</td>
<td>49°55' N, 14°16' W</td>
<td>115</td>
<td>231.0</td>
<td>77.0</td>
<td>262 (88)</td>
<td>16 (5)</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>-39 (9)</td>
<td>-304 (87)</td>
<td>251 (59)</td>
<td>3.6 - 31.8, 1981</td>
<td>94Bns</td>
</tr>
<tr>
<td>Blue Glacier</td>
<td>47°48' N, 123°17' W</td>
<td>2010</td>
<td>346.0</td>
<td>59.0</td>
<td>258 (87)</td>
<td>50 (1)</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>-13 (29)</td>
<td>-116 (70)</td>
<td>273 (61)</td>
<td>127.3 - 30.8, 1958</td>
<td>59LAC</td>
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<tr>
<td>Vennegiarden</td>
<td>46°30' N, 8°45' E</td>
<td>2660</td>
<td>265.0</td>
<td>39.0</td>
<td>269 (66)</td>
<td>32 (1)</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>-93 (23)</td>
<td>-312 (76)</td>
<td>272 (59)</td>
<td>157.7 - 18.8, 1971</td>
<td>79Wag, 80Wag, 86Tan</td>
</tr>
<tr>
<td>Rongéglacières</td>
<td>49°35' N, 0°24' E</td>
<td>2820</td>
<td>278.0</td>
<td>64.0</td>
<td>254 (89)</td>
<td>81 (19)</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>-167 (35)</td>
<td>-308 (65)</td>
<td>275 (59)</td>
<td>1.9 - 9.8, 1982</td>
<td>85Wun</td>
</tr>
<tr>
<td>No. 1 Glaicer, Titasøya</td>
<td>47°46' N, 8°15' E</td>
<td>3910</td>
<td>232.0</td>
<td>32.0</td>
<td>273 (50)</td>
<td>17 (4)</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>-66 (17)</td>
<td>-309 (79)</td>
<td>272 (59)</td>
<td>1.6 - 18.6, 1986</td>
<td>90Kal</td>
</tr>
<tr>
<td>Ablation area</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Horn Taserm Ice Cap</td>
<td>82°40' N, 30°13' W</td>
<td>340.0</td>
<td>242.0</td>
<td>46.5</td>
<td>277 (86)</td>
<td>27 (6)</td>
<td>-24 (18)</td>
<td>-11 (4)</td>
<td>-10 (3)</td>
<td>-71 (17)</td>
<td>-309 (73)</td>
<td>273 (61)</td>
<td>2.7 - 5.8, 1994</td>
<td>98Bns</td>
</tr>
<tr>
<td>Kongsfjordian Christian Ice Cap</td>
<td>79°53' N, 24°54' W</td>
<td>380.0</td>
<td>319.0</td>
<td>48.0</td>
<td>166 (32)</td>
<td>272 (52)</td>
<td>88 (7)</td>
<td>-30 (7)</td>
<td>-18 (3)</td>
<td>-139 (30)</td>
<td>-136 (60)</td>
<td>273 (61)</td>
<td>8.7 - 27.3, 1993</td>
<td>95Kor, 98Bns</td>
</tr>
<tr>
<td>Lower Ice Station, White Glacier</td>
<td>79°21' N, 90°39' W</td>
<td>308.0</td>
<td>171.0</td>
<td>45.5</td>
<td>298 (63)</td>
<td>52 (11)</td>
<td>32 (7)</td>
<td>-13 (3)</td>
<td>-14 (3)</td>
<td>-149 (31)</td>
<td>-317 (66)</td>
<td>273 (61)</td>
<td>8.7 - 18.8, 1960</td>
<td>64And</td>
</tr>
<tr>
<td>dito</td>
<td>dito</td>
<td>176.0</td>
<td>42.0</td>
<td>45.0</td>
<td>275 (65)</td>
<td>48 (11)</td>
<td>1 (1)</td>
<td>-2 (5)</td>
<td>-89 (21)</td>
<td>-315 (74)</td>
<td>272 (59)</td>
<td>273 (61)</td>
<td>12.6 - 18.8, 1961</td>
<td>67Mil, 90Mil</td>
</tr>
<tr>
<td>dito</td>
<td>dito</td>
<td>209.0</td>
<td>38.0</td>
<td>42.0</td>
<td>286 (62)</td>
<td>42 (9)</td>
<td>9 (2)</td>
<td>-3 (5)</td>
<td>-13 (3)</td>
<td>-316 (68)</td>
<td>273 (59)</td>
<td>273 (61)</td>
<td>16.7 - 31.7, 1982</td>
<td>62Bns</td>
</tr>
<tr>
<td>Snowcap Glacier, Devonsøya</td>
<td>76°00' N, 48°58' W</td>
<td>300.0</td>
<td>140.0</td>
<td>40.0</td>
<td>283 (61)</td>
<td>36 (10)</td>
<td>14 (4)</td>
<td>-1 (3)</td>
<td>-84 (20)</td>
<td>-310 (77)</td>
<td>273 (59)</td>
<td>273 (61)</td>
<td>9.1 - 10.8, 1963</td>
<td>64Kor</td>
</tr>
<tr>
<td>Camp IV, EGIG, Greenland</td>
<td>69°40' N, 49°38' W</td>
<td>1004.0</td>
<td>272.0</td>
<td>56.0</td>
<td>285 (66)</td>
<td>29 (7)</td>
<td>-2 (5)</td>
<td>-11 (3)</td>
<td>-90 (21)</td>
<td>-315 (72)</td>
<td>272 (59)</td>
<td>272 (59)</td>
<td>26.5 - 7.8, 1859</td>
<td>63Amb</td>
</tr>
<tr>
<td>Storglaciären</td>
<td>67°50' N, 10°14' E</td>
<td>1370.0</td>
<td>140.0</td>
<td>59.0</td>
<td>282 (89)</td>
<td>28 (7)</td>
<td>12 (3)</td>
<td>-1 (3)</td>
<td>-49 (14)</td>
<td>-304 (66)</td>
<td>271 (59)</td>
<td>271 (59)</td>
<td>7.6 - 17.9, 1993</td>
<td>98Bks</td>
</tr>
<tr>
<td>dito</td>
<td>dito</td>
<td>172.0</td>
<td>50.0</td>
<td>50.0</td>
<td>278 (67)</td>
<td>49 (12)</td>
<td>3 (1)</td>
<td>-1 (1)</td>
<td>-10 (2)</td>
<td>-309 (73)</td>
<td>272 (59)</td>
<td>272 (59)</td>
<td>5.7 - 6.0, 1994</td>
<td>98Bks</td>
</tr>
<tr>
<td>Aletschgletscher</td>
<td>46°36' N, 0°54' E</td>
<td>2220.0</td>
<td>216.0</td>
<td>27.0</td>
<td>156 (31)</td>
<td>288 (58)</td>
<td>38 (8)</td>
<td>14 (3)</td>
<td>0.00</td>
<td>-181 (36)</td>
<td>-315 (64)</td>
<td>272 (59)</td>
<td>2.8 - 27.8, 1965</td>
<td>67Jun, 87Rot</td>
</tr>
</tbody>
</table>
The largest melt on the planet is expected in a low altitude basin with high snow accumulation and in the terminus region of low altitude glaciers. It is useful to grasp the range of latent heat of fusion under the present climate, whereby the maximum observed is of the most interest. The sea ice will not be considered here, as its annual melt is the order of 1 to 2 m at most and far less than on glaciers and the seasonal snow cover in low altitudes. In the French Alps where the accumulation is larger, the glacier tongue is expected to extend to lower levels. On the terminus area of the Mer de Glace at 1610 m above sea level (a.s.l.), annual melt of 10.1 m water equivalent is observed (C. Vincent, personal communication). Further east in the Swiss Alps, melt of up to 9.5 m ice was observed in an average year on the lowest area of the Triftgletscher at the altitude of 1720 m a.s.l. (M. Funk, personal communication).

Globally viewed, the annual melt of 14 m ice is reported by (96Tak) and (97Nar) at 350 m a.s.l. on the terminus area of the Moreno Glacier in Patagonia. Therefore, 300 to 400 Wm$^{-2}$ can be considered as the upper limit for the latent heat of melt averaged for the entire melt season on glaciers. Latent heat of melt consumed on the seasonal snow cover can be larger when the site has a heavy snow fall and is located at a very low altitude. Melt of a snow cover as deep as 60 m was reported in northern Japan (76Tsu), where the winter monsoon delivers one of the heaviest snowfalls in the world and the summer temperature is relatively high. It is, however, more likely that the regionally averaged maximum melt is about 30 m (T. Shiraiwa, personal communication). Taking the mean density of wet snow of 0.5 t m$^{-3}$, the latent heat of melt averaged over the melt season is estimated at 450 Wm$^{-2}$.

### 10.5 The mean state of the radiation and energy budget of the earth

The annual mean global energy balance is presented in Fig. 10.5. The earth's surface energy balance is compared with the radiation balance at TOA (details are discussed in Chapter 4). The difference in the net radiation at the bottom of the atmosphere (BOA) and at TOA provides one of the most reliable estimates of the radiation budget of the atmosphere. The TOA radiation budget is the mean fluxes of the 52 month observation carried out from February 1984 to May 1988 by ERBE (personal communication by B. Barkstrom). The irradiances at BOA are primarily based on (93Ohm), which are interpreted based on about 900 site-observations assimilated in the Global Energy Balance Archive (GEBA), and 30 Baseline Surface Radiation Network (BSRN) site-observations. The irradiances for geographical regions with insufficient observational coverage are supplemented by the results of ERA-40, adjusted to the observed values at BSRN sites. The main features of the irradiances for the earth's surface are explained below:
The global radiation at the surface, 169 W m\(^{-2}\), accounts for 49% of the primary energy source from the sun. This component appears somewhat smaller than previous examples, such as 186 W m\(^{-2}\) by Sellers (65Sel), 185 W m\(^{-2}\) by Salby (96Sal), and 198 W m\(^{-2}\) by Kiehl and Trenberth (97Kie). The discrepancy of 15 to 30 W m\(^{-2}\) for global radiation is one of the most important findings of recent years. The present value can be compared with 168 W m\(^{-2}\) presented earlier by (88Bud). The new value is the result of the improved network of direct observation of global radiation at the earth's surface. The earth's surface albedo was also newly evaluated and found to converge at around 15%. This value is somewhat larger than for example 14% by 56Bud, 11% by 65Sel, 9% by 96Sal. The present value of 15% coincides with those proposed by 88Bud and 97Kie.

The recently obtained accurate information of the earth's surface albedo is mainly due to the accumulated data on albedo measurements over variety of surfaces by terrestrial and air-borne instruments, the long-term mapping of the distributions of sea ice and snow cover obtained by satellites, and the detailed account of the albedo of the water surface under variable solar elevation angle and cloud cover. Keen observers noticed that the longwave incoming radiation from the atmosphere in Fig.10.5 is considerably larger than previously published. The recent advance in infrared radiometry has corrected previous underestimations. A proper treatment of the water vapour continuum and the effect of clouds are also responsible for the present improvement. The longwave outgoing radiation is calculated based on the recently obtained earth's surface temperatures by (99Jon).

The difference in the net radiation at BOA and TOA is separately evaluated for solar and terrestrial radiation. The divergence of solar radiation is negative due to atmospheric absorption; it amounts to -96 W m\(^{-2}\). The absolute value of this component was earlier grossly underestimated. The most frequently quoted value for the atmospheric absorption of solar radiation is 60 W m\(^{-2}\) (or 17% of the TOA solar irradiance) by (65Sel). The absorption due to water vapour beyond 2.8 µm, aerosol and clouds supports the present result of larger absorption. The divergence of terrestrial radiation is positive and is equivalent to radiative cooling of 200 W m\(^{-2}\). The net divergence of atmospheric radiation is positive; it amounts to 104 W m\(^{-2}\). It is balanced by the convergence of vertical convective fluxes (sensible and latent). Electromagnetic radiation and convection together constitute what is known as radiative-convective equilibrium of the global atmosphere. For more details: see Chapter 4.

The present state of the observation network does not allow the evaluation of the global mean turbulent heat fluxes based solely on observations. Latent heat flux is, however, better understood than the sensible heat flux, as the evaporation is measured at more sites than sensible heat flux. The global mean evaporation has another advantage to be compared with the global precipitation. Summing up the above discussions, it is concluded that the annual mean global latent heat of evaporation is very likely to be 85 W m\(^{-2}\). The sensible heat flux of 19 W m\(^{-2}\) is calculated as the difference between the net radiation and latent heat flux. The kinetic energy, and the conversion rate of the available potential energy to

---

**Fig. 10.5** Redistribution of radiative energy in the climate system (from 98Wil and 04Ohm). All values refer to a mean solar constant of 1368 W m\(^{-2}\) and are global annual averages. Note the large amount of surplus heat at the surface, which must be removed from the surface by turbulent fluxes of latent and sensible heat. The radiation budget of the atmosphere alone is negative, requiring an amount of about 104 W m\(^{-2}\) to be added from ground by evapotranspiration and fluxes of sensible heat.
kinetic energy are taken from (88Kun). The mean energy flow through the climate system can be interpreted in the following manner: The atmosphere absorbs solar radiation in the amount of 96 Wm\(^{-2}\). The net radiation at the earth's surface (104 Wm\(^{-2}\)) is transferred to the atmosphere primarily as latent heat of vapourisation and secondarily as enthalpy flow. The sum of these two terms (200 Wm\(^{-2}\)), augmented by the net longwave radiation from the surface (40 Wm\(^{-2}\)), is emitted by the earth into space at the rate of 240 Wm\(^{-2}\).

The errors of the radiative fluxes of Fig. 10.5 differ for the individual components. Raschke and Ohmura (05Ras) estimate \(\pm 5-7\) Wm\(^{-2}\) as error for the TOA net radiation, \(\pm 15-20\) Wm\(^{-2}\) for the net radiation at ground, and \(\pm 20\) Wm\(^{-2}\) for the total atmospheric radiation flux divergence.

### 10.6 The ETH-Zürich global databases for energy balance

The energy balance of the earth's climate system must be investigated at least with three different methods. This has traditionally been a computational science. Global coverage necessitates satellite-based remote sensing. To validate computed and remote sensed data terrestrial measurements are needed. ETH Zürich (Swiss Federal Institute of Technology) has been building a global database of the instrumentally measured heat fluxes. These are intended to serve as anchor stations to be used both for cross-validating remotely sensed data (satellite and airborne) as well as for examining model computations. Ground truth at the earth’s surface, which is the most important level in the climate system, remains an indispensable part of quantitative energy budget climatology. Therefore, computations, remote sensing and direct observation of fluxes all play important roles. Out of these three methods considerations based on observed fluxes remained underdeveloped, mainly due to the lack of a database which is constructed with systematically collected high-quality observational fluxes. This was the background which necessitated the organisation of such databases.

Table 10.3 shows main specifications of the two data bases available at ETH Zurich (98Ohm). Baseline Surface Radiation Network (BSRN) is a World Climate Research Programme (WCRP) project. It has currently 36 stations covering the earth's surface from Ny Ålesund, Svalbard to Amundsen-Scott at South Pole as illustrated in Fig. 10.6a and Table 10.4. At all stations, important radiation fluxes are measured with 1 Hz with a clearly defined accuracy. One minute statistics are stored at ETH Zürich. An important aspect of this project is that the main characteristics of the atmosphere, such as clouds, atmospheric pressure, temperature, humidity and aerosol are also observed at the same sites. This setting makes it possible to judge if the error in the model computation is due to the radiation code or to the simulated atmospheric parameters. Owing to the high accuracy of the measurements at the BSRN stations, this network is also intended to detect globally important variations in radiation fluxes, such as the effects of volcanic eruptions and the increasing greenhouse effect.

**Table 10.3** Two data bases at the Swiss Federal Institute of Technology (ETH) for World Climate Programme (WCP)

<table>
<thead>
<tr>
<th></th>
<th>Global Energy Balance Archive (GEBA)</th>
<th>Baseline Surface Radiation Network (BSRN)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Information</td>
<td>All energy fluxes, methods, authors</td>
<td>Irradiances, ancillary atmospheric data, instruments and station documentation</td>
</tr>
<tr>
<td>Accuracy</td>
<td>As available</td>
<td>Specified</td>
</tr>
<tr>
<td>Instrumentation</td>
<td>As available</td>
<td>Specified</td>
</tr>
<tr>
<td>Sampling</td>
<td>As available</td>
<td>1 Hz</td>
</tr>
<tr>
<td>Statistics</td>
<td>Monthly mean</td>
<td>1 min. mean, standard deviation, extremes</td>
</tr>
<tr>
<td>Period</td>
<td>As long as the scale can be recovered</td>
<td>Since January 1st, 1992</td>
</tr>
<tr>
<td>Quality assessment</td>
<td>Soft</td>
<td>Strict</td>
</tr>
<tr>
<td>Main applications</td>
<td>Climate research, agriculture, hydrology, design</td>
<td>Climate research, weather prediction, discharge forecast</td>
</tr>
</tbody>
</table>
The second data base, Global Energy Balance Archive (GEBA) contains monthly mean values of all energy budget fluxes previously published by the scientific community (89Ohm; 99Gil). Currently the GEBA data base has 250,000 month station values for about 1,600 sites. The location of the sites (status of 2002) is shown in Fig. 10.6b. This data base was originally designed for the re-evaluation of the earth's surface energy balance. Recently a new additional role emerged in providing observed fluxes for evaluating climate model computations in larger areas of the earth's surface. The seasonal courses of the radiation and energy budget components extracted from BSRN and GEBA are graphically presented in Fig. 10.7 to view the important features of the surface energy budget for main climatic zones. A brief explanation is provided below.

Table 10.4 BSRN Stations with Monthly Statistics

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Abbreviation</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Altitude [m]</th>
<th>Köppen Zone of Climate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ny Ålesund, Spitsbergen</td>
<td>NYA</td>
<td>78°56'N</td>
<td>11°57'E</td>
<td>11</td>
<td>ET</td>
</tr>
<tr>
<td>Barrow, Alaska</td>
<td>BAR</td>
<td>71°19'N</td>
<td>156°24'W</td>
<td>8</td>
<td>ET</td>
</tr>
<tr>
<td>Lerwick, Shetland Islands</td>
<td>LER</td>
<td>60°00'N</td>
<td>1°00'W</td>
<td>84</td>
<td>Cfc</td>
</tr>
<tr>
<td>Toravere, Estonia</td>
<td>TOR</td>
<td>58°16'N</td>
<td>26°28'E</td>
<td>70</td>
<td>Dfc</td>
</tr>
<tr>
<td>Lindenberg, Germany</td>
<td>LIN</td>
<td>52°13'N</td>
<td>14°07'E</td>
<td>125</td>
<td>Cfb</td>
</tr>
<tr>
<td>Regina, Canada</td>
<td>REG</td>
<td>50°12'N</td>
<td>104°17'W</td>
<td>587</td>
<td>Dfb</td>
</tr>
<tr>
<td>Camborne, Great Britain</td>
<td>CAM</td>
<td>50°00'N</td>
<td>5°00'W</td>
<td>88</td>
<td>Cfb</td>
</tr>
<tr>
<td>Fort Peck SURFRAD, USA</td>
<td>FPE</td>
<td>48°31'N</td>
<td>105°12'W</td>
<td>634</td>
<td>BS</td>
</tr>
<tr>
<td>Payrerne, Switzerland</td>
<td>PAY</td>
<td>46°49'N</td>
<td>6°56'E</td>
<td>491</td>
<td>Cfb</td>
</tr>
<tr>
<td>Carpentras, France</td>
<td>CAR</td>
<td>44°03'N</td>
<td>5°02'E</td>
<td>100</td>
<td>Cfb</td>
</tr>
<tr>
<td>Albany, USA</td>
<td>ALB</td>
<td>42°42'N</td>
<td>73°50'W</td>
<td>100</td>
<td>Cfb</td>
</tr>
<tr>
<td>Rock Springs SURFRAD, USA</td>
<td>PSU</td>
<td>40°42'N</td>
<td>77°56'W</td>
<td>376</td>
<td>Cfb</td>
</tr>
<tr>
<td>Boulder SURFRAD, USA</td>
<td>BOS</td>
<td>40°13'N</td>
<td>105°24'W</td>
<td>1689</td>
<td>BS</td>
</tr>
<tr>
<td>Bondville, IL, USA</td>
<td>BON</td>
<td>40°06'N</td>
<td>88°37'W</td>
<td>213</td>
<td>Dfa</td>
</tr>
<tr>
<td>Boulder, USA</td>
<td>BOU</td>
<td>40°03'N</td>
<td>105°59'W</td>
<td>1577</td>
<td>BS</td>
</tr>
<tr>
<td>Chesapeake Lighthouse, USA</td>
<td>CHL</td>
<td>36°54'N</td>
<td>75°42'W</td>
<td>34</td>
<td>Cfa</td>
</tr>
<tr>
<td>Desert Rock SURFRAD, USA</td>
<td>DRA</td>
<td>36°38'N</td>
<td>116°05'W</td>
<td>1007</td>
<td>BW</td>
</tr>
<tr>
<td>Billings, ARM/CART, USA</td>
<td>BIL</td>
<td>36°36'N</td>
<td>97°29'W</td>
<td>318</td>
<td>Cfa</td>
</tr>
<tr>
<td>South Great Plains E13, USA</td>
<td>E13</td>
<td>36°36'N</td>
<td>97°30'W</td>
<td>318</td>
<td>Cfa</td>
</tr>
<tr>
<td>Tateno, Japan</td>
<td>TAT</td>
<td>36°03'N</td>
<td>140°08'E</td>
<td>25</td>
<td>Cfa</td>
</tr>
<tr>
<td>Goodwin Creek, USA</td>
<td>GCR</td>
<td>34°15'N</td>
<td>89°52'W</td>
<td>98</td>
<td>Cfa</td>
</tr>
<tr>
<td>Bermuda</td>
<td>BER</td>
<td>32°16'N</td>
<td>64°20'W</td>
<td>8</td>
<td>Cfa</td>
</tr>
<tr>
<td>Sede Boqer, Israel</td>
<td>SBO</td>
<td>30°52'N</td>
<td>34°46'E</td>
<td>500</td>
<td>BS</td>
</tr>
<tr>
<td>Solar Village (Riyadh), Saudi Arabia</td>
<td>SOV</td>
<td>24°39'N</td>
<td>48°46'E</td>
<td>650</td>
<td>BW</td>
</tr>
<tr>
<td>Tamanrasset, Algeria</td>
<td>TAM</td>
<td>22°47'N</td>
<td>5°51'E</td>
<td>1385</td>
<td>BW</td>
</tr>
<tr>
<td>Kwajalein, Marshall Islands</td>
<td>KWA</td>
<td>8°43'N</td>
<td>167°44'E</td>
<td>10</td>
<td>Aw</td>
</tr>
<tr>
<td>Ilorin, Nigeria</td>
<td>ILO</td>
<td>8°32'N</td>
<td>4°34'E</td>
<td>350</td>
<td>Aw</td>
</tr>
<tr>
<td>Nauru, Nauru</td>
<td>NAU</td>
<td>0°31'S</td>
<td>166°55'E</td>
<td>7</td>
<td>Af</td>
</tr>
<tr>
<td>Momote, ARM, Papua New Guinea</td>
<td>MAN</td>
<td>2°06'S</td>
<td>147°43'E</td>
<td>6</td>
<td>Af</td>
</tr>
<tr>
<td>Alice Springs, Australia</td>
<td>ASP</td>
<td>23°42'S</td>
<td>133°52'E</td>
<td>547</td>
<td>BW</td>
</tr>
<tr>
<td>Florianopolis, Brazil</td>
<td>FLO</td>
<td>27°28'S</td>
<td>48°29'W</td>
<td>11</td>
<td>Cfa</td>
</tr>
<tr>
<td>De Aar, South Africa</td>
<td>DAA</td>
<td>30°40'S</td>
<td>23°59'W</td>
<td>1287</td>
<td>BS</td>
</tr>
<tr>
<td>Lauder, New Zealand</td>
<td>LAU</td>
<td>45°03'S</td>
<td>169°41'E</td>
<td>350</td>
<td>Cf</td>
</tr>
<tr>
<td>Syowa, Antarctic</td>
<td>SYO</td>
<td>69°00'S</td>
<td>39°35'E</td>
<td>18</td>
<td>ET</td>
</tr>
<tr>
<td>Georg von Neumayer, Antarctic</td>
<td>GVN</td>
<td>70°39'S</td>
<td>8°15'W</td>
<td>42</td>
<td>EF</td>
</tr>
<tr>
<td>Amundsen-Scott, Antarctic</td>
<td>SPO</td>
<td>90°S</td>
<td>–</td>
<td>2800</td>
<td>EF</td>
</tr>
</tbody>
</table>
Fig. 10.6 Distribution of the sites where radiation and energy balance components are measured (a: top - BSRN Baseline Surface Radiation Network; b: bottom - GEBA Global Energy Balance Archive). See also Tables 10.3 and 10.4.

Fig. 10.7a Monthly mean energy fluxes at eight selected sites with long-term observations: Barrow, Cabauw, Reckenholz.
The seasonal transition of the surface energy fluxes is basically forced by the solar irradiance at the TOA. The seasonal range of the TOA solar irradiance is smallest over the equator and increases with latitude, attaining the maximum at the poles. The seasonal range is only 53 W m\(^{-2}\) over the equator, while it is 526 and 562 W m\(^{-2}\) over the North and South Poles, respectively, about ten times larger. One example of the energy balance at a tropical site, Jaru Reserve, Brazil demonstrates this point clearly. Because of the small variation of the TOA solar irradiance at 10° S, the global radiation varies little throughout the year. This astronomical condition also determines the constancy of net radiation which is about 145 Wm\(^{-2}\) in all months. Another important feature of the energy balance for tropical forests is the large terrestrial atmospheric radiation (420 Wm\(^{-2}\)) which is among the possibly largest irradiances in the world. This condition is caused by high temperature, high absolute humidity and large cloud amount. It should also be noted that the Bowen ratio over the tropical rain forests is small (0.3) and constant. This means that the latent heat flux is 3 times larger than sensible heat flux. This circumstance is due not only to large precipitation but also to the high temperature and the lack of a clear dry season in humid tropics.
The mid-latitude offers the best data for this subject. Five sites presented in this section are located either on short grass (Reckenholz, Switzerland, Rietholzbach, Switzerland and Tateno/Tsukuba, Japan) or over the forest canopy (Hartheim, Germany and Morgan-Monroe State Forest, USA). The seasonal course of global radiation is characterised by a moderately strong seasonal fluctuation of 100 to 200 W m\(^{-2}\) which is the primary cause of the fluctuation in net radiation ranging between -5 and 160 W m\(^{-2}\) in the course of a year. The change in albedo plays also an important role. Given the similar global radiation, net radiation over a forest tends to appear slightly larger due to the smaller albedo (0.10-0.15) in comparison with that of the grass-covered surfaces (0.18-0.26). The difference becomes most conspicuous during the snow cover season.

When a snow cover of more than 5 cm occurs, the albedo over the short grass increases significantly, while that over the forest often remains small, as snow flakes fall through the canopy and settle on the forest floor. When the soil moisture is sufficient the Bowen ratio tends to be larger over the forest than over the grass land. A typical Bowen ratio over the forest in mid-latitude is 0.65, while that over the grass is 0.35.

The Arctic region is represented by Barrow, Alaska and the SHEBA (Surface Heat Budget of the Arctic Ocean) site, a station on sea ice in Beaufort Sea. The Barrow site is one of the most intensely studied tundra sites, while SHEBA is the only site where the full energy balance observation was carried out for more than 11 months on sea ice. The most important features of the polar regions are the large seasonal range in global radiation and the existence of the polar night. At Barrow one year can be divided into winter (negative net radiation, October to April) and summer (positive net radiation, May to September) half years. The energy exchange in winter is dominated by net radiation as the sole and powerful sink, and by sensible heat flux as the main heat source. This pattern is universal for all land surfaces in high latitudes in winter.

The energy exchange in summer is characterised by net radiation as the only heat source and by both turbulent fluxes with a similar magnitude. In June there is a brief period of snow melt. The melt period on the tundra is usually short owing to a snow cover of small water equivalent. The Bowen ratio in the polar region remains high during the summer despite higher temperature, mainly as a result of the dryness of the near-surface soil. On the sea ice the winter energy balance is also dominated by the negative net radiation. The place of the main heat source, however, is taken by the sub-surface heat flux from the ocean. The summer energy balance on the sea ice is dominated by net radiation as the source and the latent heat of melt as the sink, except for the early summer as the downward heat conduction into the sea ice is briefly a significant sink. This feature has tremendous application. The sheer net radiation on the sea ice surface can be converted into the heat of melt as a good approximation.

Two examples of applications of the data stored in BSRN and GEBA are presented in Fig. 10.8 and 10.9. Fig.10.8a and b represent the relationships between the observed and ISCCP D2 irradiances for the period of 1991 to 1995, for global and longwave incoming radiation, respectively. The figures show that the ISCCP procedure computes longwave atmospheric radiation at the earth's surface with reasonable accuracy, while global radiation is systematically overestimated by 11%. Since the raw computational value for the ISCCP global mean solar radiation is 189 W m\(^{-2}\), the corrected value becomes 169 W m\(^{-2}\) which agrees better with recent estimates of the mean solar radiation at the earth's surface.
The computations of turbulent heat fluxes are subject to more serious uncertainties than radiative fluxes. Fig. 10.9 shows comparisons among simulated sensible heat fluxes computed by three general circulation models (ECHAM4, HadAM2b, and ARPACE) and the observations obtained from GEBA. A common feature for the six sites in Fig. 10.9 is the overestimation of the downward flow of sensible heat in winter. This problem should be interpreted together with the comparisons for latent heat flux presented in Fig. 10.10. In Fig. 10.10 monthly mean latent heat fluxes computed by two general circulation models (ECHAM3 and ECHAM4) are compared against long-term observations. Sites in Fig. 10.9 and Fig. 10.10 are rare cases where turbulent heat flux measurements have been carried out long enough to be climatological. At Rietholzbach, a weighing lysimeter is used, while the aerodynamic profile method is used at Hartheim. Cabauw is one of the few sites in the world where sensible and latent heat fluxes have been measured by the eddy covariance method for a number of years.

This comparison depicts one of the most common and serious problems in model simulations, in which the computations tend to overestimate the magnitude of sensible and latent heat fluxes during the colder seasons. The overestimation of the downward flow of sensible heat is caused by the overestimation of the turbulent exchange coefficient under stable stratification. The overestimation of evaporation is caused similarly by an overestimation of the turbulent exchange coefficient for water vapour under the stable condition. This tendency is further accentuated by the overestimated downward sensible heat flux which becomes additional available energy for evaporation. This problem is not limited to higher latitudes but observed wherever a stable boundary layer appears, for example in the tropics at night.
The overestimation of evaporation in winter and early spring causes an artificial desiccation of the surface layer of soil, hence an underestimation of evaporation in late spring and early summer. This situation creates artificially dry air and a suppression of convection, resulting in the lack of precipitation during warmer seasons. This positive feedback often creates a sustained period of artificial drought and an overestimation of the surface temperature in climate simulations. If such models are used for the simulation of the enhanced greenhouse climate, they forecast an overestimated temperature rise, which is the most controversial point in climate impact analyses. The present comparison revealed the limits of our knowledge on a stable boundary layer and contributed largely to launching several projects dedicated to investigating fundamental transport processes in a stable boundary layer. It is hoped that the leading GCMs will soon be equipped with more realistic codes for computing turbulent fluxes which are crucial for estimating surface temperature accurately.

BSRN and GEBA can be accessed through INTERNET with the following addresses: http://bsrn.ethz.ch/ for BSRN and http://bsrn.ethz.ch/gebastatus/ for GEBA.
10.7 Conclusion

The earth’s surface energy balance was traditionally computed with empirical or highly simplified physical methods. In recent decades, however, the advances in instruments and observation networks have made it possible to evaluate some fluxes on a global scale, such as global radiation based on observations. Other fluxes, such as sensible heat flux, are still rarely observed. Climate modelling and satellite radiometry have become advanced enough to fill in gaps left by observations. Combining all information and data which became available from these advanced methods, the earth’s surface energy balance is recalculated. The present chapter summarises concepts on non-radiative fluxes and evaluates each method from the viewpoint of compatibility with field experiments.

Further, the geographical and seasonal distributions of the energy balance on the earth's surface are discussed for major climatic zones, based on observed fluxes. Certain differences from previously published fluxes become evident, and are explained in the present chapter. Examples are also given to indicate how computationally simulated fluxes compare against those observed. These comparisons demonstrate the indispensable nature of instrumentally observed fluxes in order to realistically simulate climates and to accurately construct regional and global energy balance.

Acknowledgement

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