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Laten heat flow (He) involves energy lost (acquired) in the stream during evaporation (condensation) as the water moves to a higher lower energy state (or vice versa). From: Earth-Science Reviews, 2017Shiro Imawaki, ... Bo Qiu, international geophysics, 2013Latent and intelligent heat currents increase three to five times the warm waters of the Agulhas current system, and there is a deepening of the marine-atmospheric boundary layer, and increased formation of convective clouds (Jury and Walker, 1988; Lee-Thorp and his mts, 1998; Rouault and its mts, 2000). The Agulhas Current system influences storm course positions and storm development, as well as regional atmospheric orbital patterns (Reason, 2001; Nakamura and Shimpo, 2004) and have been linked to extreme rainfall events and tornadoes in South Africa (Rouault et al., 2002). Uniquely wbcs, the Agulhas Current system is thought to be an important source of continental moisture (Gimeno and al., 2010). Rainfall over Africa is linked to SST anomalies in the larger Agulhas current system, which is linked to the Indian Ocean Dipole and El Niño/Southern Oscillation cycles. General warming of the system since the 1970s may increase the sensitivity of rainfall in Africa to these cycles (Behera and Yamagata, 2001; Zinke and his mts, 2004). Lisan Yu, in Encyclopedia of Ocean Sciences (Third Edition), 2019Latent and intelligent heat flows are estimated to be wind speed and sea-to-air humidity and temperature differences in the following bulk parameters (Fairall and its mtsai., 2003): where The latent heat of distillation and the function of the SST (Ts), le = (2,501 to 0,00237 × T) × expressed in 1,06. cp is the specific heat capacity of the air at constant pressure, ce and ch are turbulent exchanges depending on stability and height for latening and reasonable heat. Ta and qa are the temperature and specific humidity at a reference altitude of 2 m above the surface of the sea. qs is the saturation humidity Ts.The calculation of laten and reasonable heat streams requires input of four variables, W, Ts, qa, and Ta. The satellite provides direct retrievals for w and ts. Infrared radio clocks, such as the five-channel Advanced Very High Resolution Radiometer (AVHRR), use wavelength bands of 3.5-4 μm and 10-12 μm, which represent a high transmission in a cloud-free atmosphere. On these wavelength bands, clouds are opaque to infrared radiation and effectively mask radiation from the surface of the ocean. Because of the cloud effect or 2 weeks to complete a global SST field AVHRR although the satellite orbits the Earth 14 times every day 833 km above the surface, and each pass of the satellite provides 2,399 km of wide swaths. Clouds, on the other hand, have little effect on microwave radio clocks. Microwave SST retrieval can be provided in all weather conditions except rain. The first satellite sensor capable of measuring SST through clouds was the TRMM microwave imager (TMI) launched in 1997, which has channels between 10.7 and 85 GHz. However, the low-angle equatorial trajectory limits the coverage of the TMI to a width of approximately 38 degrees. The measurement of global cloud ssts was made possible by the NASDA Advanced Microwave Scanning Radiometer (AMSR) on board NASA's EOS Aqua spacecraft (AMSR-E, 2002-2011) and the AMSR-2 on board the JAXA GCOM-W1 (since 2012). The subsurface wind speed can only be retrieved in a microwave electromagnetic system. The emission capacity of the ocean surface at infrared wavelengths is so high that it is not susceptible to changes in sea surface ripening caused by wind or humidity fluctuations in the lower atmosphere. Microwave wind speed recoveries are provided by the Advanced Sensor Microwave/Imager (SSM/I), which has been flying on polar orbiting operations spacecraft from the Defense Meteorological Space Program (DMSP) since July 1987. The SSM/I has a wide swate (~ 1,400 km) and covers 82% of the Earth's surface in 24 hours. Unlike the scatterometer that measures both wind speed and direction, SSM/I has a passive microwave sensor and provides wind speed recovery only (Wentz, 1997). However, the bulk flux parameterization of laten and meaningful heat streams requires wind speed, but not wind direction. With high space-time resolution, near global daily coverage, and a continuous record it made SSM/I the primary input dataset to calculate surface-turbulent heat streams. Since 2003, SSM/I instruments have been replaced by a new series of passive microwave conically scanning images and sound indicators (SSMIS). SSMIS brightening temperature measurements are calibrated to the level of SSM/I measurements, and SSMIS continues its SSM/I data record. Recovering Ta and QA a few metres above the surface is proving difficult for space government technology, as radiation is emitting relatively thick atmospheric layers, not from a single level. Using the total water vapour column (PW) measured by satellite to obtain QA and Ta, Liu (1988) pioneered, who showed that most of the spatial and temporal variability of water vapour in the lower part of water vapour Column. The separation of the atmospheric boundary layer from the higher atmosphere creates a correlation between the surface qa level and all observations of pw. Since then, various algorithms have been developed to take advantage of measurements provided by passive microwave radio meters, including SSM/I, SSMIS, AMRE-E and AMSR-2. In May 1998, the introduction of the Advanced Microwave Sounding Unit (AMSU) on a series of NOAA polar orbiting meteorological satellites made further measurements. AMSU is a microwave radiometer that can detect temperature and moisture at different levels of the atmosphere by passively recording atmospheric microwave radiation at 15 wavelengths. The AMSU channel at 52 GHz is close to the characteristic absorption peak of oxygen and contains information about the physical temperature and perspective profiles of atmospheric molecules. This channel is the most useful for Ta retrieval. The algorithm development of retrieval of Ta and QA can be divided into three approaches. The first approach uses SSM/I measurements to determine water vapour at the bottom 500 m of the atmosphere, and then predicts qa. This approach stems from the fact that water vapour in the well-mixed atmospheric boundary layer may be associated with qa a linear regression. The second approach creates qa retrievals from SSM/I Tb measurements on 19, 22, 37 GHz channels (Figure 3A-C), as these channels are most sensitive to water vapour. The weak water vapour absorption line of microwave spectroscopy is located at 22 GHz, which allows the signal from water vapour to be more distinct from the signal from the cloud (i.e. condensed water). The third approach is to make use of atmospheric sound devices such as AMSU (Fig.3D), a high-resolution infrared radiation Sounder (HIRS), and a special sensor microwave water vapor Profiler (SSM/T and SSM/T-2). The detailed profile data observed by the sound alarms helps remove variability that is not related to the surface from all column measurements. The recovery of SSMIS 19, 22, 37 GHz and AMSU-A 52 GHz Tb shows estimates of ta and qa (Figure 3), and the estimate is based on a multivariate regression with qa and Ta buoy and Ta measurements as training data.3. On 1 January 2010, the daily mean brightness temperature fields (Tb) (A) 19 GHz (B) 22 GHz(C) at 32 GHz and the AMSU-A channel (D) at 52 GHz. Daily average fields are the average of ascending and descending (E) qa and (F) Ta functions, estimated using a multivariate regression algorithm. Taken over by Yu, L. and Jin, X. (2018). Retrieving the humidity and temperature of groundwater by means of a system-specific regression model. It's remote. Environment. 215(9).L. Yu, in the Encyclopedia of Ocean Sciences (second 2009Latent and intelligent heat streams are the primary mechanism by which the ocean transmits much of the absorbed solar radiation back into the atmosphere. The two flows cannot be directly monitored by space sensors, but it is possible to estimate wind speed and sea-air humidity/temperature differences using the following mass parametrizations: where down to latening heat evaporation and the function of sea surface temperature (SST, Ts) expressed as Le=(2.501−0.002 37×Ts)×1.06. cp is the air's specific heat capacity at constant pressure; ce and ch are stability- and height-dependent turbulent exchange ingibles for laten and reasonable heat, respectively. Ta/qa is the temperature/specific humidity above the sea surface at a reference height of 2 m. qs is the saturation humidity of Ts and multiplied by 0,98 to take account of the reduction in vapour pressure caused by salt water. The two variables, Ts and W, are de-satellited between the ts in [2] and [3], so qs is known. Remote sensing of Ts is based on techniques by which infrared and microwave radiometers spread through space detect heat-emitted radiation from the surface of the ocean. Infrared radio clocks, such as the five-channel advanced high-resolution radiometer (AVHRR), use wavelength bands of 3.5-4 and 10-12 μm, which have a high transmission of a cloud-free atmosphere. The downside is that clouds are opaque to infrared radiation and effectively mask radiation from the surface of the ocean, and this affects temporal resolution. Although the AVHRR satellite orbits the Earth 14 times each day 833 km above the surface, and each pass of the satellite provides a 2,399 km-wide swath, it is usually 1 or 2 weeks, depending on the actual cloud coverage to provide total global coverage. Clouds, on the other hand, have little impact on microwave radio clocks, so microwave Ts retrieval can be done under full cloud cover, except in rainy conditions. Launched in 1997, the TRMM microwave imager (TMI) has a complete package of channels between 10.7 and 85 GHz and was the first satellite sensor capable of accurately measuring SST through clouds. However, the low-angle orbit limits TMI coverage to a latitude of only 38°. Following TMI, the first polar orbiting microwave radiometer capable of measuring global cloudY SST was made possible by NASDA's advanced microwave reading radiometer (AMSR), which flew aboard NASA's EOS Aqua mission in 2002. This is because the emissivity of the ocean surface at a wavelength of around 11 μm is so high that it is not sensitive to changes in the rippling of the wind-induced sea surface or humidity fluctuations in the lower atmosphere, wind speed recoveries are provided by the special sensor microwave/image (SSM/I) of polar orbiting operational spacecraft of the Defence Meteorological Space Programme (DMSP) since July 1987. The SSM/I has a wide swate (1,400 km) and covers 82% of the Earth's surface within 1 day. But unlike scatterometers, the SSM/I is a passive microwave sensor and cannot provide information about the direction of the wind. This is not a problem in the calculations of eqns [2] and [3], which only require wind speed monitoring. In fact, the high space-time resolution and good global coverage of this SSM/I made it a serving primary database to calculate the climate's medium and variability in oceanic laten and reasonable heat streams over the past 20-year period. Currently, high-precision wind speed measurements are also available on many NASA satellite platforms, including TMI and AMSR. The most difficult problem with satellite flux estimation is the recovery of air humidity and temperature, qa and Ta, levels several meters above the surface. This problem is inherent in all passive radiometers in space, as the measured radiation comes from relatively thick atmospheric layers, not from a single level. One of the general practices of the extraction of the qa satellite is the relationship of qa with water vapour (IWV, also known as total stormwater) integrated into the column observed from SSM/I. However, the approach results in systematic bias above 2 g kg−1 in the tropics and in the mid- and high latitudes in summer. This is due to the impact of water vapour convergence that is difficult to gauge in regions where surface air is almost saturated, but the total IWV is small. In such situations, the IWV cannot reflect actual vertical and horizontal humidity fluctuations in the atmosphere. Various remedies have been proposed to improve the QA-IWV relationship and make it applicable to synoptic and shorter deadlines. There are methods that include additional geophysical variables, replacing IWV with IWV at the lower 500 m of the planetary layer and/or empirical orthogonal functions (EO). Although overall improvements have been made, accuracy remains poor as there is no detailed information on atmospheric humidity profiles. Recovering Ta from satellite observations is even more challenging. Unlike humidity, there is no coherent vertical temperature structure in the atmosphere. Satellite temperature-sounding radio meters are of little help as they are usually designed to be scans in wide vertical layers. Low information content of the sounder the lower atmosphere does not allow the groundwater temperature to be recovered with sufficient accuracy. Various methods were tested to deduce Ta from the in conclusion qa, but all showed limited success. Due to the difficulties in determining QA and TA, laten and reasonable flows estimated from satellite measurements involve great uncertainty. Three methods have been tested to obtain better QA and Ta to improve estimation of laten and reasonable flows. The first approach is to enhance the information about temperature and moisture in the lower troposphere. This is achieved by combining SSM/I data with additional sub-microwave data from instruments such as advanced microwave sounding units (AMSU-A) and microwave humidity sounders (MHS) flown aboard the National Oceanic and Atmospheric Administration (NOAA) polar orbiting satellites, and advanced sensor microwave temperature sounders (SSM/T) and (SSM/T-2) DMSP satellites. Although sound indicators do not directly provide shallow surface measurements, detailed profile data provided by the sound indicators can help eliminate variability of entire column measurements that are not related to the surface. The second approach is to capitalize on the progress made in numerical weather forecasting models that assimilate colder observations into the physical-based system. The qa and ta estimates of the models contain less ambiguity about vertical integration and large spatial averaging of different parameters, although they are subject to systematic bias due to sub-grid parametrizations of the model. The third approach is to get a better estimate of qa and ta with the optimal combination of satellite retrieval with model outputs experimented with by the Woods Hole Oceanographic Institute (WHOI) objectively analyzed Air Sea Fluxes (OAFlux) project. The effort has led to an improvement in daily estimates of latening and reasonable flows in the global air sea. Axel Kleidon, in Encyclopedia of Ecology (Second Edition), 2019The surface heat flow in the surface energy balance is directly related to evaporation, which wets the atmosphere. The atmosphere is dried with precipitation, which releases laten heat into the atmosphere, causing a strong source of heating and wet convection. The change in evaporation in the zone strongly follows the change in solar radiation in the zone, but precipitation is concentrated in the tropics and middle latitudes (Figure 4c). The difference between evaporation and precipitation indicates that moisture is transported by atmospheric circulation (this is indicated by grey shading in Figure 4c). In the tropics, condensation exceeds evaporation, as the atmosphere imports moisture from the subtropics and condenses it within the inner tropical convergence in a relatively narrow band of tropics Large-scale atmospheric circulation also transports moisture from subtropical parts to the mid latitudes, which exceeds evaporation at these latitudes. Therefore, hydrological cycling directly follows from the latent heat of the energy balance, but with a strong imprint of atmospheric circulation. Dennis L. Hartmann of Global Physical Climatology (Second Edition), 2016A latening heat flow is sensitively dependent on temperature through the dependence of saturation steam pressure on temperature. Above water or aqueous surface, it can be assumed that the mixing ratio of water vapour on the surface is equal to the mixing ratio of q* saturation at the surface temperature. The vapour mixing ratio of saturated air can be reached at reference height with a first-order Taylor series. (4,29)qa*=q*(Ts)+∂q*∂T(Ta−Ts)+... The actual humidity ratio of the air at the reference height may be expressed in relative humidity at that level. (4.31)qa≅RH(q*(Ts)+∂q*∂T(Ta−Ts))Replacing (4.31) (4.27) from the surface in terms of temperature difference evaporates from the surface and relative humidity(4.32)LE≅ρ L CDE is q*(Ts) (1−RH)+RH Be−1 cpL(Ts−Ta), where The Bowen ratio is the ratio of reasonable cooling to latening of the surface. Comparison (4.32) and (4.26) we see that when the surface is wet and the air is saturated, RH = 1, and assuming cdh = CDE, the Bowen ratio takes a special value:When the surface and air reference level are saturated, the Bowen ratio is close to the value to be given (4.33), which can be called the equilibrium Bowen ratio. It should be assumed that the flow of moisture from the boundary layer to the free atmosphere is sufficient to balance the upward flow of moisture from the surface at the saturation value. In such a equilibrium situation, the Bowen ratio is inversely proportional to the change in the saturation mixing ratio of water vapour and temperature (4.33). The speed at which the saturation-mixing ratio changes with temperature is very sensitive to the temperature itself. The approximate formula (1.9), can be shown thatthe exponential dependence on the saturation-blending ratio of the temperature far exceeds the inverse square temperature (4.35), so that the balance Bowen ratio exponentially decreases the temperature. The temperature dependencies of the saturation blending ratio and the equilibrium Bowen ratio are graphically displayed on the log-linear curve in the 4.10. The equilibrium Bowen ratio unit is about 0 °C, and is reduced to about 0.2 to 30 °C. As the relative humidity (4.32) decreases from 1 to less, evaporative cooling increases, so the balance the ratio of wet surface is possible highest Bowen ratio. The actual Bowen ratio on a wet surface will usually be less than the equilibrium Bowen ratio, as the air at the reference height is usually not saturated. Due to the strong temperature dependence of the saturation vapour pressure, latening of the surface dominates rational cooling from the wet surface at temperatures such as the tropics, but at high latitudes in winter, reasonable heat delivery may be the most important. To display this, assume that the wet surface and saturated atmosphere, so that the actual Bowen ratio is the equilibrium Bowen ratio, we can write thatAbra 4.10. Saturation-specific humidity q*(g kg−1) and equilibrium Bowen ratio On as temperature functions. Make it = 1 at approximately 278 K. At much warmer temperatures you need «t 1 and most surface cooling is evaporation. At much colder temperatures» 1 and most of the surface cooling is done with a reasonable heat transfer. The previous debate strictly applies only to conditions where the surface is wet, so evaporative cooling is not limited by the lack of surface moisture. Evaporative cooling in land areas may be significantly reduced if moisture cannot be provided quickly enough from beneath the surface to keep the air saturated with the surface. In desert areas, the surface is typically so dry that evaporative cooling is small regardless of temperature, so reasonable cooling and longwave emissions should balance solar heating. In the vegetated terrain, cooling by evaporation and flow through the leaves are controlled by the physical and biological state of the plant's canopy and the water content of the soil. The role of soil and vegetation in the balance of surface water and energy is further discussed in Chapter 5.RolandStull , atmospheric science (Second edition), 2006This subsection describes a method that can be used to estimate the flux of meaningful and laten heat on the Earth's surface. This so-called mass aerodynamic method allows us to estimate the friction pull of the surface wind as well. The reasonable heat flow between the surface and the cover air is driven by two processes. Inside the lower few millimeters of the atmosphere, very large vertical gradients in the form of temperature, causing molecular driving heat away from the surface into the air. There is no turbulent flux at the bottom of this molecular layer (e.g. as the top surface of the soil), because the surface of the soil does not usually dance the vortex dance. But from the top of this molecular layer or microlayer to the top of the boundary layer, molecular conduction is negligible, while turbulent convection takes over, moving warm air upwards to distribute meaningful heat over the boundary layer. Because the microlayer is so thin in relation to the depth of the boundary layer and because the flux of the entire microlayer is almost constant and equal to the turbulent vortex flux at the top of the microlayer, it is possible to determine the effective turbulent flux, which is the sum of molecular and true turbulent components. In practice, the effective word is often omitted, and this quantity is simply called turbulent flux on the surface. Efficient reasonable heat flow is often determined by the difference in temperature between the surface and the air. If the surface skin temperature, Ts, is known, the reasonable heat flow (kinematic units K m s21) can be parameterized from the ground to the air(9.19a)FHs =CH| V| (Ts−Tair), where CH is the heat and | V| and Tair are wind speed and air temperature at the standard surface measurement height (10 m and 2 m respectively). To switch from the kinematic thermal current to the dynamic heat flow (W m22), the FH shall be multiplied by the density of the air, multiplying the specific heat at a constant pressure (r cp). Under static neutral conditions, there is moderate turbulence on flat earth surfaces, which replaces slow moving air near the ground with air moving faster in the boundary layer and results in ch values in the range 0,001 to 0,005 (indicated as CHN to indicate neutral conditions). The exact value of chn depends on surface finish, similar to the value of the chn 9.2. In contrast, as the air becomes more statically stable, the Richardson number increases toward its critical value, and the kinetic energy of turbulence drops to zero, which also reduces CH to zero. To estimate the vertical heat flow, one might have expected Eq. (9.19a) is a function of a vertical turbulent transport speed of five times the temperature difference. But the first order closure, wT parameterized CH | V|, where it is assumed that stronger winds near the ground cause stronger turbulence, causing stronger turbulence. By combining Eqs. (9.17-9.19a), we see that surface skin temperatures above the ground on sunny days are really a response to solar heating rather than an independent driver of heat flow. For example, in a day of mild wind, the net radiation budget causes a certain energy intake to the ground, leading to a rise in surface skin temperature, according to the first law of thermodynamics. As the skin heats up, reasonable heat fluting increases with Eq. (9.19a), as well as evaporation and heat into the ground. Because the wind is light, (9.19a) shows that skin temperature should become a little warmer as the air temperature to drive enough sensible fhs to help surface heat budget. However, on a windier day, the required heat stream can be achieved by the surface skin temperature that is only slightly warmer than the air temperature. When warmer air is released on a cooler surface, or when the soil is cooled at night by long-wave radiation, it is ts < Tair, and the heat flow becomes downwards. This cools the bottom of the boundary layer and leads to subadiabatic limitation rates and reduction or suppression of turbulence. As turbulence decreases, cooling is limited to the bottom of the boundary layer, which creates a shallow stable boundary layer embedded in the old, deeper boundary layer. Similar equations, called mass aerodynamic connections, can be derived from moisture flowing through oceans, lakes, and saturated soils. We can assume that the specific humidity qs near the surface is the same as the saturation value specified in the Clausius-Clapeyron equation, based on the temperature of the air close to the surface of the sea. Namely, moisture fwater [kinematic units (kgwater vapour / kgair) (m s21)] on the surface (9.19b)Fwater =CE| V| [qsat(Ts)−qair], where ce is the bulk transfer ingesting of moisture without dimension (CE ≈ CH). This flow of moisture is directly linked to the laten heat stream (in kinematic units k m s21) on the surface and to this evaporation rate of water (mm/day) (9,20)Fwater=ÿ FEs=(ρliq/ρair) E, where g = cp/Lv = 0,4 [(gwater vapor/kgair)/K] is the psychometric constant and ρliq is the density of pure liquid water (not seawater). On the surface, the ratio of reasonable to laten heat flows is called the Bowen ratio6: B = FHs/FEs. Due to the nonlinearity inherent in the Clausius-Clapeyron equation, the Bowen ratio over the oceans decreases as the sea surface temperature increases. Typical values on the edge of the ice range from 1.0 ±.5 to 0.1 in tropical oceans, where latent heat flows dominate due to the warmth of the sea's surface. On land, the evaporation rate, and thus the Bowen ratio, depends on the availability of water in the soil and the composition of the vegetation that transports water from the soil through osmosis. Plants release water vapor into the air through the flow through the open stomata (pores) of the leaves. Thus, the Bowen ratio ranges from 0.1 tropical oceans, to 0.2 over irrigated plants, 0.5 as pasture, 5.0 as semi-arid regions, and 10 deserts above. The momentum of the bulk aerodynamic approach gives a drag law where the CD has a dimensionless drag coefficient, ranging from magnitude 1023 on smooth surfaces to 2 ÷ 1022 rough as well (Table 9.2), and *2 with the magnitude of momentum flux lost down into the ground. The CD is used not only for friction of the skin (viscous pulling), but also for form-pulling (pressure transitions wind and tailwinds, as well as the tailwinds of obstacles such as trees, buildings and mountains) and Drag. Therefore, the CD may be larger than the CH. The CD of the pull-co2 varies depending on the stability of the neutral CDN, as does ch; cdn for unstable boundary layers, and CD CDN for stable boundary layers. The increase in wind speed over the oceans leads to an increase in wave height, which also increases drag (see framed no. 9.1. Figure 9.13 shows that mass transmission ingrates of momentum, heat, and humidity vary with wind speed measured at altitude z = 10 m above the oceans. At wind speeds of more than 5m s21, heat and moisture transfer coefficients are gradually reduced by increasing the wind speed, while the DRAG COeffic CD increases. At wind speeds of less than 5 m s21, bulk formulas are not applicable as vertical turbulent transport between surface and air depends more on convective thermals than wind speed.9.1A Surface gusts that pass through a water surface create noticeable tiny capillary waves perpendicular to the coat of arms of the surface wind vector. Because capillary waves are short-lived, their distribution at a given time reflects the current distribution of surface wind. Remote sensing of capillary waves with satellite instruments, known as scatterometers, is based on observing surface winds over the oceans on a global scale. If they are forced by surface winds during periods ranging from hours to days, waves with different wavelengths and directions interact with each other to create a continuous spectrum of ocean waves that extend to a wavelength of hundreds of meters. The stronger and more durable the wind, the greater the longer the wavelength aamp. Wind waves with shorter wavelengths usually propagate in the same direction as the wind. In contrast, faster propagating long wavelengths tend to radiate outward regions as strong winds become swollen and thus the first sign of an approaching storm. The occurrence of a ripple increases with wind speed. At speeds exceeding 50 m s21, the ripple becomes so intense and extensive that the air-sea interface becomes diffuse and difficult to determine. Changes in the mass flow of air resistance (CD), heat (CH) and moisture (CE) at wind speeds above the ocean. [Adapted from the undissolved manuscripts of M. A. Bourassa and J. Wu (1996).] Copyright © 1996 Chapter 7 showed how wind can be predicted, taking into account the amount of total force acting in the air. Turbulent pull, as we have just discussed, is one of those forces that is always opposite to the wind direction (i.e. to slow down the wind). More importantly, we see in (9.19c) that the force of the pulling force is proportional to the space of the wind speed, thus doubling the wind speed quadrupling the drag. Through the flows at the bottom of this atmosphere, the characteristics of the underlying surface impressed me air within the atmospheric boundary layer, but not in the air in the overlying free atmosphere. Daily changes in river flows on land are spread byturbulence at the bottom of the boundary layer, resulting in diurnally variable vertical profiles as described in the next section. Take a column of air initially vertically uniform u cold above ground, capped with a very strong temperature inversion that prevents the boundary layer from growing. This air column advects at a speed U on a warmer ocean surface with potential temperatures for us. (a) How does the temperature change from a distance to shore x? (b) At any fixed distance away from the coast, how does the air temperature change from wind speed? [Tip: Use Taylor7 hypothesis: ∂θ/∂t =U∂θ/∂x.] If the only heat in the air column from the surface, then change of the time found in the heat budget Eq. (9.10) integrated with the depth of the boundary layer: ∂u/∂t = FHs / zi, where zi is constant. Combined with Taylor's hypothesis ∂θ/∂x=1UFHszi Then estimates the surface heat flux with bulk aerodynamic methods (Eq. 9.19a) and approximates (Ts - T) (we - u) gives separate variables and integrates: u = us - (us - u0) exp[-CH x/zi], where u0 is the initial potential temperature of air on land. Thus, the air temperature rises tailwind distance x from the coastline, quickly to the front, but gradually further down as the air temperature asymptotically approaching the sea surface temperature. (b) Surprisingly, the air temperature at a distance determined from the coast is independent of wind speed. This is because, while faster winds cause higher heat flows and faster warming of the boundary layer, faster winds also reduce the time available to warm up before the air reaches the distance x distance from shore. Kristina B. Katsaros, in Encyclopedia of Ocean Sciences (Third Edition), 2019 Estimating evaporation/laten heat flow from the ocean using satellite data also relies on the bulk formula. To calculate laten heat flow using the bulk aerodynamic method, sst, wind speed (U 10N) and humidity are required within the surface layer qa', as eq. (3). Therefore, the three variables need to be evaluated from space. Over the ocean, the U10N and SST were obtained directly from satellite data, but qa did not. In the 1980s (Liu, 1984), a method of estimating qa and laten heat flow from the ocean was first proposed in the 1980s using microwave radiometer data from satellites (Liu, 1984) and further developed (e.g. Bentamy et al., 2003). It is based on an empirical connection between the built-in water vapour W (measured in space microwave radiometers) and qa' monthly timescale. The physical reason is that the vertical distribution of water vapor across the depth of the atmosphere longer than a week. Weekly. relationship does not work well in the synoptic and shorter term, and in summer it also fails in some regions. Changes to this method by including additional geophysical parameters have been proposed with some overall improvement, but the inherent limitation is the lack of information on the vertical distribution of q to the surface. The two possible improvements in retrieval E include obtaining information on the vertical structure of the humidity distribution and the flow of direct contact between E and the brightness temperature (TB) for radiometers. The latest developments provide an algorithm for the direct retrieval of water vapour from the boundary layer of radiation observed by the special sensor microwave/imager (SSM/I) on satellites in the Defence Meteorological Satellite Program since 1987. This sensor has four frequencies, 19.35, 22, 37 and 85.5 GHz, except for the 22 GHz operates in both horizontal and vertical polarizations. The 22 GHz channel for vertical polarization is in the middle of a weak water vapor absorption line without saturation, even at high atmospheric humidity. Measurements are only possible over the oceans, as the ocean surface acts as a relatively uniform encyclical and background. On land, signals from the ground flood water vapor information. Since all three geophysical parameters, U 10N, W and SST, can be called up on a multi-channel microwave radiometer (SMMR) nimbus-7 (similar to the SSM/I) at frequencies measured by the older microwave radio meter launched in 1978 and operating until 1985, 10,6 and 6,6 GHz channels) and not 85 GHz channels) the feasibility of retrieving E from the measured values can also be obtained. The measures measured for the SMMR 10 channel, but only 6 channels, have been identified as significantly useful for the functioning microwave radio meter following SMMR, The Tropical Rainfall Measuring Mission (TRMM) launched in 1998 contained low-frequency SST-sensitive measurements and may therefore allow direct estimation of evaporation rates. Article 3(3) shall be replaced by the following Global distribution of monthly average laten heat flow in W m−2 in September 1987. Reproduced with the permission of Schulz, J., Meywerk, J., Ewald, S. and Schlüssel P. (1997). Assessment of latening heat flows by satellite. Journal of Climate 10, 2782-2795. polar orbiting satellites. Accurate random timing is not so important for the SST, as the SST changes slowly due to the high heat capacity of the water, and this method can only provide useful accuracy if averages are taken from 5 days to a week. Wind speeds can best be achieved from scattering, not microwave radio meters, regions of heavy cloud or rain, as scatterometers (which are active radars) penetrate clouds more efficiently. Scatterometers shall be used by the European Space Agency (ESA) and the Us National Aeronautical and Space Administration (NASA) (1991 and 1995 for the 1st World Aeronautical Satellites 1. And 2 – nasa scatterometer,NSCAT) launched it on a Japanese short-lived satellite in 1996 and the QuikSCAT satellite in 1999 (see, for example, Bentamy et al., 2003). In addition, microwave radiometers provide wind speed, which is sufficient (since vectors are not necessary to estimate evaporation), so the US Navy and NOAA's operational instrument, Windsat, provide these new additional data on a polar orbiting satellite (Gaiser et al., 2004).L. Jia, ... M. Menenti, comprehensive remote sensing, 2018The exchange of energy between the earth's surface and the atmosphere is a major determinant of processes in the atmospheric boundary layer and terrestrial ecosystems. It is extremely important to accurately determine the distribution of available energy to a reasonable thermal flue density (for heating or cooling the surface) and to the density of laten heat flow across a wide range of spatial and temporal scales. The sum of water leaving the soil surface and entering from soil, vegetation and water surfaces through evaporation, which is also called laten heat flow from the point of view of water, energy and biogeochemical cycles. Accurate estimates of land et are critical for applications such as global and regional climate change research, drought monitoring, water resource management and the sustainable development of agriculture. In recent decades, a number of global and regional ET estimation methods and surface models have been proposed and data products have been produced, and Earth observation is expected to provide continuous spatial and temporal information on earth's surface variables/parameters, which are essential for large-scale ET estimation. However, ET is still the component with the greatest problem with water cycle processes due to the complex control factors and heterogeneity of the landscape (Lettenmaier and Famiglietti, 2006; McCabe and his mts, 2015). Observation and modelling of turbulent heat flows on the surface has been a very active area of research, at least since Bowen's (1926) work on dry and wet surfaces (Monteith, Feddes, 1971; Verma and its mts, 1976; Hall and mtsai., 1979; Price, 1982; De Bruin and Jacobs, 1989; Beljaars and Holtslag, 1991; Lhomme and its mts, 1994). Most conventional techniques that use point measurements to estimate the components of the energy balance are specific to local scales and cannot be extended to large areas due to the heterogeneity of the land, dynamic nature and spatial distribution of heat transfer. Radiometric measurements taken away from evaporating bodies have been used for about 50 years to measure water use: qualitatively or quantitatively. Observable variables do not fully characterize the state of the evaporating body, and research in this field of science and applications has evolved along different lines focusing on the various biophysical processes affected by evaporation. This leads to signals available through remote sensing observations and alternative parameterization of processes. Some global actual evapotranspiration (ET) data products have been developed and evaluated. Remote sensing is the ability of large and frequent coverage to become a widely used and effective tool for tracking land and environmental changes. From radiometric data in the optical range, i.e. visible near infrared to thermal infrared, surface albed, fractionated vegetation cover and land surface temperature (LST) have been used for decades in many models dealing with hydrological, meteorological and ecological processes, which have been used from radiometric data from the optical range, i.e. visible infrared near infrared to thermal infrared. Following the principle of energy saving, the available energy, i.e. net radiation energy available on the surface of the soil, minus energy delivered to the soil, should be dissipated as a reasonable heat and latent heat flow if heat storage and photosynthesis energy are neglected. Theoretically close surface meaningful and laten heat streams can be determined by gradient measurements of wind speed, air temperature, and air humidity when these measurements are available. Algorithms for estimating reasonable and latent heat flows by remote sensing fall into two broad categories: analytical and (semi)empirical (see, inter alia, Menenti, 2012 and Menenti et al., 2012). All methods share a two-step approach:(1)Estimate the maximum and minimum flux density during the given force;(2)Estimate the actual flux density using a remote-sensing state variable to scale the maximum and actual values; The three biophysical processes of the Earth's system are significantly affected by land-based evacuation:–land surface energy balance (Norman et al., 1995; Kustas and al, 1996; Anderson and his mts, 1997, 2004; Bastiaanssen et al., 1998a,b; Su, 2002; Jia et 2003a,b; Jia, 2004; van der Kwast and his mtsai. mtsai. water balance (Miralles et al., 2011; Cui, 2015);–plant growth (Mu al al, 2007, 2011; Hu and Jia, 2015; Zheng and its mts, 2016). All three have been exploited to develop methods for estimating and monitoring land avoidance. The energy balance of land is always regarded as constraint, often formed by the Penman-Monteith combination equation or its simplified form (e.g. Miralles et al., 2011). Regardless of the biophysical process, the rate of evapotranspiration should be determined in the case of unlimited water availability, the rate of reduction below the maximum speed, due to the coercion of radiation and boundary layer. (1) Terrestrial surface energy balanceThe basic physics defining the principle of water use is energy saving on the evaporating surface. However, the density of the energy flow shall be parameterated by means of variables which can actually be caught by radiometric measurements. The relative amount of laten and reasonable heat flow (see, for example, Menenti, 2012) determines the surface temperature of the land. This causal relationship led to previous experimental studies (Fuchs and Tanner, 1966, 1968) and actual evapotranspiration (ET) by radiometric surface temperature measurement (see, for example, Jackson et al.) Initially, experiments were carried out on homogeneous patches and a direct link between evaporation and surface temperature was established. Over the next 20 years, methods were used to apply and extend such direct links to larger, heterogeneous areas (see, for example, Menenti, 2000). However, both field trials and remote sensing studies have led to the conclusion that the spatial heterogeneity of radiative coercion and boundary layer conditions had to be taken into account first in order to scale land evaporation with the surface temperature of the land (Menenti, 2000; Kalma and al, 2008). Under specific net radiation coercion and boundary layer conditions, observations of the earth's surface temperature fully characterize the earth's surface response to water, providing an effective means of determining actual land evaporation, as opposed to maximum. (2) Groundwater balanceA dynamic balance of water inflow and outflowdetermines the amount of water stored in the soil and deeper groundwater reservoirs. The feedback on the soil's water content in land evaporation is significant. For time scales from day to week, the groundwater tank determines the actual evaporation. Feedback mechanisms are fundamentally different when considering evaporation from bare soil or transpiration from plants. For simplicity, in both cases, a top 1-m-thick soil layer can be taken into account. The water content of the soil in this soil layer is characterized by the – as far as water evaporation and transpiration balances with the soil are concerned. Matter. content in this layer , after surface (excess energy) coercion and taking into account soil and plant properties (see, for example, Miralles et al., 2011). A stress factor based on the actual and maximum groundwater content of the root zone is used to parameterize the actual evaporation, ETact reduction, below the maximum ET ratio, with unlimited water availability. The total groundwater content in the 1m deep soil layer cannot be observed directly from a remote platform. However, continuous monitoring of the water content of the soil can be used in the topsoil to enforce the soil's water balance model and thus to estimate soil evaporation. (3) Plant growthPlant transpiration depends on the leaf-level energy balance and leaf area, and it is controlled by stomatal conductivity. The dependence of the latter on environmental conditions can be given in the form of semi-empirical relationships and can be used to parameterise the accumulation of water, carbon dioxide and biomass by photosynthesis (Jarvis, 1976). Mu et al. (2007, 2011) developed an algorithm to estimate the actual ET along these lines. The energy balance constraint is the penman-monteith combination equation. The reduction in actual evaporation below the maximum speed is determined by the dependence of stomatal and canopy driving on environmental conditions, in particular air temperature, vapour pressure deficiency and absorbed photosynthetic active radiation. The canopy conductivity scales the leaf area index. Photosynthesis both transpiration and photosynthesis. Photosynthesis.

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