ESTIMATION OF AREAL EVAPOTRANSPIRATION BY REMOTE SENSING AND GIS TECHNIQUES

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ABSTRACT

Remote Sensing data with ancillary ground-based meteorological data provides the capability of computing three of the four surface energy balance components(i.e. net radiation, soil heat flux and sensible heat flux) at different spatial and temporal scales. As a result, this enables the estimation of the remaining term, latent heat flux. One of the practical applications with this approach is to provide evapotranspiration maps over large areas. This results could estimate and reproduce areal evapotranspiration over large area as much as several hundred square kilometers. Moreover, some calculating simulations for the effects of the land use change on the surface heat flux has been mad by this method, which is able to estimate evapotranspiration under arbitrary presumed conditions.

From the simulation of land use change, the result suggests that the land use change in study area can be produce the significant changes in surface heat flux. This preliminary research suggests that the future research should involve development of methods to account for the variability of meteorological parameters brought about by changes in surface conditions and improvements in the modelling of sensible heat transfer across the surface-atmosphere interface for partial canopy conditions using remote sensing information.

1. Introduction

An important application of remote sensing information is in the evaluation of the energy and water budgets of natural and agricultural land surfaces. One of the main objectives of some recent hydrometeorological studies has been to test the feasibility of evaluating the surface energy and water balance of regional scales with models using satellite data. Because the surface energy and water budgets have important implications in modeling a wide range of geophysical processes, particular focus has been placed on obtaining reliable estimates of evapotranspiration at various spatial and temporal scales.

The magnitude of evapotranspiration has a broad range of applications in plant physiology and irrigation practices and to regional and global scale hydrology and meteorology. For many applications, accurate values of daily evapotranspiration are necessary from field science of order $10km^2$ to mesoscale of order $10km^2$. The evaporation of water from soil and plant surface is a component of the surface energy balances that is of both theoretical and practical interests. Conventional ground-based methods for estimating evapotranspiration, such as Bowen ratio, provide accurate measurements over a homogeneous area surrounding the instruments but the results are not applicable to large diverse areas.

Although in recent years some sophisticated models are being used in an operational mode, the amounts of information required and the fine tuning of some of the geophysical parameters may make it difficult to implement in area where there is limited meteorological, plant physical, and soils information. . The only visible means of mapping the spatial distribution of evapotranspiration on regional or local scales is using of remote sensing data.

Evapotranspiration is generally estimated by using thermal infrared data acquired by satellite based sensors and conventional ground-based meteorological data as inputs to a one-dimensional boundary layer model. Although remote sensing may provide information relating to evapotranspiration at the field scale(e.g., vegetation index, surface temperature, albedo), local meteorological conditions such as wind speed, air temperature, water vapor pressure etc., cannot be assessed with remote sensing. In addition, remote sensing data from satellite are essentially instantaneous and may only be available once a day. This has resulted in the development of methods that integrate instantaneous value of estimated evapotranspiration over the whole day.

Recently, the land use/land cover change is very intense in around urban due to concentration of population to urban. This land cover change gives large impact to existing process of water or heat in natural. In short, land cover change changes the amount of incoming energy on surfaces, the thermal environment and it affects energy balances of surfaces.

Because surface energy balances is related to the evapotranspiration through latent heat flux, the changed energy balances will effect the local water balances. The change of water balances over area means the ratio change of direct runoff, infiltration, evapotranspiration from precipitation. Decreasing evapotranspiration means decrease of escaping energy flux from surface as latent heat, and it induces the increase of local temperature like heat island(Kondoh, 1991).

The main objective of this research is to calculate the latent heat flux commonly called evapotranspiration and expressed in mm of water, over an area. In this paper, an attempt is made to calculate instantaneous and daily values of the latent heat flux by solving for LE as a residual in the surface energy balance equation 1) using GIS techniques, and simulating of the change of energy balance according to land use change such as land reclamation

2. Description of energy balance model and data collection

The surface energy balance equation is usually composed of four terms assuming that advection and storage of heat in the vegetated surface is negligible(Brutsaert, 1982):

$$Rn = LE + H + G \quad \text{------------------------} \quad 1)$$

where Rn is net radiation, H is the sensible heat flux, G is the soil heat flux and LE is the latent heat flux. The unit of all is W/m^2.

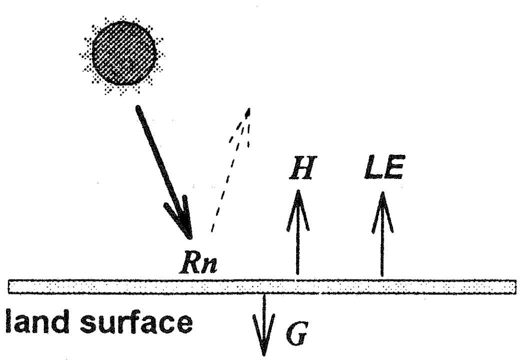

Figure 1. The concept of surface energy balance model.

Estimates of Rn and G either come from micrometeorological measurements or come from a combination of commonly available meteorological data and remote sensing data. Recent developments in the latter approach may provide estimates of Rn and G with primarily remotely sensed information(Jackson, 1985). From the equation 1), three component of the four surface energy balance components(i.e. net radiation, soil heat flux and sensible heat flux) were estimated with remote sensing data(LANDSAT TM) and ground-based meteorological data. As a result, this enables the estimation of the remaining term, latent heat flux. Table 1 summarized the used meteorological data and extracted data from remote sensing and topological data. Figure 2 depicts the general flow of data processing employed in this study.

Table 1. Used Data in this study

Remote Sensing data	albedo, α surface temperature, T_s [°C] vegetation index, NDVI
Meteorological data	air temperature, T_a [°C] water vapor pressure, e_a [mb] wind speed, u [m / s] cloudiness, n/N
Topological data	elevation, E[m] slope, β [degree] azimuth, h' [degree]

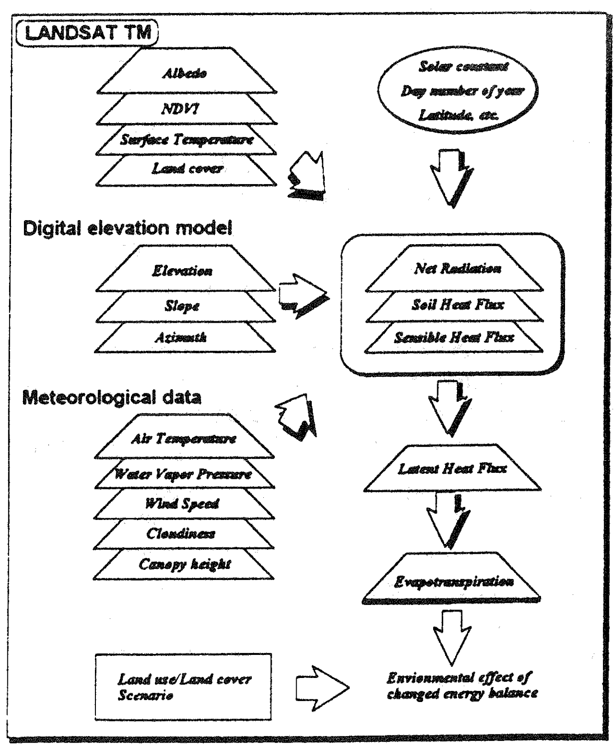

Figure 2. Data processing flow employed in this study.

Figure 3 shows the study area from the LANDSAT TM(Path/Row:116/34) which obtained on 15 Apr. 1986 and 22 Sept. 1992, respectively. From this figure, the large land cover change due to the land reclamation is visible to naked eye.

3. PARAMETERIZATION OF SURFACE ENERGY BALANCE COMPONENTS

3-1. Net Radiation, Rn

The net radiation at the land surface is the sum of incoming and outgoing spectral radiant fluxes integrated over all wavelengths and given by

$$Rn = (1 - \alpha) \cdot S\downarrow + L\downarrow - L\uparrow \quad \text{---------} 2)$$

where Rn is the net radiation(W/m^2), α is the surface albedo, S$\downarrow$ is incoming solar irradiance, L$\downarrow$ is incoming longwave radiation from atmosphere, L$\uparrow$ is the outgoing longwave radiation from the land surface.
Each components of equation 2) were computed as follows:

Shortwave radiation(W/m^2): For each grid, incoming shortwave radiation(0.15~4μm) was estimated from the cloudiness, digital elevation model, and solar geometry. In this procedure, the topographic effects were considered: direct irradiance(*Rsdir*), diffuse irradiamce(*Rsdif*) and reflected irradiance(*Rsref*). The total irradiance on a tilted surface can be estimated by integrating those three components as follows:

$$Rsi = (a + b \cdot n / N) \cdot (Rsdir + Rsdif + Rsref) \quad \text{------} 3)$$

$$Rsdir = Isc \cdot Eo \cdot Pt^m \sin h' \quad \text{---------} 4)$$

$$Rsdif = 0.5 \cdot Isc \cdot Eo \cdot \sin h \frac{1 - Pt^m}{1 - 1.4 \ln Pt} \cdot \frac{1 + \cos\beta}{2} \quad \text{-----} 5)$$

$$Rsref = \alpha\,[Isc \cdot Eo \cdot Pt^m \cdot \sin h' + 0.5 \cdot Isc \cdot Eo \cdot \sin h \frac{1 - Pt^m}{1 - 1.4 \ln Pt}] \cdot \frac{1 - \cos\beta}{2} \quad \text{----} 6)$$

where *Isc* is the solar constant, *Eo* is the eccentricity correction factor of earth, *h* is the solar altitude, *h'* is the altitude of the sun for a sloping surface, *Pt* is the atmospheric transmission coefficient, β is the slope angle of the surface, *m* is the relative optical air mass, *n/N* is the cloudiness(the relative duration of sunshine).

Albedo: Albedo was calculated using spectral reflectance factors for representative bands in the visible and near infrared bands of LANDSAT TM(Brest and Goward, 1987). For vegetated surfaces Brest and Goward calculated α as

$$\alpha = 0.526\rho(TM2) + 0.362\rho(TM4) + 0.112\rho(TM7) \quad \text{--} 7)$$

for non-vegetated surfaces

$$\alpha = 0.526\rho(TM2) + 0.474\rho(TM4) \quad \text{---------} 8)$$

where ρ(TM) is the reflectance of each TM band.

Longwave radiation(W/m^2): Incoming and outgoing longwave radiation(> 4μm)also was calculated using air temperature, surface temperature which extracted from TM band 6, and vapor pressure data in the following formula(Satterlund, 1979)

incomming longwave radiation:

$$L\downarrow = 1.08\,[1 - exp(-e_a\,T_a / 2016)]\,\sigma \cdot T_a^4 \quad \text{---} 9)$$

outgoing longwave radiation:

$$L\uparrow = \varepsilon_S \cdot \sigma \cdot T_S^4 \quad \text{---------} 10)$$

where ε_S is the emissivity of the surface, σ is the Stefan-Boltzman constant(= $5.67 \cdot 10^{-8}$ W m^{-2} K^{-4}), and T_s is the surface temperature(°K). T_a is the air temperature(° K) and e_a is water vapor pressure(mb), respectively.

3-2. Sensible Heat Flux, H

The one-dimensional bulk resistance equation computes the sensible heat flux(H) with the surface(T_s)-air temperature difference(T_a) and windspeed(u), i.e.,

$$H = \frac{\rho C_p (T_s - T_a)}{r_a} \quad \text{---------} 11)$$

where ρ is the density of air(kg m^{-3}), C_p is the specific heat of air at constant pressure(=1004.7 J kg^{-1} K^{-1}), r_a is the aerodynamic resistance(sm^{-1}). r_a can be computed using surface roughness(z_o; m), displacement height (d; m) as follows:

$$r_a = \frac{ln[(z - d + z_o) / z_o]^2}{\kappa^2\,u} \quad \text{-------} 12)$$

where κ is the von-Karaman's constant(= 0.4), Z is the measurement height(m). The surface roughness and displacement height were can be estimated using canopy height(Brutsaert, 1982).

3-3. Soil Heat Flux, G

The empirical data has shown that there is a reasonable relationship between the ratio G / Rn and normalized

a) 15 April, 1986

b) 22 September, 1992

Figure3. LANDSAT TM images of study area acquired on 14 Apr. 1986 and 22 Sept. 1992.

difference vegetation index(Kustas and Daughtry, 1990). The empirical equation estimating soil heat flux can be expressed as below:

$$G = (0.325 - 0.208 \frac{TM4 - TM3}{TM4 + TM3}) \cdot Rn \text{ --------13)}$$

3-4. Compute of instantaneous evapotranspiration, ET_i

From the above energy balance equation, the instantaneous evapotranspiration(mm / hour), ET_i can be computed as follows:

$$ET_i = 3600 \cdot \frac{Rn - G - H}{L} \text{ - ------------------------14)}$$

where L is the latent heat of vaporization $(= [2501 - 2.37\, T_a] \cdot 10^3; \mathrm{J/kg})$.

4. COMPUTE OF DAILY EVAPOTRANSPIRATION

As above described, the instantaneous evapotranspiration, ET_i can be calculated from LANDSAT TM data at one time of day and in conjunction with other meteorological data. The daily change of radiation and evapotranspiration in the field assume a sine form(Jackson and Hatfield, 1983).

$$ET_i = ET_{max} \cdot sin(\pi t / N) \text{ ------------------------ 15)}$$

where ET_{max} is the maximum evapotranspiration of a day, N is the day length between sunrise and sunset, t is the time of LANDSAT overpass after sunrise. Therefore, the daily evapotranspiration(mm/day) was computed as follows:

$$ET_d = \int_0^N ET_i = \int_0^N ET_{max} \cdot sin(\pi t / N) \text{ -----------16)}$$

$$= 2N \cdot ET_{max} / \pi = 2N \cdot ET_i / \pi \cdot sin(\pi t / N)$$

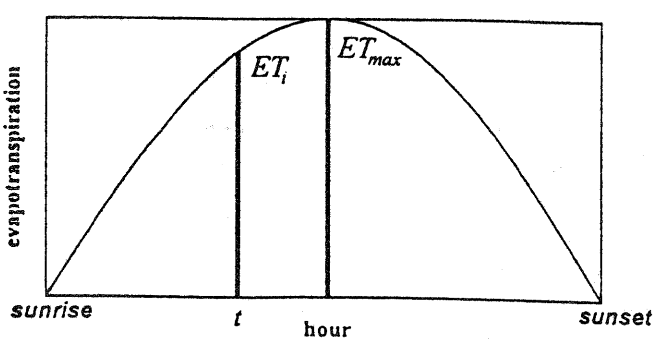

Figure 4. Graphic representation of the estimation of daily evapotranspiration from instantaneous evapotranspiration.

5. RESULTS AND DISCUSSION

Table 2 summarized daily averaged surface energy balance at each land cover and composition ratio of land cover at study area from LANDSAT TM.

Table 2 Daily averaged surface energy balance at each land cover

land cover	ratio(%)	Rn(Wm^{-2})	H(Wm^{-2})	LE(Wm^{-2})
tidal flat	20	513	155	333
vegetated	39	506	290	191
nonvegetated	41	470	145	83

Figure 4 shows the estimated evapotranspiration according to method of chapter 3 and 4. The latent heat flux which equal to evapotranspiration in water balance is greater on the tidal flat where contains lots of water in the soil. nonvegetated surfaces such as urban and reclamated area shows the half amount of latent heat rather than vegetated surfaces(Table 2). In nonvegetated areas, there is a little amount of evapotranspiration, but vegetated area especially in forest, shows the high evapotranspiration rate (> 2mm/day).

According to the national development program, in the study area, west coasts of Korea, the land cover/land use change is very severe. This situation can be seen in Figure 3. The tidal flat(or intertidal zone) was reclamated to use of industrial complex, construction of new residence section.

As a result, those land use/land cover changes will induce the potential change of thermal environment. For example, when existing tidal flat will reclamated, and changes to non vegetated land surface, the effect on surface energy balance become:

ΔRn = - 43$\cdot$ reclamated surface area

ΔH = - 10$\cdot$ reclamated surface area

ΔLE = -250$\cdot$ reclamated surface area

As known this results, the change of latent heat flux due to change of land use/land cover is greater than that of net radiation and sensible heat flux. Decreasing net radiation(increase of urban area, decrease of tidal flat or vegetated surfaces) due to increase of surface albedo works on lowering surface temperature. In the other hand, decrease of latent heat which is greater at least six times than decreasing net radiation and sensible heat flux works on highering surface temperature.

In this study, the author estimates of the evapotranspiration using LANDSAT TM data with ancillary ground-based meteorological data and topolo-

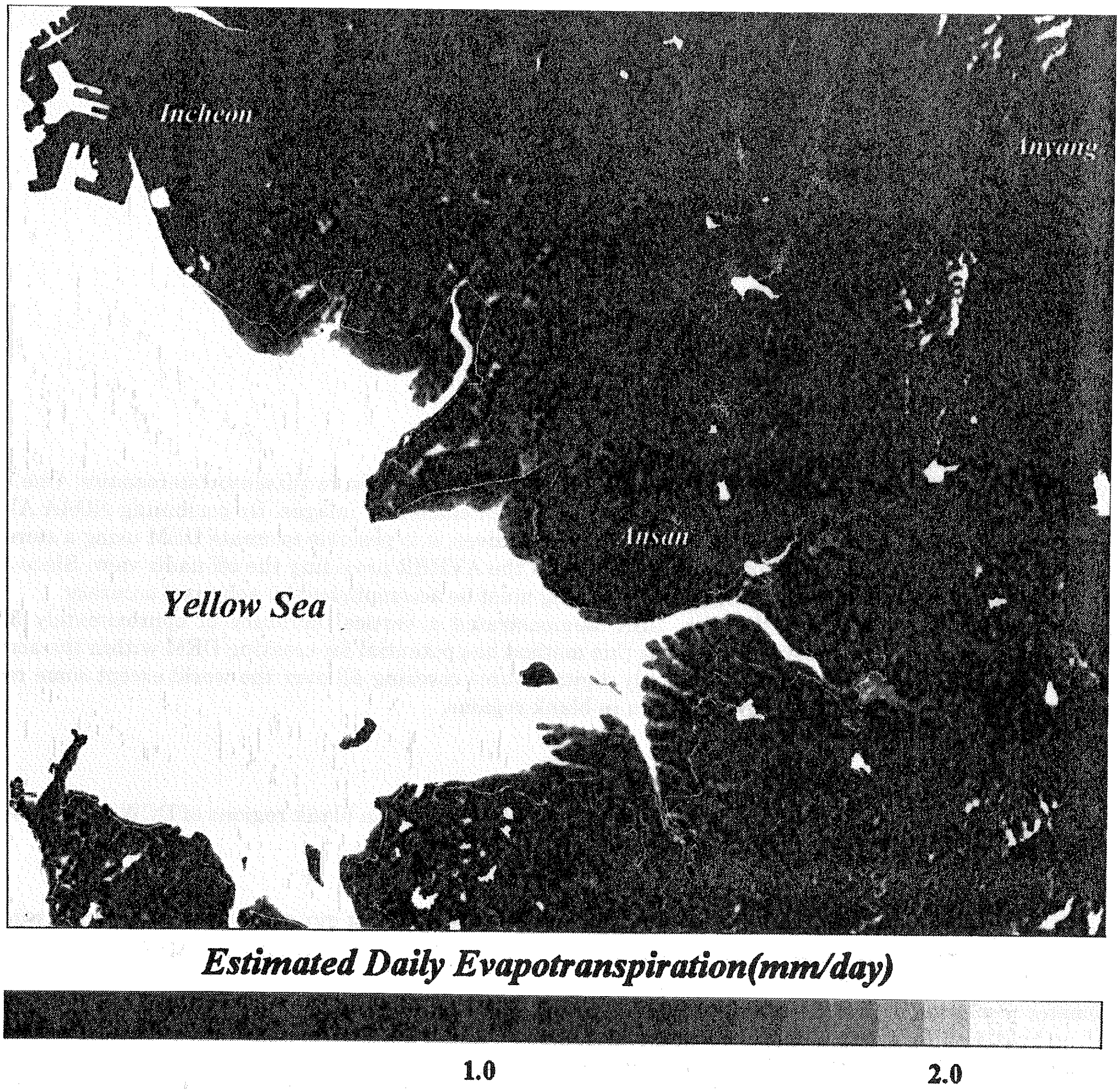

Figure 5. Estimated daily evapotranspiration from LAN DSAT TM with ancillary meteorological data.

gical data. The surface parameters such as albedo, vegetation index, surface temperature, was calculated deom LANDSAT TM data. Unfortunately, the comparision of estimated results with measurements could not accompolished. To evaluate the accuracy of proposed method and employed empirical formulations, collection of a number of in-situ data in test area and further research will be required.

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