# Surface Temperature Decrease, NDVI and Humidity Deficit, and Latent Heat Flux in Forested Region from TM and Routine Meteorological Data – a Summary –

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**Abstract**

A new method for estimating directly the latent heat flux in regional forested area from remotely sensed data and routine data measured at a meteorological observatory is proposed by applyng the similarity theory of Monin-Obukhov. The key idea (Eq.19) of the estimation method is to express the specific humidity deficit at forest canopies by an analogy with the relation derived above between the temperature decrease, NDVI and the specific humidity deficit in terms of the newly defined evapotranspiration indices.

## 1. INTRODUCTION

This paper attempts to derive a prediction relationship of the surface temperature decrease in forested area,and then to develop a method to estimate the latent heat flux directly from remotely sensed satellite data, applying the Monin-Obukhov theory on the atmospheric boundary layer flow. The present paper developes the augmented summary of our method presented in the previous preliminary papers (Kaneko and Hino (1994)).

As the evapotranspiration increases, the surface temperature falls by the effect of latent heat flux. Hall et al. (1991) showed that the canopy temperatures estimated from remote sensing correlated well with the measured canopy temperature. Nemani and Running (1989a) related the surface temperature to NDVI and several researchers has discussed this $T_s$/NDVI relationship so far (Goward and Hope (1989), Larsson (1993)).

It is shown in this paper that the surface temperature decrease over the forested area is represented as a function of the surface (leaf) temperature, NDVI and the specific humidity deficit at the leaf temperature, these factors being estimated by TM data.

In the second part of this paper, a method of remote sensing direct estimation of latent heat flux in regional area of forests is proposed applying the Monin-Obukhov similarity theory along with the three factors; leaf temperature fall due to transpiration estimated from TM, the value of NDVI derived also from TM and routine meteorological data obtained at an observatory.

Most methods of remote-sensing estimation of the latent heat flux reported so far are, to the authors' knowledge, the indirect method which estimates the latent heat flux by subtracting from the net radiation the sensible heat flux evaluated by remote sensing (for instance, Carlsonet al.(1981), Seguin and Itier (1983), Pierce and Congalton (1988), Moran et al.(1989)). Sugita and Brutsaert (1992) recently calculated the latent heat fluxes by indirect estimation, showing that the estimated values were generally in good agreement with the mean fluxes obtained by field measurement. Taconet, Bernard and Vidal-Madjar (1986) developed a methodology to predict surface fluxes by a one-dimensional boundary layer/vegetation/soil model over dense vegetation, applying surface temperature obtained by remotely sensed data.

Generally speaking, the method of direct estimation of latent heat flux is essentially superior to the indirect one because the latter method calculates the latent heat flux as residual and is apt to include undesirable errors.

Since the values of forest biomass are not uniform seasonally and spatially, its effect must be considered in the regional estimation of latent heat flux. The authors' method takes the forest biomass into account by defining an effective leaf area index (LAI).

The authors' present method combines the technique of remote sensing to estimate the forest evapotranspiration with the well-known similarity theory of Monin-Obukhov.

## 2. SURFACE TEMPERATURE FALL BY FOREST EVAPOTRANSPIRATION

### 2.1 Land Use in the Test Area

Fig.1 shows the horizontal distribution of land uses in the test region, Shimane Prefecture which is situated in the western part of Japan. The land-use data are derived from Digital Land Information supplied from the Geographical Survey Institute, Ministry of Construction.

Matsue City, at about the center of the area is the administrative center of the Prefecture, with population of 142,000, and is surrounded by dense forests as shown in Fig.1 by symbol of tree.

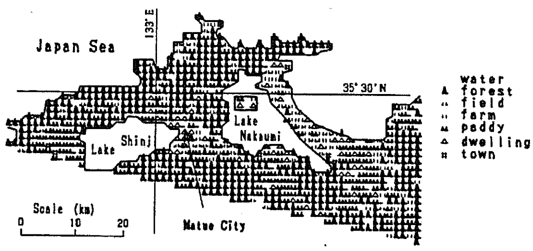



Figure 1. Land use of the test region in Shimane Prefecture.

### 2.2 Relation between Surface Temperature and NDVI

The surface temperature of forested regional area decreases almost linealy with the increase of NDVI by the effect of latent heat flux caused by evapotranspiration of forest. Fig.2 shows the relation between the surface temperature and NDVI for the whole areas of Shimane Prefecture with a variety of land-uses.

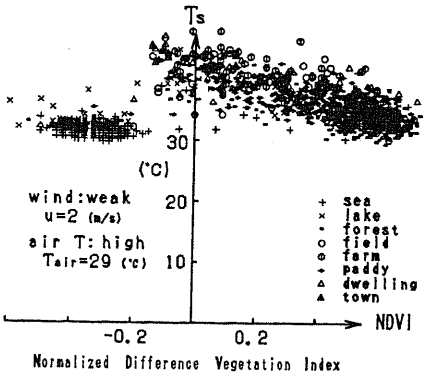



Figure 2. Relation between the surface temperature derived from Landsat TM and NDVI for all kinds of land-use.

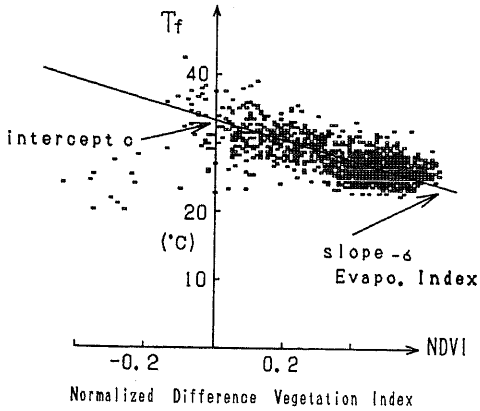



Figure 3. Definition of the normalized evapotranspiration index on the scene of 9 May 1990.

As Nemani and Running (1989a) reported firstly, the surface temperature is related to NDVI by Eq.(1),

$$T_f = \sigma \cdot NDVI + c \qquad (1)$$

where $T_f$ is the surface temperature at forest sites, $\sigma$ a proportionality constant, and c a constant.

## 2.3 Evapotranspiration Index $(-\sigma)$

The intercept c on the ordinate (temperature axis) in Fig.3 represents the surface temperature at the sites where vegetation is scarce. The temperature decrease $(c-T_f)$ is due to evapotranspiration.
The rate of temperature decrease

$$-\sigma = \frac{c-T_f}{NDVI} \qquad (1a)$$

can be defined as the normalized evapotranspiration index to indicate the degree of the evapotranspiration effect of vegetation (Kaneko and Hino (1993)). The gradient of the relation changes seasonally, showing that the evapotranspiration is affected by the variation of meteorological as well as plantphysiological conditions.

It is soon conceivable that the evapotranspiration is effected not only by the vegetation biomass(NDVI) but also by such factors especially as the leaf temperature and specific humidity deficit. We have examined a variety of combinations of these factors with the mean values of $-\sigma$ for each scene. A candidate for the most appropriate factor explaining $-\sigma$ was $(q_c-q_a)$;i.e. the specific humidity deficit between the saturation specific humidity $q_c(T_f)$ at the leaf temperature $T_f$, and the humidity of air flow $q_a$. The value $q_c$ means the humidity inside the stomata.

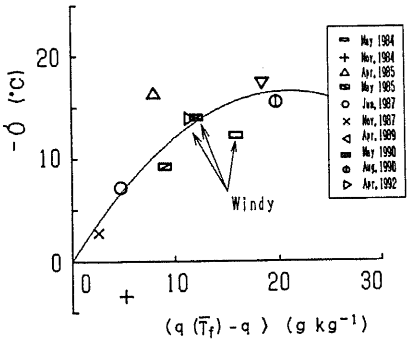



Figure 4. Relation between the normalized evapotranspiration index $-\sigma$ and the specific humidity deficit $(q_c(\overline{T_f})-q_a)$ between the saturation specific humidity $q_c$ at average leaf temperature $\overline{T_f}$ and the value q of air flow.

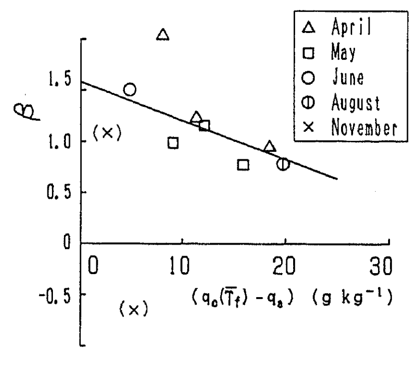



Figure 5. Relation between the stomatal opening index $\beta$ and the specific humidity deficit $(q_c(\overline{T_f})-q_a)$.

The index $-\sigma$ increases initially with the increase in the humidity deficit (Fig.4). While as the value of $(q_c(T_f)-q_a)$ approaches near 20(g $kg^{-1}$), the index $-\sigma$ saturates and begins to decline with the increase of the deficit $(q_c(T_f)-q_a)$. As a first approximation, the relationship is described by Eq.(2).

$$-\sigma = \beta \cdot [q_c(T_f) - q_a] \tag{2}$$

## 2.4 Stomatal Opening Index ($\beta_s$)

As the forest transpiration may be proportional to the specific humidity deficit $(q_c-q_a)$ as well as the stomatal opening, the proportionality constant $\beta$ in Eq.(2) may define a stomatal opening index. In other words, if the surrounding air becomes excessively dry, the plants would close the stomatal opening.

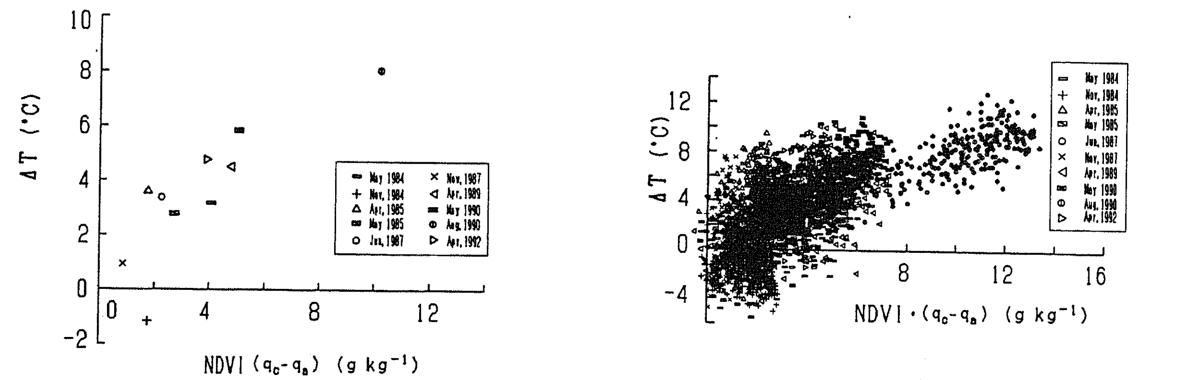
Data of the index $\beta$ in Eq.(2) obtained from ten TM scenes at different seasons show that the index $\beta$ decreases, as shown in Fig.5, linealy with the increase of the specific humidity deficit $(q_c-q_a)$ except for the winter season of November when the evapotranspiration of leaves ceases completely. The tendency may be explained by the fact that the large values of the deficit $(q_c-q_a)$ cause the stomatal closure to restrain the transpiration against excessive loss of water from plant. From the regression analysis, the index $\beta$ is expressed by Eq.(3).

$$\beta = -0.0378 \cdot (q_c - q_a) + 1.58 \tag{3}$$

The correlation coefficient for the present analysis for a wider area is r=0.731. In narrower area already reported (Kaneko and Hino (1994)), the value 0.71 of correlation coefficient increases to 0.91. The two constants of the regression formula (3) depended slightly on the places where vegetation and geology are different.

The index of stomatal opening $\beta_s$ is defined so that its value is limited within a range of $(0 \leqq \beta_s \leqq 1)$, as

$$\beta_s = \frac{\beta}{1.58} = 1 - \frac{0.0378}{1.58}(q_c(T_f)-q_a) \tag{4}$$




(a) Relation between the temperature decrease $\Delta T$ $(=c-\overline{T_f})$ and the average values of $NDVI \cdot (q_c(\overline{T_f})-q_a)$. The ten data especially the case on 18 June 1987 of the low leaf temperature approach on a single curved line.

(b) Relation between the temperature decrease $\Delta T$ and the values of $NDVI \cdot (q_c(\overline{T_f})-q_a)$ at all pixels of 10 TM scenes.

Figure 6. Relation between the temperature decrease $\Delta T$ and the values of $NDVI \cdot (q_c(\overline{T_f})-q_a)$.

## 2.5 Formulas on $-\sigma$ and $T_f$

To summarize, the evapotranspiration index $(-\sigma)$ and the surface temperature $T_f$ are expressed, respectively, by

$$-\sigma = [1.58-0.0378(q_c(T_f)-q_a)](q_c(T_f)-q_a) \quad (5)$$

$$T_f = c-[1.58-0.0378(q_c-q_a)](q_c-q_a)\cdot NDVI \quad (6)$$

$$\Delta T = \beta(q_c-q_a)\cdot NDVI \quad (7)$$

where $\Delta T(=c-T_f)$ is the temperature decrease.

## 3. ESTIMATION OF LATENT HEAT FLUX

### 3.1 Basic Equation

In this section, a new method for the estimation of latent heat flux in forested areas is developed based on the the above discussions and the Monin-Obukhov similarity theory. The idea has been presented in Kaneko and Hino (1994).

According to the Monin-Obukhov theory (originally Monin and Obukhov (1954), for reference at hand see for instance, Lumley and Panofsky (1964), Raupach and Thom (1981)), the latent heat flux is expressed as

$$E = -\rho l U_* q_* \quad (8)$$

where $\rho$ is the density of air, $U_*$ means the friction velocity, l is the latent heat of evaporation, $q_*$ is the friction specific humidity, defined by Eq.(9) as

$$q_* = -\overline{wq}/U_* \quad (9)$$

where $\overline{wq}$ is the specific humidity flux.

Integral of the universal function $\phi_h$ in the Monin-Obukhov theory yields the following equation between $q_*$, $q_a$ and $q_f$ as

$$q_a - q_f = \frac{q_*}{\kappa}[\Psi_h(\zeta)-\Psi_h(\zeta_0)] \quad (10)$$

where

$$\Psi_h = \int_{\zeta_0}^{\zeta} \frac{\phi_h(\zeta)}{\zeta} d\zeta, \quad \zeta = (z-d_0)/L, \quad \zeta_0 = z_0/L \quad (10a)$$

where z is the height at which the values of wind speed $U_a$, atmospheric temperature $T_a$, and specific humidity $q_a$ are measured; $z_0$ the roughness length, $d_0$ the zero-plane displacement, $\zeta$ the dimensionless height, and $\kappa$ the von Karman constant, $q_f$ the specific humidity on ground surface and/or over forestcanopy. As far as the ground surface is not wet, $q_f < q_c$.

The wind speed and the temperature are written in terms of the integral universal functions $\Psi$ as

$$U_a = \frac{U_*}{\kappa}[\Psi_m(\zeta)-\Psi_m(\zeta_0)], \quad (11)$$

$$T_a - T_f = \frac{\theta_*}{\kappa}[\Psi_h(\zeta)-\Psi_h(\zeta_0)]. \quad (12)$$

In Eq.(10a), L is the Monin-Obukhov length defined by the following form,

$$L = \frac{\overline{\theta} U_*^2}{\kappa g \theta_*} \quad (13)$$

where the friction temperature $\theta_*$ is defined by

$\theta_* = -H/U_*$ (14)

where H is the sensible heat flux = $\overline{w\theta}$ , and g the acceleration of gravity. The value of $\overline{\theta}$, the representative potential temperature, is defined by the following expression so as to include the convection effect of the average surface temperature $\overline{T_s}$ in surroundings into the leaf temperature $T_f$ of the forest site concerned,

$$\overline{\theta} = (\overline{T_s} + T_f)/2 \quad (15)$$

The surface temperature $T_f$ is derived from Landsat TM. The temperature $\overline{T_s}$ is determined either also from Landsat TM or an iterative computation to be explained subsequently.

The functional form of $\phi$ proposed by Businger (1988) and Dyer (1974) was applied.

## 3.2 Estimation of atmospheric stability $\zeta$ by TM and routine meteorological data

Equations (10a),(11), and (13) lead to the following well known relationship,

$$\zeta = R_i \, \phi_m^2(\zeta)/\phi_h(\zeta) \quad (16)$$

where $R_i$ means the Richardson number which is approximated by the bulk Richardson number B,

$$B = \frac{g(T_a - T_f)}{\overline{\theta} \; U_a^2} (z - d_0) \; . \quad (17)$$

## 3.2 Meteorological Data Available

The routine meteorological data are measured at the Matsue meteorological observatory; i.e. the air temperature and the humidity at the height of 1.5m, and the wind speed $U_a$ at z = 26.7m from the ground.

Since the area concerned is relatively flat, the atmospheric stability condition is unstable and the air masses are well mixed by forced thermal convection. The same meteorological conditions are expected to prevail at the higher level in the mixing (surface boundary) layer over the study area. Moreover, nearly homogeneques land use spreads widely as shown in Fig.1. Consequently, the similarity theory of Monin-Obukhov in the surface boundary layer can be applied with the common use of $U_a$, $T_a$, and $q_a$, over the test area. The value of B which stands for $R_i$ is calculated from TM and routine meteorological data. Consequently the value of $\zeta$ can be deduced from Eqs.(16) and (17).

## 3.4 Definition of Effective Leaf Area Index ($\alpha_{NDVI}$)

The values of NDVI have already recognized by Nemani and Running (1989b) that those have a logarithmic relationship with leaf area index (LAI).

In urban areas where no vegetation exists, the value of NDVI approaches a limit of -0.1. On the other hand, in densely forested areas, the surface temperature scarcely falls with the increase of NDVI, after the value of NDVI reaches a value of 0.5 and NDVI becomes nearly constant in dense forests. Consequently, an effective leaf area index is defined as follows,

$$\alpha_{NDVI} = \begin{cases} 0 & (NDVI < NDVI_0) \\ \dfrac{NDVI - NDVI_0}{NDVI_{100} - NDVI_0} & (NDVI_0 \leqq NDVI \leqq NDVI_{100}) \\ 1 & (NDVI > NDVI_{100}) \end{cases} \quad (18)$$

where $NDVI_0$ takes a value of -0.1 for no vegetation areas, and $NDVI_{100}$ is given a value of 0.5

in dense forests.

From the definition of NDVI, the meaning of the word "effective" reflects not only the leaf area density in a unit pixel but also the richness of chlorophyll in forest leaves which are actively transpirating.

### 3.5 Expression of Specific Humidity Deficit on Ground Surface

Considering the facts that the profile of q(specific humidity) over the bare soil surface is approximately constant because both of nearly zero vapour flux from the ground and of violent turbulent mixing, and that the value of q over the upper part ($z=z_a$) of the mixing layer is nearly constant (i.e. $q_{a.bare} = q_{a.bare} = q_{a.forest}$), and that the humidity deficit at the forest canopy ($q_f - q_a$) is proportional to the evaporation flux from it derived in the preceding section, the authors propose the relationship of Eq.(19) from the analogy with Eq.(7).

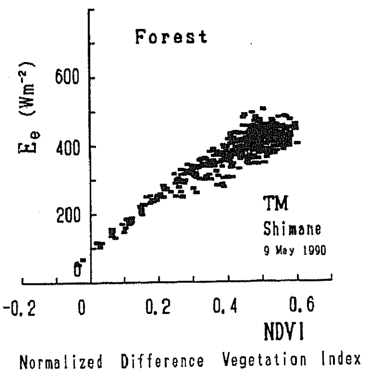



Figure 7. Relation between NDVI and latent heat flux $E_e$, estimated by means of Eq.(20) together with Eqs.(4),(10), (16) and (18).

$$q_f - q_a = a_c \cdot \alpha_{NDVI} \cdot \beta_S (q_c(T_f) - q_a) \tag{19}$$

where $\alpha_{NDVI}$ is the index of effective leaf area, $\beta_S$ means the index of stomatal opening (Eq.(4)), and $a_c$ is the conversion coefficient which is selected so that the coefficient makes the latent heat flux E at dense forests (NDVI>0.5) equal to the net radiation.

### 3.6 Estimation Formula of the Latent Heat Flux

Insertion of Eqs. (10),(11) and (19) into Eq.(8) yields the final expression for the latent heat flux as Eq.(20)

$$E_e = -\frac{\rho l \kappa^2}{[\Psi_m(\xi)-\Psi_m(\xi_0)]\cdot[\Psi_h(\xi)-\Psi_h(\xi_0)]} \cdot U(z)\cdot(a_c\,\alpha_{NDVI}\,\beta_S)\cdot(q_c - q_a). \tag{20}$$

In other words, $E_e$ is a function of the atmospheric stability L or $\xi$, the wind speed $U_a$, the effective leaf area index $\alpha_{NDVI}$, the humidity deficit ($q_c - q_a$), and the stomatal opening $\beta_S$,

$$E_e = \mathrm{fun}\{U_a, L(\xi), \alpha_{NDVI}, \beta_S, (q_c - q_a)\} \tag{21}$$

where $\alpha_{NDVI}$ is a function of NDVI and $\beta_S$ which is given by Eq.(4) depends on ($q_c - q_a$) as well as $T_f$ (leaf temperature) through $q_c(T_f)$.

## 4. CONCLUSION

The surface temperature falls with the increase of NDVI and it is also affected by the humidity deficit between the saturation humidity at leaf temperature and that of air flow. The relation is derived as Eq.(6) from the analysis of TM data. Since $q_c$ is a function of $T_f$, and $\Delta T = \beta(q_c - q_a)\cdot NDVI$, solution of Eq.(6) will give explicitly the value of $T_f$ with parameters $q_a$ and NDVI. The decrease in the surface temperature in forested areas reflects the effects of evapotranspiration.

Based on the above result, a new method for estimating the latent heat flux over forested

Based on the above result, a new method for estimating the latent heat flux over forested areas from the surface temperature data derived by the remote sensing technique as well as routine data observed at a meteorological observatory has been proposed by applying the Monin-Obukhov similarity theory. The latent heat flux is shown to depend on such factors as the value of NDVI, the specific humidity deficit, the leaf temperature which effects the stomatal opening for forest to control transpiration.

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