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ESTIMATING EVAPORATION FROM LAKE NAIVASHA KENYA, USING REMOTELY SENSED LANDSAT THEMATIC MAPPER (TM) SPECTRAL DATA

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ABSTRACT: Evaporation from the lake Naivasha. Kenya is a major component of water balance. This study examines the evaporation losses from this lake by the energy balance approach. Landsat Thematic Mapper (TM) spectral data have been used. From the satellite images, negligible spatial variation of evaporation surface temperature and albedo was found in the lake, For daily estimation of evaporation from instantaneous value, evaporative fraction approach was used considering the sensible heat flux, H=0 above the lake surface and for daily net logwave radiation calculations, a simple equation was followed. For this study TM images of 21 st January 1995 were used and the daily total evaporation from the lake was found 5.95 mm. while from pan data the estimated average evaporation was 5.46 mm for 21st January with a standard deviation of 1.28 mm for the period of 1957-1990

KEYWORD : Evaporation, Latent heat flux, Sensible heat flue, Water heat flux, Radiation, Radiance, Reflectance, Albedo, Emissivity, Transmittance.

INTRODUCTION

Lake Naivasha is one of the important lakes of Kenya, Over the last fifteen to twenty years, there has been tremendous agricultural and geothermal power development based on the extraction of water from the lake. The lake is the major source for irrigation, domestic and industrial water supply, and also a water resource for wildlife. It is very important to estimate evaporative losses from the lake in order to calculate the availability of fresh water for the agricultural, domestic and industrial sector.

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Evaporation losses are normally calculated by the empirical methods like pan evaporation, evapo-meters but these are not necessarily correct and also these data are not always available. Therefore a physically based solution i.e. energy balance method has been used in this study for estimating lake evaporation using Thematic Mapper spectral data. Also this method is considered to be the most accurate method for estimating lake evaporation by many researchers and generally used as a reference method (Keijman, 1974; Stewart and Rose, 1976; Keijman, 1981; de Bruin, 1982; Robertson and Berry, 1985; Simons and Mero, 1985; Sturrock et. al. 1992 Assouline, et. al. 1993; Rosenberry et. al. 1993: Choudhury. B.J. 1994: Reis and Dias, 1998).

STUDY AREA

The lake Naivasha is located in the southwest of Kenya, at 0°45. S and 36°20' E, 80 kms south of the equator and 70 kms northwest of Nairobi the capital of Kenya (Figure. 1). It is situated in the bottom of the eastern or Gregory Rift Valley (1885m above m.s l) which is stretched from Jordan in the Middle East to Mozambique in southeast Africa. The lake has the highest attitude and considered to be the freshest of all Rift Valley Lakes in eastern Africa. In the most resent history the lake has shown tremendous change in depth, area and volume. From 1909 to 1969 the lake's area has varied from 216.27 to 88.08 Km^2. Although depth and area changes with season, average surface area of lake Naivasha is 154 km^2 and the mean depth is 5m.

The climate of the region is semi-arid but locally the climate in the valley varies due to variation in attitude. Mean monthly temperature ranges from 15.9 to 17.8° C with the highest temperatures in January and February and lowest in July and August. Rainfall is bimodal with main pulses in April/May and again in November. The annual average rainfall in the lake area for the period 1931-1960 was 608 mm with a variation round the mean from 443 to 939 mm (East African Meteorological Dept. 1966). Winds over the Lake Naivasha are generally weak and come from varying directions in the mornings. In the afternoon, winds of 1-2.5 m /sec are typical. Winds are strongest in August to October when they reach a speed of 6 m/sec. Monthly means of daily insolation at the surface (Ndabibi 0°45'S, 36°15'E) vary from 150 W/m^2 to 250 W/m^2 (during 1998-99). Highest levels were in December - January and lowest in May-June.

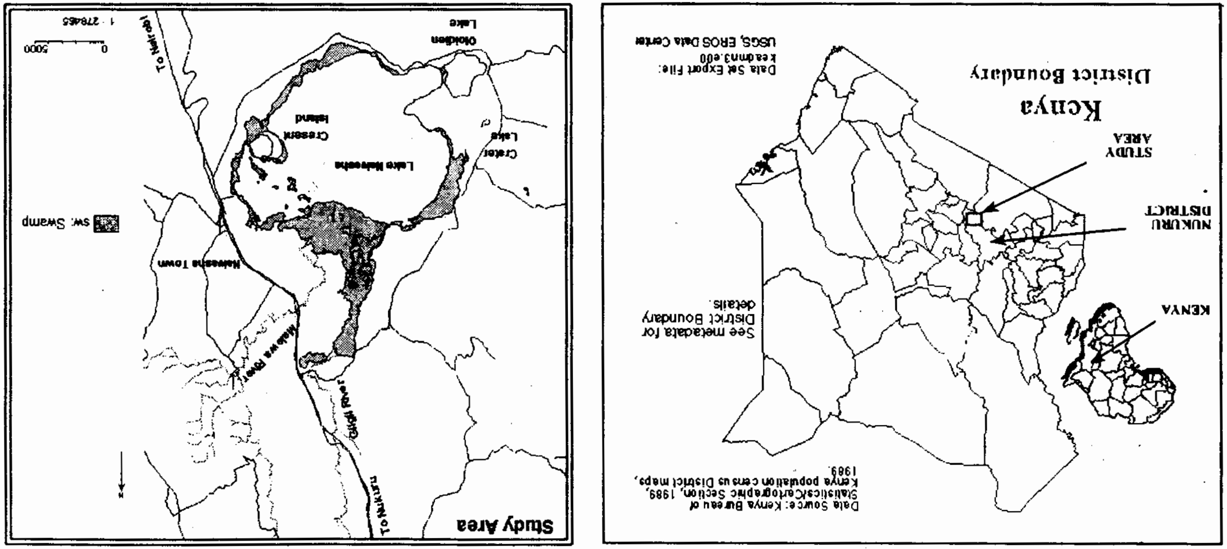

Fig 1. Location Map

RESIDUAL METHOD

Instantaneous estimation

In the energy balance approach, latent heat flux, λE or LE can be calculated as residual according to the following equation:

$$\lambda E = R_n - G_o - H \; [W.m^{-2}] \quad (1)$$

Where λE is latent heat flux [W. m^{-2}] required for evaporation (positive during evaporation) E is evaporation and λ is the latent heat of vaporization (2.45×10^6 J.Kg^{-1} at 23°C),R_n is net radiation flux [W.m^{-2}] at the water surface, G_o is water heat flux [W. m^{-2}] (positive if water is warming) and *H* is sensible heat flux [W.m^{-2}] from water surface to air (positive of air is warming). Many studies have shown that daily net radiation is closely related to daily rate of evaporation from shallow water bodies, especially in warm to hot sub-humid and humid climates (Allen. et al., 1996). The net rediation is made up of the following components:

$$R_n = K_n + L_n = K\downarrow - K\uparrow + L\downarrow - L\uparrow \; [W.m^{-2}] \quad (2)$$

Where K is shortwave (visible radiation, 0.3 and 3.0 μm) and L is long wave radiation (thermal radiation, >3.0 μm) and the arrow ↓ denotes incoming and the arrow ↑ outgoing and the subscript 'n' stands for the net radiation. The R_n can be further decomposed into its constituent parameters:

$$R_n = [(1-r_o)\, K\downarrow] + [\{\sigma\varepsilon_a.T_a^4 - (1-\varepsilon_a)\, \sigma.\varepsilon_a.T_a^4\} - \sigma.\varepsilon_0.T_0^4] \; [W.m^{-2}] \quad (3)$$

Where r_o is albedo [-]; $\sigma.\varepsilon_a.T_a^4$ [W.m^{-2}) is the incoming long wave radiation emitted by the atmosphere: $\sigma.\varepsilon_a.T_a^4$ [W.m^{-2})] is the outgoing long wave radiation emitted by the surface; $(1-\varepsilon_0)$. $\sigma.\varepsilon_a.T_a^4$ [W.m^{-2}] is the part of incoming longwave radiation reflected from the surface: σ.[5.67×10^{-8} W.$m^{-2}K^{-4}$] is the Stefan Boltzmann constant and ε_o, ε_o and T_a [K], T_o [K] are the emissivity and temperature of air and water surface, respectively.

The water heat flux (G_o) is defined as the product of thermal conductivity and the gradient of the water temperature. In case of soil surface, results from empirical studies have shown that the day-time ratio of G_o/R_n can be related to parameters that can be derived from satellite data. In case of shallow water, many studies showed that G_o is negligibly small compared to R_n e.g.; tewart & Rose, 1976; Ashfaque, 1999) and can be neglected when longer time average values of fluxes are considered. However in this study, G_o is considered as 1% of R_n for instantaneous case and zero for daily total. This value was assumed based on the study by Ashfaque (1999).

The sensible heat flux H is the part of the surface energy that is used to heat up the planetary boundary layer. Mathematically the sensible heat flux can be expressed as:

$$H = \frac{\rho_a C_p}{r_{ah}} (T_o - T_a) \quad [W.m^{-2}] \tag{4}$$

Where, ρ_a is air density [kg.m^{-3}], C_p is specific heat of moist air, 1004 [J.kg^{-1}, K^{-1}], r_{ah} is aerodynamic resistance to heat transport [s.m^{-1}]. T_o is water surface temperature [K] and T_a is the air temperature at observation height [K]. In case of water, the temperature at 2m height is almost same as the temperature of water surface (i.e. $T_o \approx T_a$) and r_{ah} is very high because of negligibly small water surface roughness (Penman assumed Z_o= 0,00137m, Thom and Oliver, 1977); hence H=0. This value was also confirmed from the study by Farah and Bastiaanssen, (1999) and Ashfaque, (1999) [see Fig. 2].

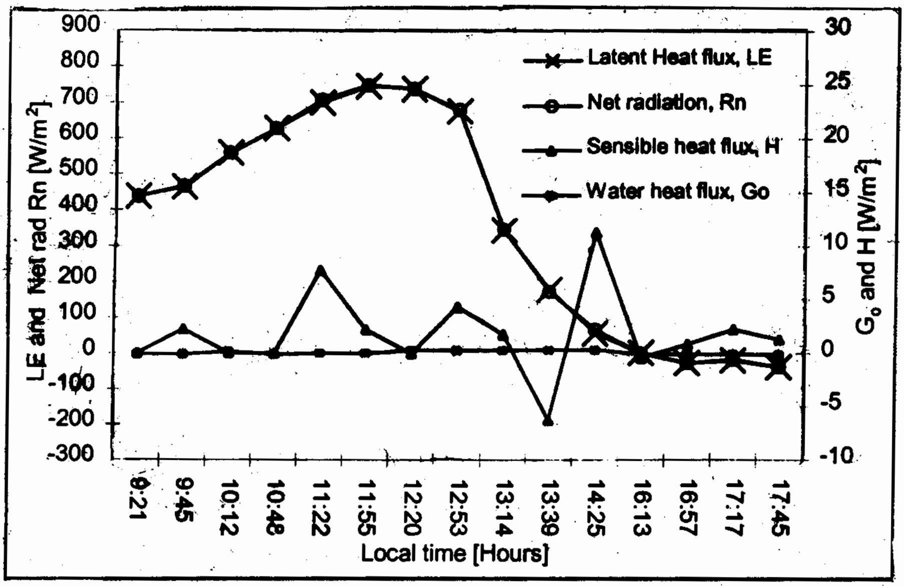

Fig 2. Variation of heat fluxes $\lambda E, Rn, G_o$ and H above Lake Naivasha on 8th October 1998 (Adopted from Ashfaque. 1999)

DAILY TOTAL FROM INSTANTANEOUS ESTIMATION

To find the daily total evaporation from instantaneous latent heat flux, the evaporative fraction approach (Brutsaert and Sugita, 1992) was used where evaporative fraction is the ratio of energy used for the evaporation process and the total amount of energy available for evaporation:

$$\Lambda = \frac{\lambda E}{\lambda E + H} = \frac{\lambda E}{R_n - G_o} \quad [-] \tag{5}$$

Where Λ is evaporative fraction. Although the sensible heat and the latent heat fluxes are fluctuating strongly on a daily basis, evaporative fraction behaves steadily during day-time (Bastiaanssen et al. 1996; Crago, 1996; Shuttleworth et al. 1989). Thus the instantaneous and the integrated daily evaporative fractions can be hold similar i.e. $A_{inst} = A_{24}h$. For shallow lakes, daily average water heat flux G_o can be ignored. This will introduce all error which can be compensated when accumulating evaporation over time. Again in water H is zero hence $A_{inst} = A_{24}h.= 1$(see Fig. 3) So the latent heat flux is approximately equals to the net radiation ($\lambda E_{24h} = R_{n24h}$). So the daily evaporation, E_{24h} can be expressed as:

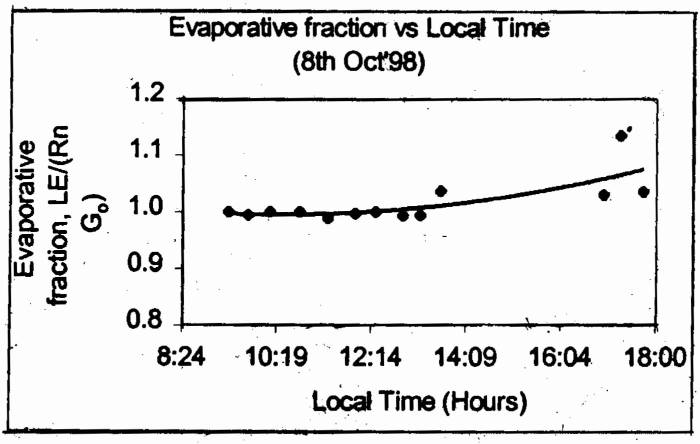

Fig 3. Variation of evaporative fraction with day-time above Lake Naivasha (Adopted from Ashfaque. 1999)

$$E_{24h} = \frac{8.64.10^7 . Rn_{24h}}{\lambda \rho w} \quad [\text{mm. day}^{-1}] \tag{6}$$

Where ρ_w is the density of water 1000 [kg.m^{-3}], and R_{n24h} is the average net radiation for the day [W.m^{-2}], which can calculated as:

$$R_{n24h} = (1 - r_o). K{\downarrow}_{24h} + L_{n24h} \quad [\text{W.m}^{-2}] \tag{7}$$

Where $K{\downarrow}_{24h}$ and L_{n24h} are the daily average incoming solar radiation and net longwave radiation, respectively. For daily net long wave radiation, L_n Slob (after de, Bruin, 1987) developed and empirical formula for "potential" conditions, which is:

$$L_n = -110, \frac{K{\downarrow}_{24h}}{R_{a24h}} \qquad [W.m^{-2}] \qquad (8)$$

Where R_{n24h} is the daily average extraterrestrial short-wave solar radiation $[W.m^{-2}]$ at the top of atmosphere and $\frac{K{\downarrow}_{24h}}{R_{a24h}}$ is the daily average atmospheric short-wave transmittance.

REMOTE SENSING EVAPORATION

Equation (3) can be solved using partial remote sensing date $(r_o, \varepsilon_o, T_o)$ in combination with ancillary ground data on $K{\downarrow}$ and $L{\downarrow}$ (i.e. Bastiaanssen, 1995, Tuzet, 1990; Jackson et al., 1985; Moran et at., 1989; Kustas et al., 1994; Pelgrum and Bastiaanssen, 1996) 1996) The net short-wave radiation K_n is estimated as:

$$K_n=(1-r_o).K{\downarrow} \qquad [W.m^{-2}] \qquad (9)$$

The broadband (0.3-3. 0μm) incoming solar radiation at the earth surface $K{\downarrow}$ can be estimated using the general algorithm of Bastiaassen (1995):

$$K{\downarrow}=R_a.\ \tau_{sw} \qquad [W.m^{-2}] \qquad (10)$$

Where τ_{sw} is the atmospheric short-wave transmittance, R_a is the incoming solar radiation at the top of atmosphere (Extraterrestrial solar radiation). In the absence of data on $K{\downarrow}$ or degree of cloudiness to find τ_{sw} from local measurements, two way transmittance τ_{sw} for broadband solar radiation (0.3-3 0μm) Eq. (11) (Zhong and Li, 1988) can be used to determine τ_{sw}:

$$\tau_{sw} = (\tau sw'')^{0.5} = \left(\frac{r_p - r_a}{r_o}\right)^{0.5} \qquad [W.m^{-2}] \qquad (11)$$

Where r_p [-] is the boardband planetary albedo, r_o [-] is the surface albedo and r_a [-] is the lowest planetary albedo at all pixels (i.e. $r_p{}^{min}$). If r_o is measured at a few specific locations in the study area, τ_{sw} can be obtained by combining r_p from remote sensing measurements and r_o from these in situ measurements. If τ_{sw} is measured, r_o can be derived from r_p and τ_{sw}'' can be obtained directly. The broadband planetary albedo r_p can be calculated as:

$$r_p = \int_{0.3}^{3.0} r_p(\lambda)d\lambda = \sum_{i=1}^{n} C(b)_i . r_p(b)_i \; [W.m^{-2}] \qquad (12)$$

Where n in the total number of spectral bands, C (b) is the weighing factor accounting for the uneven distribution of spectral incoming solar radiation at different bands.

At surface incoming longwave radiation $L\downarrow$ can be estimated as $L\downarrow$ $\sigma.\varepsilon_a.T_a^4$. The average emissivity of the atmosphere ε_a can be calculated from the empirical relationship (Bastiaanssen, 1995):

$$\varepsilon_a^{avg} = 1.08\ (-Ln\tau^{avg}{}_{sw})^{0.265} \qquad [-] \qquad (13)$$

Where $\tau^{avg}{}_{sw}$ is the average atmospheric shortwave transmittance. For water body we can assume the air temperature at 2m height to be the same as water surface temperature and the water surface temperature can be estimated using remote sensing technique. The outgoing longwave radiation $L\uparrow$ from the water surface can be calculated as $L\uparrow = \sigma.\varepsilon_a T^4{}_0$ where $L\uparrow$ is a function of surface temperature T_o and the surface emissivity $.\varepsilon_o$. A water body is normally considered as a blackbody, so we can assume $\varepsilon_o=1$. To get surface temperature from the satellite measurements, different algorithms are proposed for Thematic Mapper (TM). The thermal channel TM6 measures the spectrally emitted radiance between 10.6 to 12.4 μm at the top of atmosphere, $L_6{}^{TOA.}$ The spectral radiances at the top of atmosphere measured by the satellite are related to the spectrally emitted radiances at the land surface $L_6{}^{surf}$ (Schmugge et al. 1998):

$$L_6{}^{TOA.} = L_6{}^{surf} \ . \ \tau_6 + L_6{}^{atm} \qquad [W.m^{-2}] \qquad (14)$$

Where $L_6{}^{atm}$ is the longwave radiation emitted from the top of atmosphere upward (i.e. thermal path radiance) and τ_6 is the atmospheric transmittance in the region of wavelength 10.6 to 12.4 μm. $L_6{}^{atm}$ and τ_6 can be determined by the atmospheric radiation models or from a limited number of surface temperature T_o field measurements taken at the same moment of the satellite overpass. The spectral radiances at surface level, after atmospheric correction were converted into radiometric surface temperature through the inversion of Planck's law.

$$To = \frac{14388}{11.5 \ln\left[\frac{\varepsilon_o \times B \times 3.7427\times10^8}{.L_6{}^{surf}(11.5)^5} + 1\right]} \; [k] \qquad (15)$$

Where B is the bandwidth of thermal channel (12.4-10.6=1.8 μm ε_0[-] is the thermal infrared surface emissivity in the spectral range of TM band 6 and T_0 [K] is the radiomatric surface temperature corrected for grey body effects.

THEMATIC MAPPER APPLICATION

To estimate evaporation from remotely sensed spectral data a TM image of 21st January 1995 was used in the analysis. In the calculations, water heat flux G_0 was taken as 1% of net radiation in the instantaneous case and for daily total G_0 was considered as zero and sensible heat flux, H was considered as zero in both cases.

Landsat Thematic Mapper (TM) measures the spectral radiance in the visible, near middle and thermal infrared spectrum at the top of the atmosphere. TM has 3 bands in the visible, 3 bands in the near and middle infrared and 1 band in the thermal infrared spectral region. The digital values of each pixel are converted first to spectral radiance at the top of atmosphere using a radiometric calibration procedural. Digital Numbers (DN) of each pixel can be converted to spectral radiance at the top of the atmosphere as:

$$L\lambda i = a + \frac{b-a}{255} * DN$$

Where $L_{\lambda i}$ [mW.cm^{-2}, sr^{-1}, μm^{-1}] is the spectral radiance in band I of Thematic Mapper. The constants 'a' and 'b', given by Markham and Barker (1987), are as follows.

Band no:	1	2	3	4	5	7
a	-0.15	-0.28	-0.12	-0.15	-0.037	-0.015
b	15.21	29.68	20.43	20.62	2.72	1.44

The thermal channel TM_6 measures the spectrally emitted radiance between 10.6 to 12.4 μm at the top of the atmosphere, L_6^{TOA}, which can be interpreted from the raw digital numbers (DN_6) in band 6:

$$L_6^{TOA} = \{ 0.1238+(1.560-0.1238)\times DN_6/255\} \times \pi \times B \times 10 \quad [W.m^{-2}] \qquad (17)$$

Where B [μm] is the bandwidth of the thermal channel (12.4-10.6=1.8μm) and DN_6 is the digital number of TM band 6. The broadband planetary albedo at the top of the atmosphere (r_p) was calculated with a weighing scheme of the six visible and near infrared bands (TM 1,2,3,4,5 and 7).The band wise spectral reflectance at the top of the atmosphere was calculated as:

$$r_p(\lambda i) = \frac{\pi L_{\lambda i}\, d_s^2}{K(\lambda i).\cos\varnothing_{su}} \qquad [-] \qquad (18)$$

Where r_p (λi) is spectral reflectance at the top of the atmosphere of band i, d_s is the earth-sun distance in astronomical units (Iqbal, 1983), k (λ_i) is the spectral incoming solar radiation and $\varnothing_{su}$ is the solar zenith angle. The spectral incoming solar radiation at the top of the atmosphere for the Thematic Mapper bands is as follows:

Band no:	1	2	3	4	5	7
K (λi)	195.8	182.8	155.9	104.5	21.9	7.5

Finally the broadband reflectance at the top of the atmosphere was estimated as:

$$r_p = \Sigma W_i r_p \ (\lambda_i) \qquad [-](19)$$

The weights, w_i for the different bands are computed as the ratio of the amount of incoming shortwave radiation from the sun in a particular band and the sum of incoming shortwave radiation for all the bands. Using the spectrally emitted radiance of band 6, surface temperature T_o was estimated from Eq. (14) and Eq. (15). T_o solve Eq. (15) a trial and error procedure was followed.

Incoming solar radiation $K\downarrow$ in Eq. (3) was determined on the basis of standard astronomical equations (Iqbal, 1983) which leads to an instantaneous value of 1180 [W.m^{-2}] at the top of the atmosphere above the lake area during Landsat overpass at 9:45 a.m. The solar zenith angle at that moment was 33.3°. From Eq. (11) a value for the single way transmittance of τ_{sw}= 0.59 was obtained considering water surface albedo, r_o =0.06 and this transmittance value was found reasonable for the lake area from a study by Ashfaque (1999). Hence, a portion of 1180 × 0.59=696 [W.m^{-2}], after atmospheric absorption, scatter and transfer will reach the land surface. The apparent emissivity of the atmosphere, ε_a=0.91 was obtained from Eq (13). Incoming long wave radiation was estimated by using air temperature (To = 24.8°C) and the apparent emissivity of the atmosphere yielding a value of $\varepsilon_a.\sigma.\ T_a^4 = L\downarrow$ 407[W.m^{-2}]. The screen height air temperature during satellite overpass was T_a=24.8°C, being tentatively estimated from the minimum surface temperature of Lake Naivasha (T_a = 24.8°C, Ashfaque, 1999) assuming that $T_a \approx T_o$ above water.

From instantaneous to daily total value, evaporative fraction method was used. Daily average incoming solar radiation at the top of atmosphere was calculated on the basis of standard astronomical equations (Iqbal, 1983), which was 424.68 [W.m^{-2}] and applying transmittance = 0.59, the solar radiation reaching the surface was $K\downarrow = 0.59 \times 424.68 = 250.56$ [W.m^{-2}]. For daily net longwave radiation Slobs approach was followed which leads to a value = $110 \times \tau_{sw} = 64.9$[W.m^{-2}].

The instantaneous latent heat flux, Temperature and daily total evaporation from the lake were calculated using energy balance equation. The spatial variation of evaporation, lake surface temperature, latent heat flux calculated are presented in Figures 4, 5 and 6 respectively. The daily total evaporation E_{24h} = 5.95 mm/ day, obtained from the analysis for 21st January 1995, was reasonable.

Fig 4. Spatial Variation of daily evaporation [mm] from lake surface derived from TM image (21st January 1995)

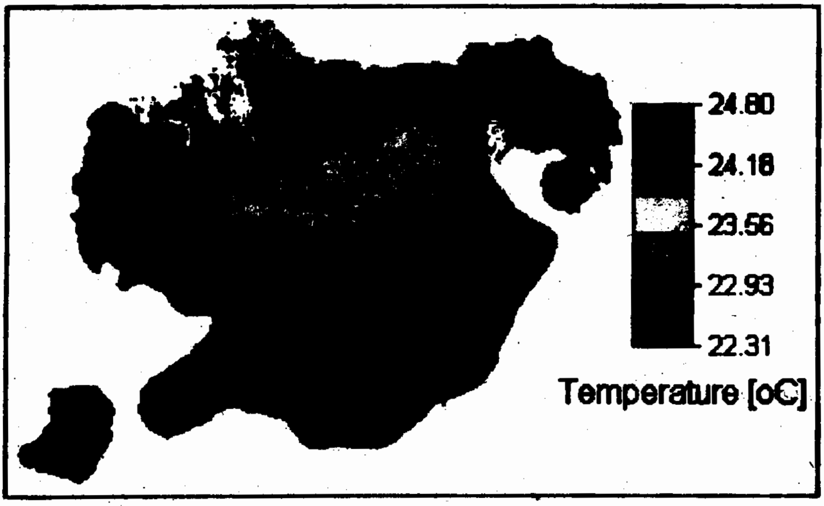

Fig 5. *Suface temperature of lake Naivasha derived from TM image (9:45 AM 21st January 1995)*

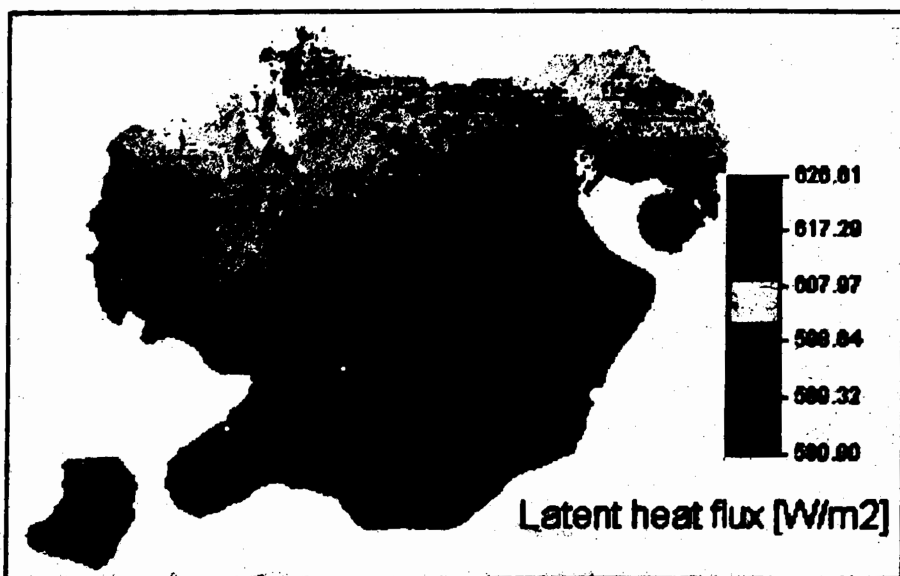

Fig 6. *Instantaneous latent heat flux from the lake using energy balance equation (TM image of 9:45 AM, 21St January 1995).*

CONCLUSIONS

Using energy balance residual approach instantaneous latent heat flux from the lake was calculated from high resolution. Thematic Mapper spectral data and this approach give very close value. The daily total evaporation of 5.95 mm from the lake using evaporative fraction approach and Slob's equation for daily net longwave radiation seems to be good estimation, because from pan data the estimated average evaporation was 5.46mm for 21st January with a standard deviation of 1.28 mm for the period of 1957-1990. Also we get 5.83 mm evaporation for 21st January 1999 from the lake using data from nearby meteorological station (Ashfaque, 1999) From the satellite images and field observations, negligible spatial variation of evaporation, surface temperature and albedo was found in the lake. However, when it is necessary to know the spatial variation of evaporation from a lake or from an irrigated area with high resolution, this approach can be used effectively.

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REFERENCES

Allen, R. G. Smith. M. Perrier, A and Pereira L.S. (1994), "An update for the definition of reference Evapotranspiration," ICID Bulletin, 43(2): 1-92.

Allen, R. G., Pruit, W. M, Businger J. A., Fritschen, L. J., Jensen, M.E., Quinn, F.H.(1996), "Evaporation and Transpiration, Chapter 4, Hydrology Handbook, second Edition. ASCE, 125-234.

Ashfaque, A (1999), Estimating Lake Evaporation Using Meteorological Data and Remote sensing. A Case Study of Lake Naivasha, Central Rift Valley, Kanya, M. Sc. thesis, Division of water Resources and Environmental Studies, International Institute for Aerospace Survey and Earth Sciences (ITC), Enschede. The Netherlands.

Assouline, S. Mahrer, Y. (1993), "Evaluation from lake Kneret 1, Eddy correlation system measurements and energy budget estimates," Water Resources Research, 29 (4), 901. 910.

Bastiaanssen, W. G. M. (1995). "Regionalization of surface flux densities and moisture indicators in composite terrain, A remote sensing approach under clear skies in mediterranean climates". Doctoral thesis, Wageningen Agricultural University, Wageningen, The Netherlands.

Bastiaanssen, W. G. M., Pelgrum, L., Menenti, M. and Feddes, R.A, (1996a), "Estimation of surface resistance and Priestley-Taylor α -parameter at different scales", In J.B Stewart, et. al. (eds). Scaling up in hydrology using remote sensing, Institute of Hydrology, Wallingford, 93-111.

Bruin, H.A.R, De (1987), "From Penman to Makkink, Committee on Hydrological Research". TNO, The Hague Proceedings and Information 39-5-31.

Bruin, H.A.R. De (1982), "Temperature and energy balance of water reservoir determined from standard weather data of a land station," Journal of Hydrology. 59-261-274.

Brutsaert, W. and Sugita, M. (1992), "Application of Self-preservation in evaluation of the surface energy budget of determine daily evaporation," Journal of Geophysical Research, 37:18377-18382.

Choudhury, B.J. (1994), "Synergism of Multispectral Satellite Observation for Estimating Regional Land Surface Evaporation," Remote sensing Environment, 49: 264-274.

Crago, R.D (1996), "Daytime evaporation from conservation of surface flux ratios," In, J.B. Stweart et, al. (eds). Scaling up in hydrology using remote sensing. Institute of Hydrology, Wallingford, 235-244.

East African Meteorological Department (1996), Monthly and annual rainfall in Kenya during the 30 years 1931-1960. 172pp.

Farah, O.H. and Bastiaanssen W. G. M. (1999), "Estimation variability of watershed surface parameters and evaporation from remote sensing measurements without access to ground data," J. Of Hydrology (press)

Iqbal, M. (1983), An introduction to solar radiation, Academic press, Canada.

Jackson, R.D.P. J. Printer, and R.J. Reginato. (1985), "Net Radiation calculated from remote multispectral and ground station meteorological data," Agricultural and Forest Meteorology, 35: 153-164.

Keijman, J.Q. (1981), "Theoretical background of some methods of determination of evaporation," Comm. Hydro Research, TNO, The Hague, Proceedings and Information.28. 12.24

Keijman, J.Q. (1974). "The estimation of energy balance of lake from simple weather data." Bounday- layers Meteorology. 7: 399-407.

Kustas, W.P. M.S Moran, K.S Humes, D.I. Stannard, P.J. Printer. L.E. Hipps, E. Swiatek. and D.C. Goodrich (1994), "Surface energy balance estimates at local and regional scales using optical remote sensing from aircraft platform and atmospheric data collected over semiarid rangelands," Water Resource Research 30 (5): 1241-1259.

Markham, B.L. and J. L. Barker (1987), "Thematic Mapper bandpass solar exo-atmospherical irradiances," Int. J. of Remote Sens. 8 (3): 517-523.

Moran, M.S. R.D. Jackson, L. H. Raymond. L.W Gay. and P.N. Slater (1989), "Mapping surface energy balance components by combining Landsat Thematic Mapper, and ground-based meteorological data," Remote Sensing of Environment 41:169-184.

Pelgrum, H. and W. G .M Bastiaanssen(1996), "An inter comparison of techniques to determine the area averaged latent heat flux from individual in situ observations: A remote sensing approach using the European Field Experiments in a Desertification. Threatened Area data," Water Resources Research 32 (9): 2775-2786.

Reis, J.R, and Dias, N.L. (1998), "Multi-season lake evaporation : energy budget estimates and CRLE model assessment with limited meteorological observations." Journal of Hydorlogy, 208 (1998) 135-147.

Robertson, E. and P.J. Barry. (1985), "The water and energy balances of perch lake (1969-1980)." Atmo. Ocean, 23, 238-253.

Rosenberry. D. O. A. M . Sturrock and T.C. Winter. (1993), "Evaluation of the Energy budget method of determining evaporation at Williams lake, Minnesota, using alternative instruments & study approach," Water Resources Research, 29(8), 2473-2483.

Schmugge, T., S. J. Hook and C. Coll (1998), "Recovering surface temperature and emissivity from thermal infrared multispectral data." Rem. Sens. Env. 65:121-131.

Shuttleworth, W.J. Gurney, R.J. Hsu, A.Y. and Ormsby, J. P. (1989). FIFE: "The variation in energy partition at surface flux sites," IAHS Pub. No. 186, 67-74.

Simons, E. and Mero. F. (1985), "A simplified procedure for the evaluation of lake Kinneret evaporation," Journal of Hydrology, 78, 291-304.

Stewart, R. Rose W. (1976), "A simple method for determining the evaporation from lakes and ponds," Water Resources Research, 12, 623-628.

Sturrock, A.M. Winter, T.C. Rosenberry (1992). "Energy budget evaporation from William lake: A closed lake in North central Minnesota," Water Resources Research, 28 (6), 1605-1607.

Thom, A. S. and H. R. Oliver (1977), "On Penman's equation for estimating regional evaporation," Qurt. J. Roy. Met. Soc. 96: 67-90

Tuzet, A (1990), "A simple method for estimating downward longwave radiation from surface and satellite data by clears sky," International Journal of Remote sensing 11: 125 -131.

Zhong. Q. and Y. H. Li (1988), "Satellite observation of surface albedo over Qinghai-Xinzang Plateau region, Adv. In Atm. Sciences 5: 57-65.