The Application of the

Surface Energy Balance Algorithm for Land (SEBAL)

For Estimating Daily Evaporation

A Handbook of Published Methodologies

Version 0.6.7

Monday, December 18, 2023

Yann Chemin

Foreword

TABLE OF CONTENTS

TABLE OF FIGURES

Chapter 1

Chapter 2

Chapter 3

Chapter 4

Chapter 5

Introduction

1.Overview

1.1.Overview of the Structure of this Manual

SEBAL processing is involving a certain number of prerequisite images to be processed before starting to run the different parts of the model.

The prerequisite images for SEBAL processing are:

Images	Code	Prerequisite	Used in SEBAL
Broadband Surface Albedo	ρ_0	Original data	Pages 79, 87
Normalized Difference Vegetation Index	$NDVI$	Original data	Pages 79, 87
Emissivity	ε_0	$NDVI$	Pages 79
Surface Temperature	T_0	Original data	Pages 79, 87, 132

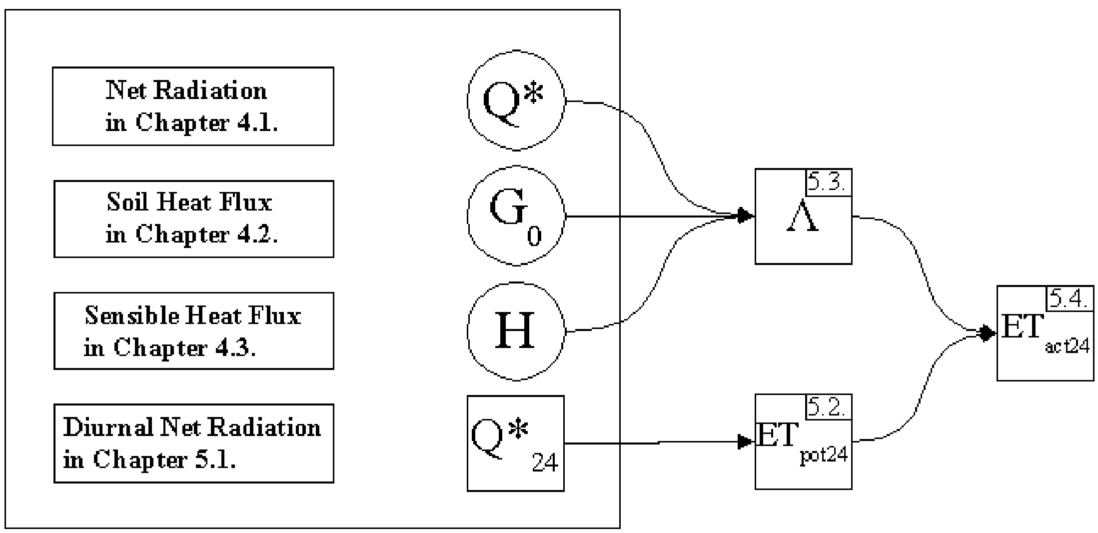


Figure 1: Overview of the structure of this manual

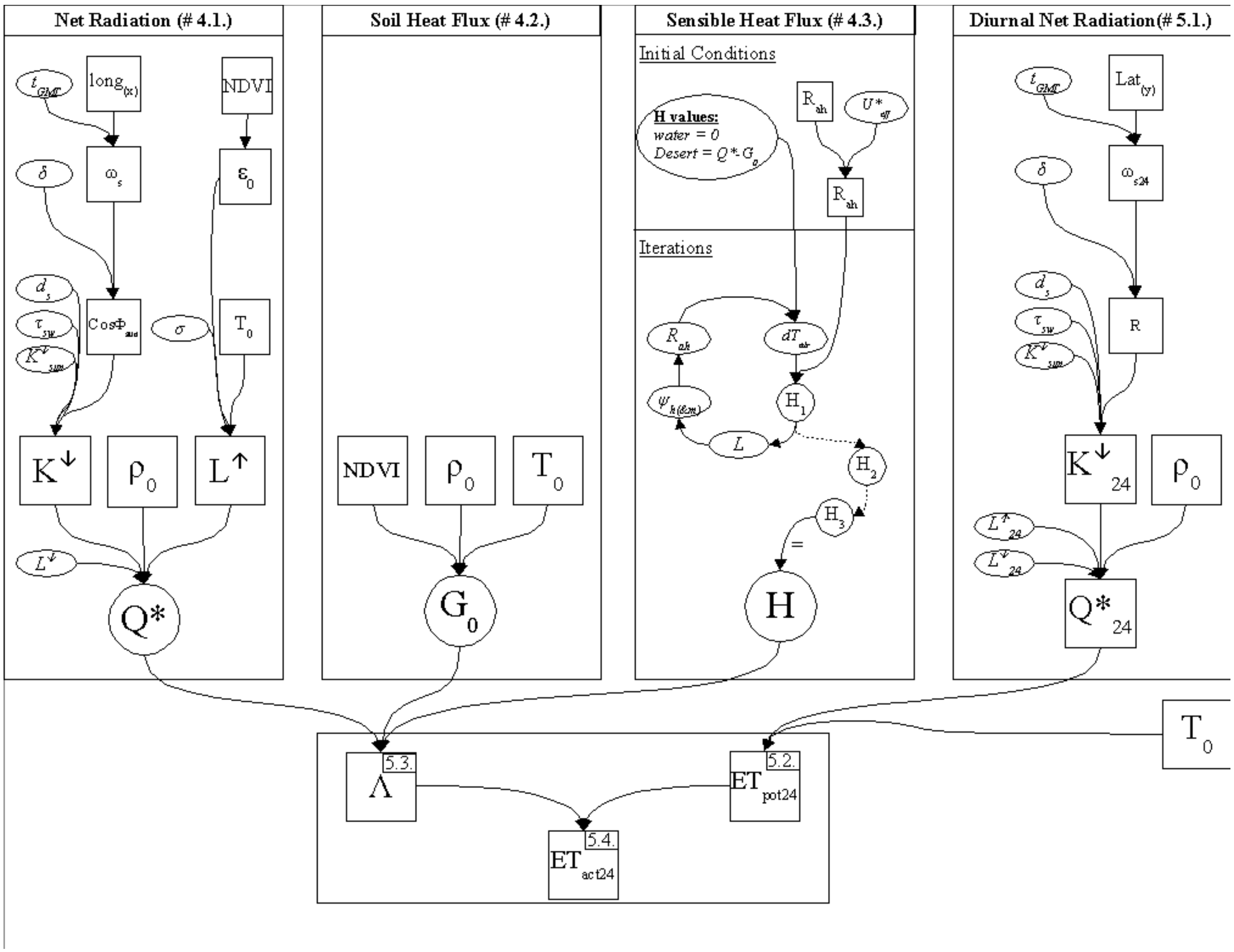
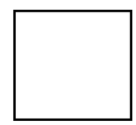


Figure 2: Overview of SEBAL processing

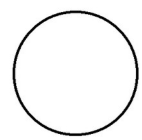
1.2.Flow-chart keys Glossary

This section is dedicated to browse extensively the flow-charts keys used in this manual, providing a definition for each of their components.

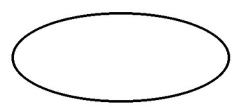
- Raster images are symbolized by Squares



- Special raster images forming the energy-balance terms are symbolized by Circles



- A single value, or a succession of simple processing steps are symbolized by Ellipsoids



1.2.1.Raster Images Keys

Chapter 3. PRE-PROCESSING on Page 20

Band i	Original data coming from the satellite image for a specific band number i
ρ_0	Broadband Surface Albedo
ρ^{TOA}	the top of atmosphere broadband Albedo
ρ_i^{TOA}	the top of atmosphere broadband Albedo for Band number i
ε_0	The surface emissivity (-)
NDVI	The Normalized Difference Vegetation Index (-)
T_{Rad}^{TOA}	The Radiance Temperature at top of atmosphere (K)
T_0	The Surface Skin Temperature (K)

Chapter 4. Energy Balance Terms on Page 77

Q^*	The Instantaneous Net Radiation (W/m^2)
G_0	The Soil Heat Flux (W/m^2)
H	The Sensible Heat Flux (W/m^2)
ε_0	The surface emissivity (-)
ρ_0	Broadband Surface Albedo (-)
ρ^{TOA}	the top of atmosphere broadband Albedo (-)
NDVI	The Normalized Difference Vegetation Index (-)
$K^{\downarrow}$	The Incoming Short-wave Solar Radiation (W/m^2)
T_0	The Surface Skin Temperature (K)

T_{Rad}^{TOA} The Radiance Temperature at top of atmosphere (K)

R_{ah} The Aerodynamic Resistance to Heat Transport (s/m)

z_{0m} The Aerodynamic Roughness Length for Momentum Transport (m)

ω_s The Instantaneous Solar Angle Hour (rad)

$long_{(x)}$ The Longitude Coordinate (dd)

$Lat_{(y)}$ The Latitude Coordinate (dd)

Chapter 5. Daily Evaporation on Page 132

Q^* The Instantaneous Net Radiation (W/m^2)

G_0 The Soil Heat Flux (W/m^2)

H The Sensible Heat Flux (W/m^2)

$Lat_{(y)}$ The latitude co-ordinates (dd)

R The Solar Angle Range for the Diurnal Sun Exposition (rad)

ω_{s24} The Solar Angle Hour for Diurnal Exposition (rad)

$K_{24}^{\downarrow}$ The Incoming Shortwave Solar Radiation (W/m^2)

ρ_0 Broadband Surface Albedo (-)

Q_{24}^* The Net Radiation for 24 hours (W/m^2)

T_0	The Surface Skin Temperature (K)
ET_{Pot24}	The 24 hours Potential Evapotranspiration (mm/day)
Λ	The Evaporative Fraction (-)
ET_{act24}	The 24 hours Actual Evapotranspiration (mm/day)

1.2.2.Single Values / Simple Processing

Chapter 3. PRE-PROCESSING on Page 20

$Calib_i$ The Calibration Processing for Band number i

τ_{sw} The Single-Way transmittance of the Atmosphere (-)

r_a The In-Band Path Radiance (-)

$Inv.Planck$ The Processing of the Inverse Planck Function for Temperature Estimation

$SplitWindow$ The Processing of the Split-Window Method for Temperature Estimation

$T_0=a+bT_{rad}^{TOA}$ The Linear Regression Estimation of Surface Temperature

Chapter 4. Energy Balance Terms on Page 77

t_{GMT}	The time at Greenwich Meridian Time (t)
δ	The Solar Declination (rad)
d_s	The Distance Sun-Earth (A.U.)
τ_{sw}	The Single-Way transmittance of the Atmosphere (-)
$K^{\downarrow}_{sun}$	The Sun External Atmosphere Radiation (W/m^2/µm)
σ	The Stefan-Boltzman Constant (W/m^2/K^4)
$L^{\downarrow}$	The Broadband Incoming Long-Wave Radiation (W/m^2)

H values: water = 0 Desert = Q^*-G_0	The Processing of the regression of the first Approximation of the Sensible Heat Flux Image (H).
U^*_{eff}	The Effective Friction Velocity (m/s)
dT_{air}	The Skin Surface / Air temperature difference for dry pixel (K)
R_{ah}	The Processing of the Aerodynamic Resistance to Heat Transport (s/m)
$\psi_{h(\&m)}$	The Stability Correction for the Atmospheric Heat Transport (& Momentum Transport) (-)
L	The Monin-Obukov Length (m)

Chapter 5. Daily Evaporation on Page 132

δ The Solar Declination (rad)

d_s The Distance Sun-Earth (A.U.)

τ_{sw} The Single-Way transmittance of the Atmosphere (-)

$K^{\downarrow}_{sun}$ The Sun External Atmosphere Radiation (W/m^2/µm)

$L^{\uparrow}_{24}$ The Diurnal Broadband Outgoing Long-Wave Radiation (W/m^2)

$L^{\downarrow}_{24}$ The Diurnal Broadband Incoming Long-Wave Radiation (W/m^2)

2. Theory of SEBAL and its links to common satellites

3.PRE-PROCESSING

3.1.Visible Remote Sensing

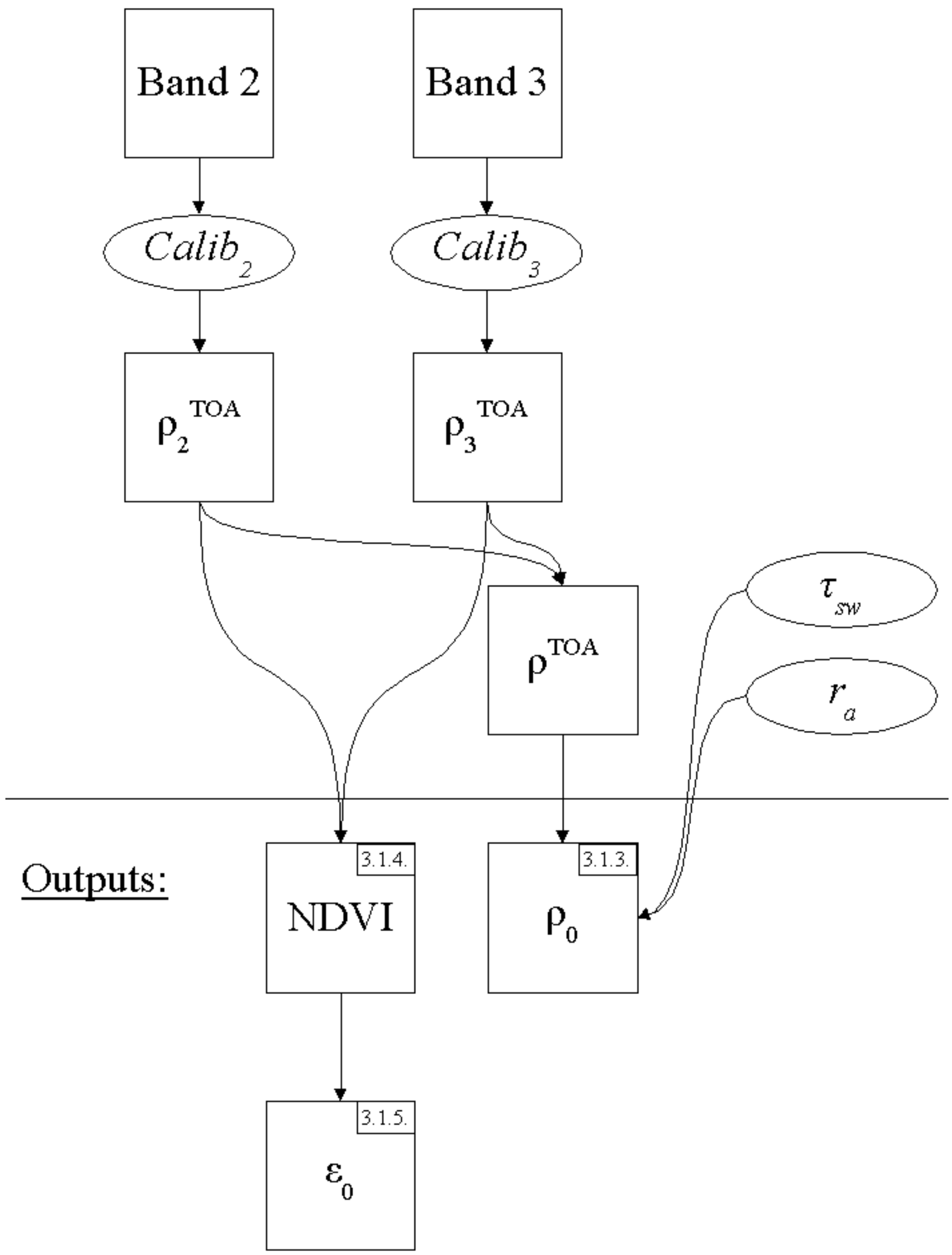


Figure 3: Production of NDVI, Surface Albedo and Surface Emissivity for NOAA

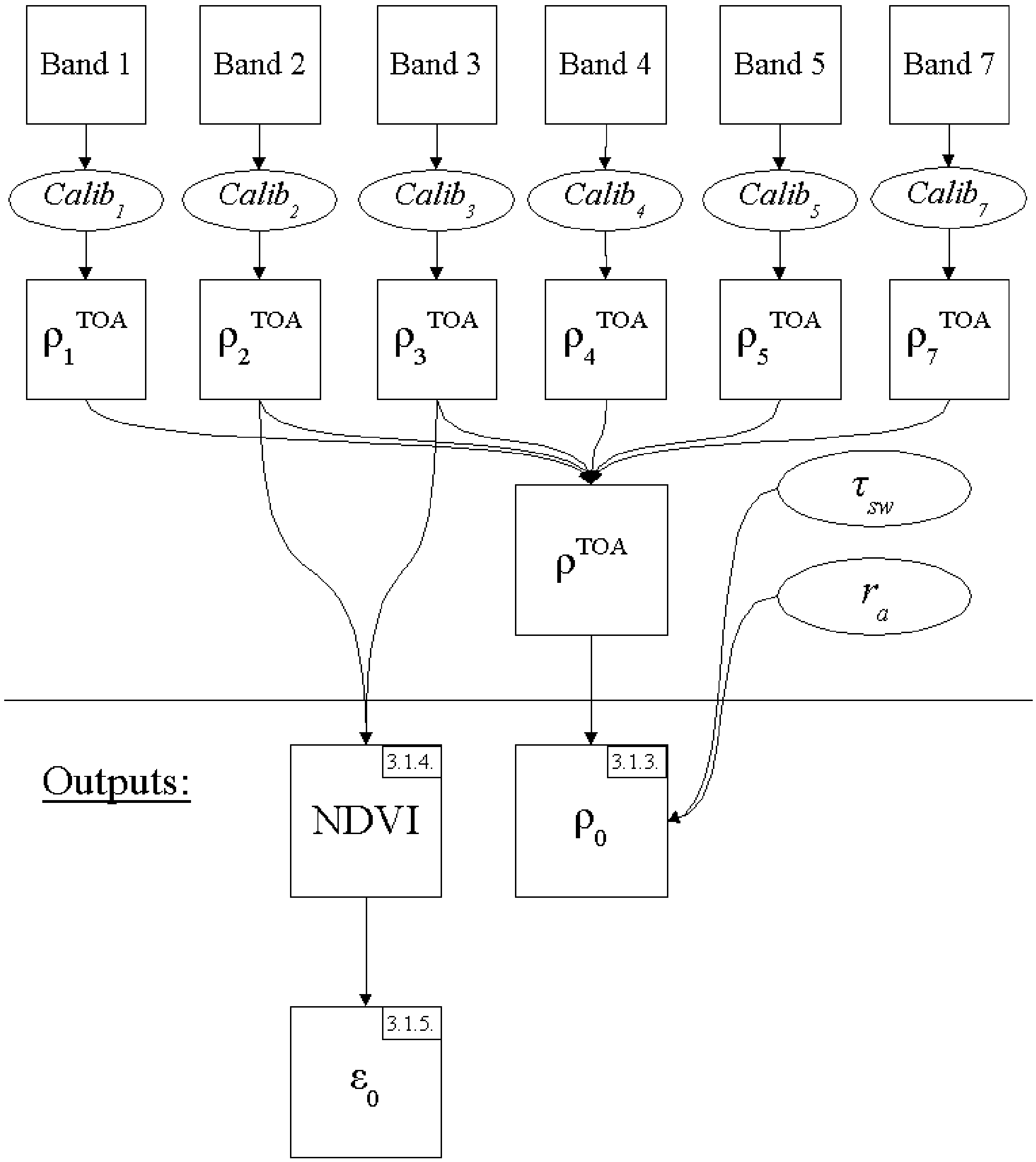


Figure 4: Production of NDVI, Surface Albedo and Surface Emissivity for Landsat

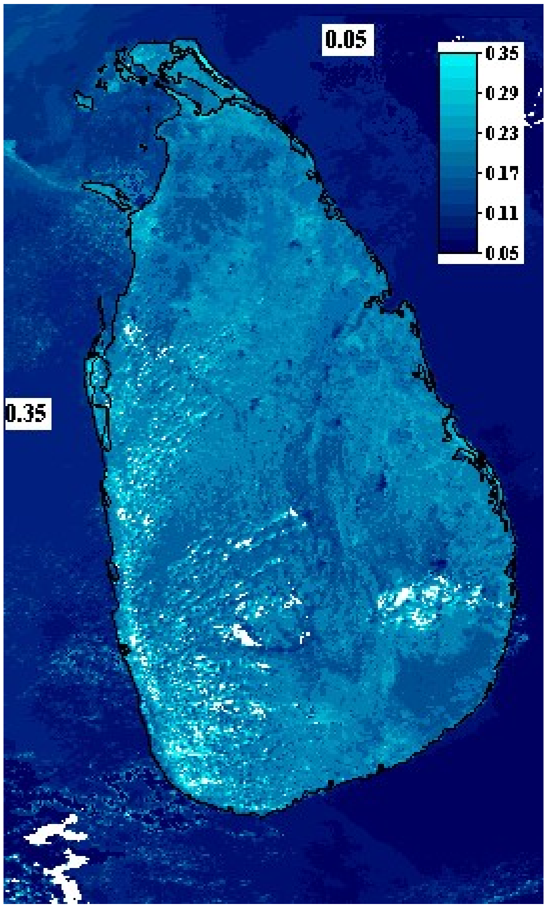


Figure 5: Surface Albedo over Sri Lanka (-)

INPUTS:

Raster:
Landsat: Band 1, 2, 3, 4, 5 and 7.
NOAA: Band 1 and 2

Tabular data:
Calibration coefficients for each Band

Optional tabular data:
Transmissivity of the atmosphere
In-band path radiance
Ground calibration of Surface Albedo in Red and InfraRed bands.

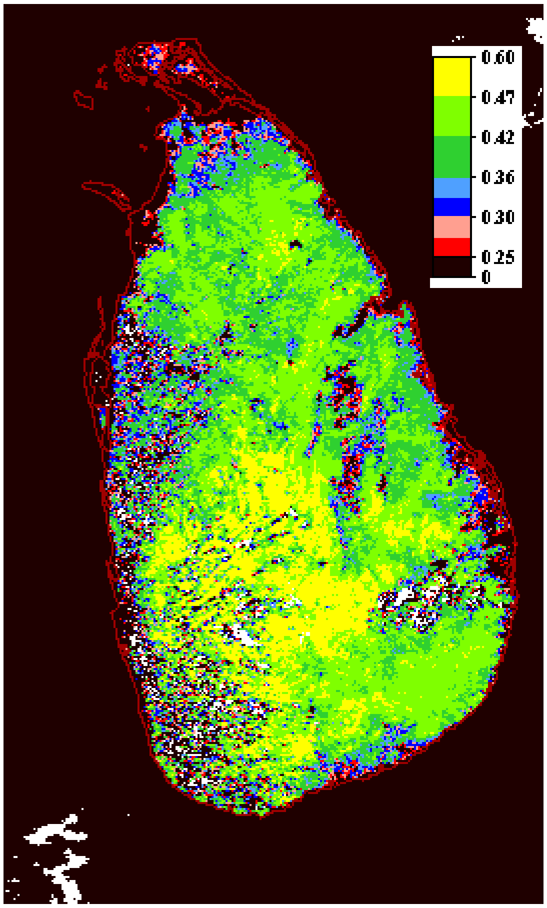


Figure 6: NDVI over Sri Lanka (-)

<u>OUTPUTS:</u>

<u>Raster:</u>
Broadband Albedo
NDVI
Emissivity

3.1.1.Calibration of the images

The aim of the calibration of the bands in the visible, is to get radiance values at the Earth Skin Surface as per Power in a certain Area (W/m^2), being the physic base for all the Remote Sensing processing to be implemented. Once this is done, a reflectance value is extracted, as being the reflected amount of radiation over the total radiation arriving. Reflectance values are the basis of most of the Remote Sensing calculation in the Visible bands.

	Q: Which method am I to use? A: If you have a Remote Sensing Software that imports any image and does geometric and radiometric corrections automatically, then skip this Calibration Part. Q: I do not have this import module for my image... A: Then do a geometric correction (georeferencing) and go to the corresponding question below. Q: I have acquired a raw NOAA AVHRR from a receiving station. A: then use Method 1 or 3. Q: I did download NOAA AVHRR Level 1B from Internet. A: Method 2 is yours. Q: I do have a CDROM of LandSat (5 TM or 7 ETM+). A: Method 4 for LandSat 5TM and Method 5 for LandSat 7ETM+.

The Top of atmosphere reflectance is:

$$\rho_i^{TOA} = \frac{L_i^{TOA}}{K_i^{\downarrow}} \qquad (-)$$

With:

ρ_i^{TOA} being the planetary spectral reflectance at the top of the atmosphere (-)

L_i^{TOA} the spectral radiance at the top of the atmosphere $(W.m^{-2}.\mu m^{-1})$

$K_i^{\downarrow}$ the incoming spectral radiance $(W.m^{-2}.\mu m^{-1})$

where $_i$ is the band number of the satellite (band 1 and 2)

In order to get to ρ_i^{TOA}, it is first to go to the two components of the ratio. Let us start with L_i^{TOA}, then finishing with $K_i^{\downarrow}$.

The spectral radiance at the top of the atmosphere, L_i^{TOA}.

$$L_i^{TOA} = \left(\frac{DN_i - I_i}{G_i} \right) \times \pi \qquad (W.m^{-2}.\mu m^{-1})$$

With:

L_i^{TOA} the spectral radiance at the top of the atmosphere $(W.m^{-2}.\mu m^{-1})$

G_i the gain (-) and I_i the offset $(W.m^{-2}.\mu m^{-1})$

DN_i the Digital Numbers (-)

where $_i$ is the band number of the satellite (band 1 and 2)

The gain and offset values for NOAA satellites are varying according to the NOAA satellite number, as explained below.

Calibration parameters for G_i and I_i.

The gain and offset for NOAA satellites are following the set of equation:

$$G_i = a_i t + b_i \qquad (-)$$
$$I_i = c_i t + d_i \qquad \left(W.m^{-2}.\mu m^{-1}\right)$$

With:

t being the number of days after launch of NOAA

$_i$ the band number (band 1 and 2)

a_i, b_i, c_i, d_i coefficient varying with t after satellite launch and with the band number

After Kerdiles (95?), coefficients a, b, c and d for the calculation of NOAA-11 and NOAA-14 AVHRR calibration coefficients (gain and offset) of channels 1 and 2[1]. The coefficients given for date D (line n) are valid up to the next date (line n+1). In other words these a, b, c and d coefficients must not be interpolated with time.

NOAA-11 Channel 1.

Date	Day post launch	A	b	C	d
24/09/88	0	-2.333e-04	1.704	1.220e-04	39.98
01/01/89	99	-3.079e-05	1.684	5.815e-05	39.99
01/01/90	464	5.412e-05	1.646	-4.236e-05	40.04
01/01/91	829	1.300e-06	1.689	-1.429e-04	40.12
01/01/92	1194	2.010e-05	1.666	-2.435e-04	40.24
01/01/93	1560	2.783e-05	1.654	3.832e-04	39.26
01/01/94	1925	-5.601e-05	1.815	0.000e-00	40.00
01/01/95	2290	-5.534e-05	1.814	0.000e-00	40.00

NOAA-11 Channel 2.

Date	Day post launch	A	b	C	d
24/09/88	0	-9.578e-04	2.606	1.623e-04	39.97
01/01/89	99	-3.049e-05	2.542	4.929e-05	39.99
01/01/90	464	9.108e-05	2.358	-1.284e-05	40.07
01/01/91	829	4.419e-06	2.397	-3.062e-04	40.20
01/01/92	1194	3.288e-05	2.410	-4.842e-04	40.43
01/01/93	1560	-4.630e-05	2.534	8.914e-04	38.28
01/01/94	1925	-1.331e-04	2.701	0.000e-00	40.00
01/01/95	2290	-1.305e-04	2.695	0.000e-00	40.00

Calibration of NOAA AVHRR: Method 1

[1] Source: Cihlar and Teillet, 1995 for NOAA-11 and Rao (NOAA/NESDIS) for NOAA-14

NOAA-14 Channel 1.

Date	Day post launch	a	b	c	d
30/12/94	0	0.000e-04	1.795	0.000e-04	41.0
01/01/95	2	-3.527e-04	1.795	0.000e-04	41.0
01/01/96	367	-3.047e-04	1.778	0.000e-04	41.0

NOAA-14 Channel 2.

Date	Day post launch	a	b	c	d
30/12/94	0	0.000e-04	2.364	0.000e-04	41.0
01/01/95	2	-6.161e-04	2.364	0.000e-04	41.0
01/01/96	367	-5.088e-04	2.324	0.000e-04	41.0

Dates of launch of NOAA satellites

NOAA 11
September 24, 1988

NOAA 14
December 30, 1994

Drifts correction of the spectral radiance with $K_i^{\downarrow}$.

The gain and offset for band 1 and 2 are subject to drift. Calibration studies have shown that the sensor response drifts versus its own lifetime. The gain and the offset have to be corrected for this feature by the number of days between the launch (i.e. NOAA 14, December 12, 1994) and the date of acquisition. Bands 1 & 2 lie in the spectral range where reflectance is more pronounced than emittance. The spectral reflectance can be obtained after specifying the incoming radiance K_i

$$K_i^{\downarrow} = \frac{K_{exo_i}^{\downarrow} \cos\phi_{su}}{\pi d_s^2} \qquad (W.m^{-2}.\mu m^{-1})$$

With:

$K_i^{\downarrow}$ being the incoming radiance for band *I* $(W.m^{-2}.\mu m^{-1})$

$K_{exo\,i}^{\downarrow}$ the exo-atmospheric irradiance for band *i* $(W.m^{-2}.\mu m^{-1})$

Φ_{su} the Zenith angle *(rad)*

d_s Earth-Sun distance *(-)*

Note on calculations of $K_{exo\,i}^{\downarrow}$, ϕ_{su} and d_s

- $K_{exo\,i}^{\downarrow}$, the exo-atmospheric irradiance $(W.m^{-2}.\mu m^{-1})$ is fixed for NOAA satellites as following:

	NOAA 11	NOAA 14	Units
For band 1: $K_{exo1}^{\downarrow}$	1629	1605	$(W.m^{-2}.\mu m^{-1})$
For band 2: $K_{exo2}^{\downarrow}$	1053	1029	$(W.m^{-2}.\mu m^{-1})$

- ϕ_{su} , the Zenith angle *(rad)* is determined following this equation:

$$Cos\phi_{su} = Sin(\delta) \times Sin(latitude) + Cos(\delta) \times Cos(latitude) \times Cos(\omega_s) \qquad \textit{(rad)}$$

With:

1. δ (rad) being the solar declination, the angular height of the sun above the astronomical equatorial plane.

$$\delta = 0.4093 \times \sin(\frac{2\pi}{365} J - 1.39) \qquad \text{(rad)}$$

   J being the Julian day number.

2. ω_s the solar angle hour varying following the time of the day.

$$\omega_s = \pi \left(\frac{t_{GMT} - 12}{12} \right) \qquad \text{(rad)}$$

   the time of the day *t* being in decimal hour, as per $t_{GMT} = t_{local} - long\left(\frac{24}{2\pi}\right)$

- d_s the distance Earth-Sun varying with the Julian day number *J*.

$$d_s = 1 + 0.01672 \times \sin\left(\frac{2\pi(J - 93.5)}{365} \right) \qquad (\text{-})$$

attention should be taken to calculate the sine of the terms in parentheses in Radian mode, ***<u>not in Degree mode!</u>***

FOR WWW.SAA.NOAA.GOV

Calibration coefficients are included in the header of the image file, any software can include these coefficients in their import procedure as they are offset and intercept of a linear regression. Erdas/Imagine import module does provide with this calibration. However, reflectance values outputs are ranging from 0 to 100. To use them in the 0.0-1.0 range, it has to be divided by 100.

MORE TO BE DESCRIBED ON THE EASY WAY TO RETRIEVE MANUALLY THOSE COEFFICIENTS!!!!

The main equation is:

$$\rho_i^{TOA} = \frac{L_i^{TOA}}{K_i^{\downarrow}} \qquad (-)$$

With:

ρ_i^{TOA} being the spectral reflectance at the top of the atmosphere (-)

L_i^{TOA} the spectral radiance at the top of the atmosphere $(W.m^{-2}.\mu m^{-1})$

$K_i^{\downarrow}$ the incoming spectral radiance $(W.m^{-2}.\mu m^{-1})$

where $_i$ is the band number of the satellite (band 1 and 2)

Spectral radiance at the top of the atmosphere

The data received is corrected by a linear equation involving the onboard calibration instruments of the sensor. This is following the "traditional" equation:

$$L_i^{TOA} = a_i \times DN_i + b_i \qquad (W/m^2/m)$$

With:

L_i^{TOA} the spectral radiance at the top of the atmosphere for band *i* $(W/m^2/m)$

a_i the gain in band i (-) and b_i the offset in band i $(W/m^2/m)$

DN_i the Digital Numbers (-)

where $_i$ is the band number of the satellite

Definition of the gain a_i and offset b_i

The definition of the gain a_i and offset b_i is dependent on calibration instruments parameters **(available in the header file?)**

$$a_i = \frac{\left(R_{sp}^i - R_{bb}^i\right)}{\left(DN_{sp}^i - DN_{bb}^i\right)} \qquad (-)$$

$$b_i = R_{sp}^i - a_i \times DN_{sp}^i \qquad \left(W.m^{-2}.sr^{-1}.\mu m\right)$$

With:

a_i the gain in band i (-) and b_i the offset in band i $\left(W.m^{-2}.sr^{-1}.\mu m\right)$

R_{sp}^i the space radiation in band I $\left(W.m^{-2}.sr^{-1}.\mu m\right)$

R_{bb}^i the blackbody radiance of band i $\left(W.m^{-2}.sr^{-1}.\mu m\right)$

DN_{sp}^i the Digital Number Space value of band i *(-)*

DN_{bb}^i the Digital Numbers Blackbody value of band i *(-)*

where $_i$ is the band number of the satellite

The Top of atmosphere reflectance is:

$$\rho_i^{TOA} = \frac{L_i^{TOA}}{K_i^{\downarrow}} \qquad (-)$$

With:

ρ_i^{TOA} being the planetary spectral reflectance at the top of the atmosphere (-)

L_i^{TOA} the spectral radiance at the top of the atmosphere $(W.m^{-2}.\mu m^{-1})$

$K_i^{\downarrow}$ the incoming spectral radiance $(W.m^{-2}.\mu m^{-1})$

where $_i$ is the band number of the satellite (band 1, 2, 3, 4, 5 and 7)

In order to get to ρ_i^{TOA}, it is first to go to the two components of the ratio. Let us start with L_i^{TOA}, then finishing with $K_i^{\downarrow}$.

The spectral radiance at the top of the atmosphere, L_i^{TOA} .

$$L_i^{TOA} = \left(\frac{a + (b - a) \times DN_i}{255} \right) \qquad (W.m^{-2}.sr^{-1}.\mu m)$$

With:

L_i^{TOA} the spectral radiance at the top of the atmosphere for band *I* $(W.m^{-2}.sr^{-1}.\mu m)$

(b-a) the gain (-) and *a* the offset $(W.m^{-2}.sr^{-1}.\mu m)$

DN_i the Digital Numbers (-)

where $_i$ is the band number of the satellite (band 1 and 2)

The gain and offset values for Landsat 5TM satellite images have been considered constant in that study, the author taking reference from the work of Markham and Baker (1987).

	A	b
Band 1	-0.15	15.21
Band 2	-0.28	29.68
Band 3	-0.12	20.43
Band 4	-0.15	20.62
Band 5	-0.037	2.719
Band 7	-0.015	1.438

Drifts correction of the spectral radiance with $K_i^{\downarrow}$.

The spectral reflectance ρ_i^{TOA} can be obtained after specifying the incoming radiance $K_i^{\downarrow}$

$$K_i^{\downarrow} = \frac{K_{exo_i}^{\downarrow} \cos\phi_{su}}{\pi d_s^2} \qquad (W.m^{-2}.\mu m^{-1})$$

With:

$K_i^{\downarrow}$ being the incoming radiance for band *i* $(W.m^{-2}.\mu m^{-1})$
$K_{exo\,i}^{\downarrow}$ the exo-atmospheric irradiance for band *i* $(W.m^{-2}.\mu m^{-1})$
Φ_{su} the Zenith angle *(rad)*
d_s Earth-Sun distance *(-)*
where $_i$ is the band number of the satellite

Note on calculations of $K_{exo\,i}^{\downarrow}$, ϕ_{su} and d_s

- $K_{exo\,i}^{\downarrow}$, the exo-atmospheric irradiance $(W.m^{-2}.\mu m^{-1})$ is fixed for Landsat 5TM as following:

	Landsat 5TM	**Units**
For band 1: $K_{exo1}^{\downarrow}$	195.8	$(W.m^{-2}.\mu m^{-1})$
For band 2: $K_{exo2}^{\downarrow}$	182.8	$(W.m^{-2}.\mu m^{-1})$
For band 3: $K_{exo3}^{\downarrow}$	155.9	$(W.m^{-2}.\mu m^{-1})$
For band 4: $K_{exo4}^{\downarrow}$	104.5	$(W.m^{-2}.\mu m^{-1})$
For band 5: $K_{exo5}^{\downarrow}$	21.91	$(W.m^{-2}.\mu m^{-1})$
For band 7: $K_{exo7}^{\downarrow}$	7.457	$(W.m^{-2}.\mu m^{-1})$

- ϕ_{su} , the Zenith angle *(rad)* is determined following this equation:

$$Cos\phi_{su} = Sin(\delta) \times Sin(latitude) + Cos(\delta) \times Cos(latitude) \times Cos(\omega_s) \qquad (rad)$$

With:

3. δ (rad) being the solar declination, the angular height of the sun above the astronomical equatorial plane.

$$\delta = 0.4093 \times \sin(\frac{2\pi}{365} J - 1.39) \qquad \text{(rad)}$$

J being the Julian day number.

4. ω_s the solar angle hour varying following the time of the day.

$$\omega_s = \pi \left(\frac{t_{GMT} - 12}{12} \right) \qquad \text{(rad)}$$

the time of the day *t* being in decimal hour, as per $t_{GMT} = t_{local} - long\left(\frac{24}{2\pi} \right)$

- d_s the distance Earth-Sun varying with the Julian day number *J*.

$$d_s = 1 + 0.01672 \times \sin\left(\frac{2\pi(J - 93.5)}{365} \right) \qquad (-)$$

attention should be taken to calculate the sine of the terms in parentheses in Radian mode, *not in Degree mode!*

The main equation is:

$$\rho_i^{TOA} = \frac{L_i^{TOA}}{K_i^{\downarrow}} \qquad (-)$$

With:

ρ_i^{TOA} being the planetary spectral reflectance at the top of the atmosphere for band I (-)

L_i^{TOA} the spectral radiance at the top of the atmosphere $(W.m^{-2}.\mu m^{-1})$

$K_i^{\downarrow}$ the incoming spectral radiance $(W.m^{-2}.\mu m^{-1})$

where $_i$ is the band number of the satellite (band 1, 2, 3, 4, 5 and 7)

In order to get to ρ_i^{TOA}, it is first to go to the two components of the ratio. Let us start with L_i^{TOA}, then finishing with $K_i^{\downarrow}$.

Spectral radiance at the top of the atmosphere L_i^{TOA}.

$$L_i^{TOA} = a + (b \times DN_i) \qquad (W.m^{-2}.sr^{-1}.\mu m$$

With:

L_i^{TOA} the spectral radiance at the top of the atmosphere for band *I* $(W.m^{-2}.sr^{-1}.\mu m$

b the gain (-) and *a* the bias $(W.m^{-2}.sr^{-1}.\mu m$

DN_i the Digital Numbers (-)

where $_i$ is the band number of the satellite (band 1 and 2)

The gain and offset values for Landsat 7ETM+ satellite images are extracted from the header files available with each CD ROM product delivered. In case they are not available (if your image files are in geoTIFF format), you should refer to the Web site of Landsat 7ETM+, for the updated Calibration Parameters Files (CPF).

Correction of the spectral radiance

The spectral reflectance can be obtained after specifying the incoming radiance $K_i^{\downarrow}$

$$K_i^{\downarrow} = \frac{K_{exo_i}^{\downarrow} \cos \phi_{su}}{\pi d_s^2} \qquad (W.m^{-2}.\mu m^{-1})$$

With:

$K_i^{\downarrow}$ being the incoming radiance for band *i* $(W.m^{-2}.\mu m^{-1})$
$K_{exo\,i}^{\downarrow}$ the exo-atmospheric irradiance for band *i* $(W.m^{-2}.\mu m^{-1})$
Φ_{su} the sun Zenith angle *(rad)*
d_s Earth-Sun distance *(-)*
where $_i$ is the band number of the satellite

Note on calculations of $K_{exo\,i}^{\downarrow}$, ϕ_{su} and d_s

- $K_{exo\,i}^{\downarrow}$, the exo-atmospheric irradiance $(W.m^{-2}.\mu m^{-1})$ is fixed for Landsat 7ETM+ as following:

	Landsat 7ETM+	**Units**
For band 1: $K_{exo1}^{\downarrow}$	1970	$(W.m^{-2}.\mu m^{-1})$
For band 2: $K_{exo2}^{\downarrow}$	1843	$(W.m^{-2}.\mu m^{-1})$
For band 3: $K_{exo3}^{\downarrow}$	1555	$(W.m^{-2}.\mu m^{-1})$
For band 4: $K_{exo4}^{\downarrow}$	1047	$(W.m^{-2}.\mu m^{-1})$
For band 5: $K_{exo5}^{\downarrow}$	227.1	$(W.m^{-2}.\mu m^{-1})$
For band 7: $K_{exo7}^{\downarrow}$	80.53	$(W.m^{-2}.\mu m^{-1})$

- ϕ_{su} , the Zenith angle *(rad)* is determined following this equation:

$$\phi_{su} = (90 - Elevation_{sun}) \times \frac{\pi}{180} \qquad (rad)$$

The sun elevation angle is available in the header file. The area of Landsat 7ETM+ is comparatively small in proportion of the solar angle variation on Earth surface, therefore using only one value per image is significant.

- d_s the distance Earth-Sun varying with the Julian day number *J*.

$$d_s = 1 + 0.01672 \times \sin\left(\frac{2\pi(J - 93.5)}{365}\right) \qquad (-)$$

attention should be taken to calculate the sine of the terms in parentheses in Radian mode, ***not in Degree mode!***

3.1.2.Top of Atmosphere Broadband Albedo

In this part, you will calculate the Albedo for an aggregated bandwidth from different visible and near infrared bands. This computation is automatic, because of the standard weight of each band. This will enable you to get the surface Albedo in the next section, an important element of SEBAL, and a widely used element in Remote Sensing generally.

Broadband Albedo (TOA) for NOAA AVHRR

The broadband Albedo at the top of the atmosphere is required to get the surface one. Therefore the following equation leads to the first step.

$$\rho^{TOA} = 0.035 + 0.545\rho_1^{TOA} + 0.32\rho_2^{TOA} \qquad \text{(-)}$$

With:

ρ^{TOA} being the top of atmosphere broadband Albedo *(-)*
ρ_i^{TOA} the top of atmosphere reflectance for band *i*. *(-)*

Broadband Albedo (TOA) for Landsat 5 TM and 7 ETM+

The following equation leads to the top of the atmosphere broadband Albedo.

$$\rho^{TOA} = 0.293\rho_1^{TOA} + 0.274\rho_2^{TOA} + 0.233\rho_3^{TOA} + 0.156\rho_4^{TOA} + 0.033\rho_5^{TOA} + 0.011\rho_7^{TOA} \quad (-)$$

With:

ρ^{TOA} being the top of atmosphere Broadband Albedo *(-)*
ρ_i^{TOA} the top of atmosphere reflectance for band *i*. *(-)*

3.1.3. Surface Broadband Albedo

The surface Albedo, an important element of SEBAL, and a widely used element in Remote Sensing generally is to be computed here. It needs some ground data, or experience in dealing with such computation. You will use some reference points to calculate this parameter. One should be an area without vegetation and dry (a desert or a beach), and the second is an open water area (a lake or the sea). Knowing their surface Albedo values will enable a regression to stretch the Top of Atmosphere image points up to the surface values.

Broadband Surface Albedo Calculations

In order to get the broadband surface Albedo, expertise is required at this step, indeed it is to estimate the broadband path radiance (r_a) and the transmissivity of the atmosphere (τ_{sw}). How to determine them to input the following equation is explained in the Note below.

$$\rho_0 = \frac{\left(\rho^{TOA} - r_a\right)}{\tau_{sw}^2} \qquad (-)$$

With:

ρ_0 being the surface broadband Albedo	*(-)*
ρ^{TOA} the top of atmosphere broadband Albedo	*(-)*
r_a the broadband path radiance	*(-)*
τ_{sw} the transmissivity single-way crossed by radiation of the atmosphere	*(-)*

Note on broadband surface Albedo calculations

About the ranges of values for r_a and τ_{sw}, it should be stated that if the NOAA number for the image being processed is not known, or that previous stages were made with some doubts about coefficients; problems will arise at this stage. Indeed, ranges of data necessary to achieve the broadband Albedo test-point values will be extraordinary. These test areas are generally speaking, around 0.05 for the deep-water broadband Albedo and usually 0.35 for sand desert (for NOAA in Rajasthan), 0.25 (Landsat 5TM in Sri Lanka) or 0..35 for saline areas (landsat 7ETM+ in Pakistan's Punjab).

About tuning r_a and τ_{sw}, a definition of them

R_a	*The reflected part of the sun radiation on the top of the atmosphere*	from 0.02 up to 0.06
τ_{sw}	*The efficiency ratio of radiation conservation of the atmosphere between its top and the Earth skin surface.*	from 0.55 up to 0.85

The sun angle has a strong effect on r_a, being maximized at full reflection of sun on the top of the atmosphere. It can easily be assessed by a display of the NOAA image in 4/2/1 (normal color composition), showing a bright yellow zone covering some area of the image.

The vapor content and the dust particles in the atmosphere are generally responsible for lost of efficiency of the atmosphere to conserve the radiation circulating through it. In cloudy conditions, after removal of cloud, the surrounding areas (varying with the spatial distribution of the clouds and their nature) will be consequently full of water vapor. The case is the same in dry times where sandstorms are traveling from deserts up to areas of interest. In these conditions the τ_{sw} will be reduced accordingly to the density the operator is assessing.

3.1.4.NDVI calculation

The Normalized Difference Vegetation Index is by far the most famous processed output from Satellite Remote Sensing. It is giving the indication on how much percentage of the ground is covered by vegetal elements. It is used in SEBAL in order to assess the vegetation surface roughness as a component of the resistance to momentum transport. It is also used to calculate the surface emissivity, a major input to the model.

The surface NDVI calculation is taking the surface reflectance from the visible bands as expressed below, this technique could be used only in Pakistan due to wide and homogeneous areas in the desert and lakes to calibrate the surface values:

$$NDVI = \frac{\rho_{0_2} - \rho_{0_1}}{\rho_{0_2} + \rho_{0_1}} \qquad (-)$$

With:

NDVI the Normalized Difference Vegetation Index *(-)*
ρ_{0_i} being the surface Albedo for band *i* *(-)*
where $_i$ is the band number of the satellite (band 1 and 2)

By lack of homogeneous areas to calibrate the ground data to the band-wise surface Albedo from the images, the Meteorological Department of Sri Lanka (Chandrapala, 2000), processed the NDVI out of the band-wise visible reflectance at the top of atmosphere. It is also the case of Bandara (1998), who calculated the $NDVI^{TOA}$ for Landsat 5 TM and validated it as representative of the Surface NDVI under very clear sky conditions. The same method is also applied to Landsat 7ETM+:

$$NDVI = \frac{\rho^{TOA_2} - \rho^{TOA_1}}{\rho^{TOA_2} + \rho^{TOA_1}} \qquad (-)$$

With:

NDVI the Normalized Difference Vegetation Index *(-)*
ρ_i^{TOA} the top of atmosphere reflectance for band *i*. *(-)*
$_i$ is the band number of the satellite (band 1 and 2 for NOAA, band 3 and 4 for Landsat)

Band-wise Albedo calculations

The calculation of the Albedo at the Earth surface involves two parts where first the single band surface reflectance is processed (i.e. Red and NIR bands) enabling the creation of the NDVI image, and in a second the step calculating the surface broadband Albedo from top of atmosphere reflectance.

Here has to be experienced the use of having a large area comprising of various type of geographical units. In the case of irrigation systems where often are large water bodies (dams, reservoirs, lakes…) and also very dry areas like desert, it is easy to set "tie-points" of steady surface reflectance. It has to be recalled at this stage that the scale factor is important while dealing with large area applications of Remote Sensing, therefore, the following system of "tie-points" is scientifically justified and correct.

The idea is to get the standard reference for surface reflectance values of extreme bodies like lakes and desert, that are to be fitted by a regression to the top of atmosphere reflectance values of the same places.

The following table is having the reference "tie-points" and their values in band 1 and 2 for the Pakistan study case, the method might be reused in other areas having sufficient data:

Body	**Long.**	**Lat.**	ρ_{0_1}	ρ_{0_2}
Water	67.6306	26.3994	0.0589	0.0348
Desert	69.4605	26.4552	0.3267	0.3436

The table below is showing the table to be filled by the operator from the top of atmosphere reflectance images, in order to fulfil the further steps i.e. the equation system solving.

Body	**Long.**	**Lat.**	ρ_1^{TOA}	ρ_2^{TOA}
Water	67.6306	26.3994	ρ_{w1}^{TOA}	ρ_{w2}^{TOA}
Desert	69.4605	26.4552	ρ_{d1}^{TOA}	ρ_{d2}^{TOA}

In the case of Pakistan, the following set of equation solving has been used, while the top of atmosphere values were directly used in Sri Lanka:

Band 1	**Band 2**
$0.0589 = a_1 + \left(\rho_{w1}^{TOA}\right) \times b_1$	$0.0348 = a_2 + \left(\rho_{w2}^{TOA}\right) \times b_2$
$0.3267 = a_1 + \left(\rho_{d1}^{TOA}\right) \times b_1$	$0.3267 = a_2 + \left(\rho_{d2}^{TOA}\right) \times b_2$

<u>Solving these equations lead to the following:</u>

$\rho_{0_1} = a_1 + \left(b_1 \times \rho_1^{TOA}\right)$ $\rho_{0_2} = a_2 + \left(b_2 \times \rho_2^{TOA}\right)$

With:

ρ_{0_i} the surface reflectance for band *i*. *(-)*

a_i and b_i the offset and the slope of the regression between ρ_{0_i} and ρ_i^{TOA} for band *i*. *(-)*

$\rho_{w_i}^{TOA}$ the top of atmosphere reflectance of water for band *i*. *(-)*

$\rho_{d_i}^{TOA}$ the top of atmosphere reflectance of desert for band *i*. *(-)*

3.1.5.Surface emissivity calculation

The Surface emissivity of physical bodies indicates the level of absorption of energy that those grey-bodies have. It is very important to know the grey-body component of a surface element in order to assess any energy component, as simple as temperature to some more complex as the Radiative Heat (Net Radiation).

Surface Emissivity estimated in the 8-14 µm range for sparse canopies

$$\varepsilon_0 = 1.009 + 0.047\,Ln(NDVI) \qquad (-)$$

With:

ε_0 the surface emissivity *(-)*

NDVI the Normalized Difference Vegetation Index *(-)*

Bandara (1998), states that *the application of* [this equation] *is restricted to measurements conducted in the range of NDVI = 0.16-0.74*. [This equation] *is not valid for water bodies with a low NDVI and high emissivity (* ε_0 *=0.99 to 1.0). The water bodies available at the images are irrigation reservoirs, rivers and the ocean. These water bodies can be assumed as black bodies (i.e.* ε_0 *= 1.0). And also values of* ε_0 *shall not be less than 0.9. These two conditions can be given with the facility of the software.*

3.2. Thermal Remote Sensing

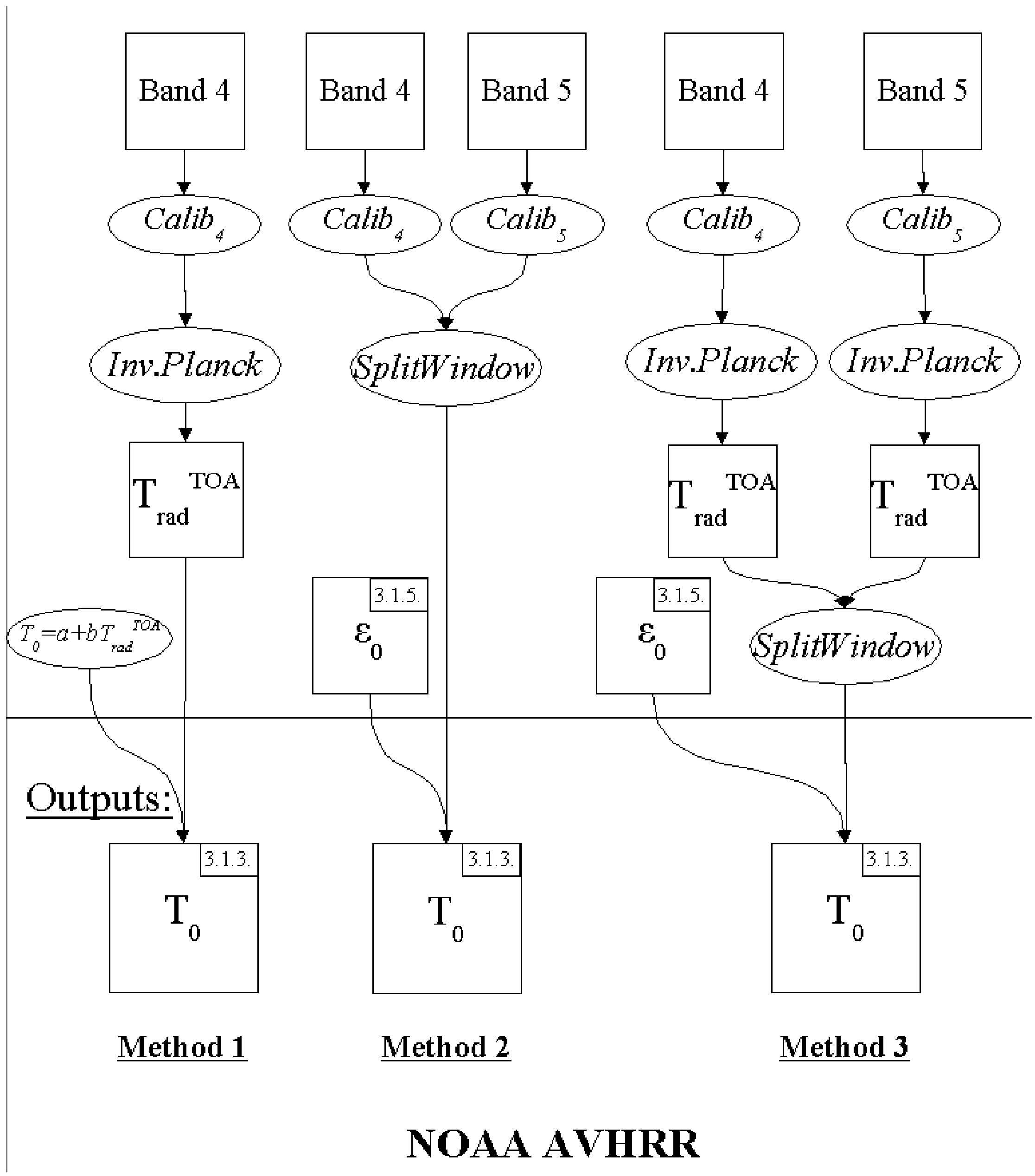


Figure 7: Production of Surface Temperature for NOAA AVHRR

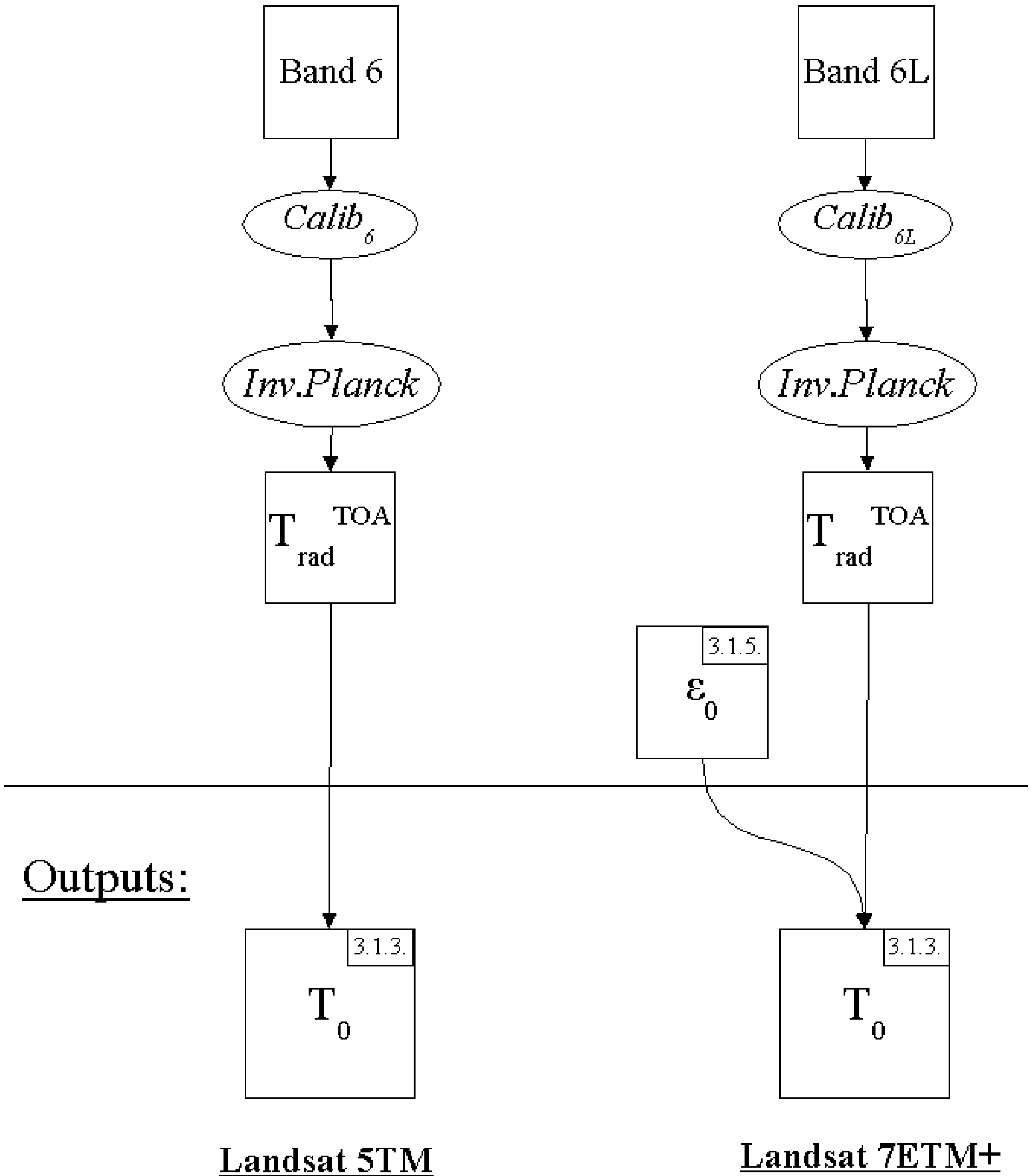


Figure 8: Production of surface temperature for Landsat

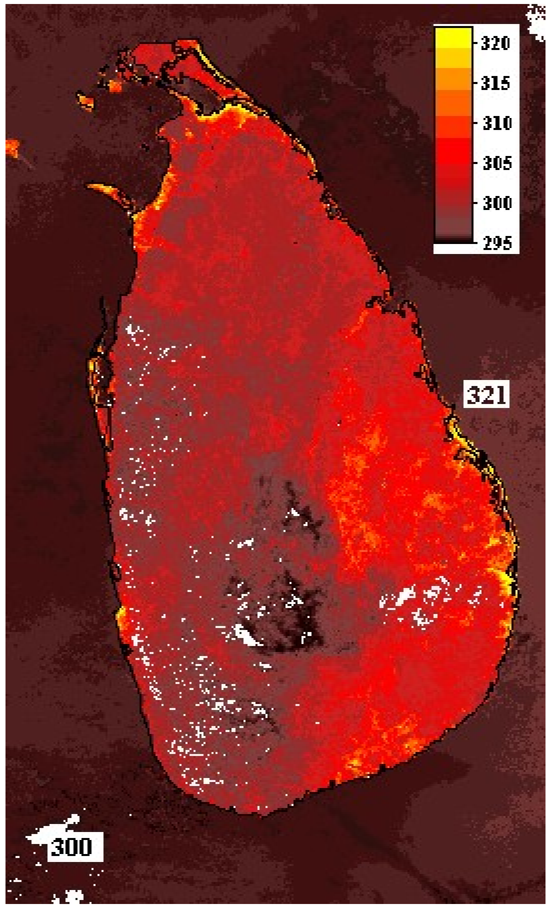


Figure 9: Surface Temperature over Sri Lanka (K)

<u>INPUTS:</u>

<u>Raster:</u>
Landsat: Band 6 for Landsat 5TM and Band 6L for Landsat 7ETM+.
NOAA: Band 4 and 5

<u>Tabular data:</u>
Calibration coefficients for each Band
Meteorological data (Tmin, Tmax, Relative Humidity)

<u>OUTPUTS:</u>

<u>Raster:</u>
Surface Temperature

Basically, the main output from the thermal remote sensing part is to have the surface temperature image. Different inputs are needed at this stage, some basic processing (correction and calibration of the thermal band) and the surface emissivity image (processed from NDVI).
The format of the data input being completely relevant to the corrections to be applied, the standard procedures are given below.

3.2.1. Calibration of the Thermal Bands

The aim of the calibration of the bands in the thermal range of the Electro-magnetic spectrum is to get radiance values at the Top of the Atmosphere, to be converted by the Inverse Planck function into Top of Atmosphere Brightness Temperature. Once this is done, several methods will be implemented in the next section to characterize the surface temperature from this information.

Q: Which method am I to use?
A: If you have a Remote Sensing Software that imports any image and does geometric and radiometric corrections automatically, then skip this Calibration Part.

Q: I do not have this import module for my image...
A: Then do a geometric correction (georeferencing) and go to the corresponding question below.

Q: I have acquired a raw NOAA AVHRR from a receiving station.
A: then use Method 1 for meteorological data supported method, or Method 3 for split window method.

Q: I did download NOAA AVHRR Level 1B from Internet.
A: then use Method 1 for meteorological data supported method, or Method 2 for split window supported method.

Q: I do have a CDROM of LandSat (5 TM or 7 ETM+).
A: Method 4 for LandSat 5TM and Method 5 for LandSat 7ETM+.

Top of Atmosphere corrections (Calculation of TOA Black Body Radiation)

Correction of band 4 drift

$$L_4^{TOA} = -0.0167 \times DN_4 + 11.93 \qquad (W/m^2/Sr)$$

With:

L_4^{TOA} the radiance at the top of the atmosphere $(W/m^2/Sr)$
DN_4 the Digital Number in Band 4 from the original data $(-)$

Black Body radiation at TOA for Band 4

$$B_4^{TOA}(\lambda_4, T) = L_4^{TOA} \times \pi \times 0.92783 \qquad (W/m^2)$$

With:

$B_4^{TOA}(\lambda_4, T)$ the Black Body Radiation at Top of Atmosphere in Band 4 (W/m^2)
L_4^{TOA} the radiance at the top of the atmosphere $(W/m^2/Sr)$

Remark: Use of Method 1 with NOAA data from the Web.

In case one wishes to use this method with NOAA data from www.saa.usgs.gov, non-Linear calibration of band 4 comes as:

$$L_4^{TOA} = 0.92378 \times DN_4^{Corr.} + 0.0003822 \times DN_4^{Corr.} + 3.72 \qquad (mW/m^2/Sr/cm)$$

With:

L_4^{TOA} the radiance in Band 4 at the top of the atmosphere with non-linear correction $(mW/m^2/Sr/cm)$

$DN_4^{Corr.}$ the digital number of Band 4 corrected linearly $(mW/m^2/Sr/cm)$

Adjustments of the L_4^{TOA} should be done to reach the Black Body Radiation at TOA in Band 4.

$$B_4^{TOA}(\lambda_4, T) = \frac{L_4^{TOA} \times 0.92783 \times \pi}{10} \qquad (W/m^2)$$

With:

$B_4^{TOA}(\lambda_4, T)$ the Black Body Radiation at Top of Atmosphere in Band 4 (W/m^2)

L_4^{TOA} the radiance at the top of the atmosphere with non-linear correction $(mW/m^2/Sr/cm)$

Surface temperature for NOAA AVHRR: Method 2
(see Chemin, Asif and Ahmad, 2000 on the Web)

For WWW.SAA.NOAA.GOV

Linear calibrations of the Band 4 and 5 are included in the import procedure of Erdas/Imagine.

Calculation of the surface temperature through the standard split-window technique should go through these inverse-Planck functions per band (POD Guide, 1998):

$$T_4^{TOA} = \frac{(1.438833 \times \lambda_4)}{Ln\left(\frac{(1.1910659 \times 10^{-5}) \times \lambda_4}{DN_4^{Corr.}} + 1\right)} \qquad (K)$$

$$T_5^{TOA} = \frac{(1.438833 \times \lambda_5)}{Ln\left(\frac{(1.1910659 \times 10^{-5}) \times \lambda_5}{DN_5^{Corr.}} + 1\right)} \qquad (K)$$

With:

T_i^{TOA} the Top of Atmosphere brightness temperature for band i *(K)*

$DN_i^{Corr.}$ the digital number corrected linearly for band i $(mW / m^2 / Sr / cm)$

λ_i The specific 290-330 K central wavelength for band i *(cm^{-1})*

(i.e. NOAA 14 = 929.5878 for band 4)

Calculation of the surface temperature through the standard split-window technique is performed through these inverse-Planck functions per band:

$$T_{bi}^{TOA} = \frac{C_2}{\lambda_i \times \left[Ln\left(\frac{C_1}{\lambda_i^5 \times L_i^{TOA}} + 1 \right) \right]} \qquad (K)$$

With:

T_{bi}^{TOA} the Top of Atmosphere brightness temperature for band *i* *(K)*

L_i^{TOA} the blackbody spectral radiance at the top of the atmosphere for band *i* $(W / m^2 / m)$

λ_i The central wavelength for band *i* *(m)*

$C_1 = 3.74 \times 10^{-16}$ W/m^2

$C_2 = 1.44 \times 10^{-2}$ mK

$_i$ is the band number of the satellite (band 4 and 5)

The top of atmosphere image has been processed following the inverse Planck function based on the outgoing spectral radiance of the band 6 at the Earth skin surface (after Roerink, 1995).

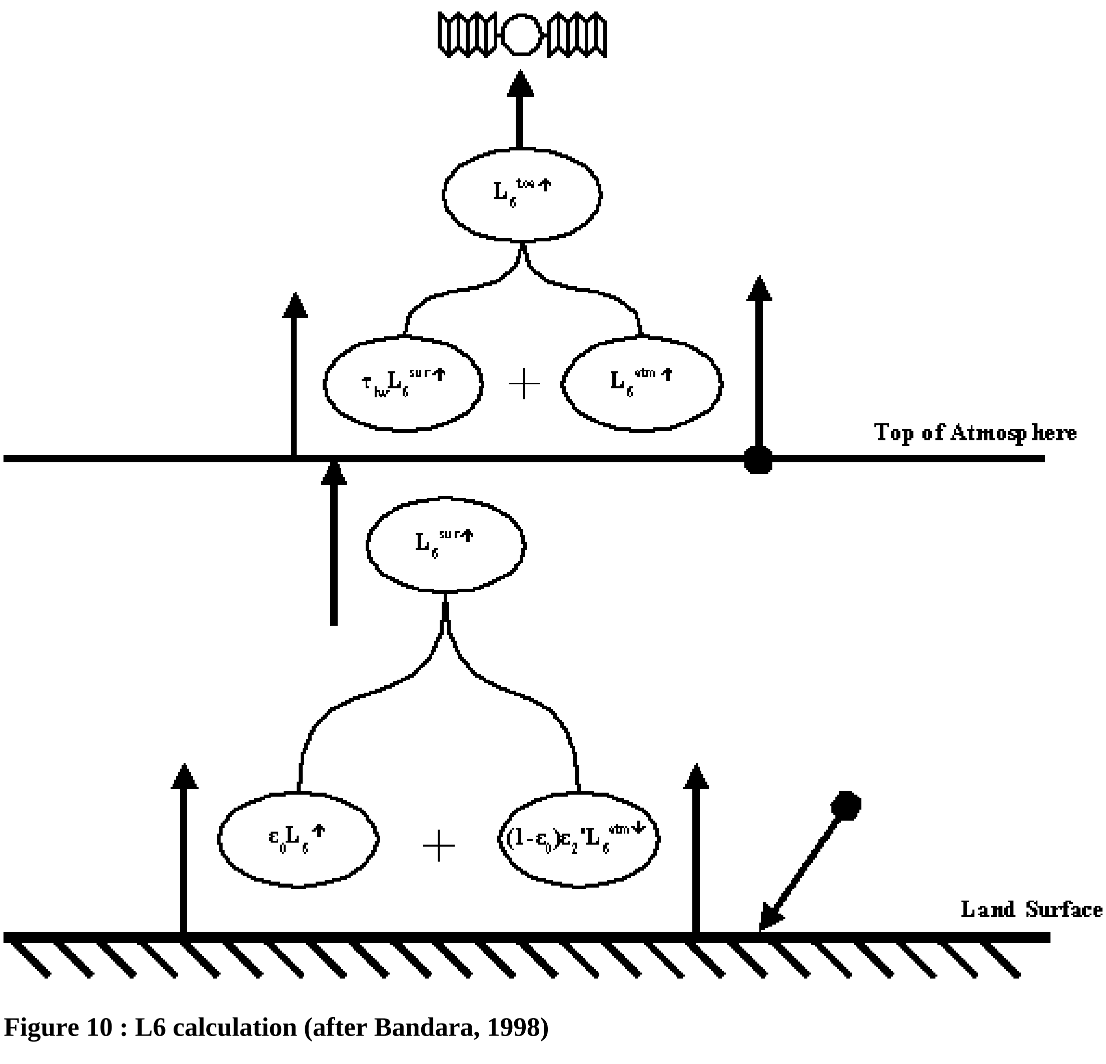


Figure 10 : L6 calculation (after Bandara, 1998)

Calculation of the emitted black body radiation in band 6 $L_6^{\uparrow}$

$$L_6^{\uparrow} = \frac{\left[\frac{L_6^{\uparrow TOA} - L_6^{\uparrow atm}}{\tau_{lw}}\right] - (1 - \varepsilon_6) \times \varepsilon_6^{'} \times L_6^{\downarrow atm}}{\varepsilon_6} \qquad (K)$$

With:

$L_6^{\uparrow}$ the emitted black body radiation in spectral width of band 6 *(W/m²)*

$L_6^{\uparrow TOA}$ the spectral radiance of band 6 at the top of the atmosphere (raster image) *(W/m²)*

$L_6^{\uparrow atm}$ the outgoing black body atmospheric long wave radiation in spectral width of band 6 *(W/m²)*

$L_6^{\downarrow atm}$ the incoming black body atmospheric long wave radiation in spectral width of band 6 (constant value) *(W/m²)*

τ_{lw} the atmospheric transmittance in spectral width of band 6 (atmospheric long wave transmittance) *(-)*

ε_6 the emissivity in spectral band 6, assumed equal to ε_0 *(-)*

$\varepsilon_6^{'}$ the apparent atmospheric emissivity in spectral band width 6, set as 0.845 for this study *(-)*

The spectral radiance of band 6 at the top of the atmosphere $L_6^{\uparrow TOA}$

$L_6^{\uparrow TOA}$ the spectral radiance of band 6 at the top of the atmosphere, used to calculate $L_6^{\uparrow}$, is in itself dependent on the following inverted Planck equation:

$$L_6^{\uparrow TOA} = \frac{3665.48}{\left[e^{\left(\frac{1235}{T_{TOA}}\right)} - 1 \right]} \qquad (W/m^2)$$

With:

$L_6^{\uparrow TOA}$ the spectral radiance of band 6 at the top of the atmosphere (raster image) *(W/m²)*

T_{TOA} the top of atmosphere temperature (for this study: the DN values in band 6 have been considered the degree Celsius values) *(K)*

The incoming black body atmospheric long wave radiation in band 6, $L_6^{\downarrow atm}$.

The incoming black body atmospheric long wave radiation in spectral width of band 6, $L_6^{\downarrow atm}$, considered as a constant in this study, is defined by this equation.

$$L_6^{\downarrow atm} = \frac{2.1 \times 3.74 \times 10^8 \times (11.65)^{-5}}{\left[e^{\left(\frac{1.44 \times 10^4}{11.65 \times T_{atm}} \right)} - 1 \right]} \qquad (W/m^2)$$

With:

$L_6^{\downarrow atm}$ the incoming black body atmospheric long wave radiation in spectral width of band 6 *(W/m^2)*

T_{atm} the air temperature (considered constant at 27C for this study) *(K)*

Determination of τ_{lw} and $L_6^{\uparrow atm}$

The determination of the atmospheric transmittance in spectral width of band 6 (τ_{lw}) and the outgoing black body atmospheric long wave radiation in spectral width of band 6 ($L_6^{\uparrow atm}$) is made by the resolution of a quadratic equation following these terms:

$$L_{6A}^{\uparrow TOA} = L_{6A}^{sur} \times \tau_{lw} + L_6^{\uparrow atm} \qquad (W/m^2)$$

$$L_{6B}^{\uparrow TOA} = L_{6B}^{sur} \times \tau_{lw} + L_6^{\uparrow atm} \qquad (W/m^2)$$

With:

$L_6^{\uparrow TOA}$ the spectral radiance of band 6 at the top of the atmosphere for points A and B (from the raster image) *(W/m^2)*

$L_6^{\uparrow sur}$ the spectral radiance of band 6 emitted from the surface, being calculated from meteorological stations measurements for points A and B. *(W/m^2)*

τ_{lw} the atmospheric transmittance in spectral width of band 6 (atmospheric long wave transmittance), being a constant. *(-)*

$L_6^{\uparrow atm}$ the outgoing black body atmospheric long wave radiation in spectral width of band 6, being a constant. *(W/m^2)*

For each of the points A and B, the following calculation of $L_6^{\uparrow sur}$ has taken place.

$$L_6^{\uparrow sur} = \varepsilon_6 \times L_6^{\uparrow} + (1 - \varepsilon_6) \times \varepsilon_6^{'} \times L_6^{\downarrow atm} \qquad (W/m^2)$$

With:

$L_6^{\uparrow sur}$ the spectral radiance of band 6 emitted from the surface, being calculated from meteorological stations measurements for points A and B. *(W/m²)*

ε_6 the emissivity in spectral band 6, assumed equal to ε_0 *(-)*

$L_6^{\uparrow}$ the emitted black body radiation in spectral width of band 6 *(W/m²)*

$\varepsilon_6^{'}$ the apparent atmospheric emissivity in spectral band width 6, set as 0.845 for this study *(-)*

$L_6^{\downarrow atm}$ the incoming black body atmospheric long wave radiation in spectral width of band 6 (constant value) *(W/m²)*

The emitted black body radiation in band 6 ($L_6^{\uparrow}$)

The calculation of the emitted black body radiation in spectral width of band 6 ($L_6^{\uparrow}$) is performed by applying the Planck's law with input from the surface temperature (T_0) for each point A and B.

$$L_6^{\uparrow} = \frac{2.1 \times 3.74 \times 10^8 (11.65)^{-5}}{\left[e^{\left(\frac{1.44 \times 10^4}{11.65 \times T_0} \right)} - 1 \right]} \qquad (W/m^2)$$

With:

$L_6^{\uparrow}$ the emitted black body radiation in spectral width of band 6 (for points A and B) *(W/m²)*

T_0 the surface skin temperature ($T_0 = T_{atm} + 3$, from meteorological station) *(K)*

Correction of the surface emissivity on the radiance brightness temperature permits to complete the inversed Planck function that was only for emissivity of water. The quality of the Thermal band calibration gives very good results on estimating surface air temperature during a clear sky day.

$$T^{TOA} = \frac{1282.71}{Ln\left(\frac{666.09}{L_{6L}^{TOA}} + 1\right)} \qquad (K)$$

With:

T^{TOA} the radiance temperature at top of atmosphere *(K)*

L_{6L}^{TOA} the emitted black body radiation in spectral width of band 6L *(W/m²)*

The top of atmosphere image has been processed following the inverse Planck function based on the outgoing spectral radiance of the band 6L at the Earth skin surface.

The emitted black body radiation in spectral width of band 6L, L_{6L}^{TOA}.

$$L_{6L}^{TOA} = a + (b \times DN_{6L}) \qquad (W.m^{-2}.sr^{-1}.\mu m)$$

With:

L_{6L}^{TOA} the emitted black body radiation in spectral width of band 6L $(W.m^{-2}.sr^{-1}.\mu m)$

b the gain (-) and *a* the bias, both extracted from the header file $(W.m^{-2}.sr^{-1}.\mu m)$

DN_{6L} the Digital Numbers *(-)*

3.2.2. Calculating the Surface temperature

The Surface Temperature is used in several sciences, especially relating to Agriculture, Ocean and climatic studies. The Long-wave radiation or the top of atmosphere brightness temperature is transformed into surface skin temperature, according to the Inversed Planck Function, with some additional refinements according to the method (sometimes using the surface emissivity). In the case of the presence of two thermal bands (NOAA), a set of Split-window equations is used to the discretion of the Scientist.

	Q: Which method am I to use? A: The following methods are split in NOAA and LandSat. Q: I have a NOAA AVHRR image, but I also have some meteorological data that I want to use. A: then use Method 1. Q: I have a NOAA AVHRR image and no meteorological data. A: then use Method 2 or 3 (Split-Window Technique). Q: I have LandSat (5 TM or 7 ETM+) images. A: Method 4 for LandSat 5TM and Method 5 for LandSat 7ETM+.

Meteorological data and spreadsheet processing

In order to get the surface temperature, a number of steps have to be undertaken from meteorological data and some images products. The solving of the radiation emittance from the Earth will enable to calculate the surface temperature from the regression fitting with the spreadsheet data of temperature and the image data of the black body emittance at the top of the atmosphere.

Inputs needed

From images:

- TIR_TOA4
- ε_0

From stations:

- Tmin air, Tmax air, Tavg air (= [Tmin air + Tmax air]/2)
- RH

Spreadsheet processing

(1)	$E_{sat} = 6.11 \times e^{\left(\frac{17.27 \times (T_{\max\_air} - 273)}{T_{\max\_air} - 273 + 237.3}\right)}$	(mbar)
(2)	$E_{act} = \left(\frac{RH \times E_{sat}}{100}\right)$	(mbar)
(3)	$\varepsilon_{atm} = 1.34 - 0.14\sqrt{E_{act}}$	(-)
(4)	$T_0 = T_{avg\_air}$ + Wim’s table	(K)
(5)	$B_0(\lambda, T_0) = \frac{C_1}{10.778^5 \times e^{\left(\left(\frac{C_2}{10.778 \times T_0}\right) - 1\right)}}$	$(W / m^2 / \mu m)$
(6)	$\varepsilon_0 B_0(\lambda, T_0) = B_0(\lambda, T_0) \times 0.92783 \times \varepsilon_0$	(W / m^2)

(7)

$$B_{atm}(\lambda, T_{air}) = \frac{C_1}{10.778^5 \times e^{\left(\left(\frac{C_2}{10.778 \times T_{\max\_air}}\right) - 1\right)}} \qquad (W / m^2 / \mu m)$$

(8)

$$\varepsilon_{tm} B_{atm}(\lambda, T_{air}) = B_{atm}(\lambda, T_{air}) \times 0.92783 \times \varepsilon_{atm} \qquad (W / m^2)$$

(9)

Total surf. Emittance =

$$\varepsilon_0 B_0(\lambda, T_0) + (1 - \varepsilon_0) \times \varepsilon_{atm} B_{atm}(\lambda, T_{air}) \qquad (W / m^2)$$

(10)

$B_x(\lambda, T) =$[Total surf. Emittance] a + b $\quad (W / m^2)$

In-band black body radiation measured by NOAA at TOA (= TIR_TOA4)
The regression helps to find a and b parameters:
$a = \tau_{(\lambda)}$ in-band atmospheric transmittance *(-)*
$b = L(\lambda, T_{TOA})$ in-band path radiance (W / m^2)

(10 bis)

Total surf. Emittance = $\dfrac{B_x(\lambda, T) - b}{a}$ $\quad (W / m^2)$

The following table is the empirical relationship used in this study to relate Air temperature to soil's one (Bastiaanssen, personal communication, 1998).

T_{air}	**T_0**
10-15	+ 1
15-20	+ 3
20-25	+ 5
25-30	+ 7
30-35	+ 10
35 +	+ 13

Figure 11: Wim's Table

Raster image Processing

(11)	$\varepsilon_0 B_0(\lambda, T_0) =$ Total_surf._emittance $- \{(1-\varepsilon_0) \times [B_{atm}(\lambda, T_{air}) \times \varepsilon_0 B_0(\lambda, T_0) \times 0.92783]\}$ Where the term inside [ ] is the average of grey-body radiation from atmosphere, from the equation 8.	(W/m^2)
(12)	$B_0(\lambda, T_0) = \dfrac{\varepsilon_0 B_0(\lambda, T_0)}{\varepsilon_0}$	$(W/m^2/\mu m)$
(13)	$T_0 = \dfrac{1.4388 \times 10^4}{10.778 \times Ln\left[\dfrac{0.92783 \times 3.7427 \times 10^8}{B_0(\lambda, T_0) \times 10.778^5} + 1\right]}$	(K)

The temperature is then corrected for altitude by the use of the DEM

$$T_0\_dem = T_0 + \left(\frac{0.627}{100}\right) \times DEM \qquad (K)$$

With:

T_0_dem the surface temperature estimation DEM corrected *(K)*

T_0 the surface temperature estimation by split-window *(K)*

DEM the Digital Elevation Model of Pakistan (1Km x 1Km) *(m)*

The approach used in this evaporation study is taking the surface skin temperature corrected by the DEM, for all the steps requiring surface skin temperature inputs.

The split-window equation for the Skin Surface Temperature without emissivity correction (considered as "A" in the following) is after Coll and Caselles, 1997:

$$T_0 = \left[\frac{(A)^4}{\varepsilon_0} \right]^{0.25} \qquad (K)$$

With A= [0.39 * $\left(T_{b4}^{TOA}\right)^2$] + (2.34 * T_{b4}^{TOA}) – (0.78 * T_{b4}^{TOA} * T_{b5}^{TOA}) – (1.34 * T_{b5}^{TOA})
+ [0.39 * $\left(T_{b5}^{TOA}\right)^2$] + 0.56

With:

T_0 the surface temperature estimation *(K)*

T_{bi}^{TOA} the Top of Atmosphere brightness temperature for band *i* *(K)*

ε_0 the surface emissivity *(-)*

The temperature is then corrected for altitude by the use of the DEM

$$T_0\_dem = T_0 + \left(\frac{0.627}{100} \right) \times DEM \qquad (K)$$

With:

T_0_dem the surface temperature estimation DEM corrected *(K)*

T_0 the surface temperature estimation by split-window *(K)*

DEM the Digital Elevation Model of Pakistan (1Km x 1Km) *(m)*

The approach used in this study is taking the surface skin temperature corrected by the DEM, for all the steps requiring surface skin temperature inputs.

Surface temperature for NOAA AVHRR: Method 3

(see Chandrapala, 2000)

The split-window technique is used (Bastiaanssen, personal communication, 1999):

$$T_0 = \left[\frac{\left(T_{b4}^{TOA} + 1.2 \times \left(T_{b4}^{TOA} - T_{b5}^{TOA}\right) + 2.2\right)^4}{\varepsilon_0} \right]^{0.25} \quad (K)$$

With:

T_0 the surface temperature estimation by split-window *(K)*

T_{bi}^{TOA} the Top of Atmosphere brightness temperature for band *i* *(K)*

ε_0 the surface emissivity *(-)*

The surface temperature thus obtained was thereafter corrected for elevation using a digital elevation model of Sri Lanka of the same resolution (1Km x 1Km) assuming an average lapse rate value of 6.27 deg.C/Km. The lapse rate value was obtained by fitting a linear regression model to the observed temperatures in Sri Lanka.

$$T_0\_dem = T_0 + \left(\frac{0.627}{100}\right) \times DEM \quad (K)$$

With:

T_0_dem the surface temperature estimation DEM corrected *(K)*

T_0 the surface temperature estimation by split-window *(K)*

DEM the Digital Elevation Model of Sri Lanka (1Km x 1Km) *(m)*

The approach used in this study is taking the surface skin temperature corrected by the DEM, for all the steps requiring surface skin temperature inputs.

The surface skin temperature is calculated as follow:

$$T_0 = \frac{1235}{Ln\left(\frac{3662.48}{L_6^{\uparrow}} + 1\right)} \qquad (K)$$

With:

T_0 the surface skin temperature *(K)*

$L_6^{\uparrow}$ the emitted black body radiation in spectral width of band 6 *(W/m²)*

The top of atmosphere image has been processed following the inverse Planck function based on the outgoing spectral radiance of the band 6 at the Earth skin surface (after Roerink, 1995).

The surface temperature, has then been extrapolated from the Wim's table (Figure 11)

$$T_{air} = \left[\frac{\left(T^{TOA}\right)^4}{\varepsilon_0} \right]^{0.25} \qquad (K)$$

With:

T_{air} the air temperature (assumed at 2m) *(K)*

T^{TOA} the radiance temperature at top of atmosphere *(K)*

ε_0 the surface emissivity *(-)*

Correction of the surface emissivity on the radiance brightness temperature permits to complete the inversed Planck function that was only for emissivity of water. The quality of the Thermal band calibration gives very good results on estimating surface air temperature during a clear sky day.

$$T^{TOA} = \frac{1282.71}{Ln\left(\frac{666.09}{L_{6L}^{TOA}} + 1\right)} \qquad (K)$$

With:

T^{TOA} the radiance temperature at top of atmosphere *(K)*

L_{6L}^{TOA} the emitted black body radiation in spectral width of band 6L *(W/m²)*

The top of atmosphere image has been processed following the inverse Planck function based on the outgoing spectral radiance of the band 6L at the Earth skin surface.

SEBAL

Processing

Chapter 4. Energy Balance terms

Chapter 5. Daily evaporation

4.Energy Balance Terms

The energy Balance can be summarized at an instant *t* by the following equation:

$$Q^* = G_0 + H_0 + \lambda E \qquad (W/m^2)$$

Where:

Q^* is the Net Radiation emitted from the Earth surface *(W/m^2)*
G_0 is the soil heat flux *(W/m^2)*
H_0 is the sensible heat flux *(W/m^2)*
λE is the latent heat flux, being the energy necessary to vaporise water *(W/m^2)*

All the interest in solving the energy balance is to get the last component of it, the latent heat flux (λE), resulting in the following equation at an instant *t*:

$$\lambda E = Q^* - G_0 - H_0 \qquad (W/m^2)$$

Where:

λE is the latent heat flux, being the energy necessary to vaporise water *(W/m^2)*
Q^* is the Net Radiation emitted from the Earth surface *(W/m^2)*
G_0 is the soil heat flux *(W/m^2)*
H_0 is the sensible heat flux *(W/m^2)*

Therefore, each component will be calculated one by one in order to solve the system.
Remark: All inputs of this part, *unless specified*, are factors taken at *t* instantaneous time. Considering *t* the instantaneous time of the satellite overpass.

4.1.Net Radiation

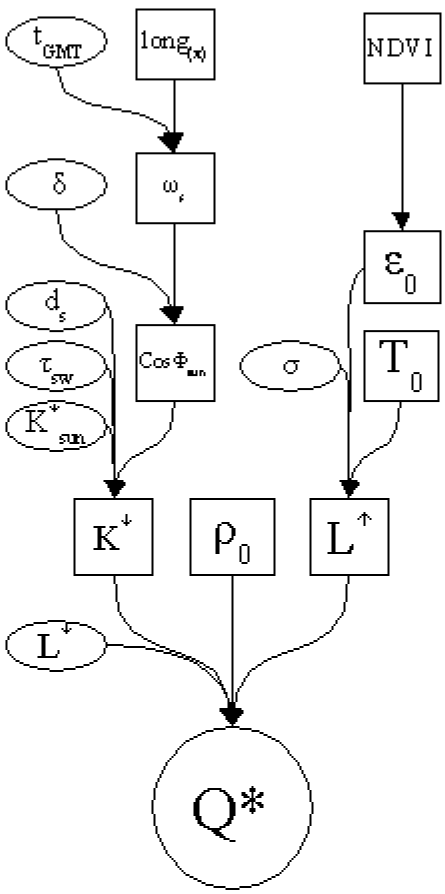


Figure 12: Overview of the Net Radiation

Figure 13: Net Radiation over Sri Lanka (W/m2)

INPUTS:

Raster:
Broadband surface Albedo
Emissivity
Longitude

Tabular data:
Sun external atmosphere radiation (band width dependent)
The atmosphere single-way transmissivity

OUTPUTS:

Raster:
Net Radiation

The net radiation Q^* is:

$$Q^* = K^{\downarrow} - K^{\uparrow} + L^{\downarrow} - L^{\uparrow}$$ *(W/m²)*

with:

Q^* being the Net Radiation *(W/m²)*

$K^{\downarrow}$ the incoming short-wave solar radiation *(W/m²)*

$K^{\uparrow}$ the outgoing short-wave solar radiation *(W/m²)*

$L^{\downarrow}$ the incoming broadband long-wave radiation *(W/m²)*

$L^{\uparrow}$ the outgoing broadband long-wave radiation *(W/m²)*

The incoming short-wave solar radiation $K^{\downarrow}$ is:

$$K^{\downarrow} = \frac{K^{\downarrow}_{sun} \times \cos\phi_{sun} \times \tau_{sw}}{d_s^2}$$ *(W/m²)*

with:

$K^{\downarrow}$ being the instantaneous solar radiation *(W/m²)*

$K^{\downarrow}_{sun}$ the sun external atmosphere radiation for the sensor's band width *(W/m²)*

$\cos\phi_{sun}$ the cosinus of the sun zenith angle (see) *(rad)*

τ_{sw} the atmosphere single-way transmissivity (determined by trial and error in the surface Albedo ρ_0 caculations, see 3.1.1) *(-)*

The outgoing short-wave solar radiation $K^{\uparrow}$:

$$K^{\uparrow} = \rho_0 \times K^{\downarrow} \qquad (W/m^2)$$

with:

$K^{\uparrow}$ being the outgoing short-wave solar radiation	*(W/m²)*
ρ_0 being the surface broadband Albedo	*(-)*
$K^{\downarrow}$ being the incoming short-wave solar radiation	*(W/m²)*

The outgoing broadband long-wave radiation $L^{\uparrow}$ is:

$$L^{\uparrow} = \varepsilon_0 \times \sigma \times T_0^4 \qquad (W/m^2)$$

with:

$L^{\uparrow}$ the outgoing broadband long-wave radiation	*(W/m²)*
ε_0 the surface emissivity	*(-)*
σ the Stefan-Boltzman constant (5.67x10^{-8})	*(W/m²/K⁴)*
T_0 the surface temperature	*(K)*

$L^{\downarrow}$ the incoming broadband long-wave radiation, appears to be defined differently according to the procedure used in the different projects, there are documented accordingly below.

	Q: Which method am I to use? A: The following methods are split in NOAA and LandSat. Q: I have a NOAA AVHRR image, but I also have some meteorological data that I want to use. A: then use Method 1. Q: I have a NOAA AVHRR image and no meteorological data. A: then use Method 2. Q: I have LandSat (5 TM or 7 ETM+) images. A: Method 3 for LandSat 5TM and LandSat 7ETM+.

The net radiation Q^* is:

This method has also been used for the Internet based data.

$$Q^* = (1 - \rho_0) \times (K^{\downarrow}) + L^{\downarrow} - (\varepsilon_0 \sigma T_0^4) - (1 - \varepsilon_0) \times L^{\downarrow} \qquad (W/m^2)$$

with:

Q^* being the Net Radiation *(W/m²)*

ρ_0 the surface reflectance *(-)*

$K^{\downarrow}$ the incoming shortwave solar radiation *(W/m²)*

$L^{\downarrow}$ the averaged incoming broadband long-wave radiation of atmosphere from different meteorological stations *(W/m²)*

ε_0 the surface emissivity *(-)*

T_0 the surface temperature *(K)*

ρ_0, $K^{\downarrow}$, ε_0 and T_0 are all raster images data.

Note on the incoming broadband long-wave radiation of atmosphere ($L^{\downarrow}$)

The incoming broadband long-wave radiation of atmosphere ($L^{\downarrow}$), is coming from the spreadsheet calculations in Surface Temperature for NOAA AVHRR: Method 1, where the Stefan-Boltzmann equation has to be applied for each meteorological station in order to be averaged for all station, giving $L^{\downarrow}$. Care should be taken in this regard to input the temperature values coming from the time being the closest to the instantaneous overpass time of the satellite.
The averaged incoming broadband long-wave radiation of atmosphere ($L^{\downarrow}$) has been calculated by determining the critical parameter ε_{atm} from the air temperatures and relative humidity values coming from different meteorological stations in the Pakistan Indus basin (see Bastiaanssen et al., 1999).

$$L^{\downarrow} = \sum_{i=2}^{i=n} \left[\varepsilon_{atm_i} \sigma T_{atm_i}^4 \right] \qquad (W/m^2)$$

with:

$L^{\downarrow}$ the averaged incoming broadband long-wave radiation of atmosphere from different meteorological stations *(W/m²)*

$\varepsilon_{atm\,i}$ the atmospheric emissivity at the meteorological station i *(-)*

σ the Stefan-Boltzman constant (5.67x10^{-8}) *(W/m²/K⁴)*

T_{atm_i} the atmospheric temperature at the meteorological station i *(K)*

Note on the calculation of the instantaneous incoming solar radiation $K^{\downarrow}$.

The incoming solar radiation at t instantaneous time is:

$$K^{\downarrow} = \frac{K^{\downarrow}_{sun} \times \cos\phi_{sun} \times \tau_{sw}}{d_s^2} \qquad (W/m^2/\mu m)$$

with:

$K^{\downarrow}$ being the instantaneous incoming solar radiation *(W/m²/µm)*
$K^{\downarrow}_{sun}$ the sun external atmosphere radiation (constant = 1358) *(W/m²)*
$\cos\phi_{sun}$ the cosine of the sun zenith angle (see 3.1.1) *(rad)*
τ_{sw} the atmosphere single-way transmissivity (determined by trial and error in the surface Albedo ρ_0 calculations, see 3.1.1) *(-)*

The net radiation $Q*$ is:

$$Q* = (1 - \rho_0) \times (K^{\downarrow}) + L^{\downarrow} - (\varepsilon_0 \sigma T_0^4) - (1 - \varepsilon_0) \times L^{\downarrow} \qquad (W/m^2)$$

with:

$Q*$ being the Net Radiation *(W/m²)*
ρ_0 the Broadband surface Albedo *(-)*
$K^{\downarrow}$ the incoming shortwave solar radiation *(W/m²)*
$L^{\downarrow}$ the averaged incoming broadband long-wave radiation *(W/m²)*
ε_0 the surface emissivity *(-)*
T_0 the surface temperature *(K)*

ρ_0, $K^{\downarrow}$, ε_0 and T_0 are all raster images data.

The averaged incoming broadband long-wave radiation of atmosphere ($L^{\downarrow}$):

$$L^{\downarrow} = \varepsilon_{atm} \times \sigma \times T_{atm}^4 \qquad (W/m^2)$$

with:

$L^{\downarrow}$ the averaged incoming broadband long-wave radiation of atmosphere (constant) *(W/m²)*
ε_{atm} the atmospheric emissivity (constant) *(-)*
σ the Stefan Boltzman Constant *(W/m²/K⁴)*
T_{atm} the average atmospheric temperature from the meteorological stations (constant) *(K)*

The atmospheric emissivity was generated out of the single-way transmissivity of the atmosphere after Bastiaanssen (1995):

$$\varepsilon_{atm} = 1.08 \times \left[- Ln(\tau_{sw}) \right]^{0.265} \qquad (-)$$

Bearing the range of application: $0.55 < \tau_{sw} < 0.82$

The incoming solar radiation at t instantaneous time is:

$$K^{\downarrow} = \frac{K^{\downarrow}_{sun} \times \cos\phi_{sun} \times \tau_{sw}}{d_s^2} \qquad (W/m^2/\mu m)$$

with:

$K^{\downarrow}$ being the instantaneous incoming solar radiation *(W/m²/µm)*
$K^{\downarrow}_{sun}$ the sun external atmosphere radiation (constant = 1380) *(W/m²)*
$\cos\phi_{sun}$ the cosinus of the sun zenith angle (see 3.1.1) *(rad)*
τ_{sw} the atmosphere single-way transmissivity (determined by trial and error in the surface Albedo ρ_0 caculations, see 3.1.1) *(-)*

Net Radiation for Landsat 5 TM and 7 ETM+
(see Bandara, 1998)

The net radiation $Q*$ is:

$$Q* = (1 - \rho_0) \times (K^{\downarrow}) + L^{\downarrow} - (\varepsilon_0 \sigma T_0^4) \qquad (W/m^2)$$

with:

$Q*$ being the Net Radiation *(W/m²)*
ρ_0 the surface reflectance *(-)*
$K^{\downarrow}$ the incoming shortwave solar radiation (see 3.1.1) *(W/m²)*
$L^{\downarrow}$ the averaged incoming broadband long-wave radiation of atmosphere from different meteorological stations *(W/m²)*
ε_0 the surface emissivity *(-)*
T_0 the surface temperature *(K)*

ρ_0, $K^{\downarrow}$, ε_0 and T_0 are all raster images data.

The averaged incoming broadband long-wave radiation of atmosphere ($L^{\downarrow}$):

$$L^{\downarrow} = \varepsilon_{atm} \times \sigma \times T_{atm}^4 \qquad (W/m^2)$$

with:

$L^{\downarrow}$ the averaged incoming broadband long-wave radiation of atmosphere from different meteorological stations *(W/m²)*
ε_{atm} the atmospheric emissivity (constant value, considered to be the averaged area effective apparent emissivity at 2m altitude) *(-)*
σ the Stefan Boltzman Constant *(W/m²/K⁴)*
T_{atm} the atmospheric temperature (raster image, $T_{atm} = T_0 - 3$) *(K)*

The atmospheric emissivity was generated out of the single-way transmissivity of the atmosphere after Bastiaanssen (1995):

$$\varepsilon_{atm} = 1.08 \times \left[- Ln(\tau_{sw})\right]^{0.265} \qquad (-)$$

Bearing the range of application: $0.55 < \tau_{sw} < 0.82$

4.2.Soil Heat Flux

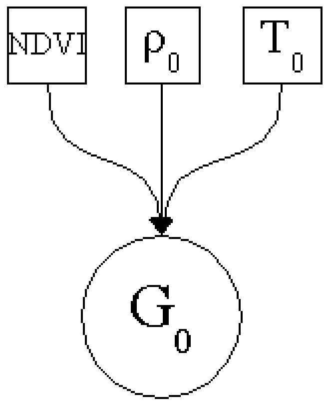


Figure 14: Overview of the Soil Heat Flux

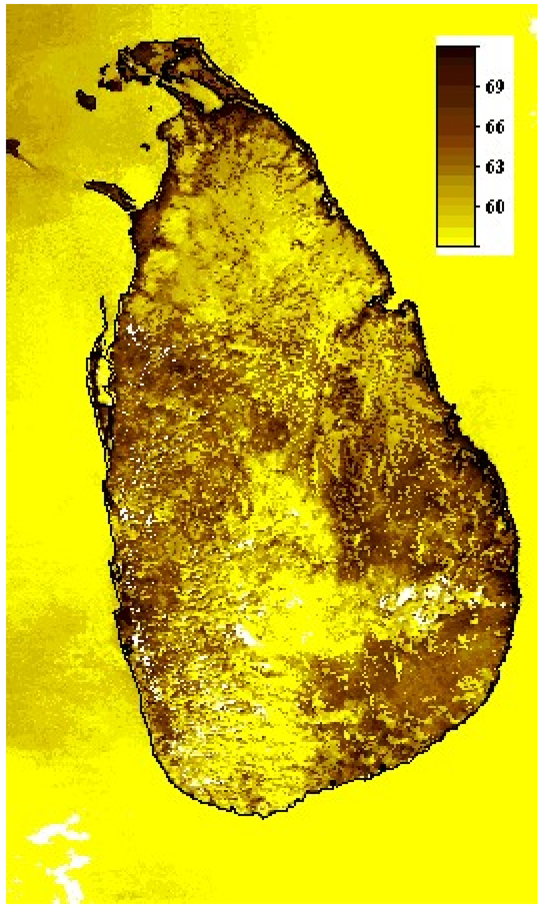


Figure 15: Soil Heat Flux over Sri Lanka (W/m2)

INPUTS:

Raster:
NDVI
Surface Albedo
Surface Temperature

OUTPUTS:

Raster:
Soil Heat Flux

The soil heat flux G_0 is:

$$G_0 = \frac{Q^* \times T_0}{\rho_0} \times \left(0.0032 r_0 + 0.0062 r_0^2\right) \times \left(1 - 0.978 \times NDVI^4\right) \qquad (W/m^2)$$

with:

G_0 being the Soil Heat Flux	*(W/m²)*
Q^* being the Net Radiation	*(W/m²)*
T_0 the surface temperature	*(°C)*
ρ_0 the surface reflectance	*(-)*
r_0 the day time average surface reflectance	*(-)*
NDVI the Normal Difference Vegetation Index	*(-)*

Note on the daytime average surface reflectance r_0.

To get r_0 while having the instantaneous surface Albedo (ρ_0), empirical relationship, valid only at large area is actually depending on the local time of overpass as shown below:

Local time of satellite overpass	**ρ_0 to r_0 relationship**
12.00	$\rho_0 = 0.9\ r_0$
14.00	$\rho_0 = r_0$
16.00	$\rho_0 = 1.1\ r_0$

4.3.Sensible Heat Flux

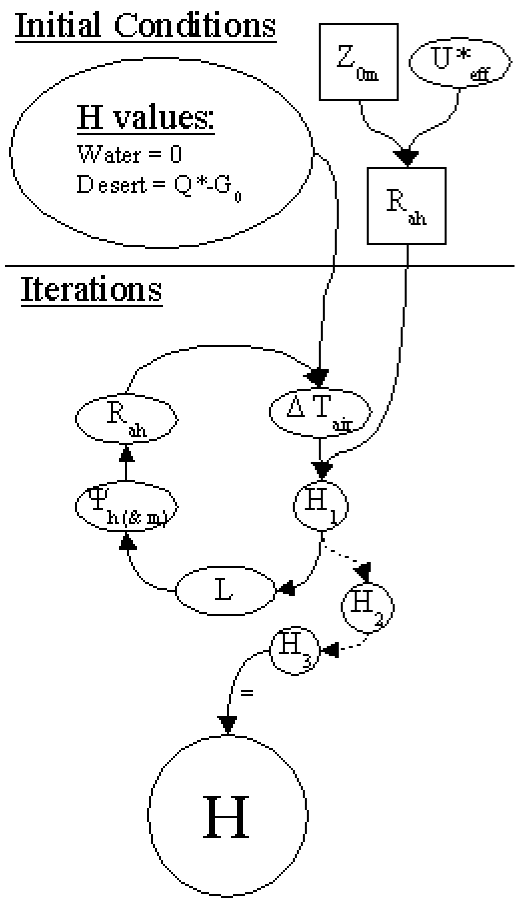


Figure 16: Overview of the Sensible Heat Flux

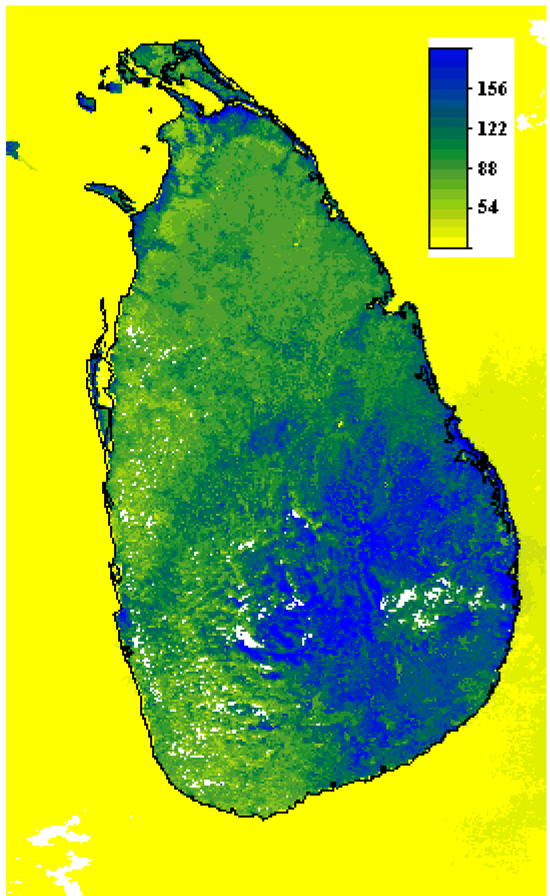


Figure 17: Sensible Heat Flux over Sri Lanka (W/m2)

<u>INPUTS:</u>

<u>Raster:</u>
Surface Roughness

<u>Tabular data:</u>
Net Radiation
Soil Heat Flux
Effective Friction Velocity

<u>OUTPUTS:</u>

<u>Raster:</u>
Sensible Heat Flux

Processing of the Sensible Heat Flux

The whole interest of iteration is to refine the value of a parameter, here it is the difference of Temperature between the Air and the surface skin (dT_{air}).
A good evaluation of this parameter is essential to assess the Sensible Heat Flux (H).

Only one equation is the core of the cycle.

$$H = \frac{\rho_{air} \times C_p \times dT_{air}}{r_{ah}} \qquad (W/m^2)$$

With:

H	being the Sensible Heat Flux	*(W/m²)*
ρ_{air}	the Atmospheric Air Density	*(Kg/m³)*
C_p	the Air Specific Heat at Constant Pressure	*(J/Kg/K)*
dT_{air}	the Temperature difference between air (2 m) and soil surface	*(K)*
r_{ah}	the Aerodynamic Resistance to Heat transport	*(s/m)*

The iterations will focus on refining the image processing of the Sensible Heat Flux (H) by repeating the determination of the soil-air temperature difference (dTair) as a linear function of the DEM adjusted soil surface temperature *(T_0_dem)*.
The Landsat study did not consider the DEM correction of the surface skin temperature, taking directly T_0 for the regressions fitting (after considering height differences and climate conditions).
The variations arising in the dTair calculations from the first iteration to the other ones are coming from the fact that r_{ah}, having psychometric buoyancy parameters for heat and momentum heat transfer can only be adjusted by iteration.

	Q: Which method am I to use? A: The following methods are sensor independent. Q: I have no meteorological data. A: then use Method 1. Q: I have meteorological data, which I want to use. A: then use Method 2. Q: I want to learn more about the sensible Heat Flux... A: Then enjoy the Method 3!

4.3.1.Calculation of the starting dTair

The difference of temperature between the soil surface and the air is evaluated by calculating the energy balance for two points, a desert or a beach, being the warmest and driest and a lake or the sea, being the coolest and most humid.

The maximum difference of temperature between the soil and the air will be assessed in those dry conditions, where the latent Heat Flux is null. The Sensible Heat Flux being the difference between the Net Radiation and the Soil Heat Flux.

In order to get a proper regression, a temperature difference on a wet point should be assessed. Theoretically a non-limited evaporation system is giving out all its energy into the latent Heat Flux, resulting in a Sensible Heat Flux of a null value.

From these two points, a regression is set after the next part to describe the variations of the dTair in function of the Surface Temperature pixels.

The number of iterations necessary to produce a proper level of Sensible Heat Flux quality is varying according to authors; it can be 5 for Tasumi and Allen (2000) to 9 for Hafeez and Chemin (2001). The latter however found another way to assess the quality that is dependent of the number of iterations and yet controls it, the difference of the dTair of the Hot pixel between two successive iterations, if inferior to 1 degree, is stopping the iteration system.

Calculation of the starting dTair

From extreme situations the following arises:

For Desert: $dT_{air} = \frac{(Q* - G_0) \times r_{ah1}}{\rho_{air} \times C_p}$ with $H = Q* - G_0$

For Water: $dT_{air} = 0$

A linear relationship can be drawn out from these two values, linking T_0_dem and dT_{air}, giving the initial H_1 image.

$Q*$ is a known raster image (calculated previously)
G_0 is a known raster image (calculated previously)
C_p is the Air specific Heat at constant pressure, $C_p = 1004$ J/Kg/K
ρ_{air} is the Atmospheric Air Density, explained below
r_{ah1} is the recipient of most of the efforts, and takes most of the calculation time from the operator. It is explained in length in the following sections.

- ρ_{air} the Atmospheric Air Density is responding to the added density of dry air and water vapor was taken as a constant ($\rho_{air} = 1.15\ Kg/m^3$) in the Sri Lankan NOAA project (see Chandrapala, 2000), same is the case for the Landsat study that used the following equation and related from a single temperature value considered representative of the study area. However, in the Pakistan case, the following equation has been used:

$$\rho_{air} = \left[\frac{P - \bar{e}_{act}}{T_{0-}\ dem \times 2.87}\right] + \left[\frac{\bar{e}_{act}}{T_{0-}\ dem \times 4.61}\right] \qquad (Kg/m^3)$$

With:

ρ_{air} the atmospheric air density *(Kg/m^3)*
P the atmospheric air pressure *(mbar)*
$\bar{e}_{act}$ mean actual water vapor pressure (from spreadsheet in) *(mbar)*
T_0_dem the surface temperature adjusted with the DEM *(K)*

Raster images inputs:

T_0*_dem* the surface temperature adjusted with the DEM altitude

From these, the Desert value of dT_{air} is calculated, making it possible to calculate the first Sensible Heat flux image (referred as H_1 in 4.3.3) from the linear relationship of dT_{air} with T_0*_dem* completed with the water and desert values of dT_{air}.

4.3.2.Calculation of the first R_{ah}

The Aerodynamic Resistance to Heat transport r_{ah1} is calculated on a first approximation in this part, though being very rough in the absence of the psychometrics parameters. Still it is helping to start the iteration cycle in order to refine subsequently all other parameters one by one.
A modified equation is used to calculate its first approximation, without correction from the psychometric parameters. See the corresponding sections about it in the following pages.

Calculation of the first R_{ah} for NOAA AVHRR: Method 1
(see Chemin and Ahmad, 2000)

A modified equation is used to calculate the first approximation of r_{ah1}, without correction from the psychometric parameters.

$$r_{ah1} = \frac{1}{U_5 \times 0.41^2} \times Ln\left(\frac{5}{z_{0m}}\right) \times Ln\left(\frac{5}{z_{0m} \times 0.1}\right) \qquad (s/m)$$

With:

r_{ah1} the Aerodynamic Resistance to Heat transport (<u>first approximation</u>) *(s/m)*
U_5 being the z = 5m wind velocity *(m/s)*
z_{0m} the aero-dynamical roughness length for momentum transport *(m)*

<u>Raster images inputs:</u>
U_5 the z = 5m wind velocity
z_{0m} the aero-dynamical roughness length for momentum transport

Calculation of z_{0m}

In the area of interest, being the irrigated Indus Basin, identify the maximum NDVI values. Once located, the crop type and stage should be determined to give the height of the crop. In this regard, agronomists and field survey officers have been very helpful, saving huge time and workload to get secondary data. The height of the max NDVI crop is related to the z_{0m} by the following relationship:

$$h_v = 7 \times z_{0m} \text{ for crop vegetation} \qquad (m)$$

Having the z_{0m} of the vegetation responding to the max NDVI, and assuming constant the Z_{0m} and NDVI values of the desert from the following table:

Land Cover	**NDVI**	**Z_{0m}**
Vegetation	$NDVI_{max}$	$\frac{h_v}{7}$
Desert	0.02	0.002

And the relating equation between NDVI and Z_{0m}, $Ln(z_{0m}) = a + (b \times NDVI)$, Ln being Neperian Logarithm.

Therefore two equations can be drawn out for each land cover type, enabling to solve the two parameters *a* and *b*:

Crop vegetation $\quad Ln\left(\frac{h_v}{7}\right) = a + (b \times NDVI_{max})$

Desert $\quad Ln(0.002) = a + (b \times 0.02)$

As being:

$$b = \frac{\left[Ln\left(\frac{h_v}{7}\right) - Ln(0.002)\right]}{(NDVI_{max} - 0.02)} \qquad \text{And} \qquad a = Ln(0.002) - \left[b \times (0.02)\right]$$

Finally, the Raster image of z_{0m} can be processed based on the NDVI one, following the equation below:

$$z_{0m} = EXP(a + b \times NDVI) \qquad (m)$$

Calculation of wind speed at z = 5m:

$$U_5 = \frac{U^*_{eff}}{0.41} \times \ln\left(\frac{5}{z_{0m}}\right) \qquad \textit{(m/s)}$$

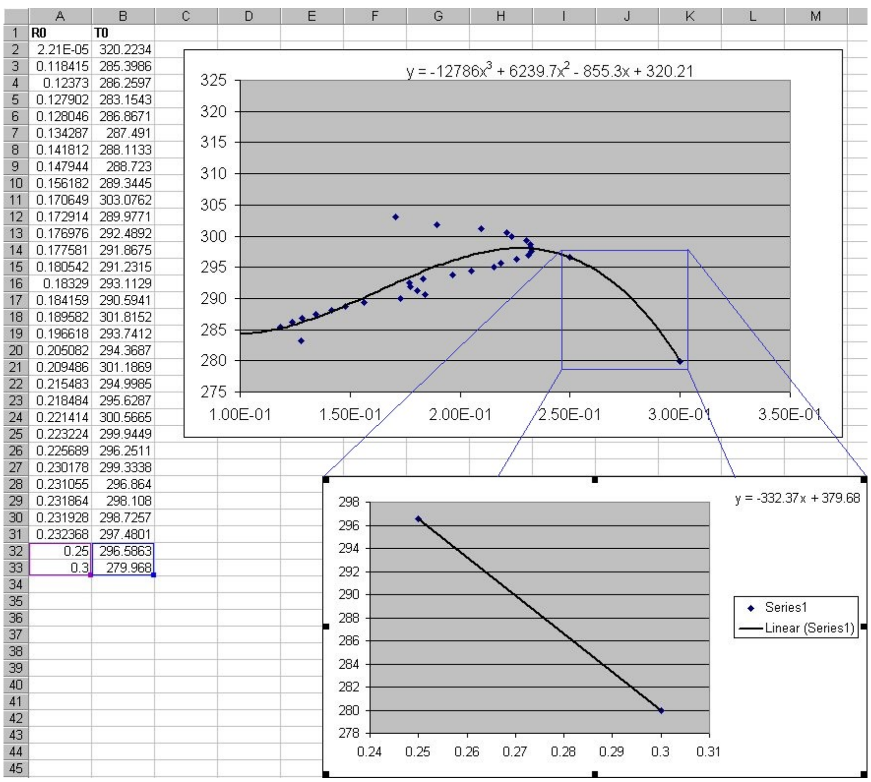
With:

U_5 being the z = 5m wind velocity *(m/s)*
U^*_{eff} the effective friction velocity *(m/s)*
z_{0m} the aero-dynamical roughness length for momentum transport *(m)*

Calculation of U^*_{eff}

$$U^*_{eff} = 1282.1 \times \left(\frac{\partial \rho_0}{\partial T_{0\_dem}} \right)^2 + 47.821 \times \left(\frac{\partial \rho_0}{\partial T_{0\_dem}} \right) + 0.45 \qquad \textit{(m/s)}$$

The parameter $\left(\frac{\partial \rho_0}{\partial T_{0\_dem}} \right)$ is calculated from the polynomial trend fitting regressive curve as an example given below:



	A	B
1	R0	T0
2	2.21E-05	320.2234
3	0.118415	285.3986
4	0.12373	286.2597
5	0.127902	283.1543
6	0.128046	286.8671
7	0.134287	287.491
8	0.141812	288.1133
9	0.147944	288.723
10	0.156182	289.3445
11	0.170649	303.0762
12	0.172914	289.9771
13	0.176976	292.4892
14	0.177581	291.8675
15	0.180542	291.2315
16	0.18329	293.1129
17	0.184159	290.5941
18	0.189582	301.8152
19	0.196618	293.7412
20	0.205082	294.3687
21	0.209486	301.1869
22	0.215483	294.9985
23	0.218484	295.6287
24	0.221414	300.5665
25	0.223224	299.9449
26	0.225689	296.2511
27	0.230178	299.3338
28	0.231055	296.864
29	0.231864	298.108
30	0.231928	298.7257
31	0.232368	297.4801
32	0.25	296.5863
33	0.3	279.968

Figure 18: U_{*eff} calculation

In this case for November 18th, 1993: $\left(\frac{\partial \rho_0}{\partial T_{0\_dem}} \right) = \frac{-1}{Gain} = \frac{-1}{-332.37} = 0.3181$

The method used to express r_0 in function of T_0 has different components.

First is to get the T_0 image corrected upward onto sea level from height variations:

$$T_0\_dem = T_0 + \left(\frac{0.627}{100}\right) \times DEM \qquad (K)$$

With:

T_0_dem the surface temperature estimation DEM corrected *(K)*
T_0 the surface temperature estimation by split-window *(K)*
DEM the Digital Elevation Model (1Km x 1Km) *(m)*

Note: Ultimately the T_0*_dem* images where used only for the sensible heat flux iteration procedure as the surface skin temperature input.

Second, is to follow this procedure in Erdas/Imagine:

1. Layer Stack the ρ_0 and T_0*_dem* images
2. Unsupervised classification with 30 classes
3. Signature editor, View/Columns/stats/means for each layer
4. Copy the two "means" columns and paste them in a spreadsheet

In the Spreadsheet, sort the columns on ρ_0 (called r_0 in Figure 18) in increasing order and draw a chart out of them, fit a polynomial (power 3) and reach Figure 18, after Roerink (1995).

Practically, U^*_{eff} is bounded from 0.1 to 0.45, any values out of these should be taken as their closest limit value.

A modified equation is used to calculate the first approximation of r_{ah1}, without correction from the psychometric parameters.

$$r_{ah1} = \frac{15.158}{U^*} \qquad \textit{(s/m)}$$

With:

r_{ah1} the Aerodynamic Resistance to Heat transport (first approximation) *(s/m)*
U^* being the nominal efficient friction velocity *(m/s)*

The friction velocity calculation, U^*.

To develop the average wind profile for Sri Lanka at the satellite overpass time, the wind speed estimated over Sri Lanka at 200 meters level was used. For this purpose, 10 m wind speed at reference site was used to estimate U_{200}.

Input: reference wind speed [m/s] at 10 m level in Sri Lanka = U_{10}

The surface roughness raster map is:

$$z_{0m} = e^{(-5.5 + 5.8 \times NDVI)} \qquad (m)$$

With:

z_{0m} the surface roughness *(m)*
NDVI the Normalized Difference Vegetation Index *(-)*

The reference site friction velocity is calculated by:

$$U^* = \frac{0.41 \times U_{10}}{Ln\left(\frac{10}{z_{0m}}\right)} \qquad (m/s)$$

With:

U^* the friction velocity for the reference site *(m/s)*
z_{0m} the surface roughness *(m)*
U_{10} the measured wind speed at height = 10 m *(m/s)*

The reference site nominal wind speed is calculated by:

$$U_{200} = \frac{U^* \times Ln \frac{200}{z_{0m}}}{0.41} \qquad (m/s)$$

With:

U_{200} the calculated wind speed at 200 m for the reference site *(m/s)*
U^* the friction velocity *(m/s)*
z_{0m} the surface roughness *(m)*

The nominal wind speed raster map is then calculated as the following:

$$U^* = \frac{0.41 \times U_{200}}{Ln\left(\frac{200}{z_{0m}}\right)} \qquad (m/s)$$

With:

U^* the friction velocity *(m/s)*
z_{0m} the surface roughness *(m)*
U_{200} the calculated wind speed at 200 m from the reference site *(m/s)*

A modified equation is used to calculate the first approximation of r_{ah1}, without correction from the psychometric parameters.

$$r_{ah1} = \frac{Ln\left(\frac{Z_{sur}}{0.01 \times Z_{0m}}\right)}{k \times U^*} \qquad (s/m)$$

With:

r_{ah1}	the Aerodynamic Resistance to Heat transport (<u>first approximation</u>)	*(s/m)*
z_{sur}	the height of study (z_{sur} = 10 m)	*(m)*
z_{0m}	the surface roughness (assuming $z_{0h} = 0.01 \times z_{0m}$)	*(m)*
U^*_{eff}	being the friction velocity	*(m/s)*

Calculation of z_{0m}

The calculation of z_{0m} is performed through an exponential relationship with the NDVI raster image.

$$z_{0m} = e^{(m_1 + m_2 \times NDVI)} \qquad (m)$$

With:

z_{0m} the surface roughness *(m)*
NDVI the Normalized Difference Vegetation Index *(-)*

The coefficients m_1 and m_2 have been used from similar studies and applied as best to this study. **(reference studies?)**

Calculation of the first R_{ah} for Landsat
(see Bandara, 1998)

Calculation of U^*_{eff}

$$U^*_{eff} = \frac{k \times \overline{U}_{Zsur}}{Ln\left(\frac{Z_{sur}}{Z_{0m}}\right)} \qquad \textit{(m/s)}$$

With:

U^*_{eff} the effective friction velocity *(m/s)*
$\overline{U}_{Zsur}$ the mean friction velocity for the height Z_{sur} (Z_{sur} = 10 m) *(m/s)*
z_{0m} the aero-dynamical roughness length for momentum transport *(m)*

The calculation of the mean friction velocity goes as:

$$\overline{U}_{Zsur} = U^*_{dry} \times \left[Ln\left(\frac{Z_{sur}}{Z_{0m}}\right) - \psi_m \right] \qquad \textit{(m/s)}$$

With:

$\overline{U}_{Zsur}$ the mean friction velocity for the height Z_{sur} (Z_{sur} = 10 m) *(m/s)*
U^*_{dry} the effective friction velocity for the “dry” pixels (see below) *(m/s)*
z_{0m} the aero-dynamical roughness length for momentum transport *(m)*
ψ_h the stability correction for the atmospheric heat transport *(-)*

Calculation of the first R_{ah} for Landsat
(see Bandara, 1998)

The stability correction for the atmospheric heat transport (φ_h) is calculated from the equation:

$$\psi_h = 2 \times Ln\left[\frac{1 + x^2}{2}\right] \qquad (-)$$

With $x = \left(1 - \dfrac{16 \times Z_{sur}}{L}\right)^{0.25}$

ψ_h the stability correction for the atmospheric heat transport *(-)*
L the Monin-Obukov Length *(m)*

The Monin-Obukov Length in this case is:

$$L = \frac{- \rho_{air} \times C_p \times \left(U^*_{dry}\right)^3 \times \overline{T_0}}{k \times g \times \dfrac{\overline{H_{dry}}}{2}} \qquad (m)$$

With:

L the Monin-Obukov Length *(m)*
ρ_{air} the Atmospheric Air Density *(Kg/m^3)*
C_p the Air Specific Heat at Constant Pressure (1004) *(J/Kg/K)*
U^*_{dry} the effective friction velocity for "dry" pixels (see below) *(m/s)*
$\overline{T_0}$ the mean surface skin temperature for all pixels *(K)*
k is the Von Karman's Constant (0.4) *(-)*
g is the gravitational acceleration (9.81) *($m.s^{-2}$)*
$\overline{H_{dry}}$ being the mean Sensible Heat Flux for "dry" pixels (here considering that $H = \dfrac{H_{dry}}{2}$) *(W/m^2)*

The Sensible Heat Flux for “dry” pixels ($\overline{H_{dry}}$):

$$\overline{H_{dry}} = \frac{1}{n} \times \left[\sum_{1}^{n} \left(Q^* - G_0 \right) \right]$$	*(W/m²)*

With:

$\overline{H_{dry}}$ being the mean Sensible Heat Flux for “dry” pixels	*(W/m²)*
Q^* being the Net Radiation for “dry” pixels	*(W/m²)*
G_0 being the Soil Heat Flux for “dry” pixels	*(W/m²)*

At this point in the calculation the only unknown is the effective friction velocity (U^*_{dry}) for the pixels considered “dry”.

A pixel was considered “dry” if it was having a surface skin temperature (T_0) superior to a given threshold value estimated by the researcher (T_0 > 310 K in that case study). Simultaneously the Surface Albedo had to be superior to another specific threshold (ρ_0 > 0.18 in this case) as it is unraveled in the coming calculation of $\left(\frac{\partial \rho_0}{\partial T_0} \right)$ described below.

Subsequently, raster maps of T_0 and ρ_0 were created, masking all values that are simultaneously answering to the two thresholds negatively. Any parameter having a subscript “dry” is processed out of the group of the “dry” pixels identified by the above-mentioned method.

Calculation of the effective friction velocity for “dry” pixels (U^*_{dry})

The calculation of U^*_{dry} is following an iteration process described in the following figure.

$$\psi_h = 0 \rightarrow U^*_{dry1} \rightarrow L_{dry1} \rightarrow x_1 \rightarrow \psi_{h1} \ (1)$$
$$\psi_{h1} \rightarrow U^*_{dry2} \rightarrow L_{dry2} \rightarrow x_2 \rightarrow \psi_{h2} \ (2)$$
$$\psi_{m2} \rightarrow U^*_{dry3} = U^*_{dry} \ (3)$$

Figure 19 : effective friction velocity for "dry" pixels (iteration)

The first run (1) is starting with the calculation of U^*_{dry1} under the consideration that ψ_h cannot be estimated at this time, therefore set to $\psi_h = 0$.

$$U^*_{dry1} = \frac{Ln\left(\frac{Z_{sur}}{\overline{Z_{oh\_dry}}}\right)}{k \times r_{ah\_dry}}$$ *(m/s)*

With:

U^*_{dry} the effective friction velocity for “dry” pixels *(m/s)*

U_{Zsur} the friction velocity for the height Z_{sur} (Z_{sur} = 100 m) *(m/s)*

$\overline{Z_{oh\_dry}}$ the mean aero-dynamical roughness height for momentum transport taken from the “dry” pixels of the Z_{0m} raster image: $\overline{Z_{oh\_dry}} = \frac{1}{n}\sum_1^n \left(\frac{Z_{0m\_dry}}{100}\right)$ *(m)*

r_{ah_dry} the Aerodynamic Resistance to Heat transport for “dry” pixels **(see xxx)** *(s/m)*

The input U^*_{dry} is then taken into the processing of the Monin-Obukov Length L_{1_dry}:

$$L_{1\_dry} = \frac{-\rho_{air} \times C_p \times \left(U^*_{dry1}\right)^3 \times T'_{dry}}{k \times g \times \overline{H}_{dry}} \qquad (m)$$

With:

L_{1_dry} the Monin-Obukov Length (first approximation) *(m)*
ρ_{air} the Atmospheric Air Density *(Kg/m³)*
C_p the Air Specific Heat at Constant Pressure (1004) *(J/Kg/K)*
U^*_{dry1} the effective friction velocity for "dry" pixels (first approximation) *(m/s)*
T'_{dry} the mean air (*Zsu r = 100 m*) to surface skin temperature for "dry" pixels *(K)*
k is the Von Karman's Constant (0.4) *(-)*
g is the gravitational acceleration (9.81) *(m.s⁻²)*
$\overline{H}_{dry}$ being the mean Sensible Heat Flux for "dry" pixels (see) *(W/m²)*

The mean air to surface skin temperature for "dry" pixels is:

$$T'_{dry} = \frac{\left(\overline{T}_{0\_dry} - \overline{T}_{Z\_dry}\right)}{2} \qquad (K)$$

With:

T'_{dry} the mean air (*Zsur = 100 m*) to surface skin temperature for "dry" pixels *(K)*
$\overline{T}_{0\_dry}$ the mean surface skin temperature for "dry" pixels *(K)*
$\overline{T}_{Z\_dry}$ the mean air temperature at *Zsur = 100 m* for "dry" pixels (considered as: *(K)*

$$\bar{T}_{Z\_dry} = \bar{T}_{0\_dry} - \frac{\bar{r}_{ah\_dry} \times \overline{H}_{dry}}{\rho_{air} \times C_p})$$

Calculation of the first R_{ah} for Landsat
(see Bandara, 1998)

Having the first approximation of the Monin-Obukov Length for the “dry” pixels (L_{1_dry}), the first approximation of x the buoyancy parameter of the first approximation of the psychometric parameter of the atmospheric momentum transport (ψ_m).

$$x_1 = \left(1 - \frac{16 \times Z_{sur}}{L_{1\_dry}} \right) \qquad (-)$$

With:

x_1 the buoyancy parameter (first approximation) *(-)*
Zsur the height of the potential air temperature *(Zsur = 100 m)* *(m)*
L_{1_dry} the Monin-Obukov Length for “dry” pixels (first approximation) *(m)*

That can be processed into the first approximation of ψ_h:

$$\psi_{h1} = 2 \times Ln \left[\frac{1 + x_1^2}{2} \right] \qquad (-)$$

With:

ψ_{h1} the stability correction for the atmospheric heat transport (first approximation) *(-)*
x_1 the buoyancy parameter (first approximation) *(-)*

Thus ending the first run of the iteration aiming to determine U^*_{dry} as explained in Figure 19. Out of this iteration, one input has not been explained, it is r_{ah_dry} the Aerodynamic Resistance to Heat transport for “dry” pixels, that in itself takes a whole set of calculations to reach.

Aerodynamic Resistance to Heat transport for "dry" pixels (r_{ah_dry})

The calculation undergone here is only to get a single value (r_{ah_dry}), each one of the derivatives determined below in order to approach the value of $\left(\frac{\partial H}{\partial T_0}\right)_{dry}$ are also single values.

$$r_{ah\_dry} = \frac{\rho_{air} \times C_p}{\left(\frac{\partial H}{\partial T_0}\right)_{dry}} \qquad (s/m)$$

With:

r_{ah_dry} the Aerodynamic Resistance to Heat transport for "dry" pixels *(s/m)*

ρ_{air} the Atmospheric Air Density *(Kg/m^3)*

C_p the Air Specific Heat at Constant Pressure (1004) *(J/Kg/K)*

$\left(\frac{\partial H}{\partial T_0}\right)_{dry}$ the partial derivative of the Sensible Heat Flux by the Surface Skin Temperature for "dry" pixels.

$\left(\frac{\partial H}{\partial T_0}\right)_{dry}$ can be determined by joining the Net Radiation equation (see 4.1) with the Energy Balance (see Error: Reference source not found), giving the following (for more detail, see Bandara, 1998, Appendix 3.4, page XXIX):

$$\left(\frac{\partial H}{\partial T_0}\right)_{dry} = -\overline{K^{\downarrow}}_{dry} \times \left(\frac{\partial \rho_0}{\partial T_0}\right)_{dry} + \left(\frac{\partial L^*}{\partial T_0}\right)_{dry} - \left(\frac{\partial G_0}{\partial T_0}\right)_{dry} \qquad (J/m^2/K)$$

With:

$\left(\frac{\partial H}{\partial T_0}\right)_{dry}$ the partial derivative of the Sensible Heat Flux by the Surface Skin Temperature for "dry" pixels *(J/m²/K)*

$\overline{K^{\downarrow}}_{dry}$ the mean incoming shortwave solar radiation for "dry" pixels *(W/m²)*

$\left(\frac{\partial \rho_0}{\partial T_0}\right)_{dry}$ the partial derivative of the Surface Albedo by the Surface Skin Temperature for "dry" pixels *(J/m²/K)*

$\left(\frac{\partial L^*}{\partial T_0}\right)_{dry}$ the partial derivative of the atmospheric Long-wave radiation balance by the Surface Skin Temperature for "dry" pixels *(J/m²/K)*

$\left(\frac{\partial G_0}{\partial T_0}\right)_{dry}$ the partial derivative of the Soil Heat Flux by the Surface Skin Temperature for "dry" pixels *(J/m²/K)*

Determination of $\left(\frac{\partial \rho_0}{\partial T_0}\right)_{dry}$.

The system used in this case study is very similar to the one in . The only differences reside in the selection of the data to be used in the regression curve.
The surface skin Temperature data was not corrected to the elevation (z).

Only data of $T_0 > 310$ K (called "dry" pixels) were computed into a raster file. The corresponding pixels where selected from the Albedo (ρ_0) as the second data set. The half bell shape is studied to get the highest curve fitting T_0 value leading to its corresponding ρ_0 value.

These two values are selected as the new references for the same exercise. Refining the values to the peak of the half-bell curve and its right end. A trend line of order 1 is then set as shown in the lower part of Figure 18, thus leading to $\left(\frac{\partial \rho_0}{\partial T_0}\right)_{dry}$.

Determination of $\left(\frac{\partial L^*}{\partial T_0}\right)_{dry}$.

$$\left(\frac{\partial L^*}{\partial T_0}\right)_{dry} = -4 \times \bar{\varepsilon}_{0\_dry} \times \sigma \times \bar{T}_{0\_dry} \qquad (J/m^2/K)$$

With:

$\left(\frac{\partial L^*}{\partial T_0}\right)_{dry}$ the partial of the atmospheric Long-wave radiation balance by the Surface Skin Temperature for "dry" pixels *(J/m²/K)*

$\bar{\varepsilon}_{0\_dry}$ the mean surface emissivity for "dry" pixels *(-)*

σ the Stefan-Boltzman constant (5.67×10^{-8}) *(W/m²/K⁴)*

$\bar{T}_{0\_dry}$ the mean surface skin temperature for "dry" pixels *(K)*

Determination of $\left(\frac{\partial G_0}{\partial T_0}\right)_{dry}$.

By fitting a linear regression T_{0_dry} on x-axis and G_{0_dry} on the y-axis, one can find the derivative of the Soil Heat Flux by the Surface Skin Temperature for "dry" pixels as being the slope of the relationship

$G_{0\_dry} = f(T_{0\_dry}) = \lfloor slope \rfloor \times T_{0\_dry} + \lfloor offset \rfloor$.

4.3.3. Calculation of the first H approximation

The first approximation of the Sensible Heat Flux is the last energy balance component to be found in SEBAL, even if it is a rough estimate, still it gives the trend of the general values of this convection parameter. The stabilization of the pixels values are maximized at the third iteration of the convection (H_3).

Calculation of the first H approximation

This first iteration of the sensible heat flux (H_1) is the application of the main equation cited in the beginning of this section () with the above-mentioned calculated events. The equation will look this way:

$$H_1 = \frac{\rho_{air} \times C_p \times \left[a_1 + \left(b_1 \times T_{0-}\, dem\right)\right]}{r_{ah1}} \qquad (W/m^2)$$

With:

H_1 being the Sensible Heat Flux (<u>first approximation</u>) *(W/m^2)*
ρ_{air} the Atmospheric Air Density *(Kg/m^3)*
C_p the Air Specific Heat at Constant Pressure *(J/Kg/K)*
T_0*_dem* the surface temperature adjusted with the DEM *(K)*
r_{ah1} the first Aerodynamic Resistance to Heat transport approximation without corrections from psychometric parameters *(s/m)*

<u>Raster images inputs:</u>
ρ_{air} the Atmospheric Air Density
r_{ah1} the Aerodynamic Resistance to Heat transport (<u>second approximation</u>)
T_0*_dem* the surface temperature adjusted with the DEM

4.3.4. Calculation of the second R_{ah}

The Aerodynamic Resistance to Heat transport r_{ah2} is calculated on a second approximation in this part, refining considerably the values, improving consequently the accuracy to an acceptable level.
The psychometric parameters are now included in the equation, even if they are roughly estimated. A second iteration of these parameters later on will improve sufficiently their accuracy. See the corresponding sections about it in the following pages.

Calculation of r_{ah2} and the psychometric parameters

The parameters of interests are ψ_h and ψ_m. They are appearing in the r_{ah2} calculations in the following form:

$$r_{ah2} = \frac{1}{U_5 \times 0.41^2} \times Ln\left(\frac{5}{z_{0m}} - \psi_{m1}\right) \times Ln\left(\frac{5}{z_{0m} \times 0.1} - \psi_{h1}\right) \quad \textit{(s/m)}$$

With:

r_{ah2} the Aerodynamic Resistance to Heat transport (<u>second approximation</u>)	*(s/m)*
U_5 being the z = 5m wind velocity	*(m/s)*
Z_{0m} the aero-dynamic roughness length for momentum transport	*(m)*
ψ_{h1} the stability correction for the atmospheric heat transport (<u>first approximation</u>)	*(-)*
ψ_{m1} the stability correction for the atmospheric momentum transport (<u>first approximation</u>)	*(-)*

<u>Raster images inputs:</u>

r_{ah2} the Aerodynamic Resistance to Heat transport (<u>second approximation</u>)
U_5 the z = 5m wind velocity
T_0*_dem* the surface temperature adjusted with the DEM

The parameters ψ_h and ψ_m are defined as follow:

$$\psi_{h1} = 2 \times Ln\left[\frac{(1 + x_1)^2}{2}\right] \qquad (-)$$

$$\psi_{m1} = 2 \times Ln\left[\frac{(1 + x_1)}{2}\right] + Ln\left[\frac{(1 + x_1)^2}{2}\right] - 2 \times A\tan(x_1) + \frac{\pi}{2} \qquad (-)$$

With $x_1 = \left[(1 - 16) \times \left(\frac{5}{L_1}\right)^{0.25}\right]$

ψ_{h1} the stability correction for the atmospheric heat transport (<u>first approximation</u>) *(-)*
ψ_{m1} the stability correction for the atmospheric momentum transport (<u>first approximation</u>) *(-)*
L_1 the Monin-Obukov Length (<u>first approximation</u>) *(m)*

<u>Raster images inputs:</u>
L_1 the Monin-Obukov Length (<u>first approximation</u>)

Calculation of the first Monin-Obukov Length

L_1 is computed from H_1, and is used into the next step, the psychometric parameters (ψ_h and ψ_m).

$$L_1 = \frac{\rho_{air} \times C_p \times \left(U^*_{eff}\right)^3 \times T_{0-}dem}{K \times g \times H_1} \qquad (m)$$

With:

L_1 the Monin-Obukov Length (<u>first approximation</u>) *(m)*
ρ_{air} the Atmospheric Air Density *(Kg/m^3)*
C_p the Air Specific Heat at Constant Pressure *(J/Kg/K)*
U^*_{eff} the effective friction velocity *(m/s)*
T_0_dem the surface temperature adjusted with the DEM *(K)*
K is the Von Karman's Constant (0.4) *(-)*
g is the gravitational acceleration (9.8) *(m.s^{-2})*
H_1 being the Sensible Heat Flux (<u>first approximation</u>) *(W/m^2)*

<u>Raster images inputs:</u>
ρ_{air} the Atmospheric Air Density
T_0_dem the surface temperature adjusted with the DEM
H_1 the Sensible Heat Flux (<u>first approximation</u>)

A simplified system to reach r_{ah2} has been designed.

$$r_{ah2} = \frac{\left(6.215 - 2 \times Ln\left(\frac{1 + x_1}{2}\right)\right)^2}{0.41 \times U^*} \qquad (s/m)$$

With:

r_{ah2} the Aerodynamic Resistance to Heat transport (second approximation) *(s/m)*
U* being the nominal efficient wind speed *(m/s)*
x_1 the buoyancy parameter (first approximation) *(-)*

The buoyancy parameter x_1 (first approximation)

$$x_1 = \sqrt{\left[1 + \left(\frac{0.278 \times H_1 \times (U^*)^{-3}}{T_0}\right)\right]} \qquad (-)$$

With:

x_1 the (first approximation) *(-)*
H_1 being the Sensible Heat Flux (first approximation) *(W/m²)*
U* being the nominal efficient wind speed *(m/s)*
T_0 the surface skin temperature *(K)*

Calculation of the second R_{ah} for Landsat
(see Bandara, 1998)

A "traditional" method is used to reach r_{ah2} and r_{ah3}, going from the Sensible Heat Flux (H) to the Monin-Obukov Length (L) then to the Momentum of Heat transport (ψ_h).
The set of equations below is describing how to get r_{ah2} from H_1 the same system has been applied to get r_{ah3} from H_2.

Calculation of r_{ah2}

$$r_{ah2} = \frac{1}{U^* \times 0.41} \times Ln\left(\frac{Z_{sur}}{z_{0m} \times 0.01} - \psi_{h1}\right) \qquad \textit{(s/m)}$$

With:

r_{ah2} the Aerodynamic Resistance to Heat transport (<u>second approximation</u>) *(s/m)*
U^*_{eff} the effective friction velocity *(m/s)*
Z_{0m} the aero-dynamical roughness length for momentum transport *(m)*
ψ_{h1} the stability correction for the atmospheric heat transport (<u>first approximation</u>) *(-)*

That can be processed by using the first approximation of ψ_h:

$$\psi_{h1} = 2 \times Ln\left[\frac{1 + x_1^2}{2}\right] \qquad \textit{(-)}$$

With:

ψ_{h1} the stability correction for the atmospheric heat transport (<u>first approximation</u>) *(-)*
x_1 the buoyancy parameter (<u>first approximation</u>) *(-)*

Calculation of the second R_{ah} for Landsat
(see Bandara, 1998)

The buoyancy parameter can be calculated as follow:

$$x_1 = \left(1 - \frac{16 \times Z_{sur}}{L_1}\right) \qquad (-)$$

With:

x_1 the buoyancy parameter (first approximation) *(-)*
Zsur the height of the potential air temperature *(Zsur = 10 m)* *(m)*
L_1 the Monin-Obukov Length (first approximation) *(m)*

The Monin-Obukov Length (first approximation) in this case is:

$$L_1 = \frac{-\rho_{air} \times C_p \times (U^*)^3 \times T_0}{k \times g \times H_1} \qquad (m)$$

With:

L_1 the Monin-Obukov Length (first approximation) *(m)*
ρ_{air} the Atmospheric Air Density *(Kg/m^3)*
C_p the Air Specific Heat at Constant Pressure (1004) *(J/Kg/K)*
U^* the effective friction velocity *(m/s)*
T_0 the surface skin temperature for all pixels *(K)*
k is the Von Karman’s Constant (0.41) *(-)*
g is the gravitational acceleration (9.81) *(m.s^{-2})*
H_1 being the Sensible Heat Flux (first approximation) *(W/m^2)*

4.3.5. Calculation of the second dTair

The difference of temperature between the soil surface and the air is evaluated by calculating the energy balance for two points, a desert or a beach, being the warmest and driest and a lake or the sea, being the coolest and most humid.
The maximum difference of temperature between the soil and the air will be assessed in those dry conditions, where the latent Heat Flux is null. The Sensible Heat Flux being the difference between the Net Radiation and the Soil Heat Flux.
In order to get a proper regression, a temperature difference on a wet point should be assessed. Theoretically a non-limited evaporation system is giving out all its energy into the latent Heat Flux, resulting in a Sensible Heat Flux of a null value.
From these two points, a regression is set in the calculation of the next Sensible Heat Flux (H_2) to describe the variations of the dTair in function of the Surface Temperature pixels.

Calculation of the second dTair

From extreme situations the following arises:

For Desert: $dT_{air2} = \frac{H_1 \times r_{ah2}}{\rho_{air} \times C_p}$

For Water: $dT_{air2} = 0$

A linear relationship can be drawn out from these two values, linking T_0_dem and dT_{air2}, being an input to the second H image approximation.

4.3.6. Calculation of the second H approximation

The second approximation of the Sensible Heat Flux is leading to the last energy balance component to be found in SEBAL, the estimation much precise at this point, even if the stabilization of the pixels values are only maximized at the third iteration of the calculation of the convection (H_3).

Calculation of the second H approximation

The calculation of the second H is dependent on the first equation of 4.3, where, this time, the components are:

$$H_2 = \frac{\rho_{air} \times C_p \times \left[a_2 + \left(b_2 \times T_{0\_}dem\right)\right]}{r_{ah2}} \qquad (W/m^2)$$

With:

H_2 being the Sensible Heat Flux (<u>second approximation</u>) (W/m^2)
ρ_{air} the Atmospheric Air Density (Kg/m^3)
C_p the Air Specific Heat at Constant Pressure $(J/Kg/K)$
T_0*_dem* the surface temperature adjusted with the DEM (K)
r_{ah2} the Aerodynamic Resistance to Heat transport with corrections from psychometric parameters (<u>second approximation</u>) (s/m)

<u>Raster images inputs:</u>
ρ_{air} the Atmospheric Air Density
r_{ah2} the Aerodynamic Resistance to Heat transport (<u>second approximation</u>)
T_0*_dem* the surface temperature adjusted with the DEM

5.Daily Evaporation

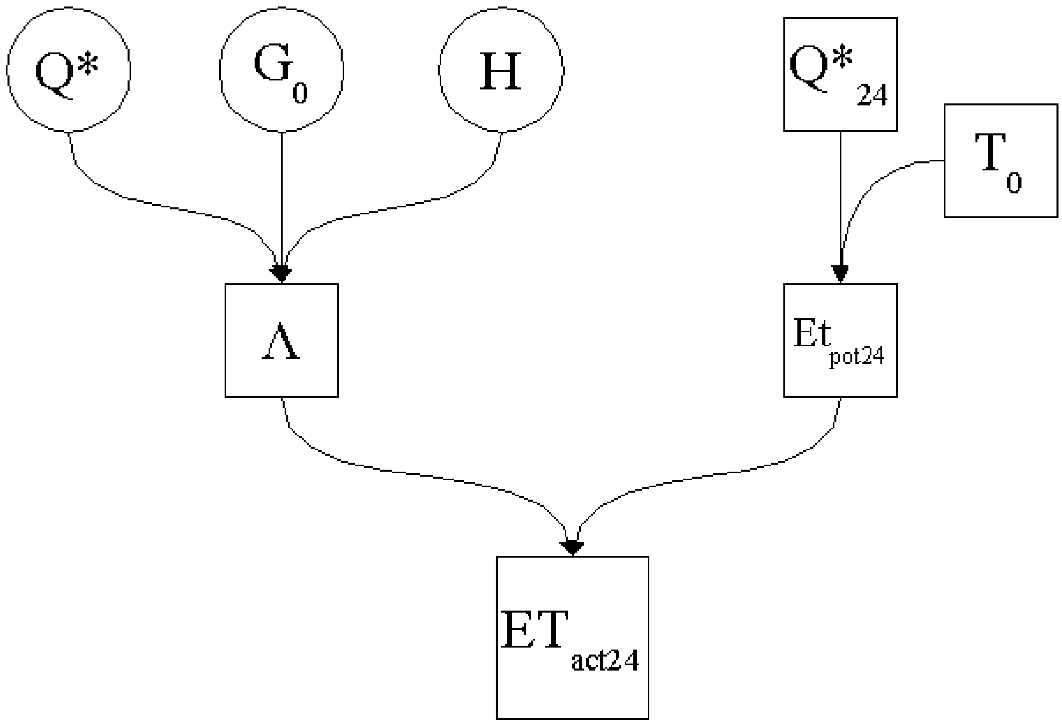


Figure 20: Daily Evaporation steps

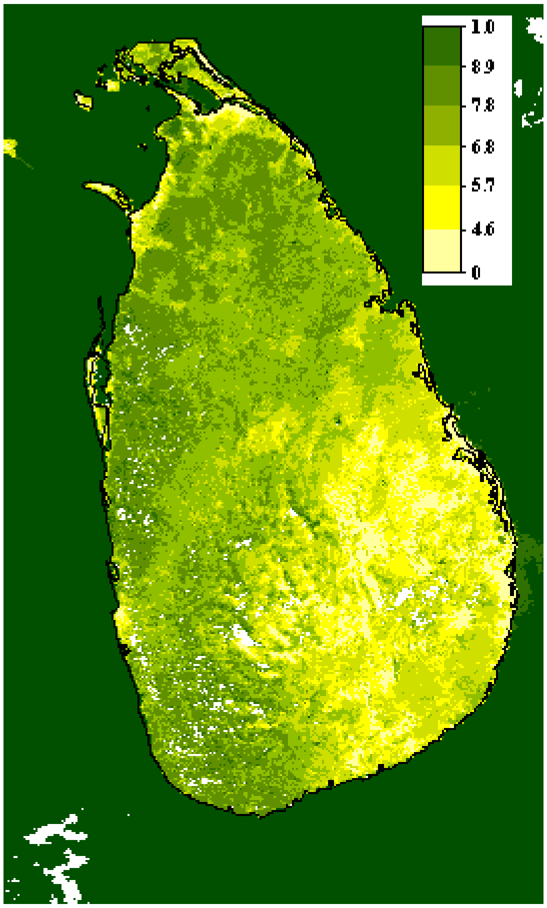


Figure 21: Evaporative Fraction over Sri Lanka (-)

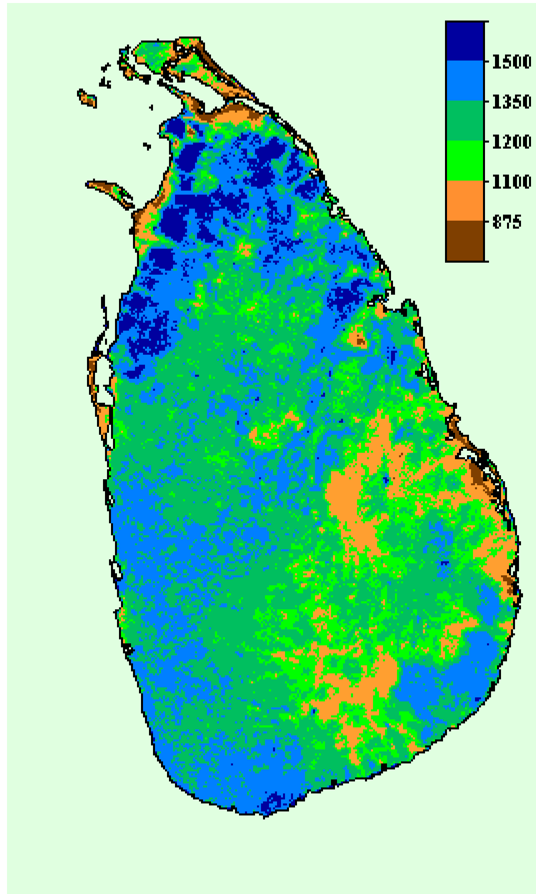


Figure 22: Annual evaporation over Sri Lanka (mm)

<u>INPUTS:</u>

<u>Raster:</u>
Surface Temperature
Net Radiation for a day
Net Radiation
Soil Heat Flux
Sensible Heat Flux

<u>Tabular input:</u>
Density of fresh water

<u>OUTPUTS:</u>

<u>Raster:</u>
Evaporative Fraction
Potential Evaporation for a day
Actual evaporation for a day

5.1.Net Radiation for 24 hours

When calculating for the 24 hours net radiation, a different procedure is implemented than to get the instantaneous net radiation. This has an important meaning, for two different ways have to lead to the two terms (Q^{*}_{24} , Λ) needed to get the actual ET on a 24 hours basis. This to avoid accuracy as well as a methodological concern that may arise on the validity of the ET actual 24 calculation from terms issued from the same sources.

Each specific team had a different way of assessing the 24 hours Net Radiation, it is available in the corresponding parts below.

	Q: Which method am I to use? A: The following methods are sensor independent. Q: I have meteorological data, that I want to use. A: then use Method 1 and 2. Q: I have no meteorological data. A: then use Method 3 and 4.

This method has also been used for the Internet based data. The 24 hours net radiation is:

$$Q_{24}^{*} = (1 - \rho_0) \times (K_{24}^{\downarrow}) + L_{24}^{\downarrow} - (\varepsilon_0 B_{24}^{\uparrow}) - (1 - \varepsilon_0) \times L_{24}^{\downarrow} \qquad (W/m^2)$$

with:

Q_{24}^{*} being the 24 hours Net Radiation *(W/m²)*
ρ_0 the surface reflectance *(-)*
$K_{24}^{\downarrow}$ the 24 hours integrated solar radiation *(W/m²/µm)*
$L_{24}^{\downarrow}$ the 24 hours averaged incoming broadband long-wave radiation of atmosphere from different meteorological stations (spreadsheet) *(W/m²)*
ε_0 the surface emissivity *(-)*
$B_{24}^{\uparrow}$ the black body radiation from averaged daily air temperatures *(W/m²)*

ρ_0, $K_{24}^{\downarrow}$, ε_0 and T_0 are all raster images data.

The outgoing Black body radiation from atmosphere $B_{24}^{\uparrow}$.

The outgoing black body radiation from atmosphere ($B_{24}^{\uparrow}$), is coming from the spreadsheet, where the Stefan-Boltzmann equation has to be applied with the averaged temperatures for all the stations.

$$B_{24}^{\uparrow} = \sigma T_{air24}^{4} \qquad (W/m^2)$$

with:

$B_{24}^{\uparrow}$ the black body radiation from averaged daily air temperatures *(W/m²)*
σ the Stefan-Boltzman constant (5.67×10^{-8}) *(W/m²/K⁴)*
T_{air24} the averaged daily temperature from several well spread meteorological stations *(K)*

The incoming broadband long-wave radiation of atmosphere ($L_{24}^{\downarrow}$):

This is coming from spreadsheet calculations, where the Stefan-Boltzmann equation has to be applied with the daily average temperature for each meteorological station in order to be averaged for all station, giving $L_{24}^{\downarrow}$. A minimum of two well-spread meteorological stations was recommended (Bastiaanssen, personal communication, 1997) to have sufficient accuracy. Nevertheless, five were used in this study.

$$L_{24}^{\downarrow} = \frac{1}{n} \sum_{i=2}^{i=n} \left[\varepsilon_{atm_i} \sigma T_{avgatm_i}^{4} \right] \qquad (W/m^2)$$

with:

$L_{24}^{\downarrow}$ the averaged incoming broadband long-wave radiation of atmosphere from different meteorological stations *(W/m²)*

ε_{atm_i} the atmospheric emissivity of the meteorological station i, already computed in the spreadsheet **(see 3.2.2)** *(-)*

σ the Stefan-Boltzman constant (5.67x10^{-8}) *(W/m²/K⁴)*

T_{avgatm_i} the daily average atmospheric temperature at the meteorological station i in the spreadsheet **(see 3.2.2)** *(K)*

The average incoming solar radiation on the surface $K_{24}^{\downarrow}$:

$$K_{24}^{\downarrow} = K_{exo24}^{\downarrow} \times \tau_{sw} \qquad (W/m^2)$$

with:

$K_{24}^{\downarrow}$ being the average solar radiation on 24 hours *(W/m²)*

$K_{exo24}^{\downarrow}$ the diurnal average sun exo-atmospheric radiation *(W/m²/µm)*

τ_{sw} the atmosphere single-way transmissivity (estimated constant for the day at 0.7) *(-)*

The calculation of $K^{\downarrow}_{exo24}$ goes this way:

$$K^{\downarrow}_{exo24} = \frac{K^{\downarrow}_{sun} \times R}{\pi \times d_s^2} \qquad (W/m^2)$$

with:

$K^{\downarrow}_{exo24}$ the diurnal average sun exo-atmospheric radiation *(W/m²/µm)*

$K^{\downarrow}_{sun}$ the sun external atmosphere radiation (constant = 1358) *(W/m²/µm)*

R the solar angle range for the diurnal sun exposition *(rad)*

τ_{sw} the atmosphere single-way transmissivity (estimated constant for the day at 0.7) *(-)*

d_s the Sun-Earth distance *(A.U.)*

The solar angle range for the diurnal sun exposition is:

$$R = \omega_{s24} \times Sin(\delta) \times Sin(Lat_{(y)}) + Cos(\delta) \times Cos(Lat_{(y)}) \times \omega_{s24} \qquad (rad)$$

with:

R the solar angle range for the diurnal sun exposition *(rad)*

ω_{s24} the solar angle hour for diurnal exposition *(rad)*

δ the solar declination *(rad)*

$Lat_{(y)}$ the Latitude *(rad)*

The solar angle hour for diurnal exposition ω_{s24} is:

$$\omega_{s24} = A\cos\left[- Tan(Lat_{(y)}) \times Tan(\delta)\right] \qquad (rad)$$

with:

ω_{s24} the solar angle hour for diurnal exposition *(rad)*

δ the solar declination *(rad)*

$Lat_{(y)}$ the Latitude *(rad)*

The 24 hours net radiation is:

$$Q_{24}^{*} = (1 - \rho_0) \times (K_{24}^{\downarrow}) - 110 \times \tau_{sw} \qquad (W/m^2)$$

with:

Q_{24}^{*} being the 24 hours Net Radiation *(W/m²)*
ρ_0 the surface reflectance *(-)*
$K_{24}^{\downarrow}$ the 24 hours integrated solar radiation *(W/m²/µm)*
τ_{sw} the atmosphere single-way transmissivity *(-)*

The average incoming solar radiation for a given day is:

$$K_{24}^{\downarrow} = \left[a + b\left(\frac{n}{N} \right) \right] \times R_{a24} \qquad (W/m^2)$$

With:

$K_{24}^{\downarrow}$ being the average incoming solar radiation on 24 hours *(W/m²)*
R_{a24} the sun external atmosphere radiation *(W/m²/µm)*

The sun external atmosphere radiation:

$$R_{a24} = 37.6 \times d_s \left[\omega_s \times Sin(lat) \times Sin(\delta) + Cos(lat) \times Cos(\delta) \times Sin(\omega_s)\right] \qquad (W/m^2)$$

With:

R_{a24} the sun external atmosphere radiation *(W/m^2)*
d_s the sun earth distance *(A.U.)*
ω_s the solar angle hour *(rad)*
δ being the solar declination, the angular height of the sun above the astronomical equatorial plane *(rad)*

1. δ (rad) being the solar declination, the angular height of the sun above the astronomical equatorial plane.

$$\delta = 0.409 \times Sin(0.0172 \times J - 1.39) \qquad \text{(rad)}$$

J being the Julian day number.

2. ω_s the solar angle hour varying following the time of the day.

$$\omega_s = Cos^{-1}\left[- \tan(lat) \times \tan(\delta)\right] \qquad \text{(rad)}$$

3. d_s the distance Earth-Sun varying with the Julian day number *J*.

$$d_s = 1 + 0.033 \times Cos(0.0172 \times J) \qquad (-)$$

Daily Net radiation for Landsat
(see Bandara, 1998)

The net radiation for 24 hours is:

$$Q_{24}^{*} = (1 - \rho_0) \times (K_{24}^{\downarrow}) + \left[\varepsilon_6^{'} \times \sigma \times T_{atm}^{4} - \varepsilon_0 \times \sigma \times T_{atm}^{4} \right] \qquad (W/m^2)$$

with:

Q_{24}^{*} being the Net Radiation for 24 hours *(W/m²)*

ρ_0 the surface reflectance (raster image) *(-)*

$K_{24}^{\downarrow}$ the incoming shortwave solar radiation (raster image) *(W/m²)*

$\varepsilon_6^{'}$ the apparent atmospheric emissivity in spectral band width 6, set as 0.845 for this study *(-)*

T_{atm} the atmospheric temperature (raster image, $T_{atm} = T_0 - 3$) *(K)*

ε_0 the surface emissivity (raster image) *(-)*

Daily Net Radiation for Landsat 7 ETM+
(see Chemin, Ahmad and Asif, 2000)

The 24 hours net radiation is:

$$Q_{24}^{*} = (1 - \rho_0) \times (K_{24}^{\downarrow}) - 110 \times \tau_{sw} \qquad \textit{(W/m}^2\textit{)}$$

with:

Q_{24}^{*} being the 24 hours Net Radiation *(W/m²)*
ρ_0 the surface reflectance *(-)*
$K_{24}^{\downarrow}$ the 24 hours integrated solar radiation *(W/m²/µm)*
τ_{sw} the atmosphere single-way transmissivity *(-)*

The average incoming solar radiation for a given day is:

$$K_{24}^{\downarrow} = K_{exo24}^{\downarrow} \times \tau_{sw} \qquad \textit{(W/m}^2\textit{)}$$

with:

$K_{24}^{\downarrow}$ being the average solar radiation on 24 hours *(W/m²)*
$K_{exo24}^{\downarrow}$ the diurnal average sun exo-atmospheric radiation *(W/m²/µm)*
τ_{sw} the atmosphere single-way transmissivity (estimated constant for the day at 0.7) *(-)*

The calculation of $K_{exo24}^{\downarrow}$ goes this way:

$$K_{exo24}^{\downarrow} = \frac{K_{sun}^{\downarrow} \times R}{\pi \times d_s^2} \qquad (W/m^2)$$

with:

$K_{exo24}^{\downarrow}$ the diurnal average sun exo-atmospheric radiation *($W/m^2/\mu m$)*
$K_{sun}^{\downarrow}$ the sun external atmosphere radiation (constant = 1358) *($W/m^2/\mu m$)*
R the solar angle range for the diurnal sun exposition *(rad)*
τ_{sw} the atmosphere single-way transmissivity (estimated constant for the day at 0.7) *(-)*
d_s the Sun-Earth distance *(A.U.)*

The solar angle range for the diurnal sun exposition is:

$$R = \omega_{s24} \times Sin(\delta) \times Sin(Lat_{(y)}) + Cos(\delta) \times Cos(Lat_{(y)}) \times \omega_{s24} \qquad (rad)$$

with:

R the solar angle range for the diurnal sun exposition *(rad)*
ω_{s24} the solar angle hour for diurnal exposition *(rad)*
δ the solar declination *(rad)*
$Lat_{(y)}$ the Latitude *(rad)*

The solar angle hour for diurnal exposition is:

$$\omega_{s24} = A\cos\left[- Tan\left(Lat_{(y)}\right) \times Tan(\delta)\right] \qquad \textit{(rad)}$$

with:

ω_{s24} the solar angle hour for diurnal exposition *(rad)*
δ the solar declination *(rad)*
$Lat_{(y)}$ the Latitude *(rad)*

With:

δ (rad) being the solar declination, the angular height of the sun above the astronomical equatorial plane.

$$\delta = 0.4093 \times \sin\left(\frac{2\pi}{365} J - 1.39\right) \qquad \text{(rad)}$$

J being the Julian day number.

5.2.ET potential for 24 hours

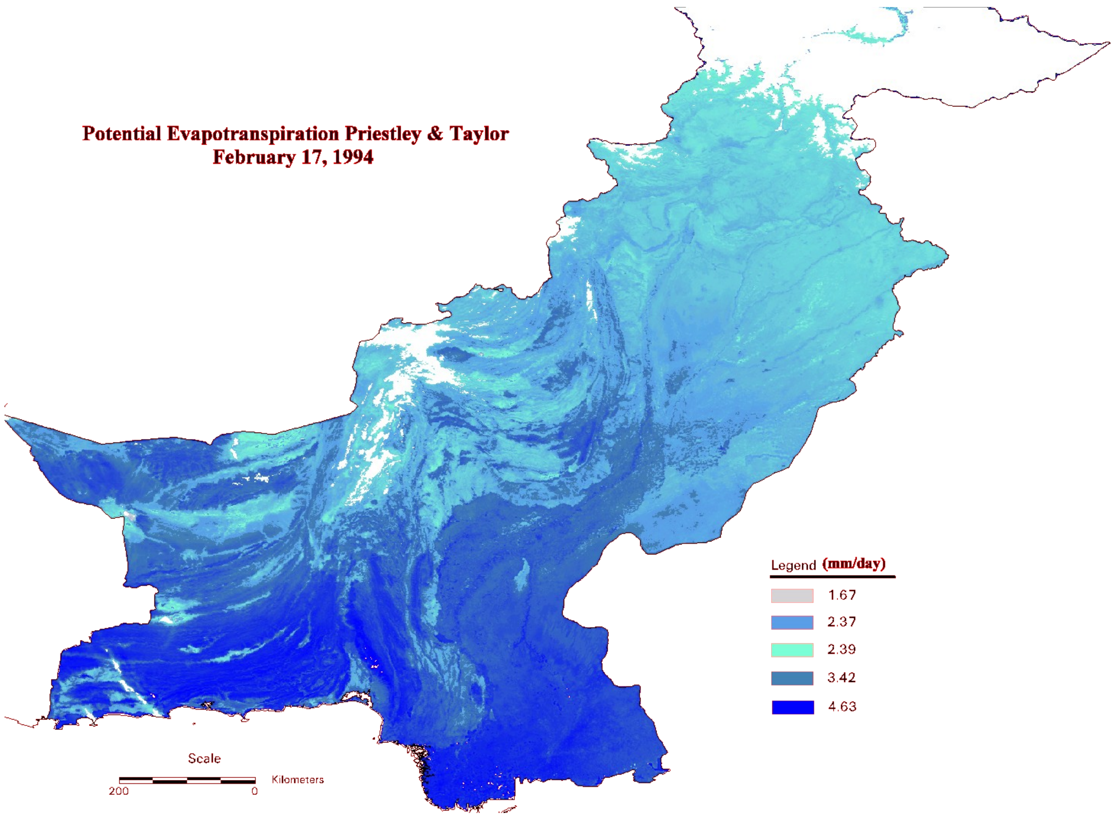


Figure 23: ET Potential over Pakistan (mm/day)

INPUTS:

Raster:
Surface Skin Temperature
Net Radiation for a Day

OUTPUTS:

Raster:
ET Potential for a Day

On converting the 24 hours Net radiation into 24 hours ET potential

$$ET_{pot_{24}} = \frac{Q_{24}^{*}}{L \times \rho_w} \times 86400 \times 10^{3} \qquad (mm/day)$$

With:

$ET_{pot_{24}}$ being the 24 hours potential evapotranspiration *(mm/day)*
Q_{24}^{*} the net radiation on 24 hours *(W/m²)*
L the latent heat of vaporization *(J/Kg)*
ρ_w the density of fresh water *(Kg/m³)*

Where *L* can be calculated from the temperature image, providing the temperature input in Celsius degree:

$$L = \left[2.501 - \left(0.002361 \times T_0\right)\right] \times 10^{6} \qquad (J/Kg)$$

With:

L the latent heat of vaporization *(J/Kg)*
T_0 the surface skin temperature *(˚C)*

Where ρ_w, the density of fresh water is 1000 *Kg/m³*, however, in the case of Pakistan, considering the high concentration of material in the water, 1010 *Kg/m³* has been used.

5.3.Evaporative Fraction

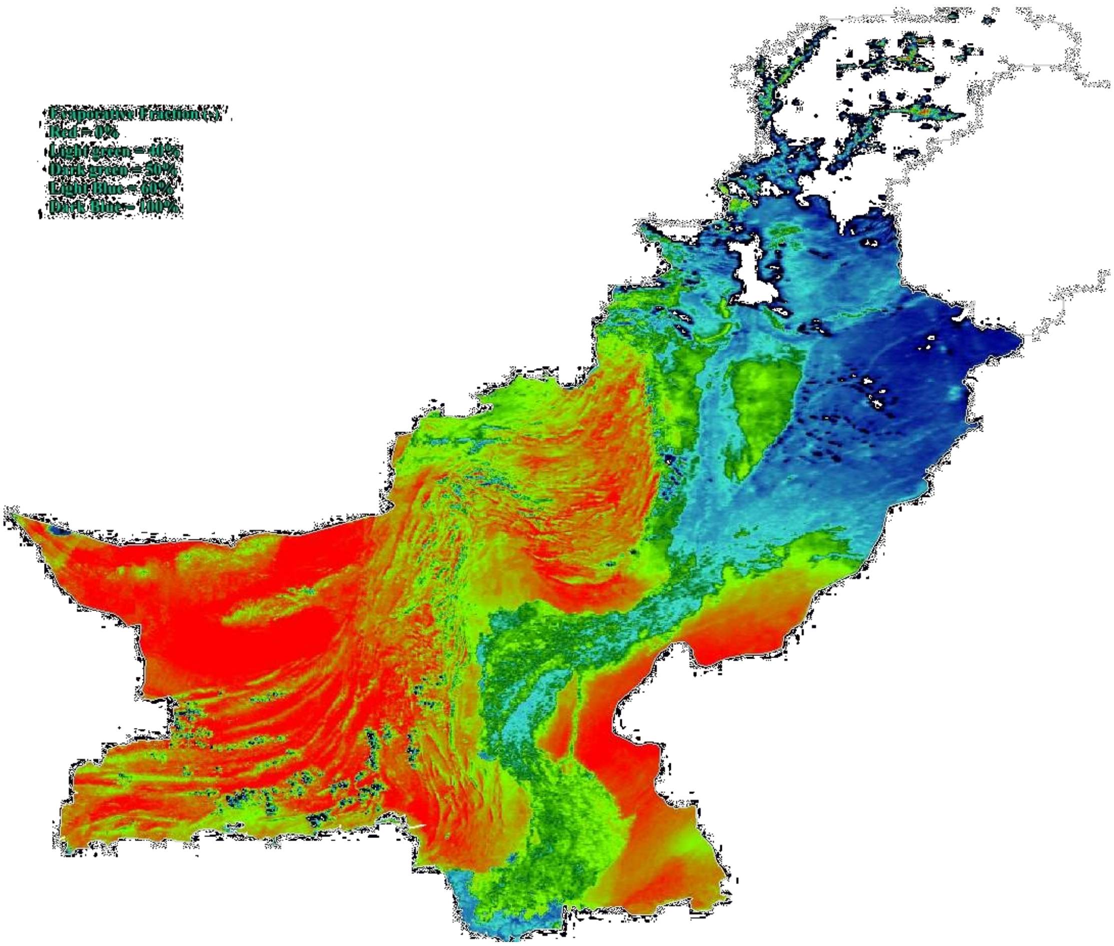


Figure 24: Evaporative Fraction over Pakistan (-)

<u>INPUTS:</u>

<u>Raster:</u>
Instantaneous Net Radiation
Instantaneous Soil Heat Flux
Instantaneous Sensible Heat Flux

<u>OUTPUTS:</u>

<u>Raster:</u>
Evaporative Fraction

Calculation of the evaporative fraction

The evaporative fraction can be defined as the fraction of the actual ET by the potential ET on an instantaneous basis. The evaporative fraction is constant other the day. In radiation terms it can be described as in the following equation.

$$\Lambda = \frac{\lambda E}{Q^* - G_0} = \frac{Q^* - G_0 - H}{Q^* - G_0} \qquad \textit{(mm/day)}$$

With:

Λ being the Evaporative Fraction *(-)*
Q^* the instantaneous Net Radiation *(W/m²)*
G_0 the instantaneous Soil Heat flux *(W/m²)*
H the instantaneous Sensible Heat flux *(W/m²)*
λE the instantaneous Latent Heat of Vaporization *(W/m²)*

5.4.Actual Evaporation for 24 hours

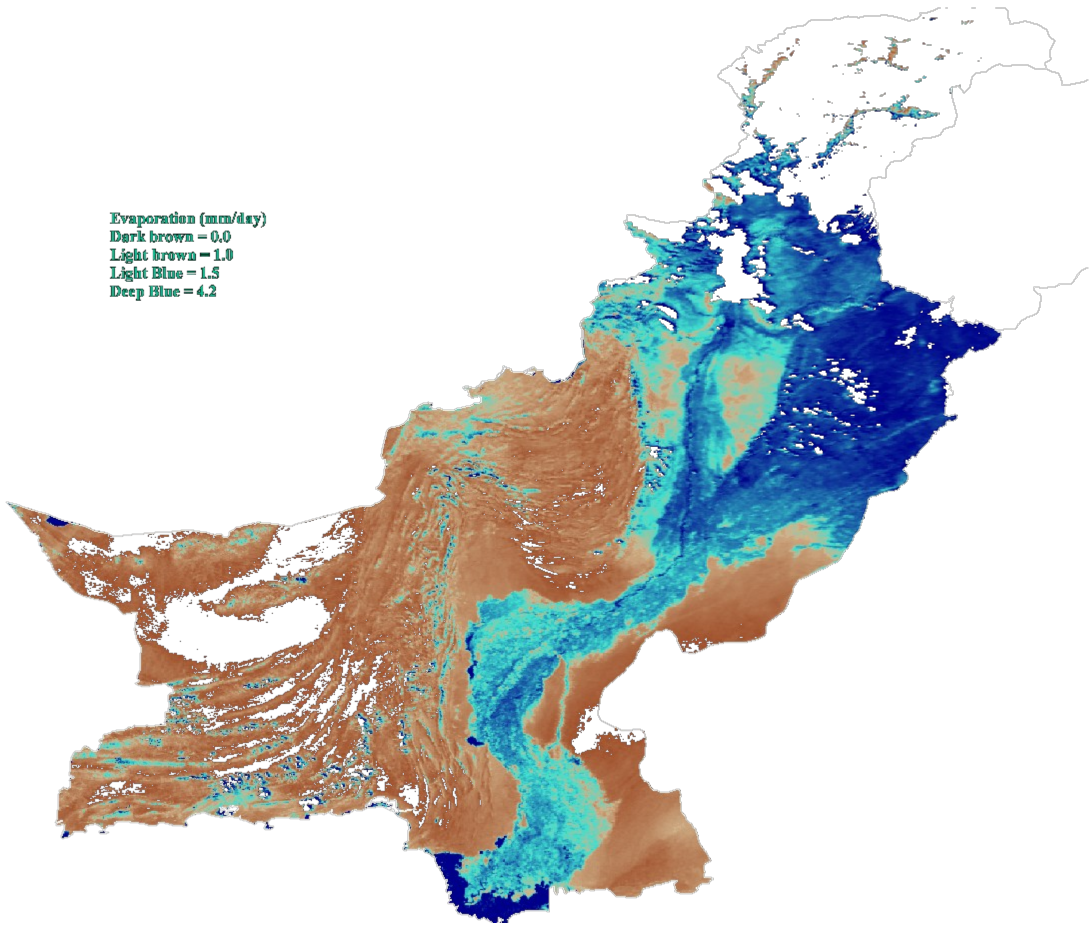


Figure 25: Actual Evaporation over Pakistan (mm/day)

<u>INPUTS:</u>

<u>Raster:</u>
Potential Evaporation for a day
Evaporative Fraction

<u>OUTPUTS:</u>

<u>Raster:</u>
Actual evaporation for a day

Calculating the actual evaporation for 24 hours $ET_{act_{24}}$:

The final part of this exercise is to calculate the actual evaporation for 24 hours by multiplying the ET potential on 24 hours by the evaporative fraction.

$$ET_{act24} = ET_{pot_{24}} \times \Lambda \qquad \textit{(mm/day)}$$

with:

$ET_{act_{24}}$ the actual daily evaporation *(mm/day)*

$ET_{pot_{24}}$ being the 24 hours potential evapotranspiration *(mm/day)*

Λ the evaporative fraction *(-)*

6.References

Alexandridis T., Asif S. and Ali S., 1999. "Water Performance Indicators Using Satellite Imagery for the Fordwah Eastern Sadiqia (South) Irrigation and Drainage Project". Pakistan National Program, International Water Management Institute, Lahore, Pakistan.

Bandara K.M.P.S., 1998. "Water needs and water use of agro-ecosystems in the Kirindi Oya watershed, Sri Lanka. A remote sensing approach." M.Sc. GFM thesis, ITC, Enschede, NL.

Bandara K.M.P.S., 2000. "Monitoring Irrigation Performance in Sri Lanka with high-frequency satellite measurements" Agricultural Water Management, special issue, under press.

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". SC-DLO Report N 109, Wageningen, The Netherlands, 271p.

Bastiaanssen W.G.M., 1998a. "Remote sensing in water resources management: the state of the art". International Water Management Institute (IWMI), Colombo, Sri Lanka: 118 pp.

Bastiaanssen W.G.M., Menenti M., Feddes R.A. and Holtslag A.A.M., 1998b. "A remote sensing surface energy balance algorithm for land (SEBAL), 1. Formulation". J. of Hydr. 212-213: 198-212

Bastiaanssen W.G.M., Thiruvengadarachi S., Sakhtivadivel R. and Molden D.J., 1999. "Satellite Remote Sensing For Estimating productivities of Land and Water". J. of Water Resources Development, Vol 15, Nos. 1/2, 181-196.

Bastiaanssen W.G.M., Chemin Y., Ahmad M.D., and Asif S., 1999. " Patterns of crop evaporation in the Indus Basin recognized from NOAA-AVHRR satellite " 17th ICID conference, Lausanne, Switzerland, 9p.

Brutsaert W. and Sugita M., 1992. "Application of self-preservation in the diurnal evolution of the surface energy budget to determine daily evaporation". J. of Geophysical Res., 97, D17: 18,322-18,377

Chemin Y. and Ahmad M.D., 2000. "Estimating Evaporation using the Surface Energy Balance Model (SEBAL) – A manual for NOAA-AVHRR in Pakistan" IIMI-Pakistan Blue Report N 102, Lahore, Pakistan, July 2000, 27p.

Coll C. and Caselles V., 1997. "A Split-window algorithm for land surface temperature from advanced very high-resolution radiometer data: Validation and algorithm comparison" Journal of Geophysical Research, Vol. 102, No D14, July 27, 1997: 16697-16713.

Cooper D.I. and Asrar G., 1990. "Evaluating atmospheric correction models for retrieving surface temperatures from the AVHRR over a tall grass prairie". Remote Sensing of Environment, 27, 93-102.

Fiedler F. and Panofsky H.A., 1972. "The geostrophic drag coefficient and the effective roughness length". Q.J.R.Met. Soc. 98: 213-220.

Gordon C. and Erasmus J.F., 1995. “Automatic registration of NOAA AVHRR satellite Images”. Institute of Soil, Climate and Water, Pretoria, South Africa.

Hafeez M.M. and Chemin Y., 2001. “Evapotranspiration Estimation using TERRA/ASTER sensor: A case study in District 1 of UPRIIS, Central Luzon, Philippines” Canadian Journal of Remote Sensing. Submitted.

Kaufman Y.J. and Holben B.N., 1993. “Calibration of the AVHRR visible and near-ir bands by atmospheric scattering, ocean glint and desert reflection”. International Journal of Remote Sensing, 14, 21-52.

Kerdiles H., 1995 (?). “Software processing AVHRR data for the communities of Europe (SPACE): algorithms used in SPACE version 2. Part 1: Calibration, atmospheric correction and cloud detection”. MARS project, TP 441, JRC – Space Applications Institute.

Markham and Baker, 1987. Landsat 5TM calibration

Mutuwatte L. and CheminY., 2000. "Vegetation growth zones of Sri Lanka interpreted from satellite data". Agricultural Water Management, special issue, submitted.

Rao C.R.N., 1993. “Non-linearity corrections for the thermal infrared channels of the Advanced Very High Resolution Radiometer: assessment and recommendations”. NOAA Technical Report NESDIS-69, NOAA/NESDIS, Washington, DC.

Roerink G.J., 1995. “SEBAL estimations of the areal patterns of sensible heat fluxes over the HAPEX-Sahel grid – a case study on 18 September 1992.” DLO-Staring center, Wageningen, The Netherlands, September 1995, 61p.

Tasumi, M., and Allen, R.G., 2000. Application of the SEBAL methodology for estimating consumptive use of water and streamflow depletion in the Bear River Basin of Idaho through Remote Sensing. Appendix A: The theoretical basis of SEBAL. Final Report. The Raytheon Systems Company, EOSDIS Project.

Valiente et al. 1995. Broadband Reflectance at top of atmosphere formula.

Weinreb M.P., Hamilton G., Brown S. and Koczor R.J., 1990. ”Nonlinearity corrections in calibration of Advanced Very high Resolution Radiometer infrared channels”. Journal of Geophysics Research, 95, 7381-7388.

World Wide Web references

http://ltpwww.gsfc.nasa.gov/IAS/handbook/handbook_toc.html. Handbook for the users of Landsat 7 ETM+ satellite data.

http://www.saa.noaa.gov/. The National Oceanographic and Atmospheric Administration Gateway for Satellite Active Archive. It is the main downloading site available for many satellites, most are free download.

http://www2.ncdc.noaa.gov/docs/podug/. The NOAA Polar Orbiter Data User's Guide (November 1998 version) is a document that describes the orbital and spacecraft characteristics, instruments, data formats, etc. of the TIROS-N, NOAA-6 through NOAA-14 polar orbiter series of satellites.

http://www2.ncdc.noaa.gov/docs/podug/html/c1/sec1-4.htm. Miscellaneous Parameters for the Polar Orbiter Satellites (NOAA AVHRR satellites). Gives the Thermal Infrared bands calibration equations and related information.