REMOTE SENSING LAB

Εκτίμηση βιο-γεωφυσικών παραμέτρων: αποτύπωση της θερμοκρασίας της επιφάνειας του εδάφους

Νεκτάριος Χρυσουλάκης

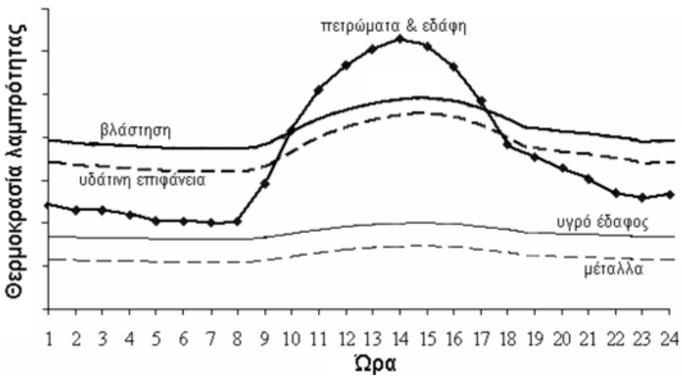
Εργαστήριο Τηλεπισκόπησης | ΙΥΜ | http://rslab.gr

FOUNDATION FOR RESEARCH AND TECHNOLOGY - HELLAS

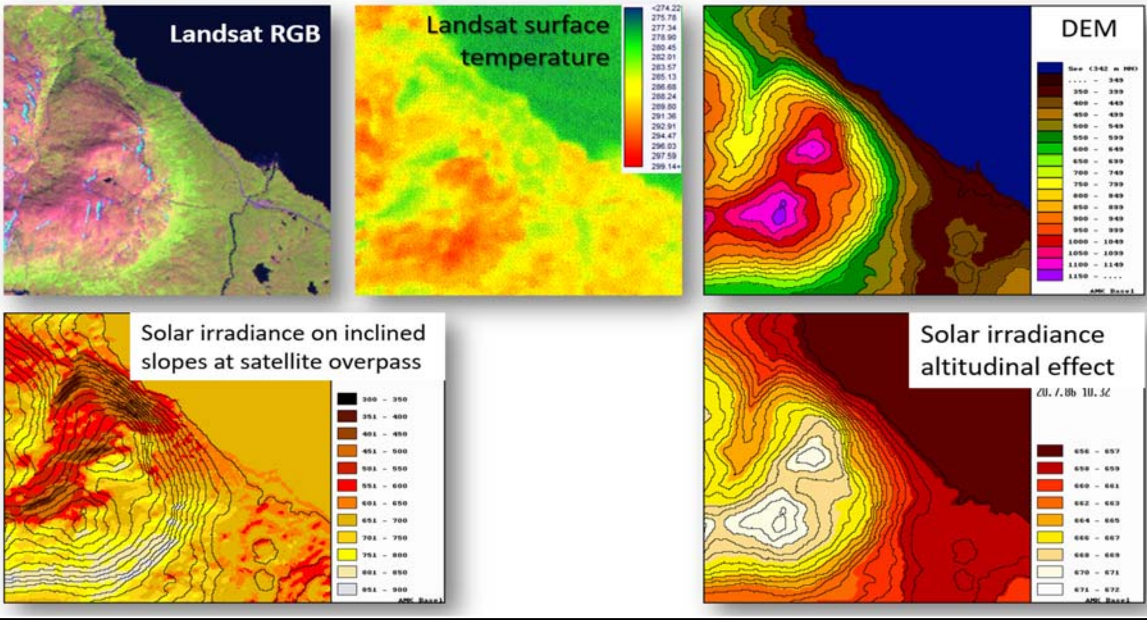
Θερμοκρασία επιφάνειας

Η θερμοκρασία των σωμάτων στην επιφάνεια της γης εξαρτάται από:

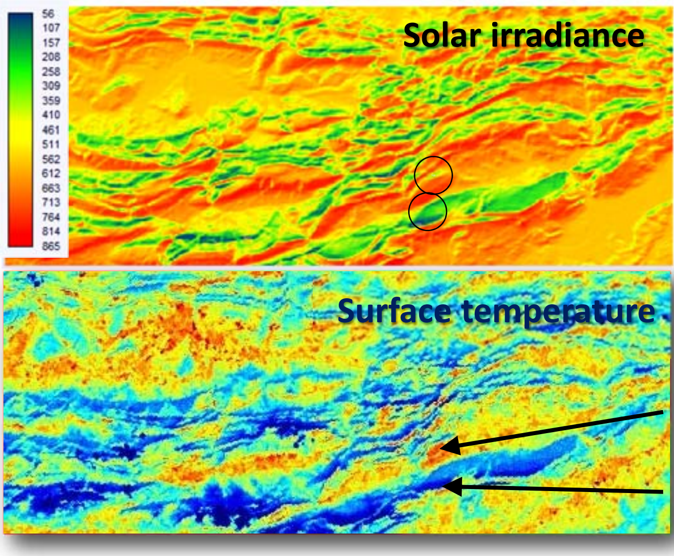
- Το ποσό της ηλιακής ακτινοβολίας που είναι διαθέσιμο για τη θέρμανση του σώματος.
- Τις θερμικές ιδιότητες του σώματος οι οποίες είναι συνάρτηση της σύστασής του:
  - Θερμοχωρητικότητα (c)
  - Θερμική αγωγιμότητα (K)
  - Θερμική αδράνεια (P): $P = (K\ c\ \rho)^{1/2}$
  - Θερμική διαχυτικότητα (κ): $\kappa = K/(c\ \rho)$



Θερμοκρασία επιφάνειας



Θερμοκρασία επιφάνειας

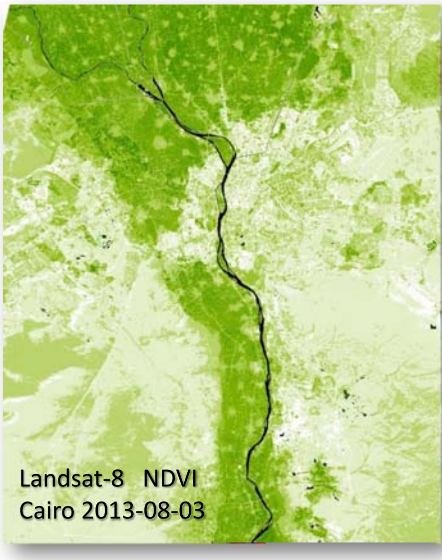


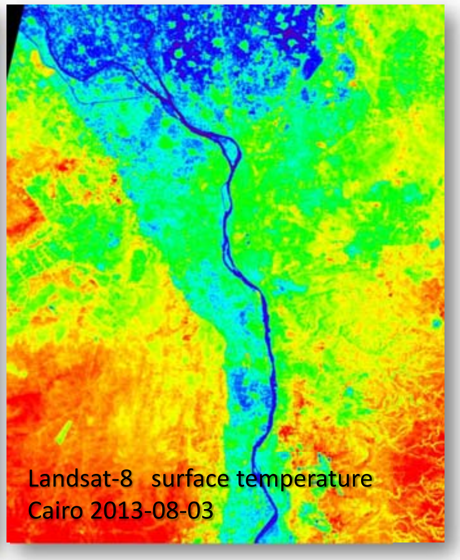
Aspect depending surface temperature in the Swiss Jura Mountains

≈ 298 K = 25 °C

≈ 278 K = 5 °C

Θερμοκρασία επιφάνειας

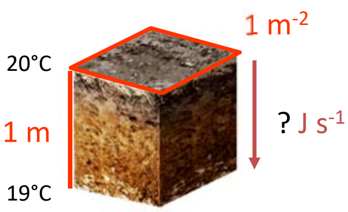




Θερμοκρασία επιφάνειας

Describes the property of a material to conduct heat.

Energy (J) conducted in the soil per m² unit surface and second if temperature gradient is 1 K m^{-1}.



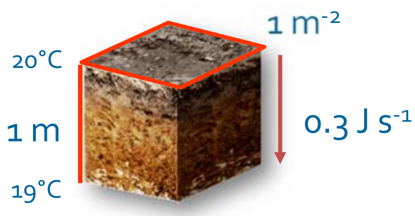
$J\ s^{-1}\ m^{-2}\ m\ K^{-1} \rightarrow W\ m^{-1}\ K^{-1}$

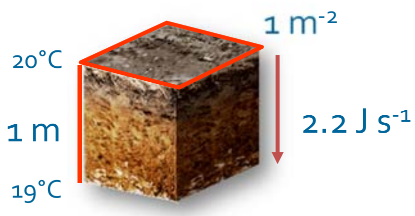
Material	Status	Heat conductivity λ ($W\ m^{-1}\ K^{-1}$)
Sandy soil	dry	0.30
(40% pore volume)	saturated	2.20
Clay	Dry	0.25
(40% pore volume)	saturated	1.58
Bog soil	dry	0.06
(80% pore volume)	saturated	0.50
Snow	old	0.08
	new	0.42
Ice	0 °C, pure	2.24
Water	4 °C	0.57
Concrete	dense	1.51

Θερμοκρασία επιφάνειας

Assuming two cases with equal energy uptake at the surface e.g. **sandy soil dry** (left, 0.3 W K^{-1} m^{-1}) and

40 % water saturated (right, 2.2 W K^{-1} m^{-1}).





- Bad heat conductivity
- Strong vertical temperature gradient
- Heat stored close under surface
- High surface temperature

- Good heat conductivity
- Medium vertical temperature gradient
- Heat stored over a bigger volume
- Lower surface temperature

Θερμοκρασία επιφάνειας

Describes the specific heat needed to raise the temperature of 1 kg of mass by 1 K = the **volumetric heat capacity** C (J m^{-3} K^{-1}) normalized by the **density** ρ (kg m^{-3})

$$c = C / \rho$$

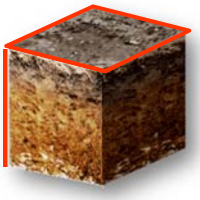
J K^{-1} kg^{-1}

Material	Status	Specific heat capacity *c* (J kg^{-1} K^{-1} x 10^3)
Sandy soil	dry	800
(40% pore volume)	saturated	1480
Clay	dry	890
(40% pore volume)	saturated	1550
Bog soil	dry	1920
(80% pore volume)	saturated	3650
Snow	old	2090
	new	2090
Ice	0 °C, pure	2100
Water	4 °C	4180
Concrete	dense	880

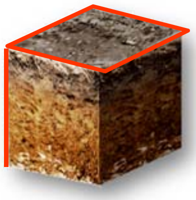
Θερμοκρασία επιφάνειας

Assuming two cases with equal energy uptake at the surface
e.g. **sandy soil dry** (left, 890 x 10^3 J K^{-1} kg^{-1}) and

40 % water saturated (right, 1400 x 10^3 J K^{-1} m^{-1})



890 x 10^3 J K^{-1} kg^{-1}



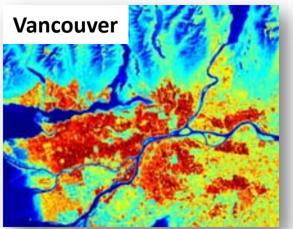
1400 x 10^3 J K^{-1} kg^{-1}

- **Low heat capacity**
- **Heat uptake leads to fast temperature increase**
- **High surface temperature**

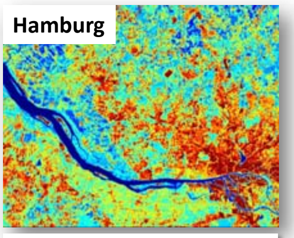
- **High heat capacity**
- **Heat uptake leads to slow temperature increase**
- **Lower surface temperature**

Due to very high heat capacity of water all materials with higher water content heat up or cool down quite slowly !

Θερμοκρασία επιφάνειας

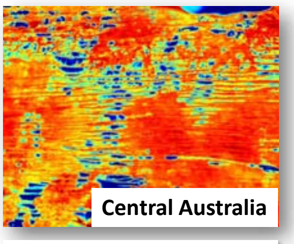


Surface properties/altitude

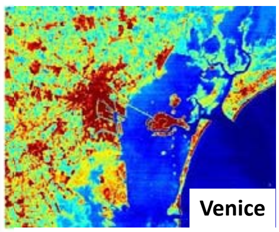


Surf. properties/soil wetness

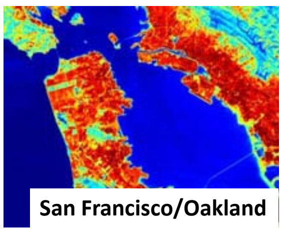

Surf. Properties/emissivity



Soil wetness

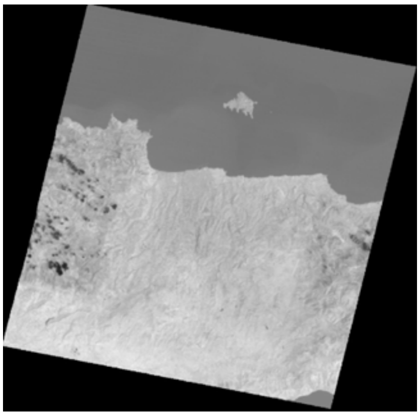


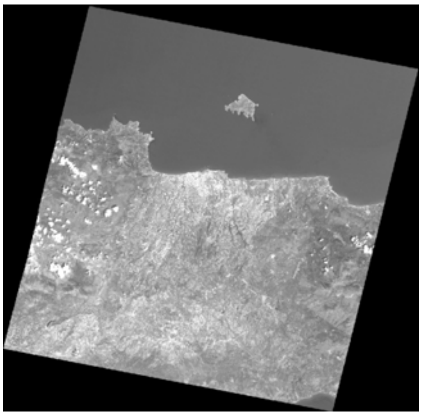
Surf. properties/soil wetness

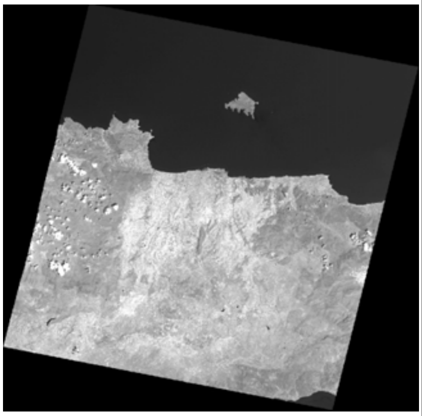


Surface properties

Θερμική απεικόνιση







Δορυφορική καταγραφή στο θερμικό υπέρυθρο

$$L_i^{sat} = \int_{\lambda_1}^{\lambda_2} f_i(\lambda)\varepsilon(\lambda)B(\lambda, T_S)\tau(\lambda)d\lambda + \int_{\lambda_1}^{\lambda_2}\int_{p=0}^{p_s} f_i(\lambda)B(\lambda, T_p)\frac{d\tau}{dp}\,d\lambda\, dp$$

$$+ \ 1/2 \int_{\lambda_1}^{\lambda_2}\int_{\theta=0}^{\pi/2}\int_{\phi=0}^{2\pi} (1-\varepsilon(\lambda))\, f_i(\lambda)L^{\propto}(\lambda,\theta,\phi)\tau(\lambda)\sin 2\theta\, d\lambda\, d\theta\, d\phi$$

- λ wavelength; i channel; λ1, λ2 lower/upper limits of spectral range;
- fi normalized channel response function;
- θ zenith angle, ϕ azimuth angle;
- p pressure, ps pressure at Earth's surface;
- τ(λ) spectral atmospheric transmissivity;
- ε(λ) surface spectral emissivity; T_S surface temperature;
- L∝(λ,θ,ϕ) downwelling irradiance divided by π;
- T_p mean air temperature at pressure level p.

RSLab

Δορυφορική καταγραφή στο θερμικό υπέρυθρο

The variation of ε(λ) with TS is negligible. For simplification purposes the downward radiation may be considered independent of azimuth and for channel i Planck's function Bi can be written as:

$$B_i(T) = \int_{\lambda_1}^{\lambda_2} f_i(\lambda) B(\lambda, T)\, d\lambda$$

If L∝hem is the downward hemispherical spectral irradiance:

$$L^{\propto}_{hem} = \pi \int_0^{\pi/2} L^{\propto}(\lambda, \theta) \sin 2\theta \, d\theta$$

$$L_i^{sat} = \tau_i \, \varepsilon_i B_i (T_s) + S_{\uparrow} + \tau_i \, (1 - \varepsilon_i) \frac{S_{\downarrow}}{\pi}$$

The first term denotes the emission from the surface, which is usually warmer than the atmosphere and is least affected by the atmosphere in an atmospheric window.

The second term varies strongly with the vertical structure of the atmosphere-warm/moist layers increase its contribution.

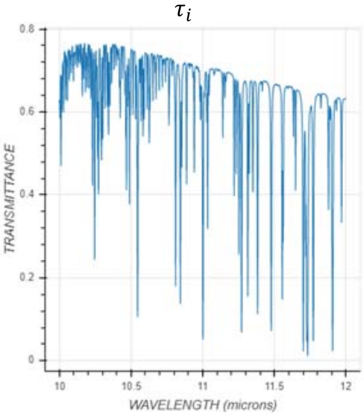
The third term describes the reflection of downwelling radiation. For a spectral emissivity close to unity it is nearly zero.

RSLab

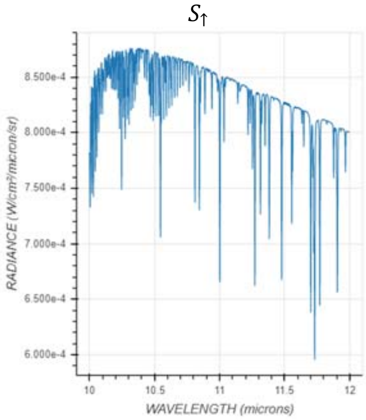
Ατμοσφαιρικές παράμετροι

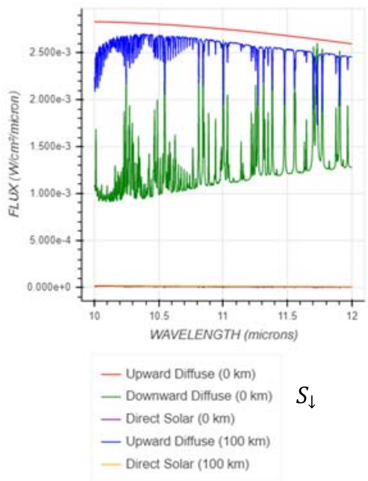
$$L_i^{sat} = \tau_i \, \varepsilon_i B_i (T_s) + S_{\uparrow} + \tau_i \, (1 - \varepsilon_i) \frac{S_{\downarrow}}{\pi}$$

τ_i



$S_{\uparrow}$



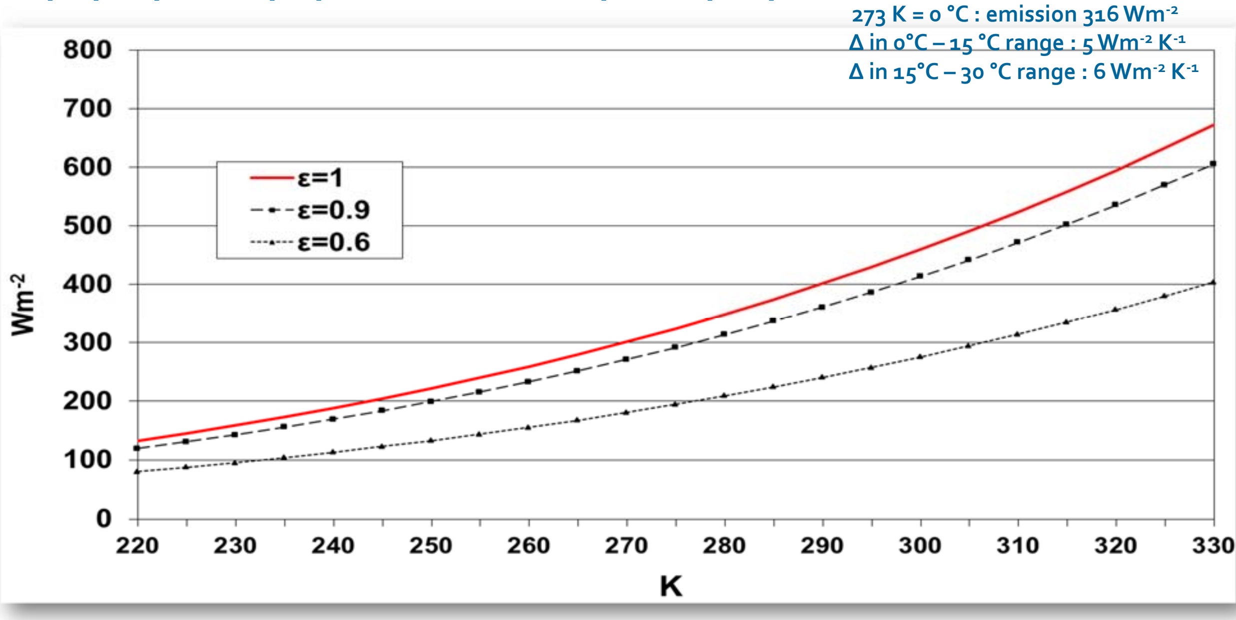


$S_{\downarrow}$

RSLab

Παραμετροποίηση του συντελεστή εκπομπής

273 K = 0 °C : emission 316 Wm^{-2}
Δ in 0°C – 15 °C range : 5 $Wm^{-2} K^{-1}$
Δ in 15°C – 30 °C range : 6 $Wm^{-2} K^{-1}$



RSLab

Παραμετροποίηση του συντελεστή εκπομπής

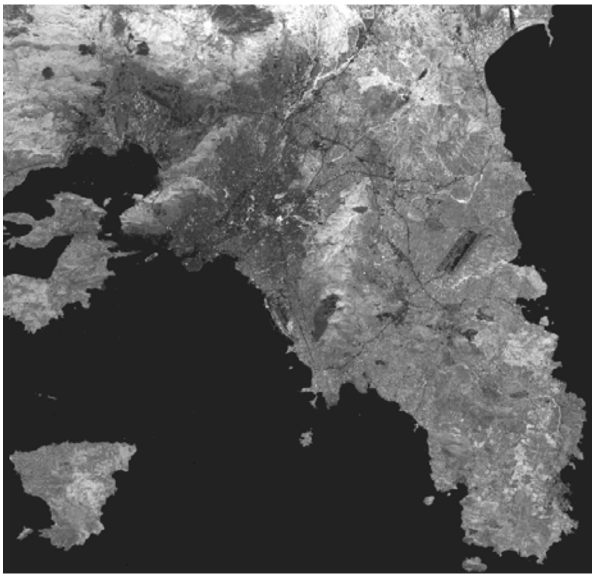
$$NDVI = \frac{\rho_{NIR} - \rho_{RED}}{\rho_{NIR} + \rho_{RED}}$$

$$P_v = \begin{cases} 0\ , & NDVI < 0.2\ (pure\ soil) \\ \dfrac{(NDVI - NDVI_s)^2}{(NDVI_v - NDVI_s)^2} & , 0.2 \le NDVI \le 0.5 \\ 1\ , & NDVI > 0.5\ (fully\ vegetative) \end{cases}$$

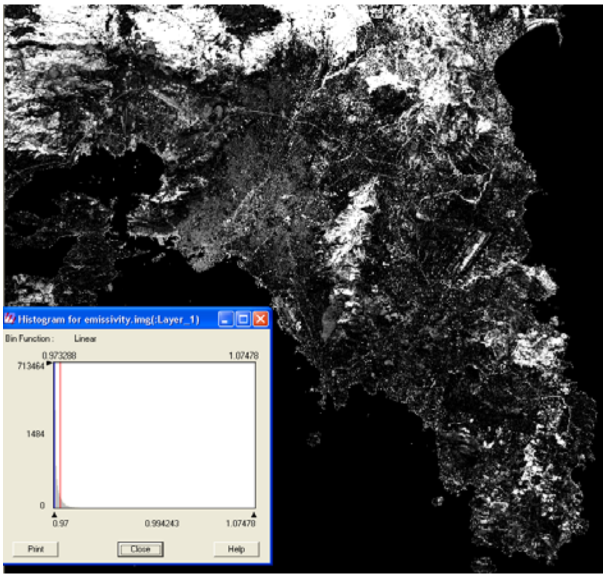
$$\varepsilon = \begin{cases} 0.97, & NDVI < 0.2\ (pure\ soil) \\ 0.97(1-P_v) + 0.99\,P_v & , 0.2 \le NDVI \le 0.5 \\ 0.99, & NDVI > 0.5\ (fully\ vegetative) \end{cases}$$

Παραμετροποίηση του συντελεστή εκπομπής

NDVI



ε



Μονοκαναλικοί αλγόριθμοι

Single-channel methods use radiance measurements in one thermal infrared channel and correct the atmospheric effects to determine LST.

LST can be retrieved from a single infrared channel through an accurate radiative transfer model if surface emissivity is known and temperature and water vapor profile is given by either satellite soundings or conventional radiosonde data.

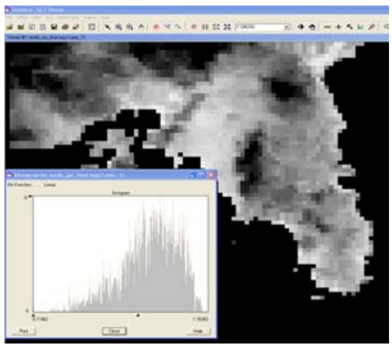
$$B_i(T_s) = \frac{L_i^{sat} - \tau_i (1-\varepsilon_i) \dfrac{S_\downarrow}{\pi} - S_\uparrow}{\tau_i \, \varepsilon_i}$$

L_i^{sat}

PW:

↓

$\tau_i, S_\uparrow, S_\downarrow$



Μονοκαναλικοί αλγόριθμοι

$$B_\lambda(T) = \frac{C_1}{\lambda^5 \left(e^{\frac{C_2}{\lambda T}} - 1\right)}$$

$$T = \frac{C_2}{\lambda \ln\left[\frac{C_1}{\lambda^5 B_\lambda(T)} + 1\right]}$$

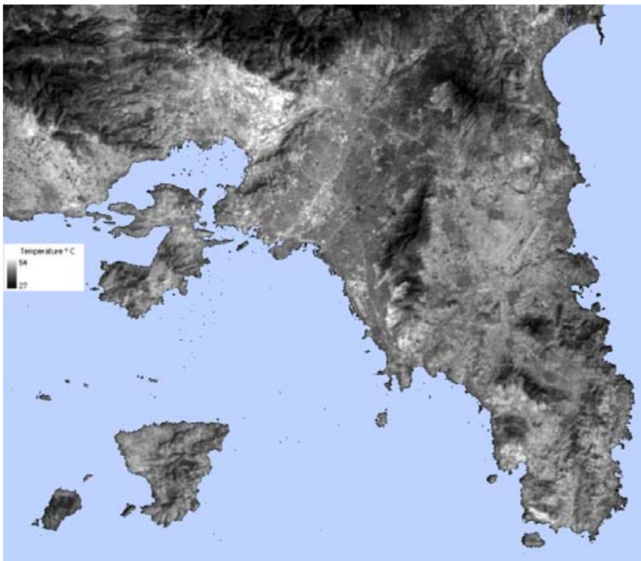
$C_1 = 1.1911 \times 10^8$ W m^{-2} sr^{-1} μm^4

$C_2 = 1.4388 \times 10^4$ K μm

$$B_i(T_s) = \frac{K_1}{e^{\frac{K_2}{T}} - 1}$$

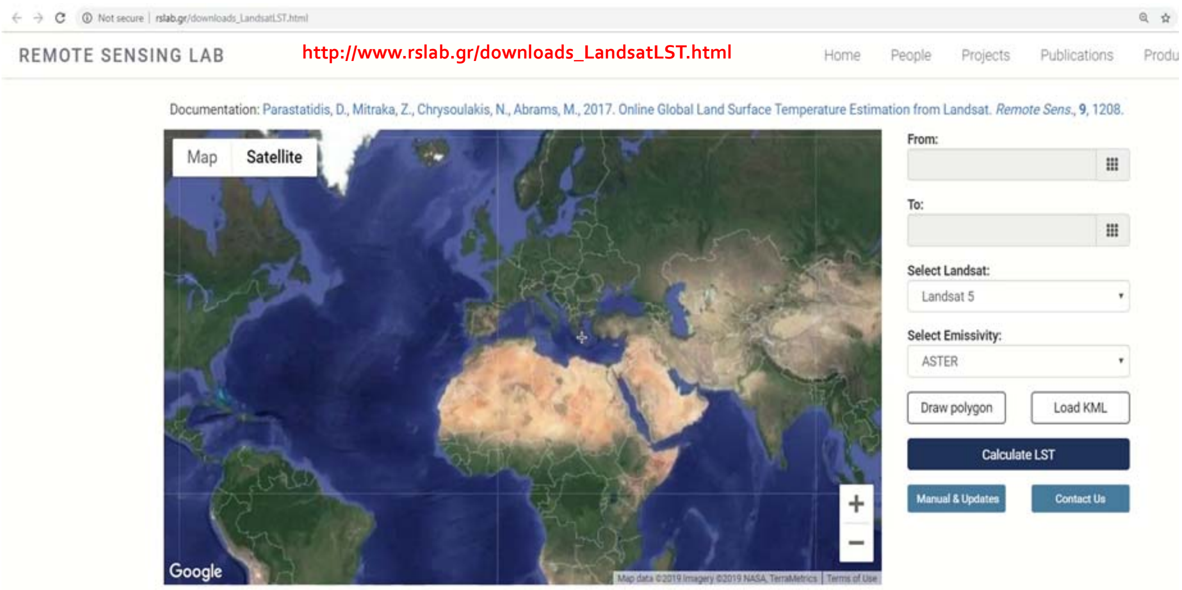
$$T_s = \frac{K_2}{\ln\left[\frac{K_1}{B_i(T_s)} + 1\right]}$$

$K_1 = 607$ Wm^{-2}μm^{-1}sr^{-1} και $K_2 = 1260{,}56$ K.

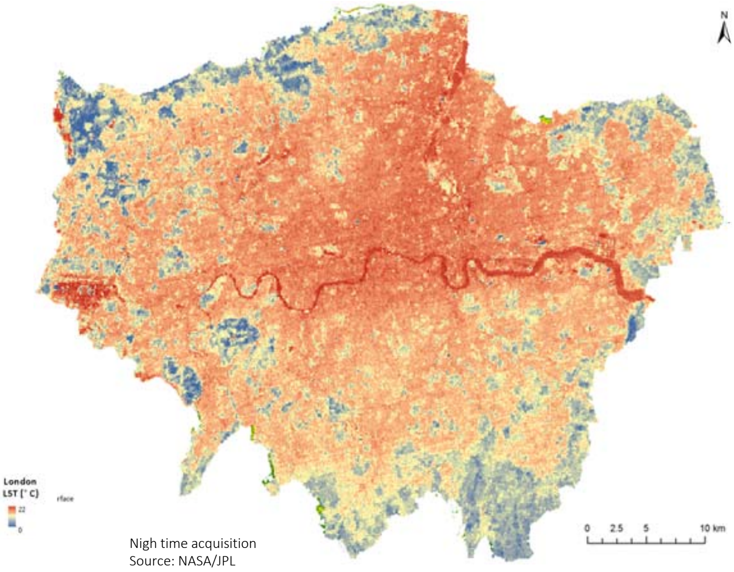


On-line εφαρμογή θερμοκρασίας επιφάνειας

http://www.rslab.gr/downloads_LandsatLST.html



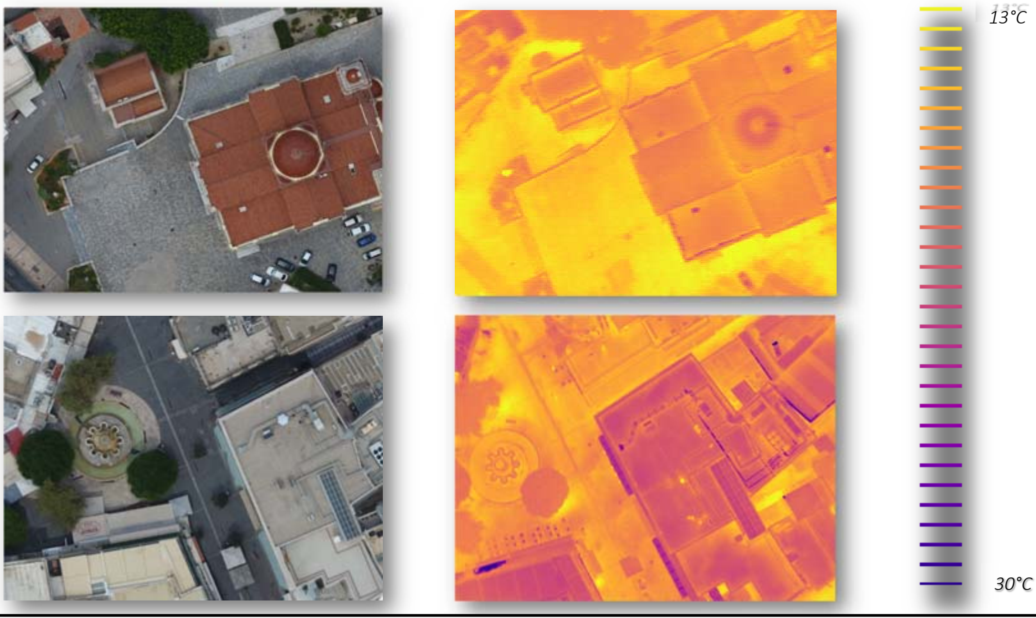
Θερμοκρασία αστικής επιφάνειας



Nigh time acquisition
Source: NASA/JPL

London Barbican

Θερμοκρασία αστικής επιφάνειας

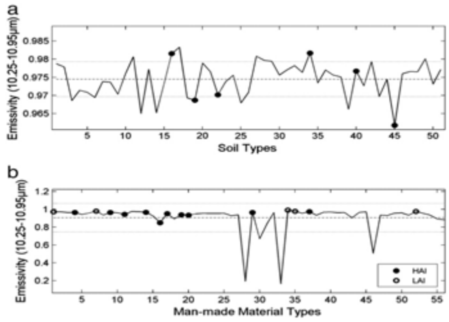


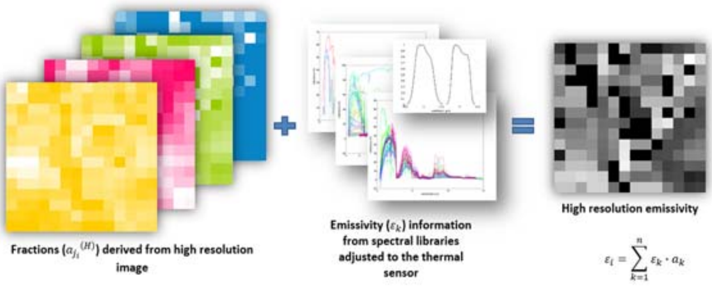
Θερμοκρασία αστικής επιφάνειας

Emissivity based on fractional land cover:

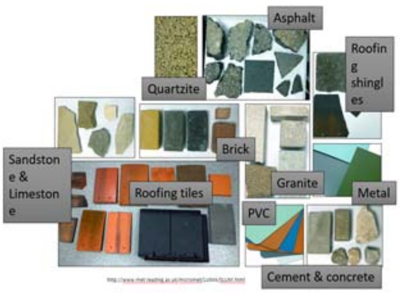
- Linear Spectral Mixture analysis
- Constraint using mean absolute deviations
- End members selection.
- Emissivity assignment to each end member.

- Emissivity estimation as: $\varepsilon = \sum_{k=1}^{n} \varepsilon_k \cdot f_k$

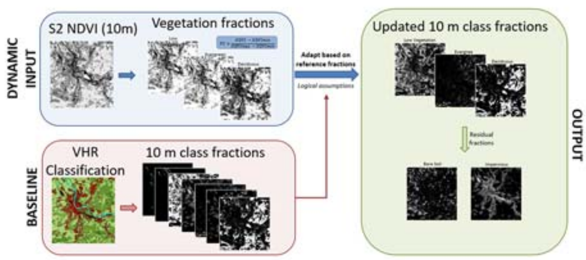


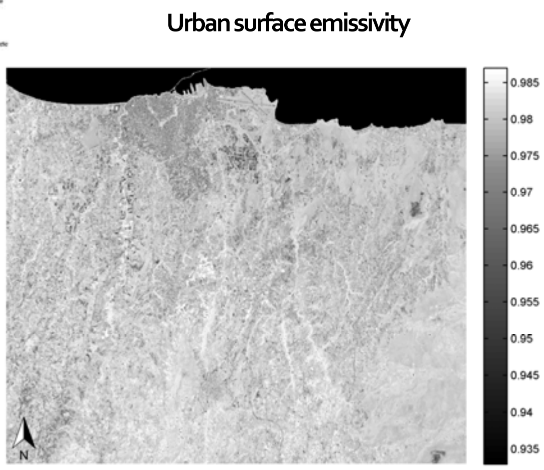


Θερμοκρασία αστικής επιφάνειας

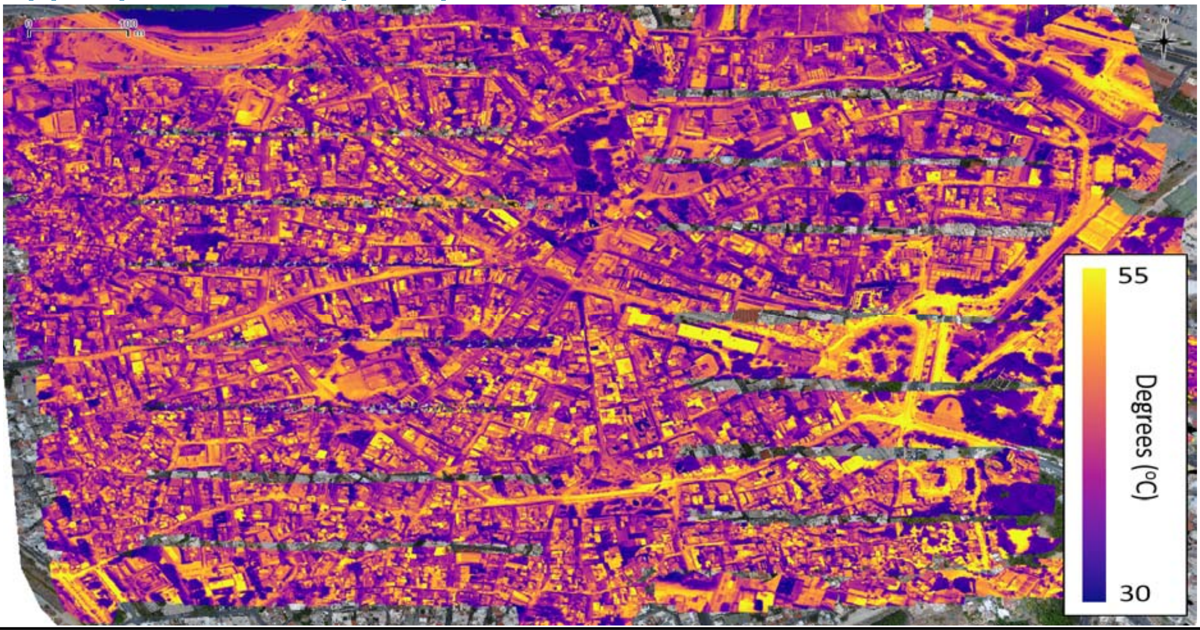


Urban vegetation



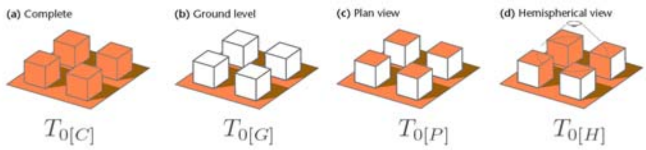


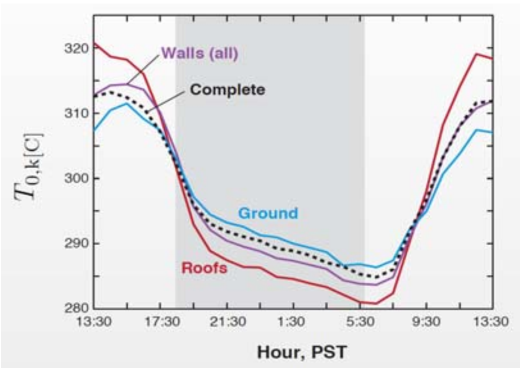
Θερμοκρασία αστικής επιφάνειας

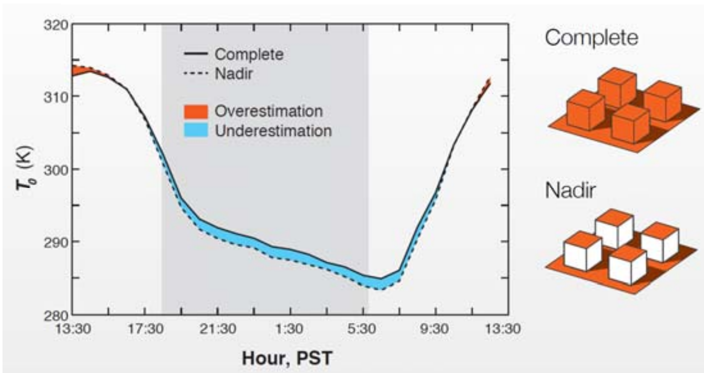


Θερμοκρασία αστικής επιφάνειας

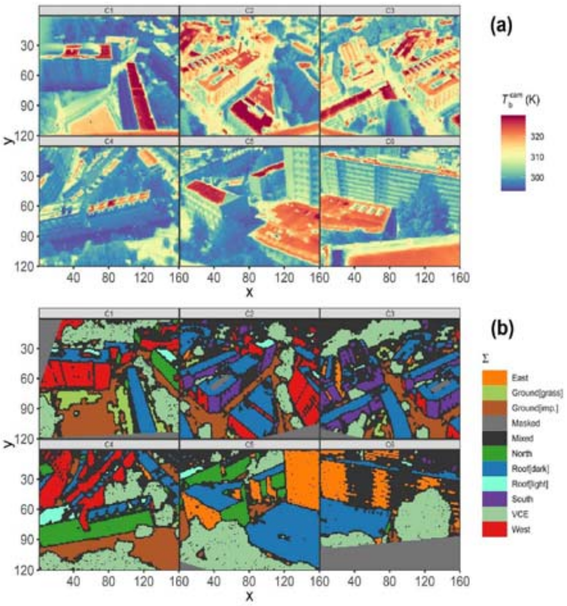
Because of the three dimensionality - different sensing systems `see' different parts of the urban surface. Which surface do we mean? How do we calculate it?

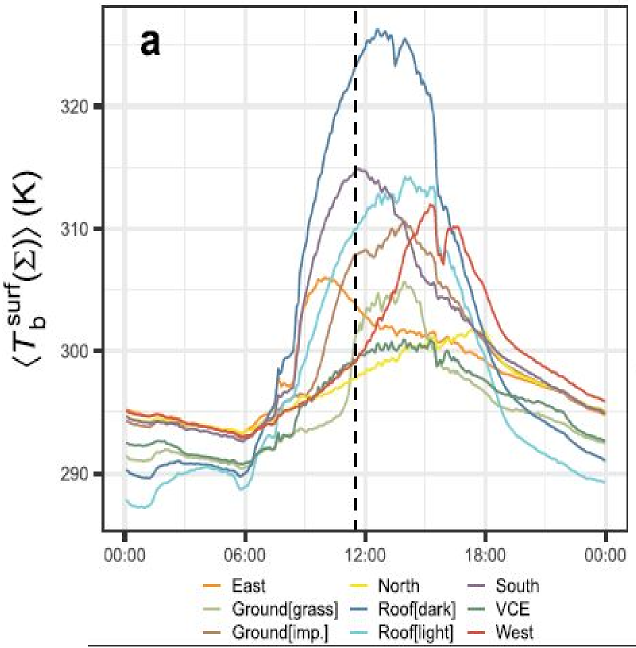




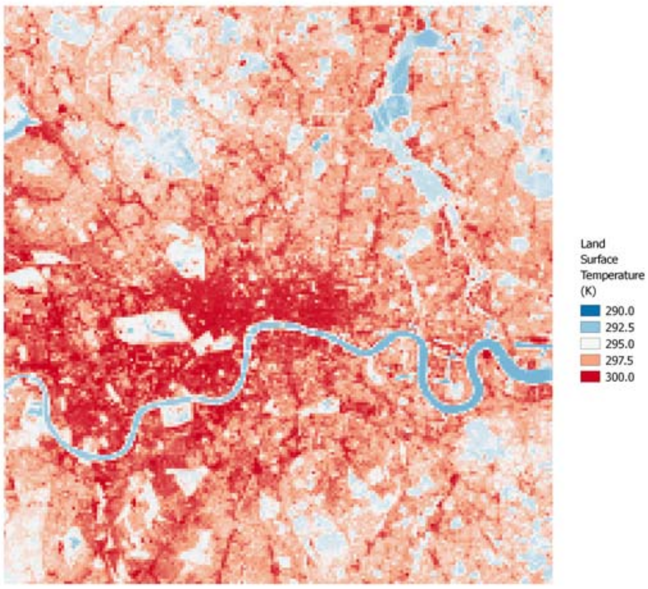


Θερμοκρασία αστικής επιφάνειας



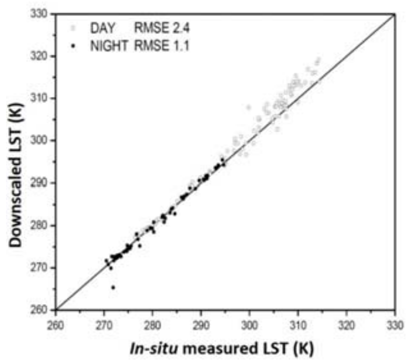


Υποβιβασμός κλίμακας



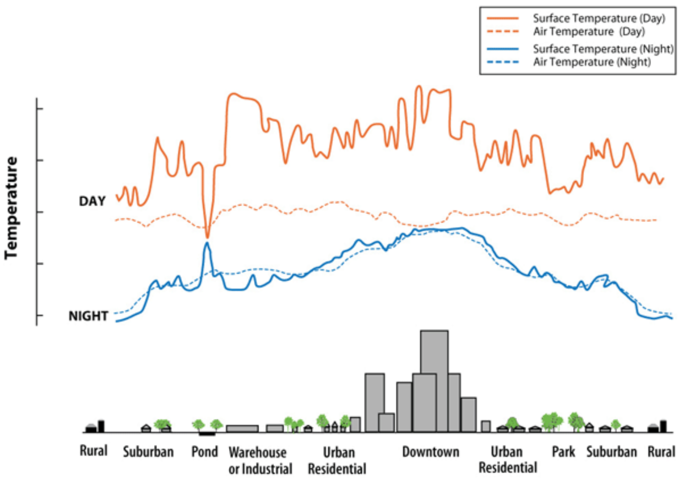
London, 19 July 2016, 22:05

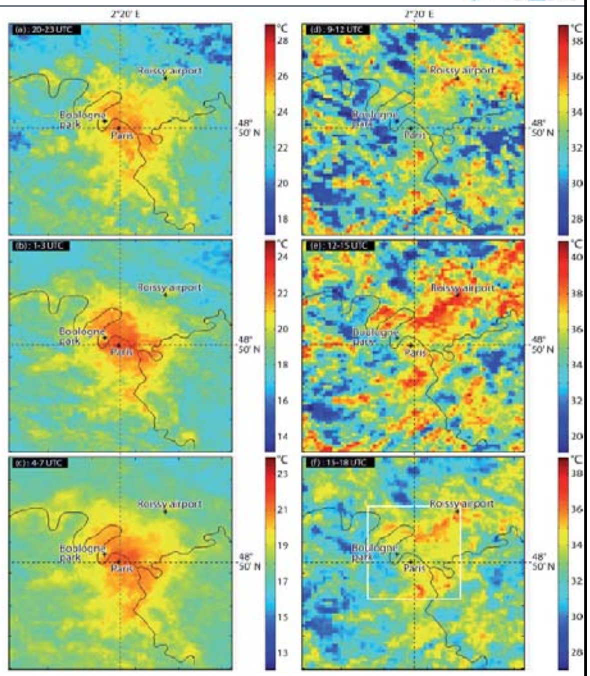
MODIS LST at 1 km x 1 km



RSLab

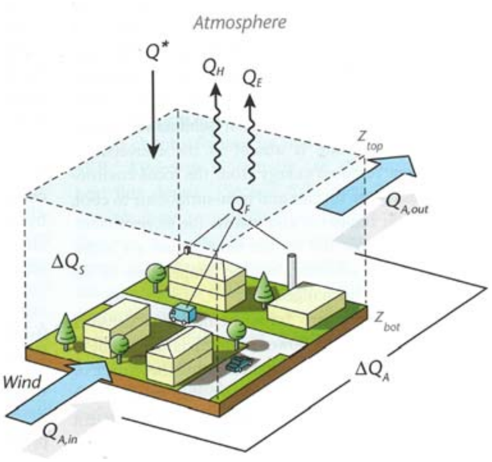
Αστική θερμική νησίδα





RSLab

Ενεργειακό ισοζύγιο



$$Q^* + Q_F = Q_H + Q_E + \Delta Q_S + \Delta Q_A$$

$$Q^* = K^* + L^* = K_{\downarrow} - K_{\uparrow} + L_{\downarrow} - L_{\uparrow}$$

$K_{\uparrow} = \alpha K_{\downarrow}$

$L_{\downarrow} = \varepsilon_a \sigma T_a^4$

$L_{\uparrow} = \alpha_{long} L_{\downarrow} + \varepsilon \sigma T_s^4 = \alpha_{long} \varepsilon_a \sigma T_a^4 + \varepsilon \sigma T_s^4$

$\mathbf{K_{\downarrow} - K_{\uparrow} = (1-\alpha) K_{\downarrow}}$

$L_{\downarrow} - L_{\uparrow} = \varepsilon_a \sigma T_a^4 - \alpha_{long} \varepsilon_a \sigma T_a^4 - \varepsilon \sigma T_s^4 = (1- \alpha_{long}) \varepsilon_a \sigma T_a^4 - \varepsilon \sigma T_s^4$

$(1- \alpha_{long}) = A_{long} = \varepsilon$

$\mathbf{L_{\downarrow} - L_{\uparrow} = \varepsilon \varepsilon_a \sigma T_a^4 - \varepsilon \sigma T_s^4}$

$$Q^* = (1-\alpha) K_{\downarrow} + \varepsilon \varepsilon_a \sigma T_a^4 - \varepsilon \sigma T_s^4$$

Ενεργειακό ισοζύγιο

http://rslab.gr/heraklion_eddy.html