Notas de Geotermia e Energia Geotérmica

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Aulas de Mestrado
Recursos Hidrogeológicos e Geoenergia

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I
The Heat Conduction Equation

Introduction

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K\,\frac{dT}{dz}$$

Fourier equation

$$\frac{\partial T}{\partial t} = \alpha \cdot \frac{\partial^2 T}{\partial z^2}$$

Heat conduction equation

$$T = T_0 + \frac{Q_0}{K} \cdot z - \frac{A'}{2K} \cdot z^2$$

Part 1

General aspects of thermal energy in the Earth

Energy balance at the Earth's surface

- The solar energy that reaches the Earth per unit time is 2×10^{17} W (4×10^{2} Wm^{-2})
- The energy from the Earth's interior that reaches the surface per unit time is 4.4×10^{13} W (8.7×10^{-2} Wm^{-2})
- The energy per unit time spent to desaccelerate Earth's rotation (tides) is about 10^{12} W
- The amount of energy per unit time released in earthquakes is about 10^{11} W.

Energy balance at the Earth's surface

(normalized to mean HFD at the Earth's surface)

• Solar energy	4000
• Mean HFD at the Earth's surface	1
• Desacceleration by tides	0,1
• Earthquakes	0,01

Geothermal energy manifestations

- Direct evidences
  - Volcanos
  - Thermal springs
  - Temperature measured in boreholes and mines

Geothermal energy manifestations

- Indirect evidences
  - Plate tectonic motions
  - Seismic activity
  - Orogeny
  - Earth’s magnetic field
  - Metamorphic processes

Thermal history of the Earth

- It is not possible to reconstruct the Earth's thermal history

- However, plausible models can be constructed based on:

  - The general heat transmission theory
  - Measurements of the heat flux from the Earth's interior
  - Radioactive studies and meaurements in rocks
  - Study of the physical properties of rocks

Thermal history of the Earth

The firt attempt to estimate the Earth's age was done by William Thomson (Lord Kelvin)

$$t = \frac{T_0{}^2}{\left(\frac{dT}{dz}\right)^2_{z=0} \cdot \pi \cdot \alpha}$$

$$T_0 = 3871\ ^{o}C \quad (= 7000\ ^{o}F)$$

$$(dT/dz)_{z=0} = 0{,}036\ ^{o}C\, m^{-1}$$

$$\alpha = 1{,}4 \times 10^{-6}\ m^2\, s^{-1}$$

$$t = 100\ Ma$$

Heat sources inside the Earth

- Uranium, thorium and potassium radioactive decay
- Conversion of gravitational energy into thermal energy
- Heat resulting from thermodynamic processes involved in the Earth's formation

Temperature evolution inside the Earth

1. In a first phase, about 4.7 billion years, planetary formation and compression processes warmed the Earth's interior to temperatures of the order of 1000 ^{o}C

2. In a second phase, heat generation by radioactive processes started and the Earth's interior temperature continued to increase

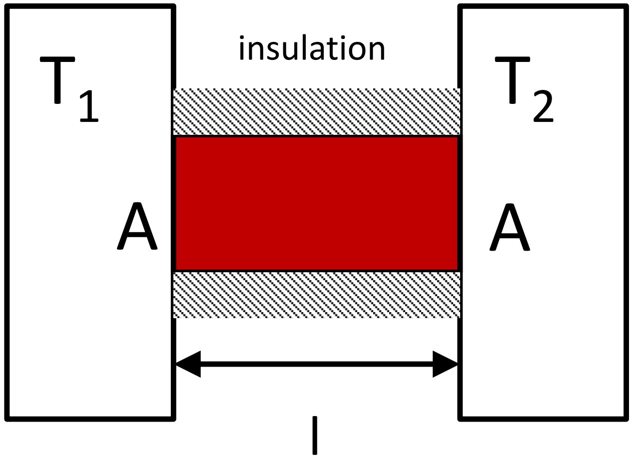
Temperature evolution inside the Earth

3. About 4 to 4.5 billion years, when the iron fusion temperature was reached, gravitational differention between the nucleous and the mantle started, which released gravitational energy of the order of 2×10^{30} J

4. All these sources of thermal energy started fusion, reorganization, and differentiation inside the Earth, and the formation of a core, a mantle and a crust began

Types of energy transfer in the Earth

- Conduction – Thermal energy transfer as a kinetic energy transfer among the atoms and molecules of a given material



$$\Delta Q = K \cdot A \cdot \frac{T_2 - T_1}{L} \cdot \Delta t$$

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K \frac{dT}{dz}$$

Types of energy transfer in the Earth

- Convection - Thermal energy transfer with motion of portions of matter with different densities resulting from temperature differences

$$Ra = \frac{\alpha \cdot \rho \cdot g \cdot \Delta T}{K \cdot \eta} \cdot D^3$$

where Ra is the Rayleigh number, α is the coefficient of thermal expansion, ρ is the density, g is the accelerarion of gravity, ΔT is the temperature difference, K is the thermal diffusivity, η is the viscosity and D^3 is a measure of the volume involved in the convection.
For Ra about 10^3 convection starts and at about 10^5 heat transfer is entirely by convection.

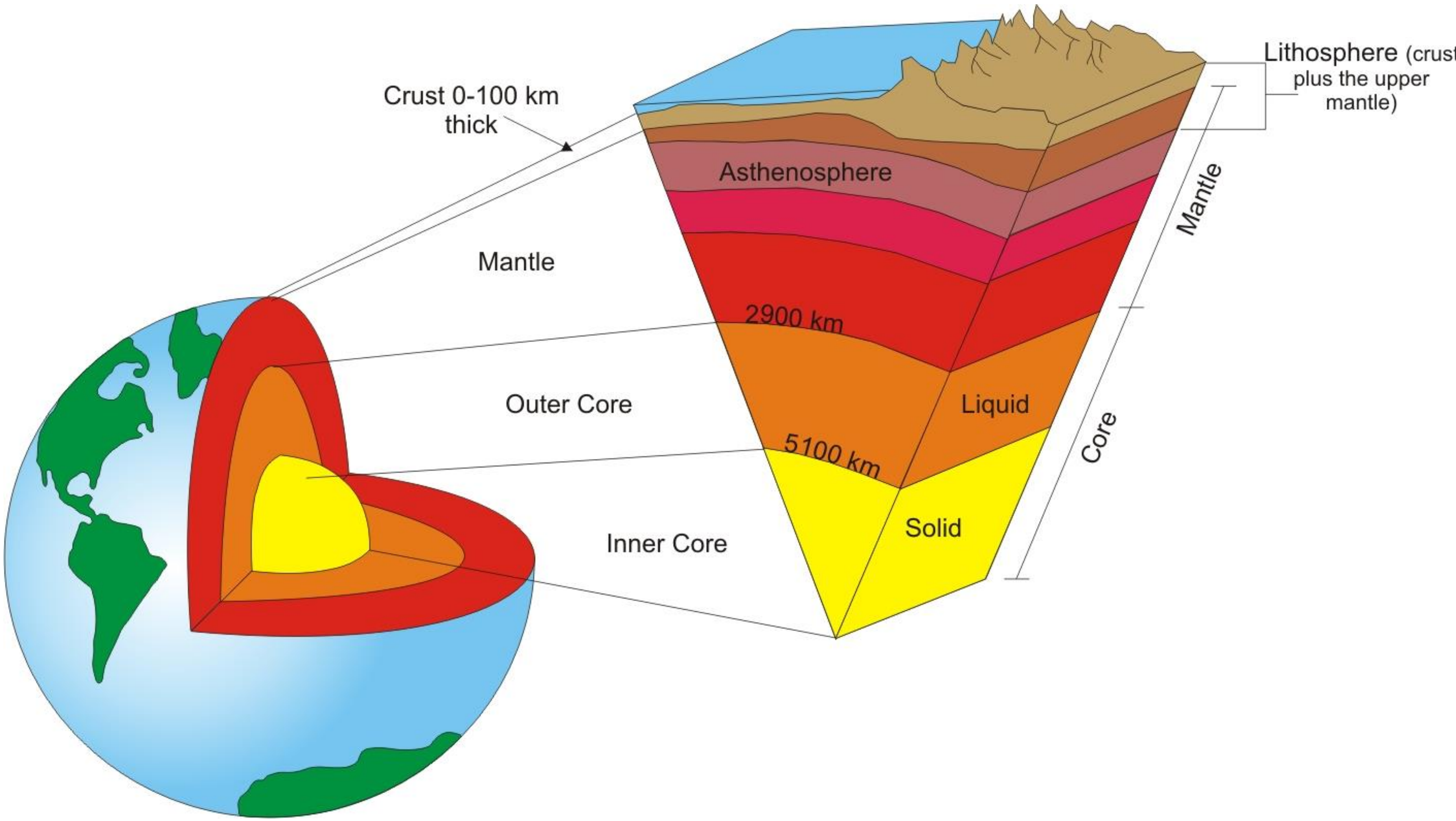
Types of energy transfer in the Earth

- Radiation – Thermal energy transfer as electromagnetic waves

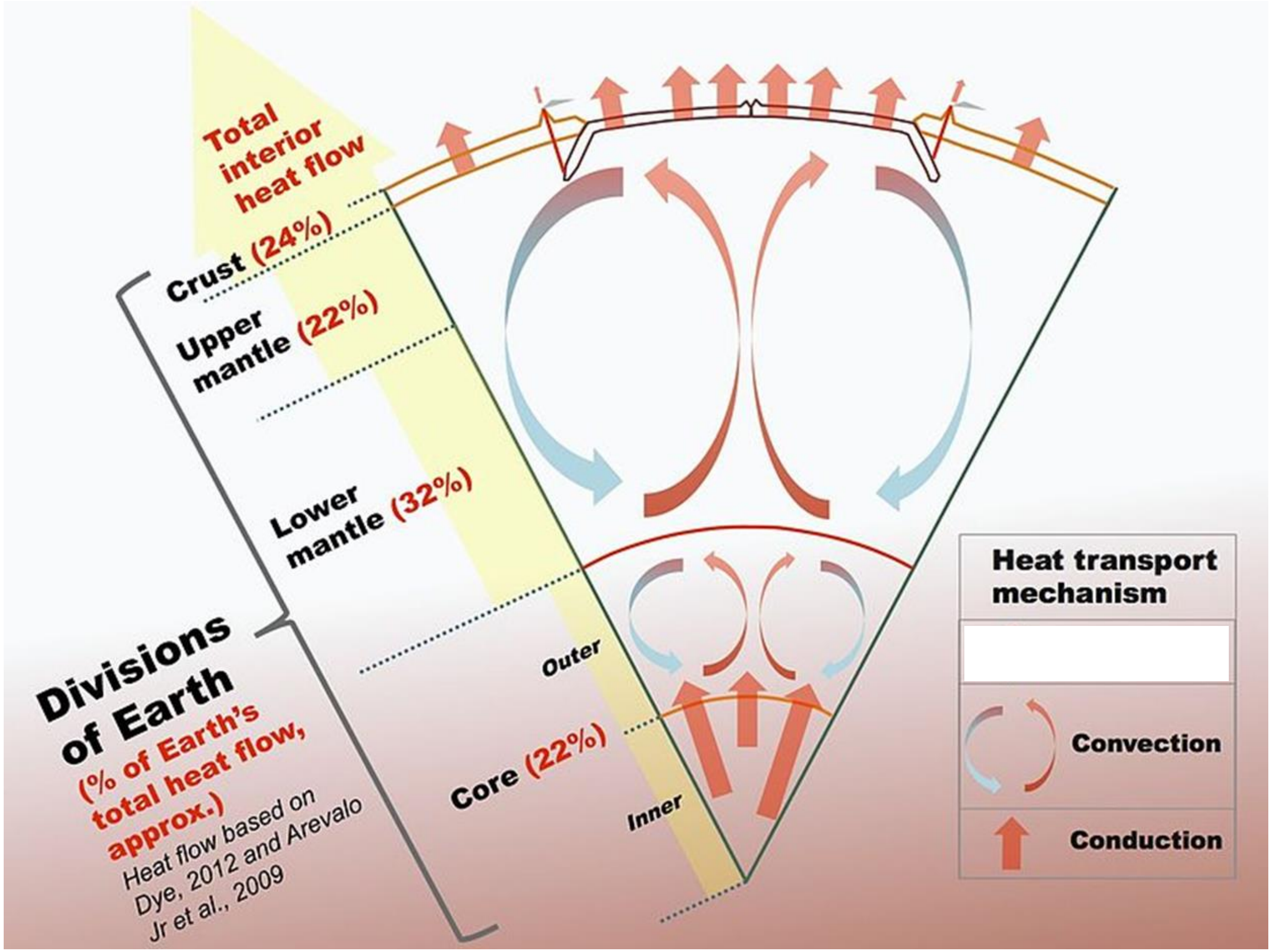
$$R = \sigma \cdot T^4$$

R represents the radiant energy per second emitted per unit area of the surface of the body at temperature T, and σ is the Stefan-Boltzman constant $\sigma = 5{,}67 x 10^{-8}$ $Wm^{-2}K^{-4}$

Types of energy transfer in the Earth



Types of energy transfer in the Earth

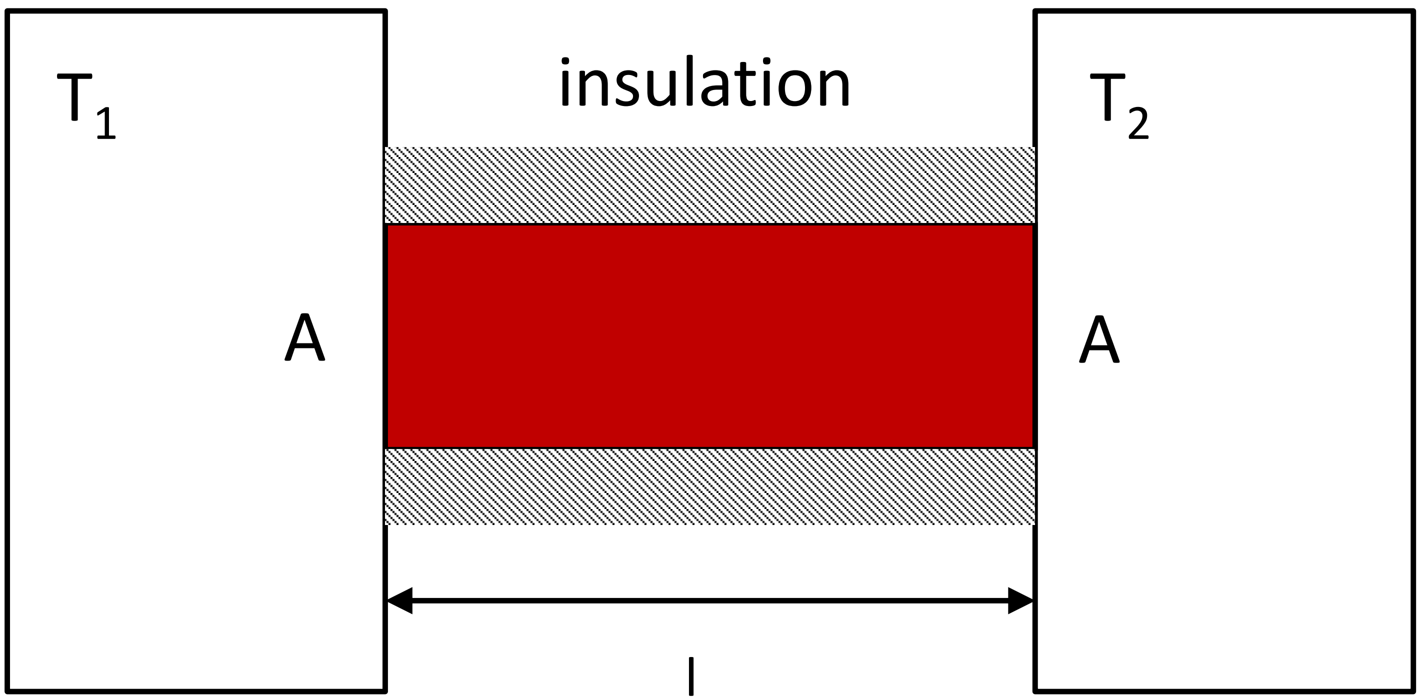


Part 2

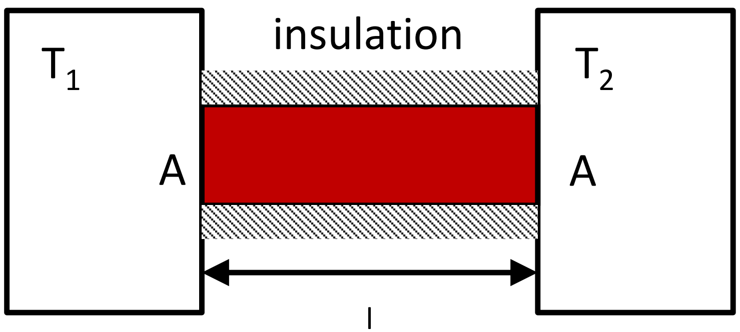
About the conduction equation and some results

Heat conduction equation

$T_2 > T_1$



Heat conduction equation



$$\Delta Q \propto A \cdot \frac{T_2 - T_1}{L} \cdot \Delta t$$

$$\Delta Q = K \cdot A \cdot \frac{T_2 - T_1}{L} \cdot \Delta t$$

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K \frac{dT}{dz}$$

Diversion about scalars, vectors and tensors

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K\,\frac{dT}{dz}$$

q_z is a vector as well as the temperature gradient dT/dz. K is a tensor and the minus sign indicates the temperature gradient has an opposite direction of q_z. We are (almost) always assuming that q_z is vertical.

Diversion about units

$$\Delta Q = K \cdot A \cdot \frac{T_2 - T_1}{L} \cdot \Delta t$$

$$K = \frac{\Delta Q \cdot L}{\Delta t \cdot \Delta T \cdot A}$$

$$[K] = \frac{[\Delta Q] \cdot [L]}{[\Delta t] \cdot [\Delta T] \cdot [A]}$$

In SI, Q is in joules or J, T is in kelvin K (or ^{o}C), L is in meter, t is in second, A in m^2. So, K should be in W/mK

$$[K] = \frac{J \cdot m}{s \cdot K \cdot m^2} = \frac{W}{m\mathbf{K}} = W/m\mathbf{K} = Wm^{-1}\mathbf{K}^{-1}$$

Diversion about units

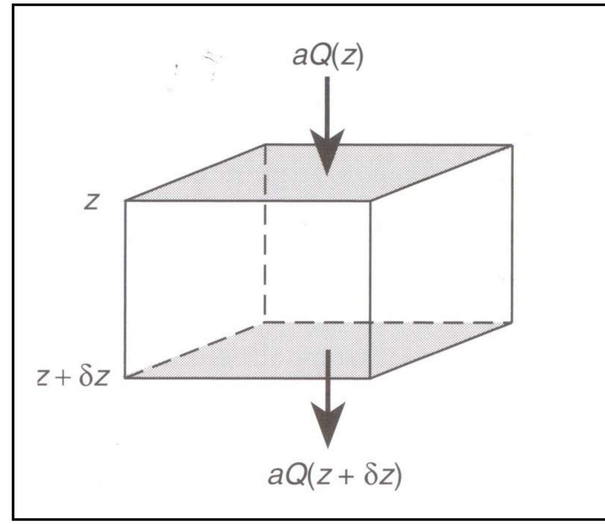
What UNITS can do or not: the Mars Climate Orbiter case

(CNN) -- ***September 30, 1999.*** NASA lost a $125 million Mars orbiter because one engineering team used metric units while another used English units for a key spacecraft operation, according to a review finding released Thursday.
For that reason, information failed to transfer between the Mars Climate Orbiter spacecraft team at Lockheed Martin in Colorado and the mission navigation team in California.

"People sometimes make errors," said Edward Weiler, NASA's Associate Administrator for Space Science in a written statement.

Heat conduction equation

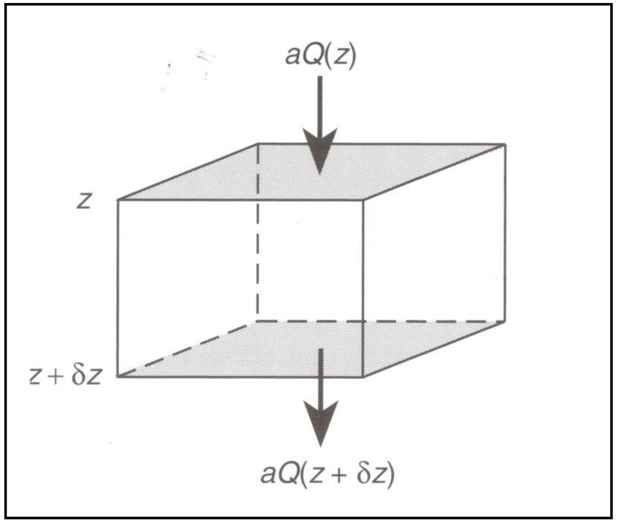
- C_p is the specific heat
- ρ is the density
- A is the heat production per unit volume
- a is the area of the element of volume
- z is the depth
- Q is the amount of heat
- T is the temperature



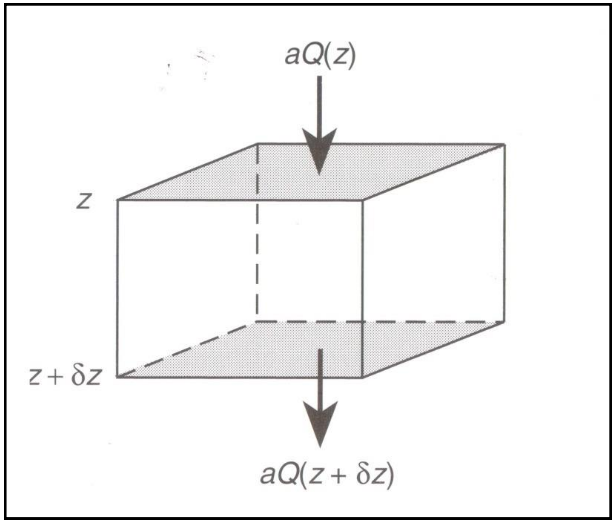
What are the units of the quantities?

Heat conduction equation

- C_p is the specific heat (J/kg.^{o}C)
- ρ is the density (kg/m^3)
- A is the heat production per unit volume (W/m^3)
- a is the area of the element of volume (m^2)
- z is the depth (m)
- Q is the amount of heat (J)
- T is the temperature (K)



Heat conduction equation



$$a \cdot Q(z) - a \cdot Q(z+\delta z) = -a \cdot \delta z \cdot \frac{\partial Q}{\partial z}$$

$$A \cdot a \cdot \delta z$$

$$c_P \cdot a \cdot \delta z \cdot \rho \cdot \frac{\delta T}{\delta t}$$

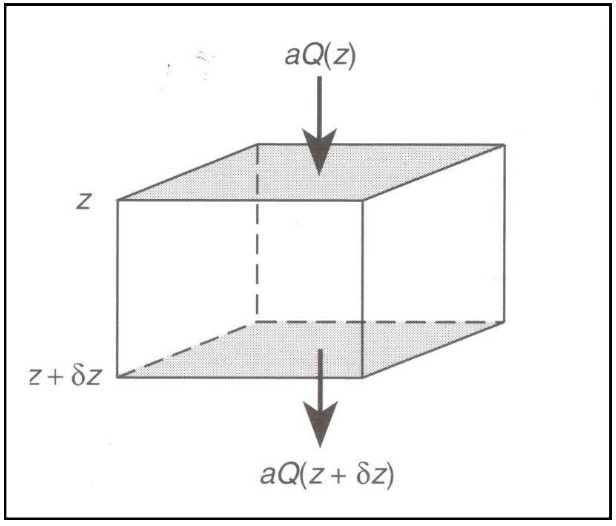
$$c_P \cdot a \cdot \delta z \cdot \rho \cdot \frac{\delta T}{\delta t} = A \cdot a \cdot \delta z - a \cdot \delta z \cdot \frac{\partial Q}{\partial z}$$

Heat conduction equation

$$c_P \cdot a \cdot \delta z \cdot \rho \cdot \frac{\delta T}{\delta t} = A \cdot a \cdot \delta z - a \cdot \delta z \cdot \frac{\partial Q}{\partial z}$$

Dividing by aδz and making δt go to 0, we obtain

$$c_P \cdot \rho \cdot \frac{\partial T}{\partial t} = A - \frac{\partial Q}{\partial z}$$



Since

$$Q_z = -K \frac{dT}{dz}$$

We obtain

$$\frac{\partial T}{\partial t} = \frac{K}{c_p \cdot \rho} \cdot \frac{\partial^2 T}{\partial z^2} + \frac{A}{c_p \cdot \rho}$$

Heat conduction equation

Heat conduction equation with heat production

$$\frac{\partial T}{\partial t} = \frac{K}{c_p \cdot \rho} \cdot \frac{\partial^2 T}{\partial z^2} + \frac{A}{c_p \cdot \rho}$$

Heat conduction equation without heat production

$$\frac{\partial T}{\partial t} = \alpha \cdot \frac{\partial^2 T}{\partial z^2}$$ with $$\alpha = \frac{K}{c_p \cdot \rho}$$

Heat conduction equation with motion of material through the region where temperature changes with depth with velocity u_z

$$\frac{\partial T}{\partial t} = \alpha \cdot \frac{\partial^2 T}{\partial z^2} + \frac{A}{c_p \cdot \rho} - u_z \cdot \frac{\partial T}{\partial z}$$

where $u_z \cdot \frac{\partial T}{\partial z}$ is the advective transfer term

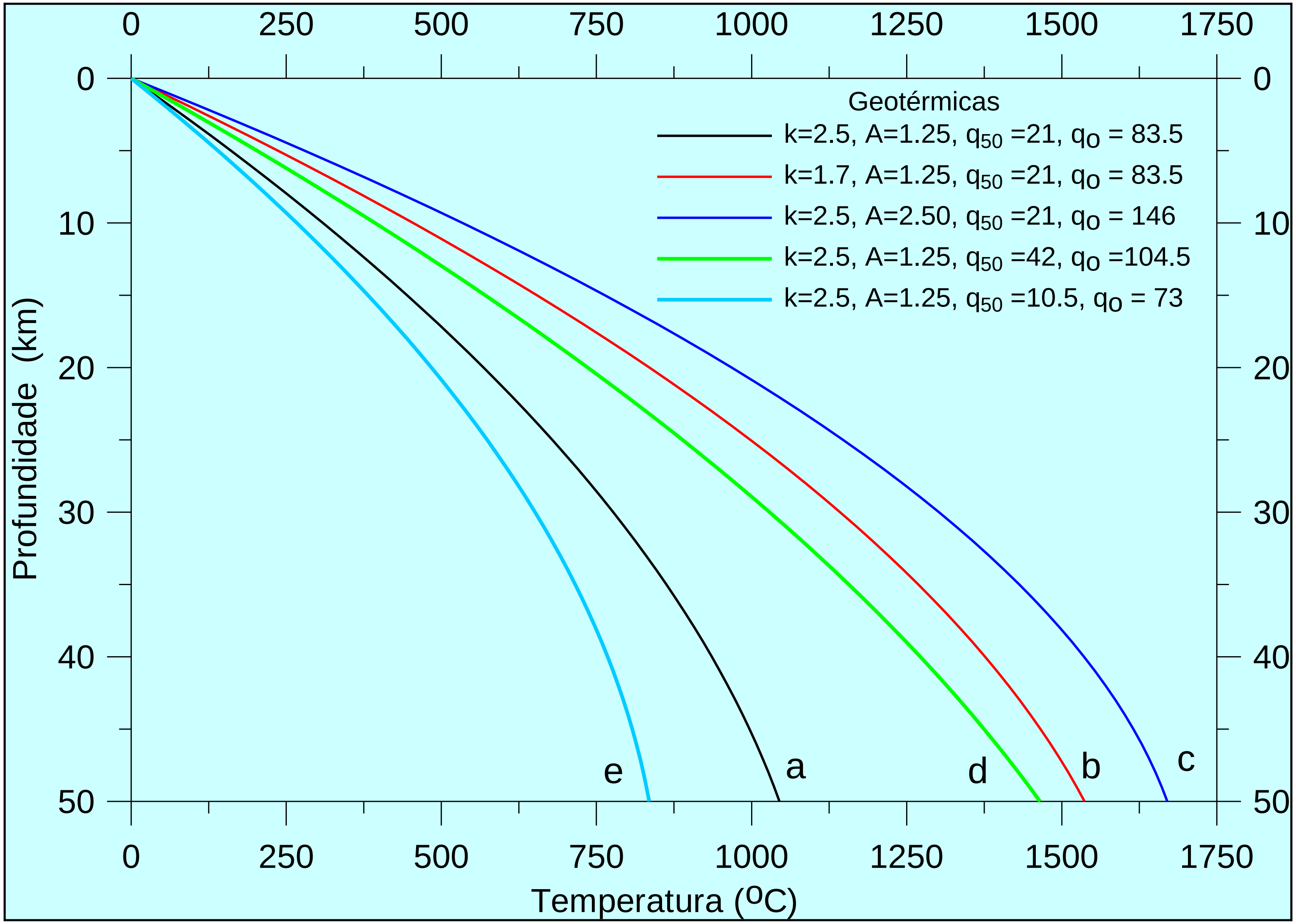
Geotherms (steady state)

$$0 = \frac{K}{\rho c_p} \cdot \frac{\partial^2 T}{\partial z^2} + \frac{A}{\rho c_p}$$

$T = T_0$ for $z = 0$ $\qquad$ $Q = Q_o$ for $z = 0$

$$T = T_0 + \frac{Q_0}{K} \cdot z - \frac{A}{2K} \cdot z^2$$

Geotherms (steady state)



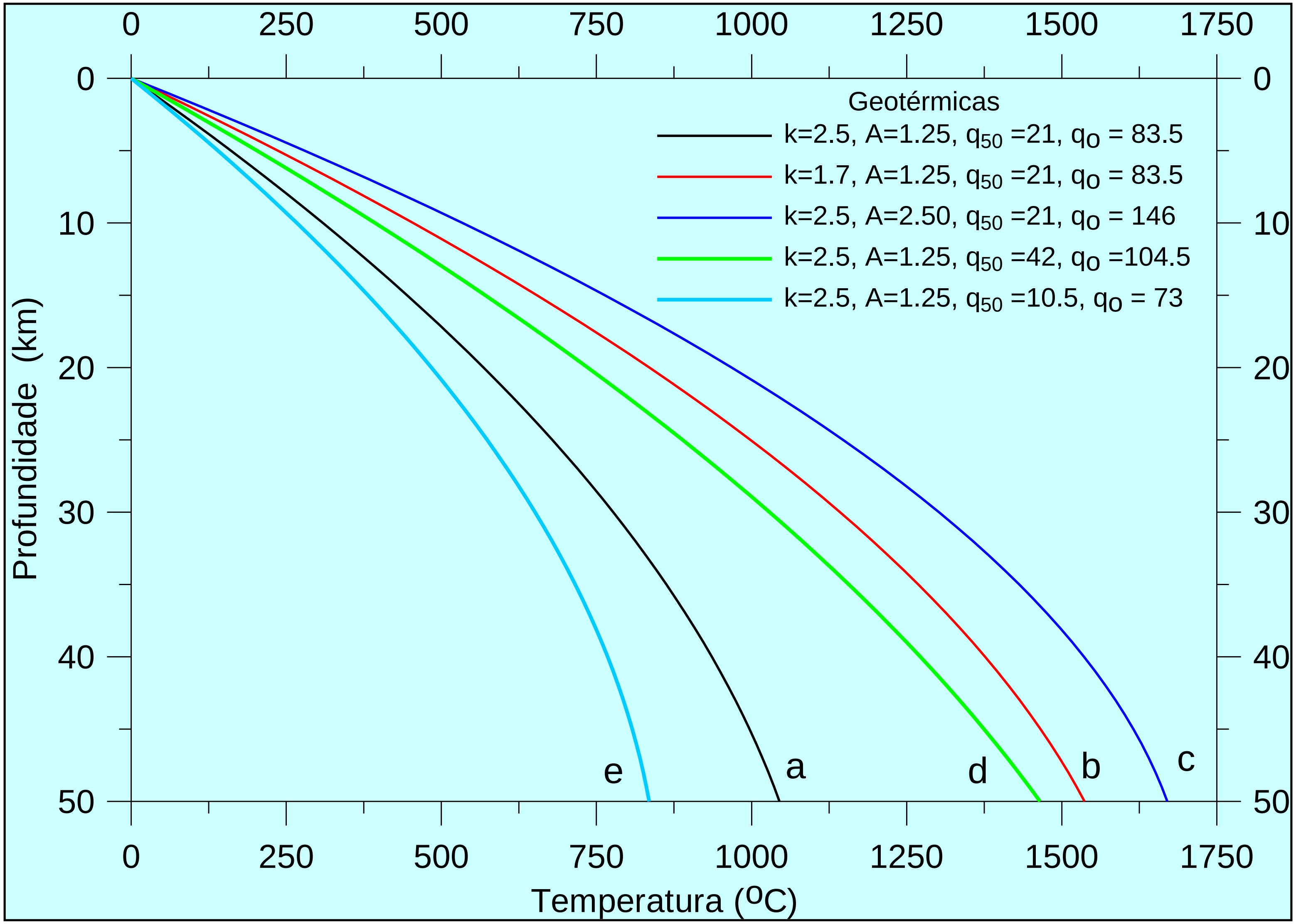
Geotherms (steady state)

$$0 = \frac{K}{\rho c_p} \cdot \frac{\partial^2 T}{\partial z^2} + \frac{A}{\rho c_p}$$

$T = 0$ for $z = 0$ $Q = Q_d$ for $z = d$

$$T = \frac{Q_d + Ad}{K} \cdot z - \frac{A}{2K} \cdot z^2$$

Geotherms (steady state)



Geotherms (steady state)

$$0 = \frac{K}{\rho c_p} \cdot \frac{\partial^2 T}{\partial z^2} + \frac{A}{\rho c_p}$$

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$Q = Q_o$ for $z = 0$

$T = 0$ for $z = 0$
$Q = Q_d$ for $z = d$

$$T = T_0 + \frac{Q_0}{K} \cdot z - \frac{A}{2K} \cdot z^2$$

$$T = \frac{Q_d + Ad}{K} \cdot z - \frac{A}{2K} \cdot z^2$$

Geotherms (steady state)

What is the interesting thing about these two equations?

$$T = T_0 + \frac{Q_0}{K} \cdot z - \frac{A}{2K} \cdot z^2 \qquad T = \frac{Q_d + Ad}{K} \cdot z - \frac{A}{2K} \cdot z^2$$

Geotherms (steady state)

Is that $Q_0 = Q_d + Ad$, i.e., a column of rock with thickness *d* and radioactive generation *A* contributes to the surface heat flow by an amount of *Ad*.

In the same way, if Q_d is the mantle heat flow, it contributes $Q_d.z/K$ to the temperature at depth z.

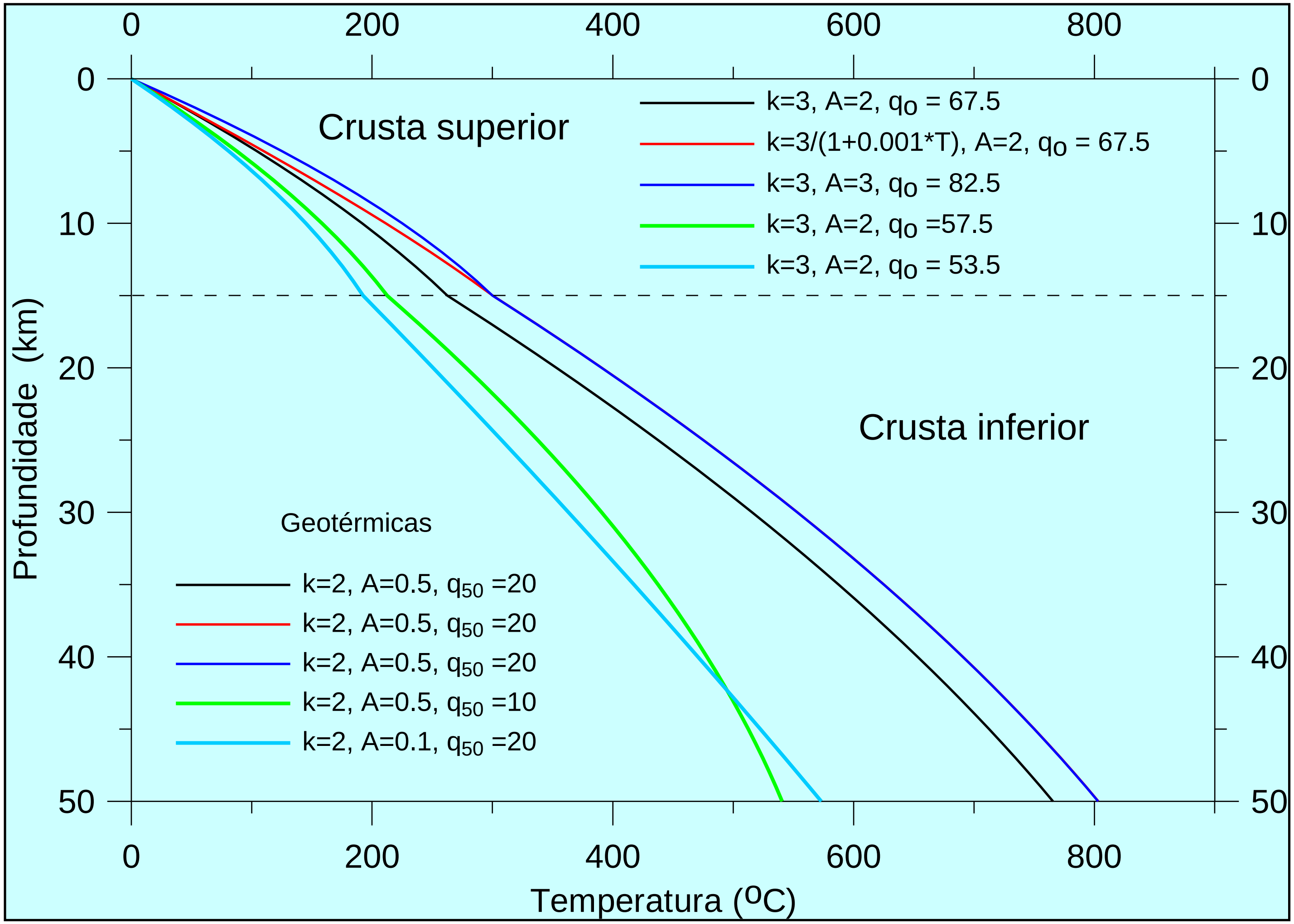
$$T = T_0 + \frac{Q_0}{K} \cdot z - \frac{A}{2K} \cdot z^2 \qquad T = \frac{Q_d + Ad}{K} \cdot z - \frac{A}{2K} \cdot z^2$$

Geotherms (steady state)

$$T = T_0 - \frac{A_1}{2K} \cdot z^2 + \left(\frac{Q_2}{K} + \frac{A_2}{K}\left(z_2 - z_1\right) + \frac{A_1 \cdot z_1}{K} \right) \cdot Z \qquad \left(0 \le z < z_1\right)$$

$$T = T_0 - \frac{A_2}{2K} \cdot z^2 + \left(\frac{Q_2}{K} + \frac{A_2 \cdot z_2}{K} \right) \cdot z + \frac{A_1 - A_2}{2K} \cdot z_1^2 \qquad \left(z_1 \le z < z_2\right)$$

Geotherms (steady state)



Heat conduction equation

Of course we can consider other complications such as non stationary temperature (variation of temperature with time) and other boundary conditions

$$\frac{\partial \mathrm{T}}{\partial \mathrm{t}} = \alpha \cdot \frac{\partial^2 T}{\partial z^2} \quad \text{with} \quad \frac{\partial \mathrm{T}}{\partial \mathrm{t}} \neq 0$$

Periodic temperature change at the Earth's surface

The heat conduction equation can be integrated to give the temperature distribution as a function of time (t) and depth (z)

$$\frac{\partial T}{\partial t} = \alpha \cdot \frac{\partial^2 T}{\partial z^2}$$

$$T(0,t) = T_0 \cdot e^{i\omega t}$$

$$T(z,t) \rightarrow 0 \quad \text{for} \quad z \rightarrow \infty$$

$$T(z,t) = T_0 \cdot \exp\left(-\sqrt{\frac{\omega \cdot \rho \cdot c_P}{2K}} \cdot z\right) \cdot \exp\left[i\left(\omega \cdot t - \sqrt{\frac{\omega \cdot \rho \cdot c_P}{2K}} \cdot z\right)\right]$$

$$L = \sqrt{\frac{2K}{\omega \cdot \rho \cdot c_P}} \qquad \Phi = \sqrt{\frac{\omega \cdot \rho \cdot c_P}{2K}}$$

Periodic temperature change at the Earth's surface

- For a sinusoidal temperature perturbation at the surface, the temperature variation for large depths tends to zero.
- For a depth of $L=(2K/\omega\rho c_P)^{1/2}$, the temperature perturbation has an amplitude of 1/e of the amplitude at the surface. L is called the skin depth.
- For a daily temperature variation L=17 cm, for an annual temperature variation L=3.3 m, and for time periods of the order of 10^5 years L>1 km.

Problem diversion

1. Calculate the angular frequency for a sinusoidal temperature perturbation of one year.

2. Show that for a depth of $z=(2K/\omega\rho c_P)^{1/2}$, the temperature perturbation has an amplitude of 1/e of the amplitude at the surface.

3. Show that for an annual temperature variation the skin depth is 3.3 m.

Problem diversion

1. Calculate the angular frequency for a sinusoidal temperature perturbation of one year.

2. Show that for a depth of $z=(2K/\omega\rho c_P)^{1/2}$, the temperature perturbation has an amplitude of 1/e of the amplitude at the surface.

3. Show that for an annual temperature variation the skin depth is 3.3 m.

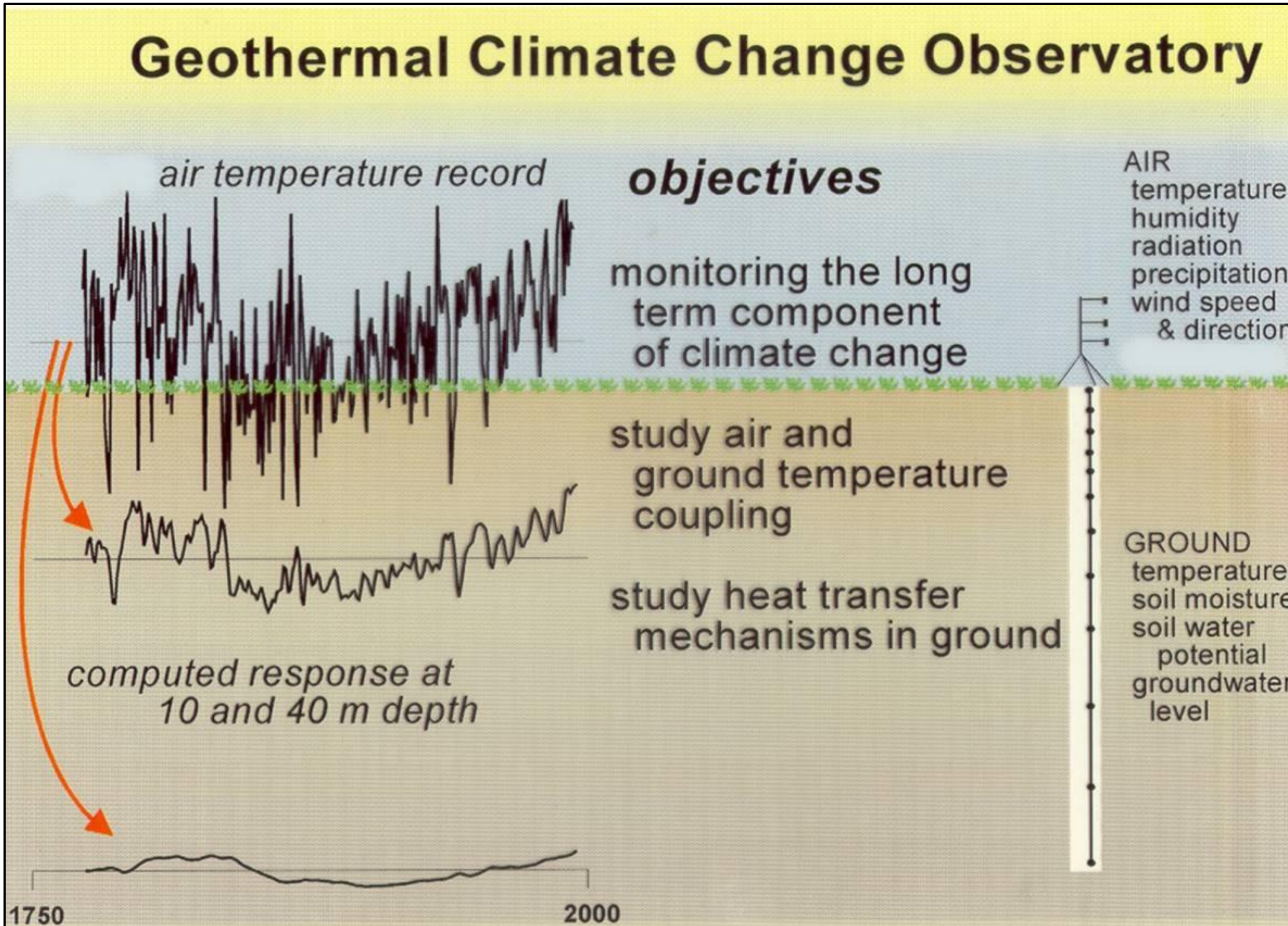
Take $K=2.5$ W/mK, $c_p=103$ J/kg.^{o}C and $\rho=2.3 \times 10^3$ kg/m^3 (approximate values for sandstone).
$\omega = 2 \cdot \pi \cdot f = 1.9 \times 10^{-7}$ s^{-1}

Geothermal Climate Change Observatory in the TGQC-1 well (borehole 190 m deep)

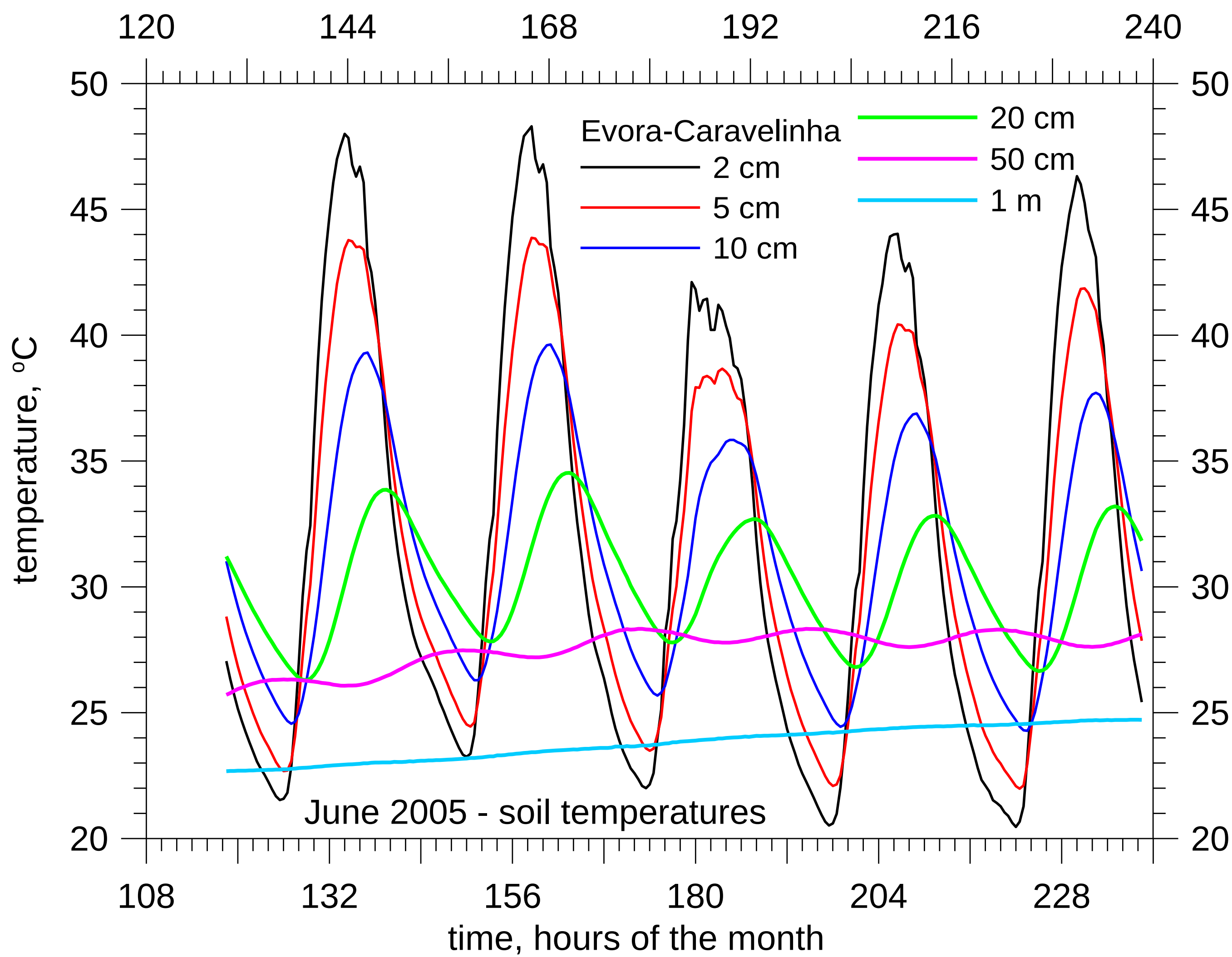
Depth (m) of sensors in the borehole:

0.02 0.05 0.10
0.20 0.50 1.0
2.0 5.0 10.0
20.0 30.0 e 40.0

Geothermal paleoclimatology

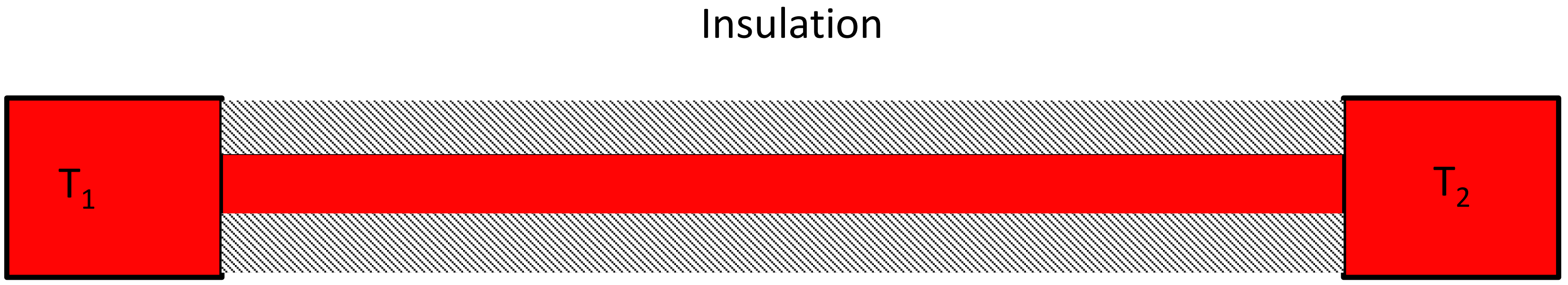


Periodic temperature change at the Earth's surface



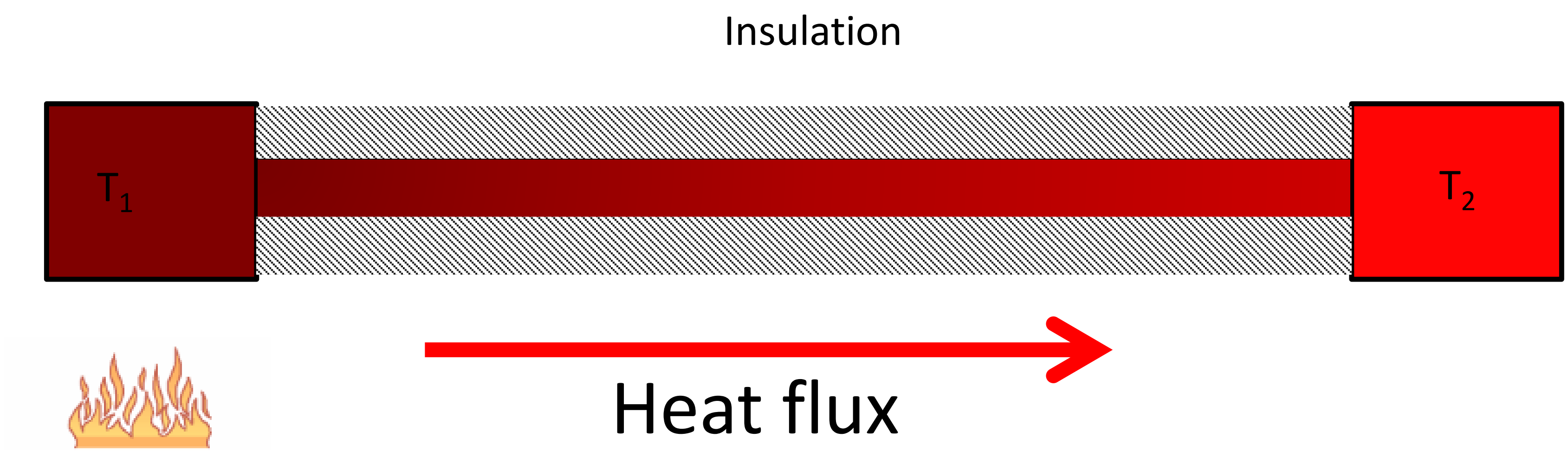
Heat conduction

$$T_1 = T_2$$



Heat conduction

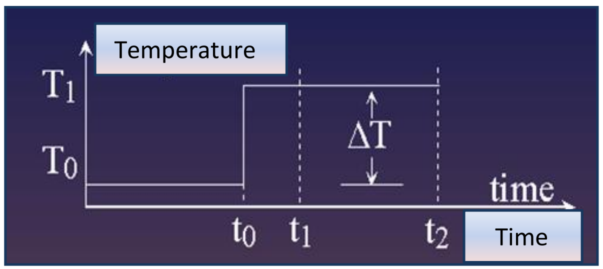
$$T_1 > T_2$$

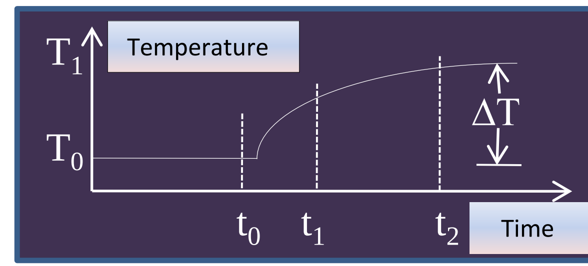


Heat conduction

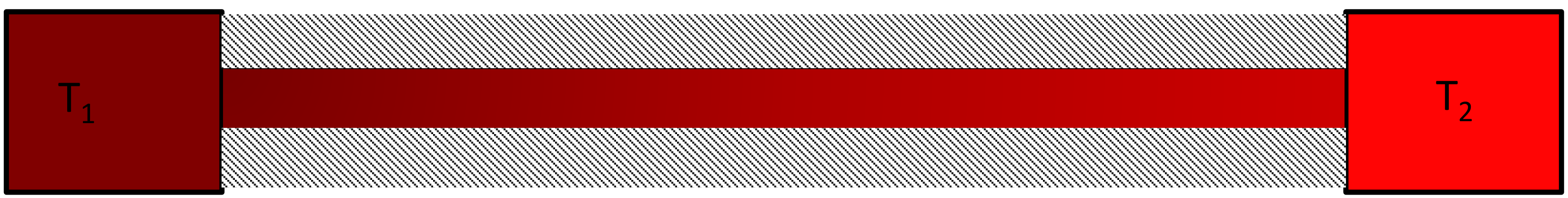
Temperature jump of ΔT

For very long times temperature tends to T_1





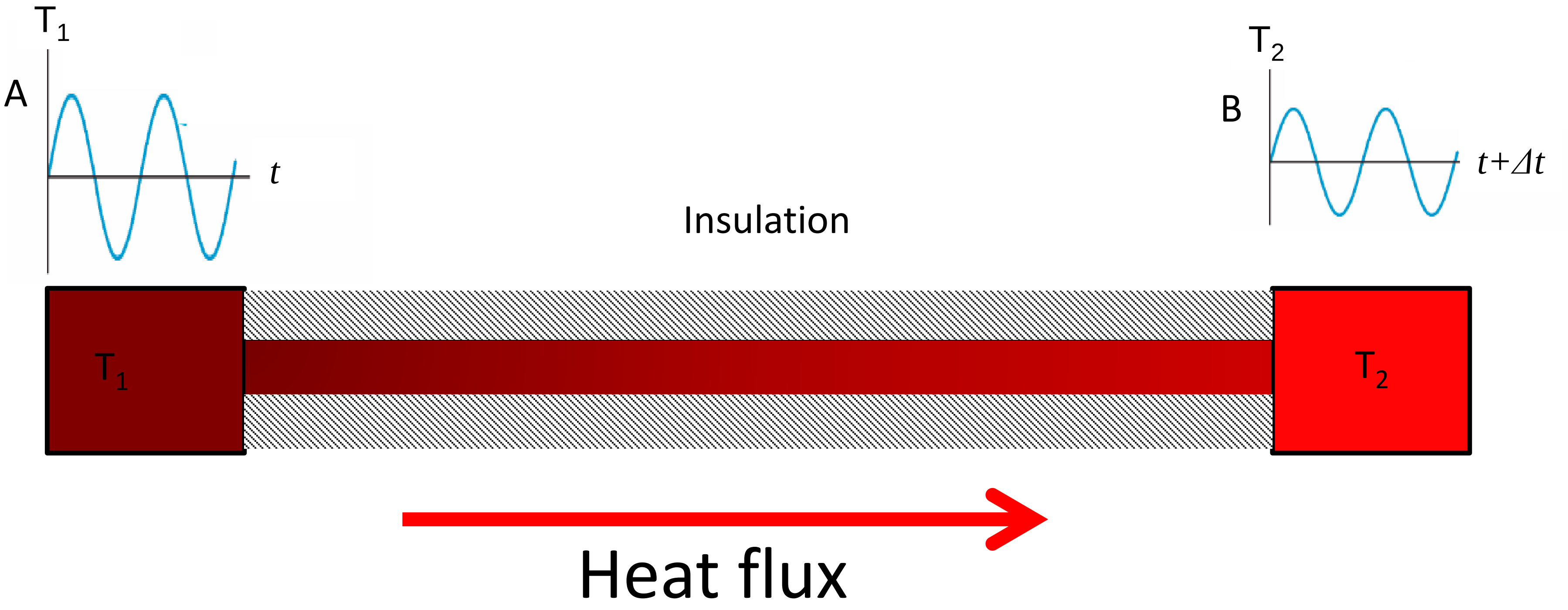
Insulation



Heat flux

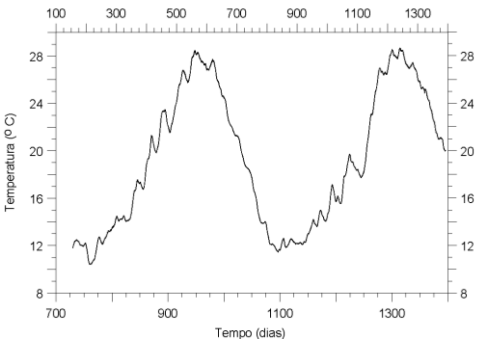
Heat conduction

$$T_1 = A \cdot sen(\omega t_1) > T_2 = B \cdot sen(\omega t_2)$$

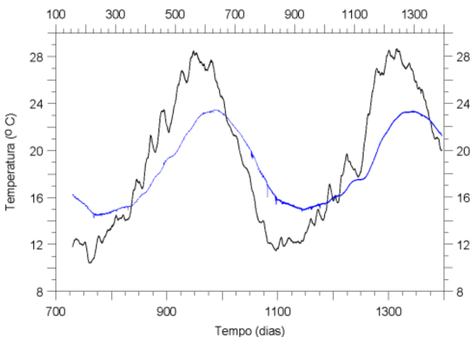


Heat conduction

Surface



Distance$_1$

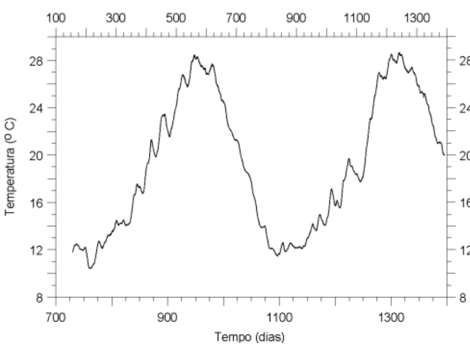


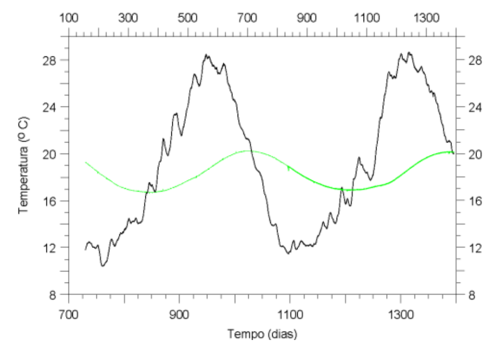
Insulation

Heat conduction

Surface

$Distance_2$



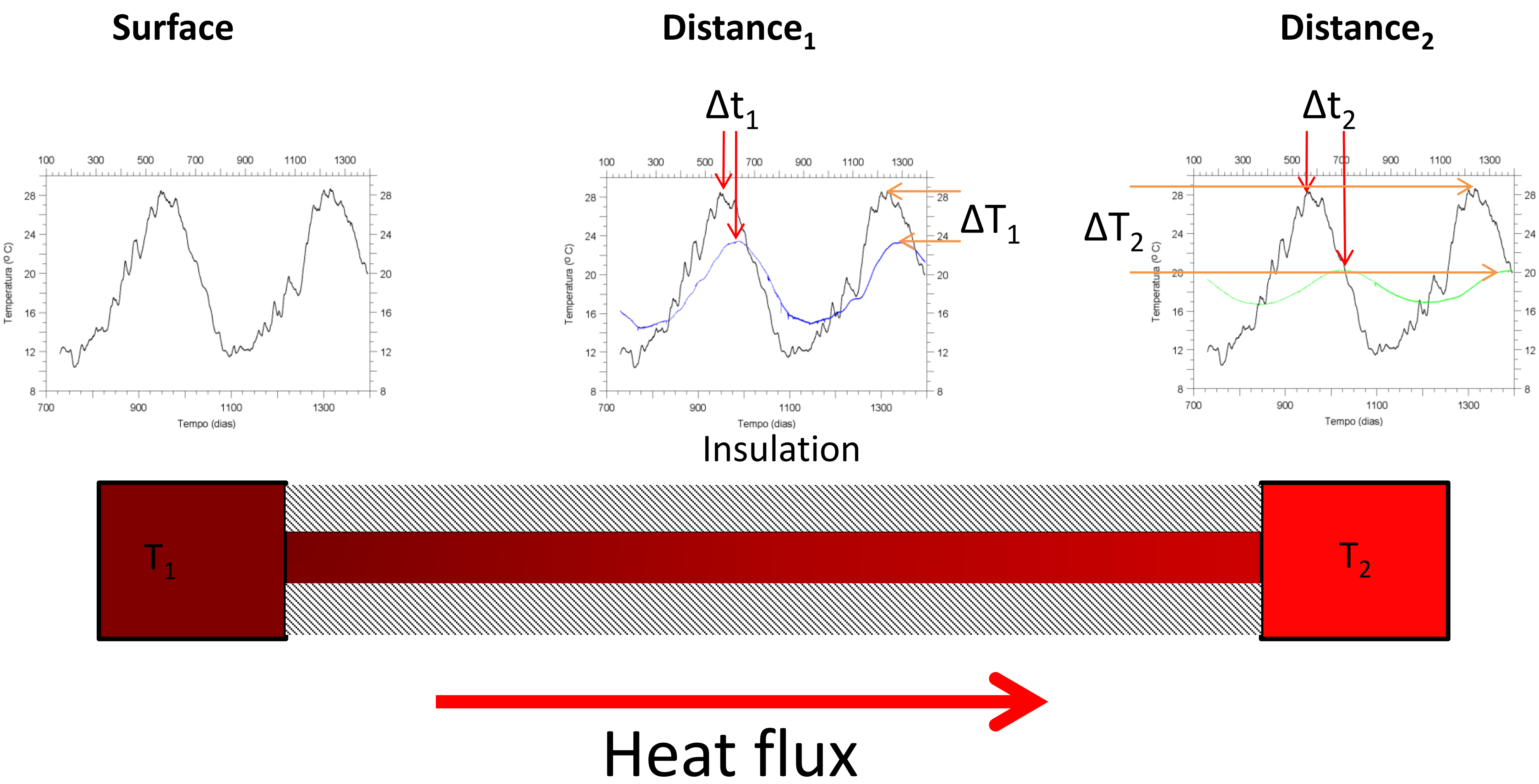


Insulation

Heat flux

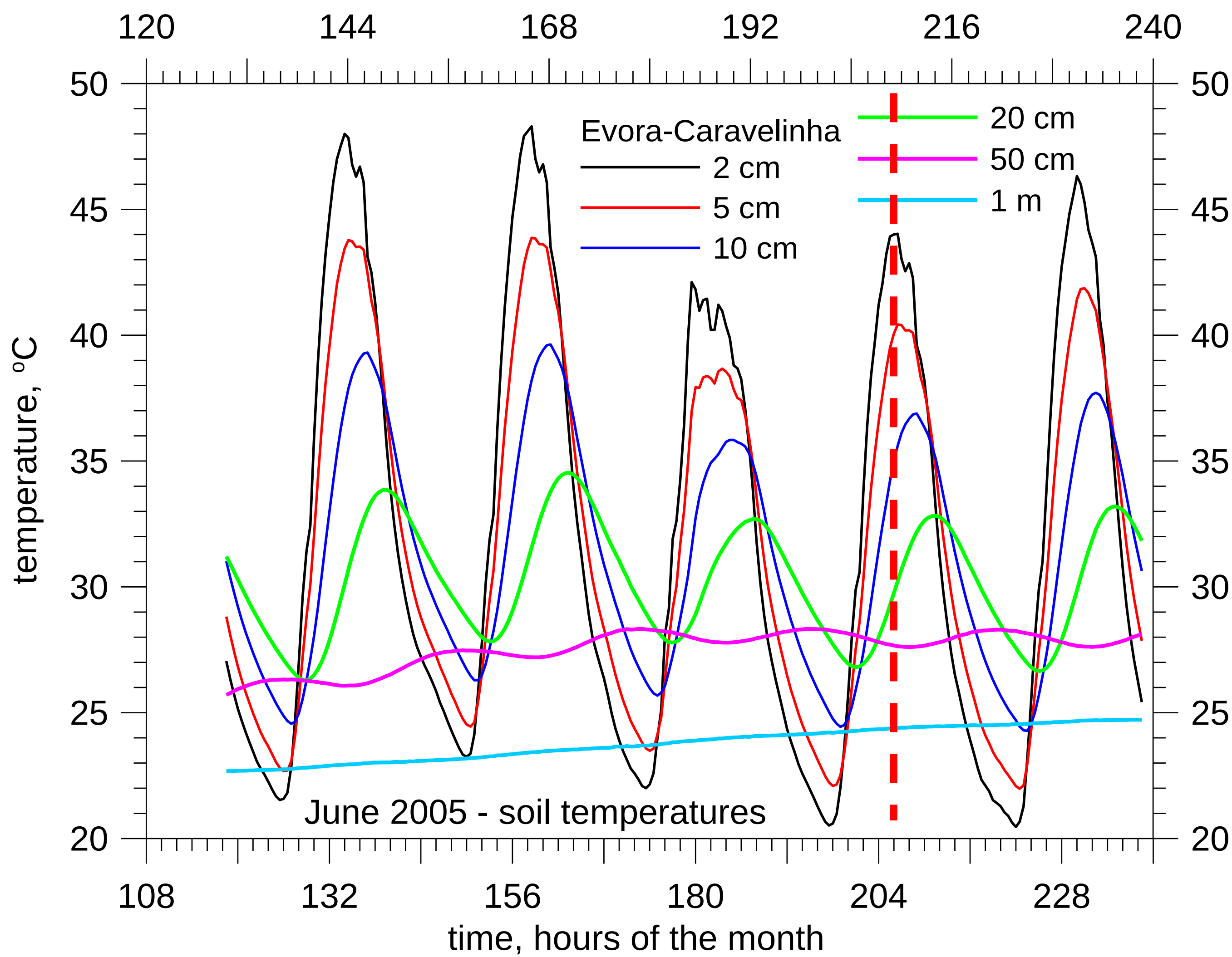
$Distance_1 < Distance_2$

Heat conduction



$\Delta t_2 > \Delta t_1$ e $\Delta T_2 > \Delta T_1$

Periodic temperature change at the Earth's surface



Some obvious conclusions

- High frequency components are attenuated
- Signal decreases in amplitude with distance (depth)
- The time lag increases with distance (depth)
- The signal's wavelength increases with distance (depth)
- As distance (depth) increases resolution decreases

Temperature change: cooling phenomena in the Earth

$$\frac{\partial \mathrm{T}}{\partial \mathrm{t}} = \alpha \cdot \frac{\partial^2 T}{\partial z^2}$$

$$\alpha = \frac{\mathrm{K}}{\rho \cdot c_p}$$

Semi-infinite solid with surface at z=0 and initial temperature $T=T_0$

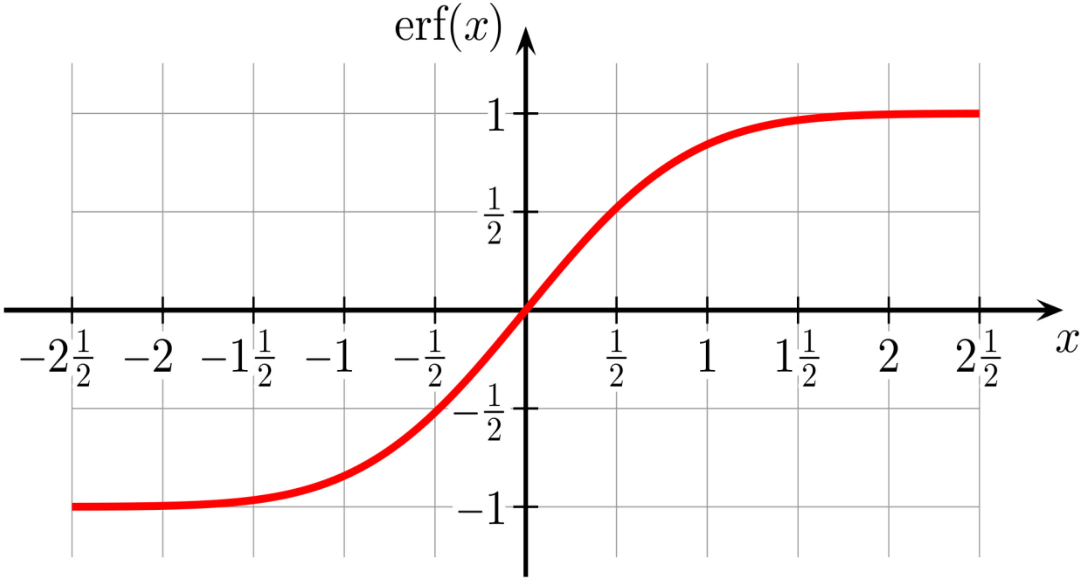
Solution for t>o with surface temperature T=0

$$\mathrm{T} = \mathrm{T}_0 \cdot \mathrm{erf}\left(\frac{z}{2\sqrt{\alpha \cdot t}}\right)$$

$$\mathrm{erf}(x) = \frac{2}{\sqrt{\pi}} \int_0^x e^{-y^2} dy$$

Temperature change: cooling phenomena in the Earth

$$\mathrm{erf}\left(x\right)=\frac{2}{\sqrt{\pi}}\int_0^x e^{-y^2}\,dy$$



Temperature change: cooling phenomena in the Earth

Differentiating in oder to z

$$T = T_0 \cdot \mathrm{erf}\left(\frac{z}{2\sqrt{\alpha \cdot t}}\right)$$

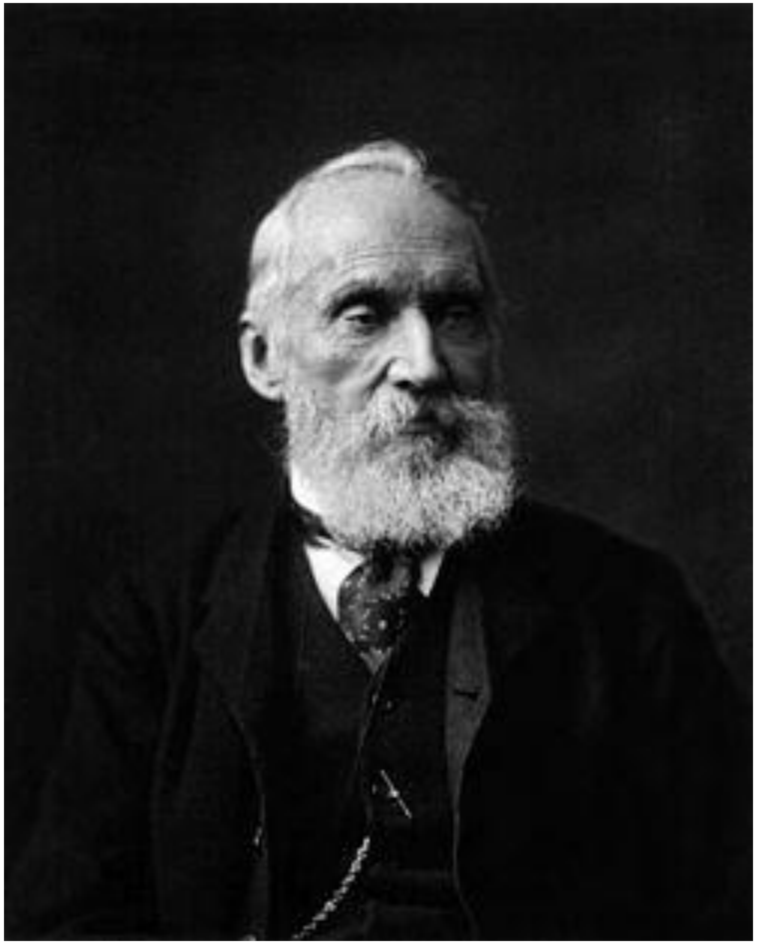
it comes

$$\frac{\partial T}{\partial z} = \frac{\partial}{\partial z}\left(T_0 \cdot \mathrm{erf}\left(\frac{z}{2\sqrt{\alpha \cdot t}}\right)\right) = \frac{T_0}{\sqrt{\pi \cdot \alpha \cdot t}} \cdot e^{-z^2/4\alpha t}$$

which for z=0 gives

$$\left(\frac{\partial T}{\partial z}\right)_{z=0} = \frac{T_0}{\sqrt{\pi \cdot \alpha \cdot t}}$$

Lord Kelvin and the age of the Earth



William Thomson

(26 June 1824 – 17 December 1907)

$$t = \frac{T_0^{\,2}}{\left(\frac{dT}{dz}\right)_{z=0}^{2} \cdot \pi \cdot \alpha}$$

$$T_0 = 3871\ ^{o}C \quad (= 7000\ ^{o}F)$$

$$\left(dT/dz\right)_{z=0} = 0{,}036\ ^{o}C\,m^{-1}$$

$$\alpha = 1{,}4 \times 10^{-6}\ m^2\,s^{-1}$$

$$t = 100\ Ma$$

Lord Kelvin and the age of the Earth

Kelvin argued that a positive thermal gradient with depth means that the Earth must be cooling. By assuming the Earth began as a sphere initially at a constant temperature, he was able to calculate when the cooling must have started. Using an average gradient of 36.5°Ckm^{-1} (l°F/50 ft), he calculated the age of the Earth to be 200 Ma if the initial temperature was 5540°C (10,000°F), or only 98 Ma if the initial temperature was 3870°C (7000°F).

Kelvin preferred the latter estimate, but included a sizeable error margin, 20-400 Ma, due to the uncertainty in the value of thermal diffusivity for crustal rocks (Thomson, 1862).

Thermal history of the Earth

The firt attempt to estimate the Earth's age was done by William Thomson (Lord Kelvin)

$$t = \frac{T_0^{\,2}}{\left(\frac{dT}{dz}\right)_{z=0}^{2} \cdot \pi \cdot \alpha}$$

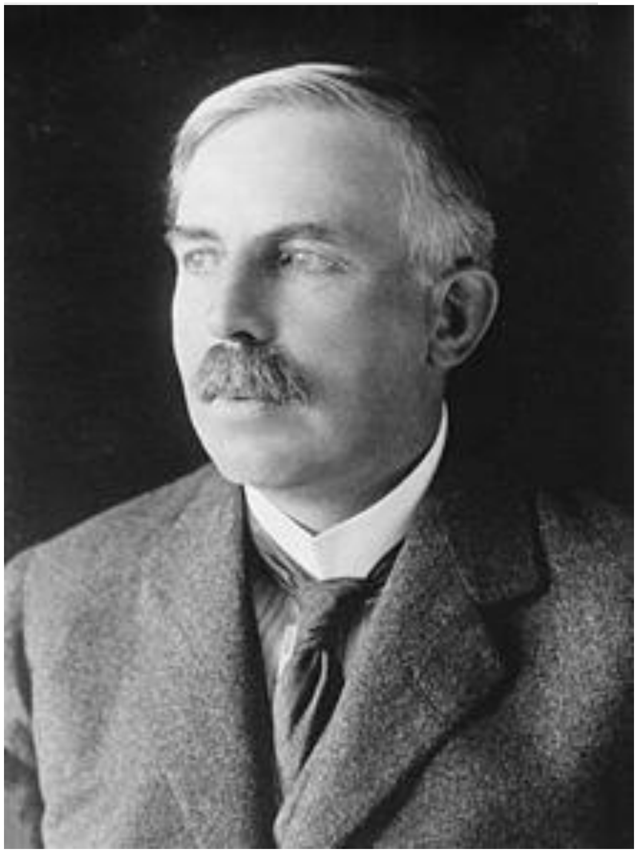
$$T_0 = 3871\ {}^{o}C \quad (= 7000\ {}^{o}F)$$

$$\left(dT/dz\right)_{z=0} = 0{,}036\ {}^{o}C\,m^{-1}$$

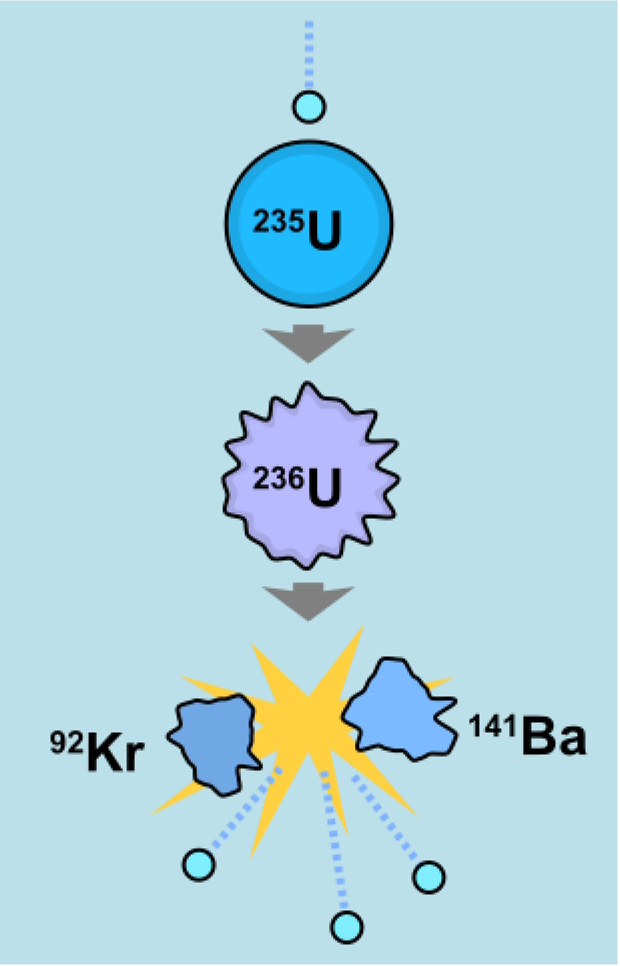
$$\alpha = 1{,}4 \times 10^{-6}\ m^2\,s^{-1}$$

$$t = 100\ Ma$$

Other ages of the Earth



Ernest Rutherford
(30 August 1871 – 19 October 1937)



Other ages of the Earth

- Major modifications had to be made to Kelvin's model after it was recognized that radiogenic heating acted to significantly extend the cooling time.

- Rutherford began the job; serious age estimates were developed after models of the distribution of radioactive isotopes within the Earth suggested by Lubimova (1958) and MacDonald (1959).

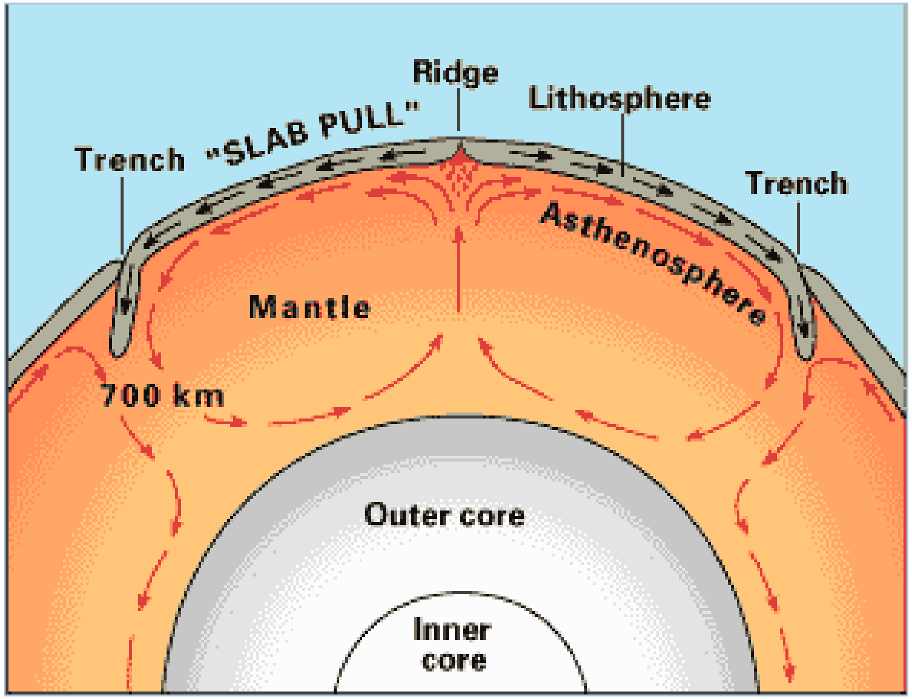
Other ages of the Earth

- Turcotte (1980) attributed 83% of the present surface heat flow to the decay of radioactive isotopes, and only 17% to the cooling of the Earth. He concluded that the mantle is presently cooling at a rate of 36 °C Ga^{-1}, and that three billion years ago it was likely 150 °C hotter than at present.

- The presently accepted age of the Earth is around 4.55 Ga.

Other ages of the Earth

John Perry
(14 February 1850 – 4 August 1920)



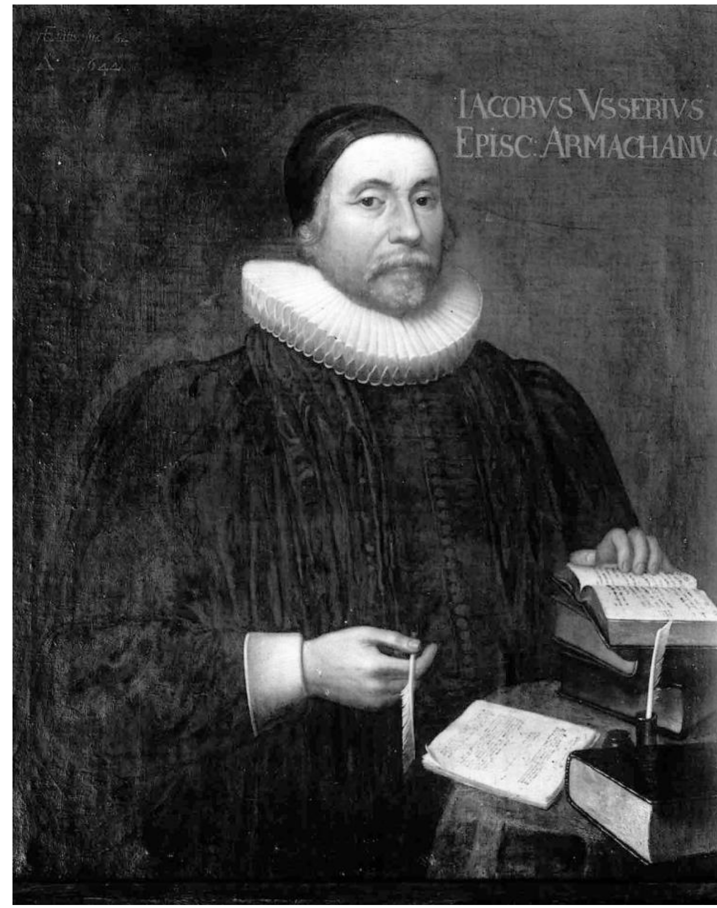
Other ages of the Earth

John Perry, a former Kelvin assistant realized that some kind of convection should be taking place inside the Earth. However, he hesitated to criticize Kelvin. Finally, long after Kelvin made his prediction, Perry went to him with a convection-based idea. Kelvin brushed him off. So, in 1895, Perry published his idea in Nature.

Kelvin' age for the Earth was wrong because radioactivity in Earth's mantle released energy. That makes the gradient steeper and Earth seems younger. However, Kelvin did not use the wrong gradient! He used a completely wrong physical model. We now know that both convection and conduction moves heat through Earth's interior.

Other ages of the Earth

In 1650, the Irish Archbishop James Ussher (1581-1656) announced that according to the Old Testament, the Earth began during the "*beginning of the night of October 22nd of the year 4004 b.C.*"



II
Geothermal gradient

Reminder

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K\,\frac{dT}{dz}$$

q_z is a vector as well as the temperature gradient dT/dz. To calculate (estimate) the heat flux we have to know thermal conductivity K and geothermal gradient dT/dz. Most of the times a q_z vertical is assumed.

Temperature logs

A temperature log is a direct measure of the actual temperature in a borehole. It can be obtained in land or offshore.

- In land, water, mining or oil wells can be used for temperature measurement (with different degrees of accuracy)
- Offshore, temperatures are measured with special probes (violin-bow type probes) inserted by gravity in soft sediments or using boreholes of the late ODP (Ocean Drillig Program) or IODP (Integrated Ocean Drillig Program)

Temperature logs

Temperature logs are used to estimate the geothermal gradient, which, by definition is given by:

$$\nabla T = \text{grad T} = \frac{dT}{dz} = \frac{T_2 - T_1}{\Delta z},$$

where T_1 (shallower) and T_2 (deeper) are the temperatures at two points separated by a distance Δz. It is vector quantity and has magnitude and direction; by convention, it's positive in the direction of increasing temperature. In SI is expressed in K/m or Km^{-1} (kelvin per meter). Practical unit is ^oC/km.

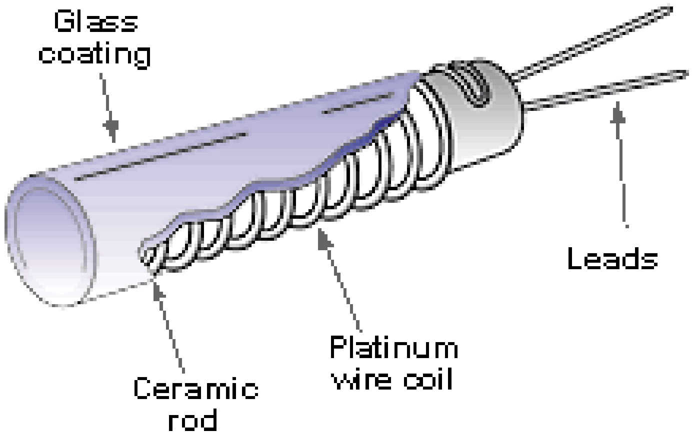
Direct temperature measurements

Precision Temperature Logs

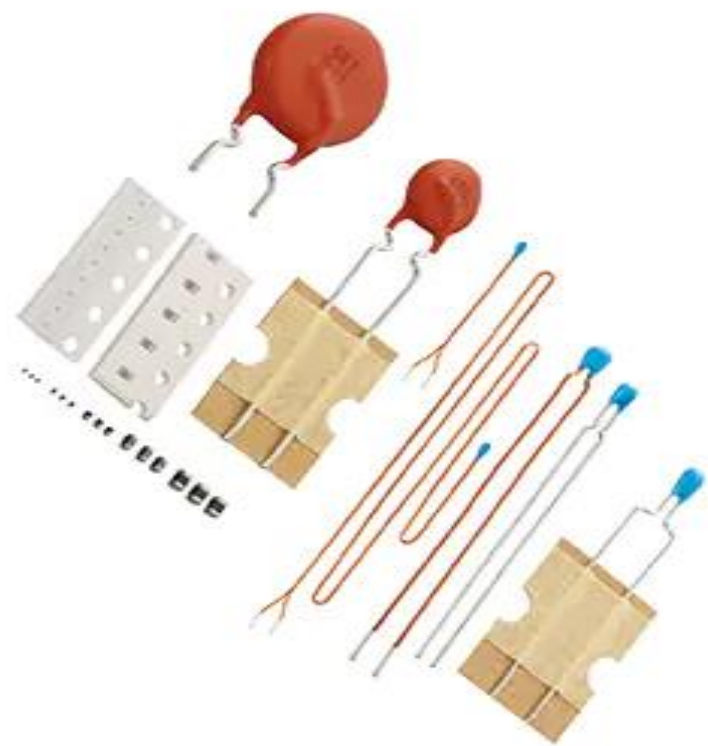
- Thermistors
- Thermo-couples (out dated)
- Platinum resistances
- DTS (Distributed optical fibre Temperature Sensing system)

They can be programmable computer loggers or probes connected to the surface by conductors

Direct temperature measurements

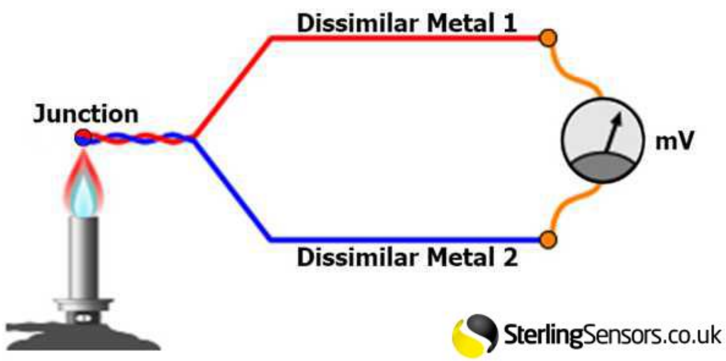


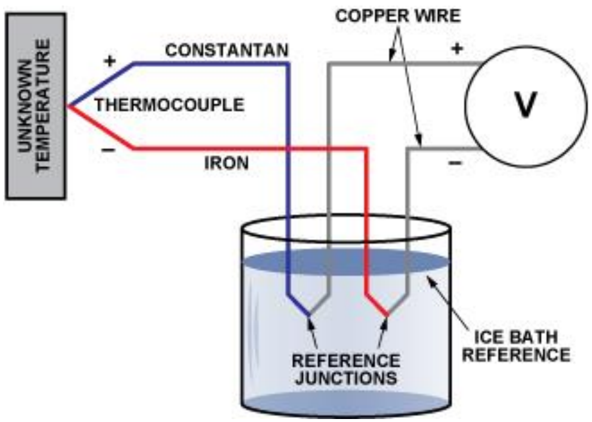
Platinum resistance



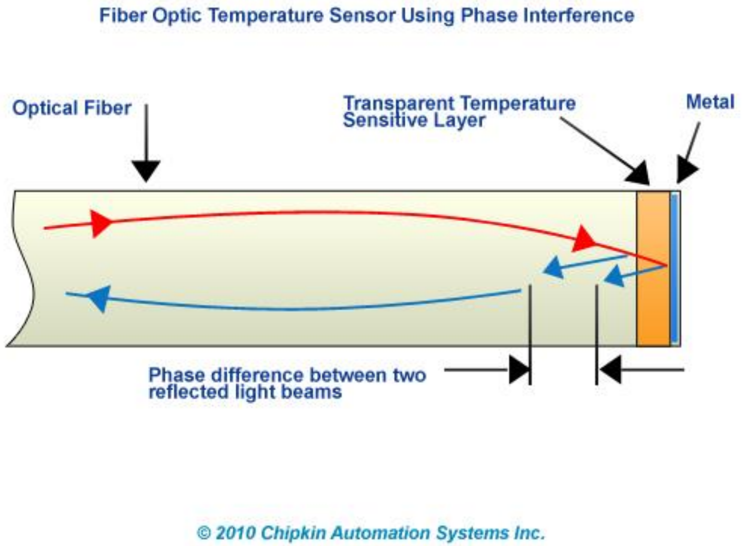
Thermistors

Direct temperature measurements





Thermocouples

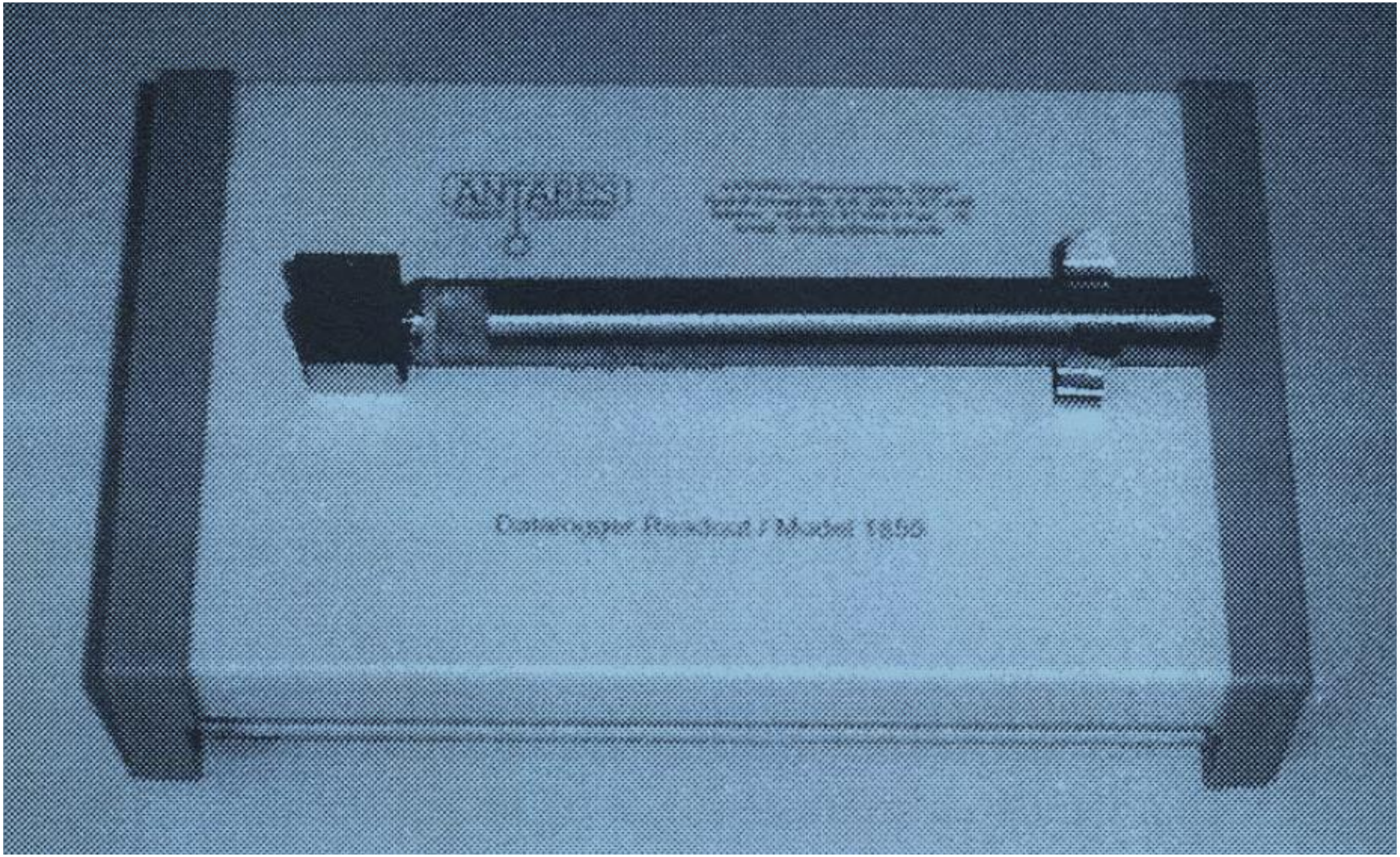


DTS

Direct temperature measurements

ANTARES Datalogger. SN-185 42 50 and protection casing

Direct temperature measurements



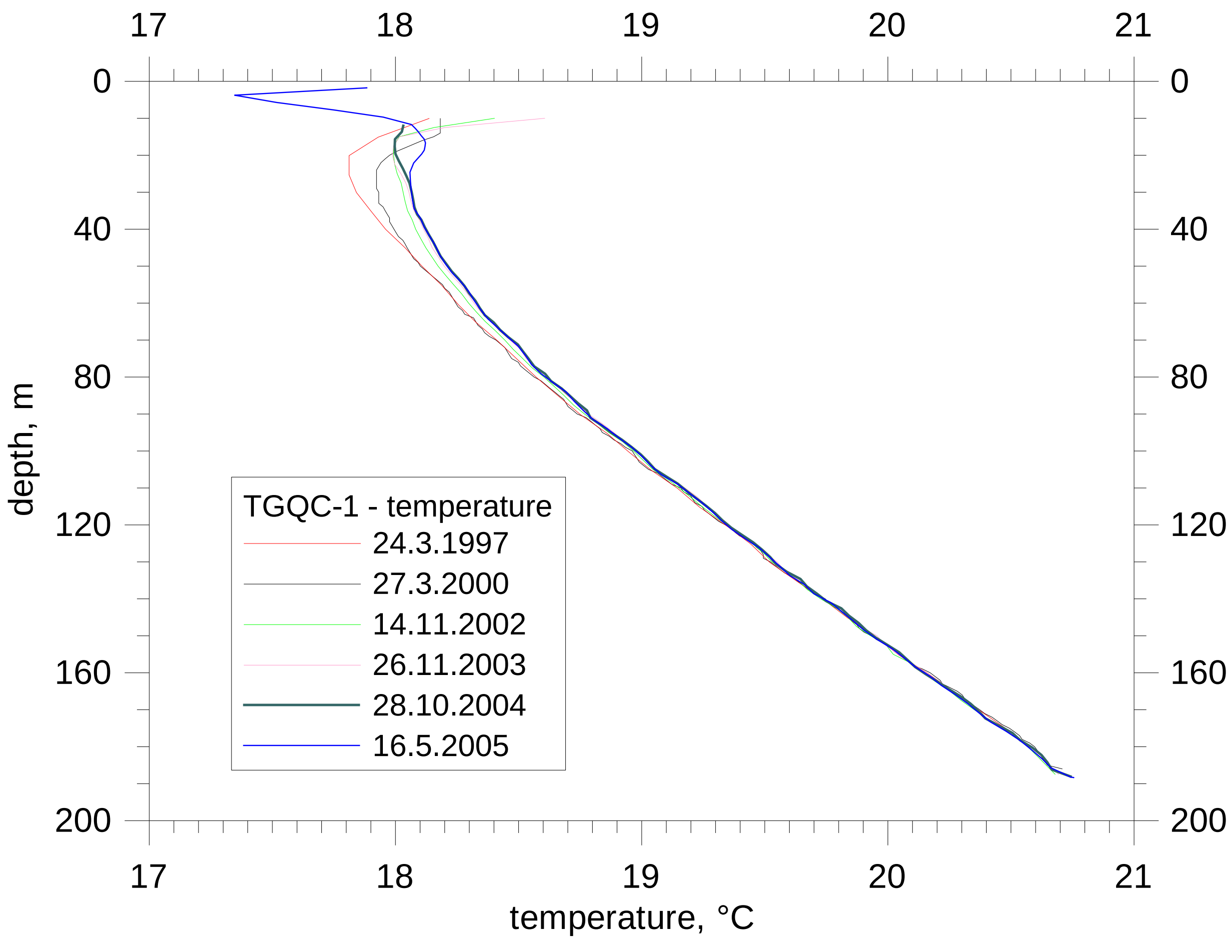
Direct temperature measurements

Geothermal Climate Change Observatory in the TGQC-1 well

Direct temperature measurements

Direct temperature measurements

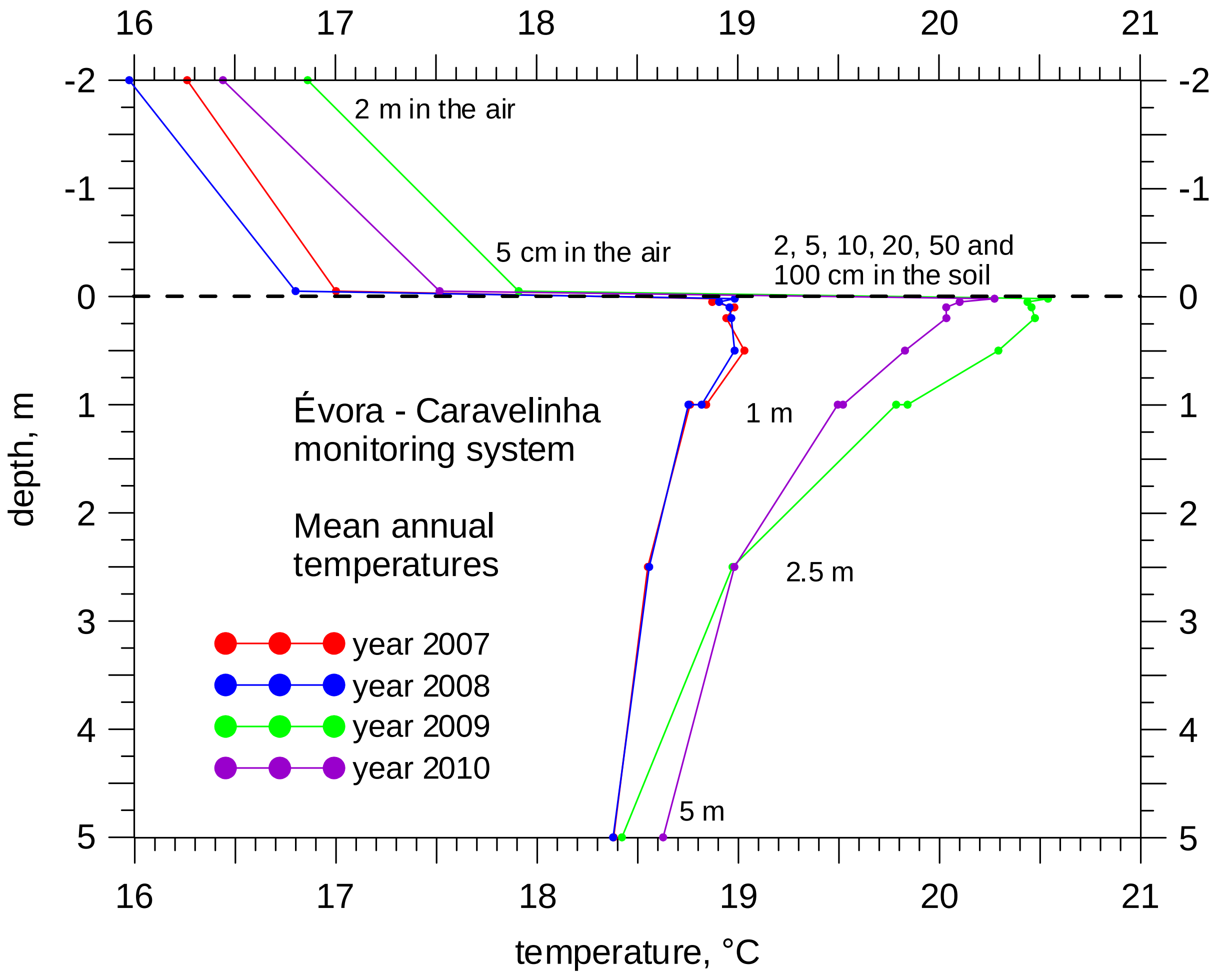
Temperature logs for different years



Geothermal Climate Change Observatory in the TGQC-1 well

Direct temperature measurements

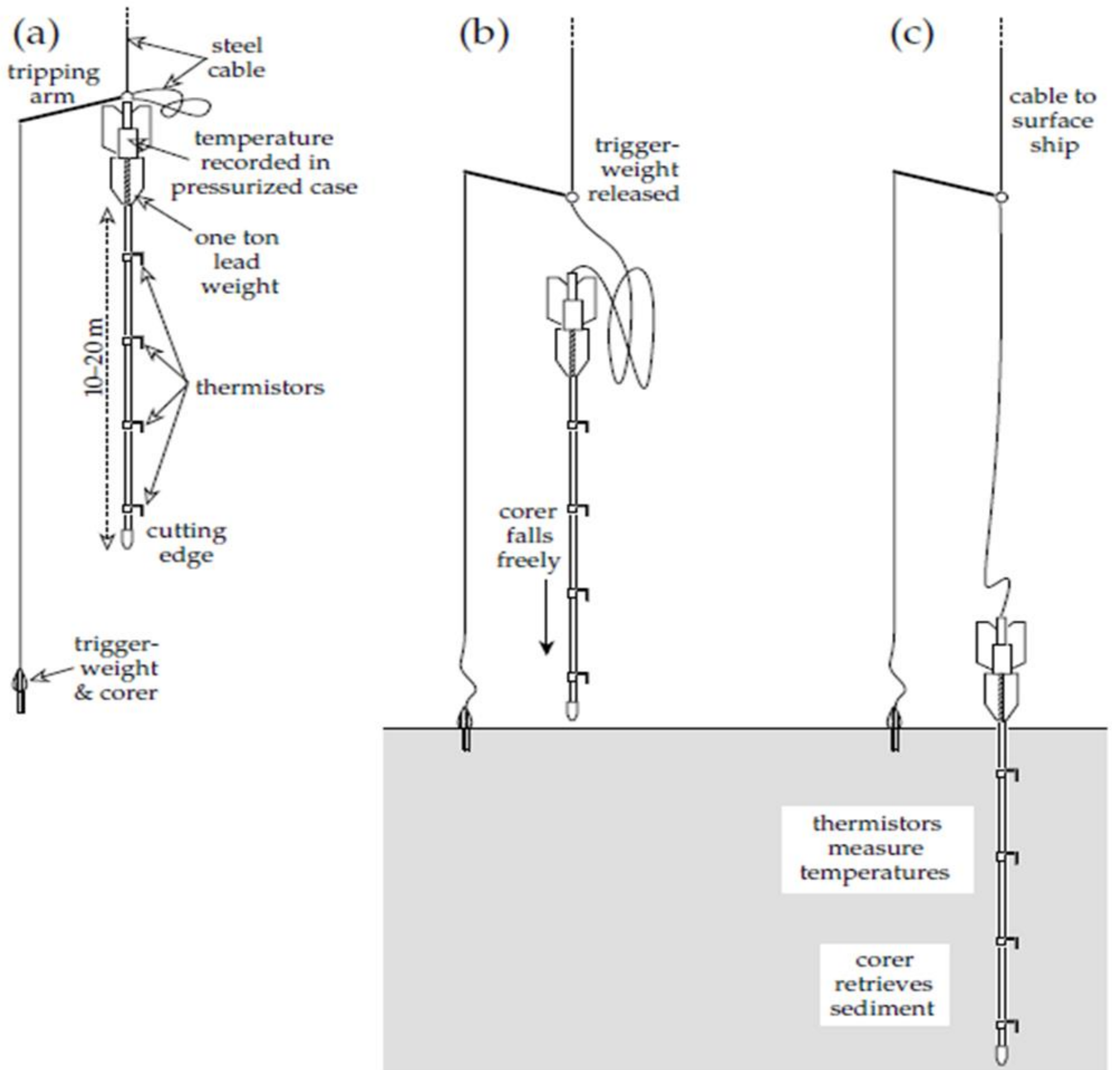
Temperatures at the surface of the ground



Geothermal Climate Change Observatory in the TGQC-1 well

Direct temperature measurements

Deep-Water Probes



Problems with temperature measurements

When mesuring temperatures in boreholes a few problems have to be considered and corrections may have to be applied:

- Borehole convection
- BHT correction

Problems with temperature measurements: borehole convection

A column of fluid under the influence of a thermal gradient may support convection cells which may cause the temperature log within the fluid to differ from that within the surrounding rocks.

Problems with temperature measurements: borehole convection

Critical thermal gradient above which convection start:

$$\frac{\partial \mathrm{T}}{\partial \mathrm{z}} = \frac{\mathrm{C} \cdot \upsilon \cdot \kappa}{\mathrm{g} \cdot \alpha \cdot \mathrm{r}^4} \qquad (^{o}\mathrm{C/m\ or\ K/m})$$

where $\frac{\partial \mathrm{T}}{\partial \mathrm{z}}$ is the critical thermal gradient; g is the acceleration of gravity; α, ν and κ are the thermal expansion coefficient, the kinematic viscosity and the thermal diffusivity of the fluid; r is the borehole radius; C is a constant equal to 2.16×10^{-4} (if SI units are used).

Problems with temperature measurements: borehole convection

For water at 95 ^{o}C

$$\frac{\partial T}{\partial z} = \frac{C \cdot \upsilon \cdot \kappa}{g \cdot \alpha \cdot r^4}$$

reduces to approximately

$$\frac{\partial T}{\partial z} = \frac{1.4 \cdot 10^{-9}}{r^4}$$

What are the units of the constant 1.4×10^{-9} in the second equation?

Diversion about units

$[g] = m \cdot s^{-2}$ $[\alpha] = K^{-1}$ $[\nu] = m^2 \cdot s^{-1}$

$[\kappa]= m^2 \cdot s^{-1}$ $[r] = m$

$$\frac{\partial \mathrm{T}}{\partial \mathrm{z}} = \frac{\mathrm{C} \cdot \upsilon \cdot \kappa}{\mathrm{g} \cdot \alpha \cdot \mathrm{r}^4}$$

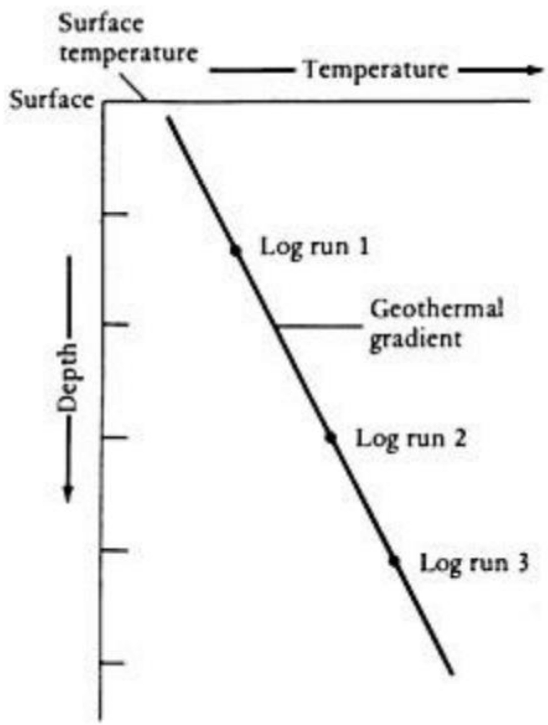
What are the units of constant C = 2.16 x 10^{-4}?

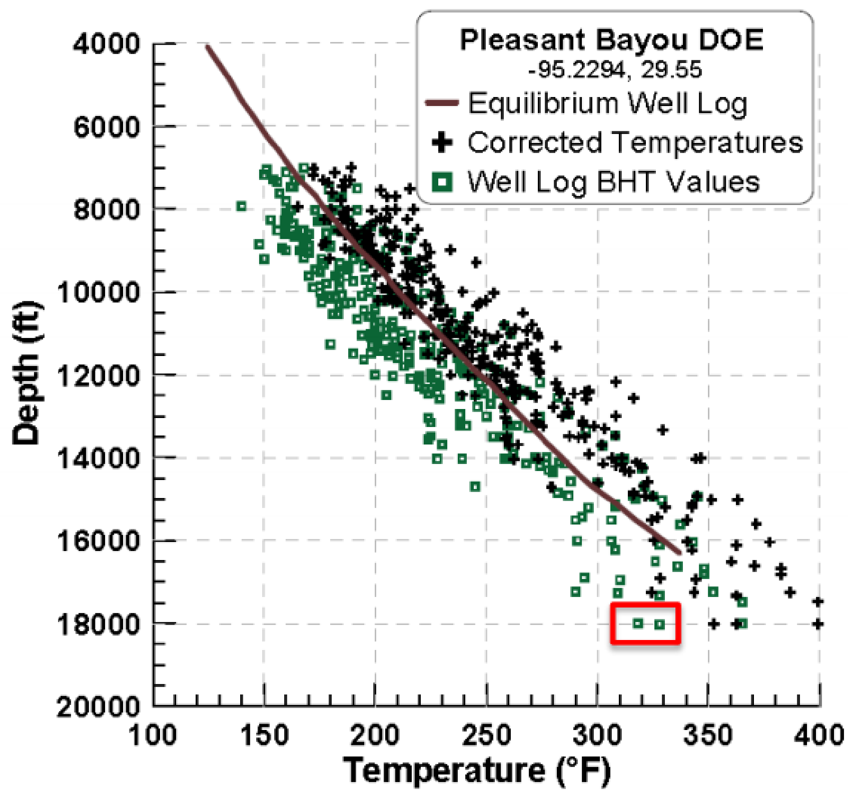
Verify the dimensional homogeneity of the equation

Direct temperature measurements

Bottom-Hole Temperatures (BHT)

They are obtained by attaching maximum thermometers
to probes in different runs in oil prospecting wells





Direct temperature measurements

The Horner plot method for correcting BHTs

1. For ***<u>each depth</u>*** there must exist, at least, 2 BHTs measured at 2 different times.

2. It is necessary to know the time of the mud circulation in the borehole after stopping drilling (t_c).

3. It is necessary to know the time elapsed between the end of the mud circulation and the measurement of temperature at the bottom of the borehole (Δt).

Direct temperature measurements

The Horner plot method for correcting BHTs (cont.)

4. Therefore, for each BHT there must be a different time of measurement (Δt).

5. A graph of BHT vs. ln $(1+t_c/(\Delta t))$ should be drawn.

6. The best fit straight line should be adjusted to the points in the graph.

7. Extrapolation of the straight line to the vertical axis (temperature axis) will give the geological formation temperature before drilling (VRT – Virgin Rock Temperature).

Example:

Borehole: Cormorant 1

Depth of the first runs of logs: 900 m.

Number of log runs: 3

Cormorant 1			
Depth – 900 m	SP/Res/GR run	Neutron run	64' N run
Circulation Time (hours)	2	2	2
BHT (°C)	120	124	127
Time of meas. (Δt) (hours)	6	12	18
$\ln(1+t_c/(\Delta t))$	0.28768	0.15415	0.10536

Cormorant 1			
Depth – 900 m	SP/Res/GR run	Neutron run	64' N run
Circulation Time (hours)	2	2	2
BHT (ºC)	120	124	127
Time of meas. (Δt) (hours)	6	12	18
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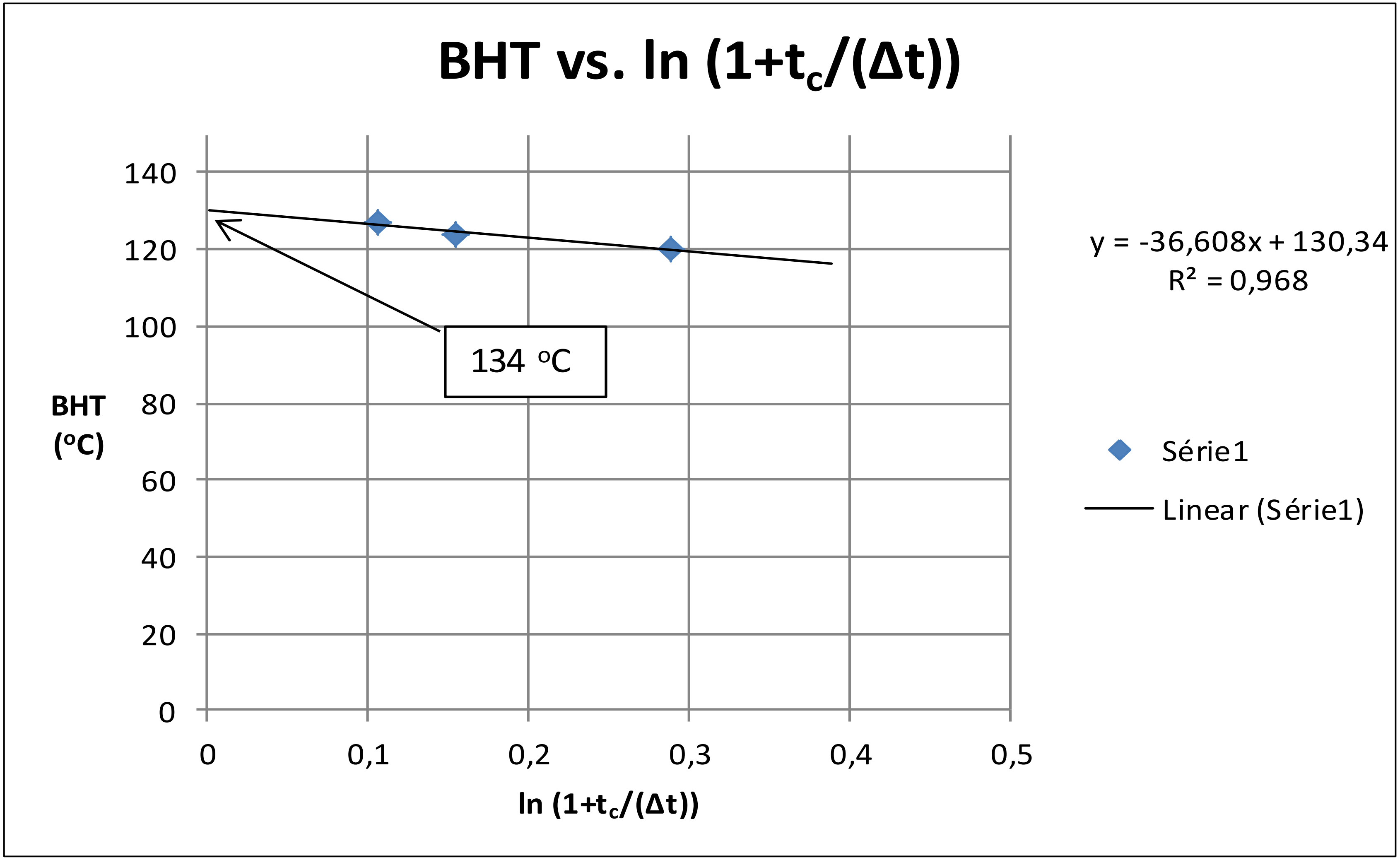


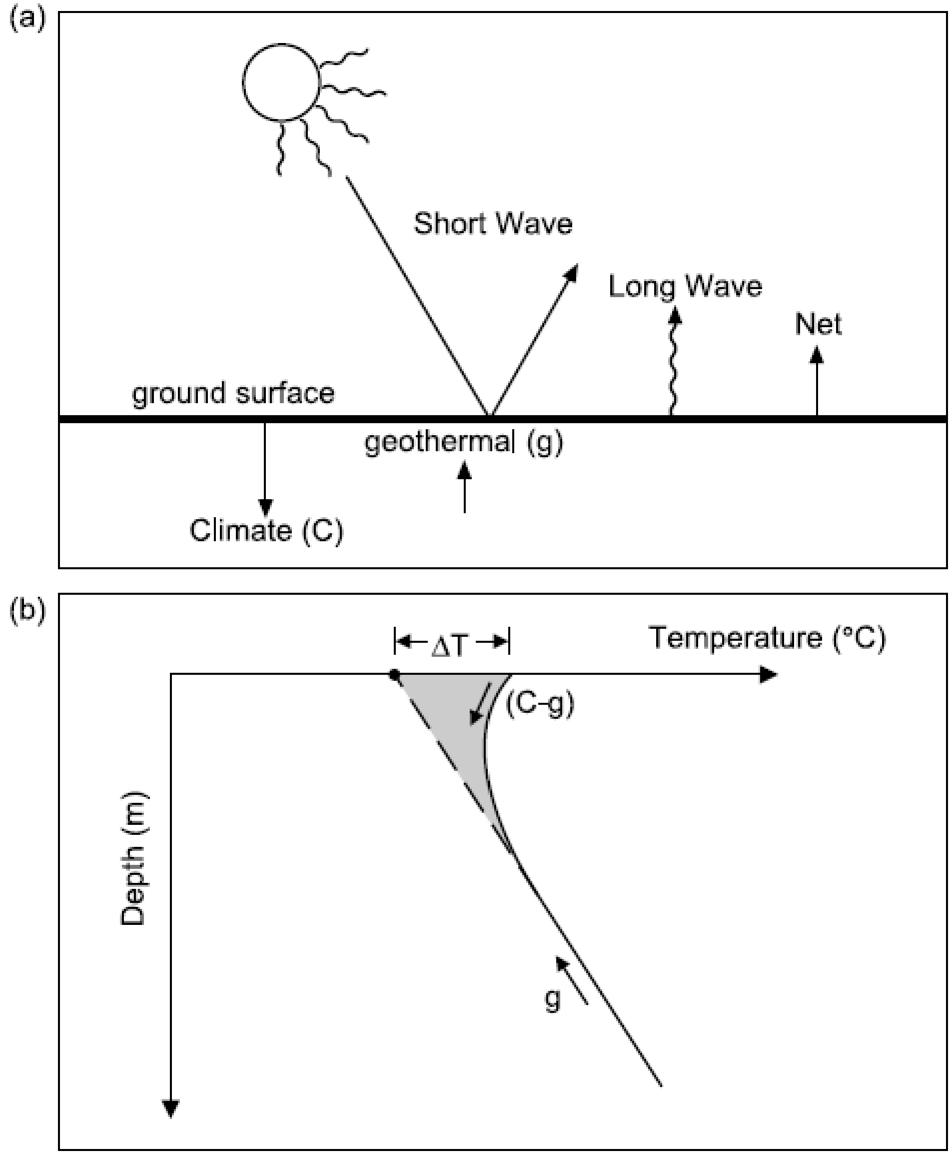
Direct temperature measurements

Bottom-Hole Temperatures

To make experiments with bottom hole temperatures you can go to this site:

http://zetaware.com/utilities/bht/horner.html

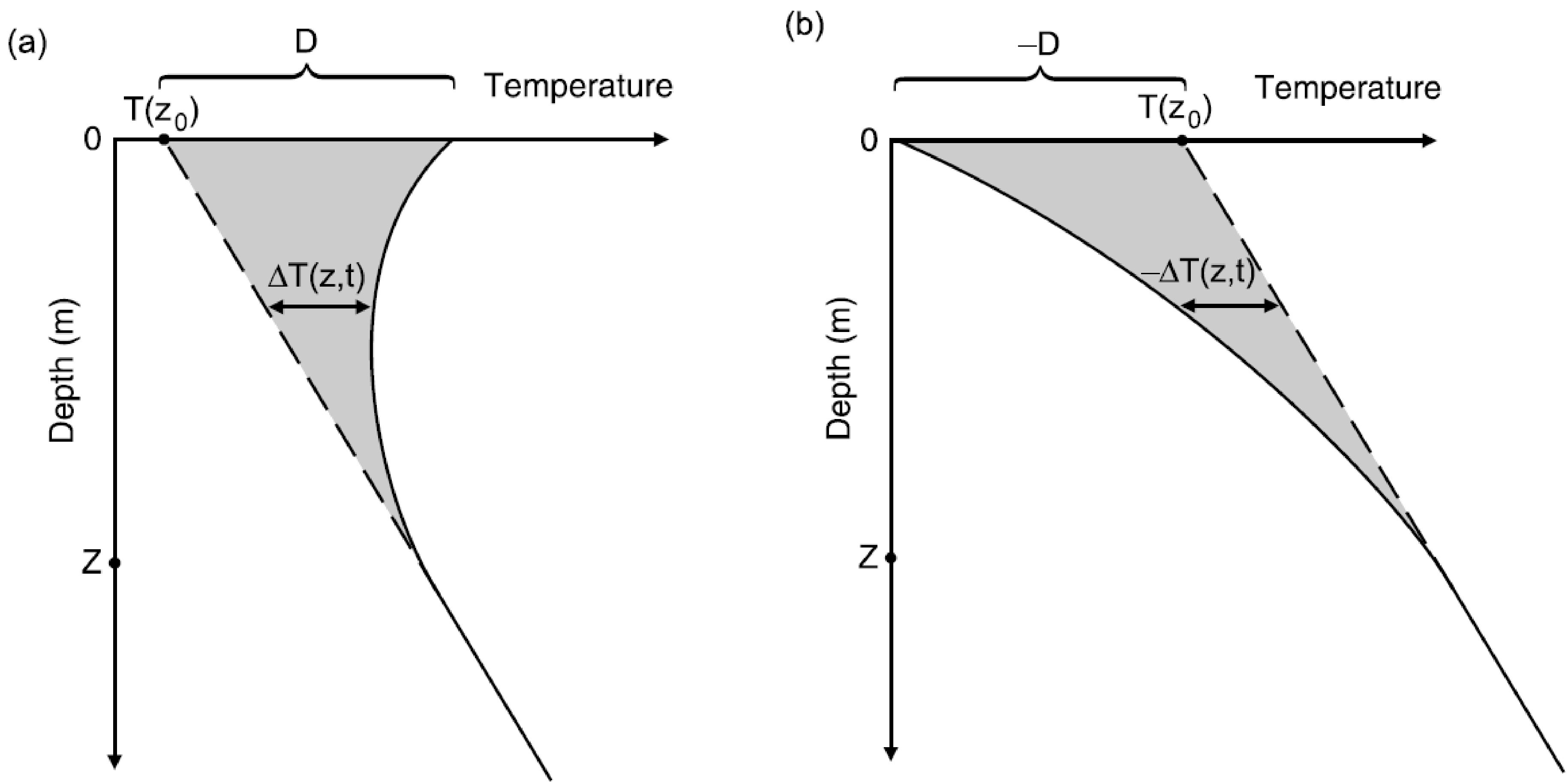
A climate signal in ground temperature



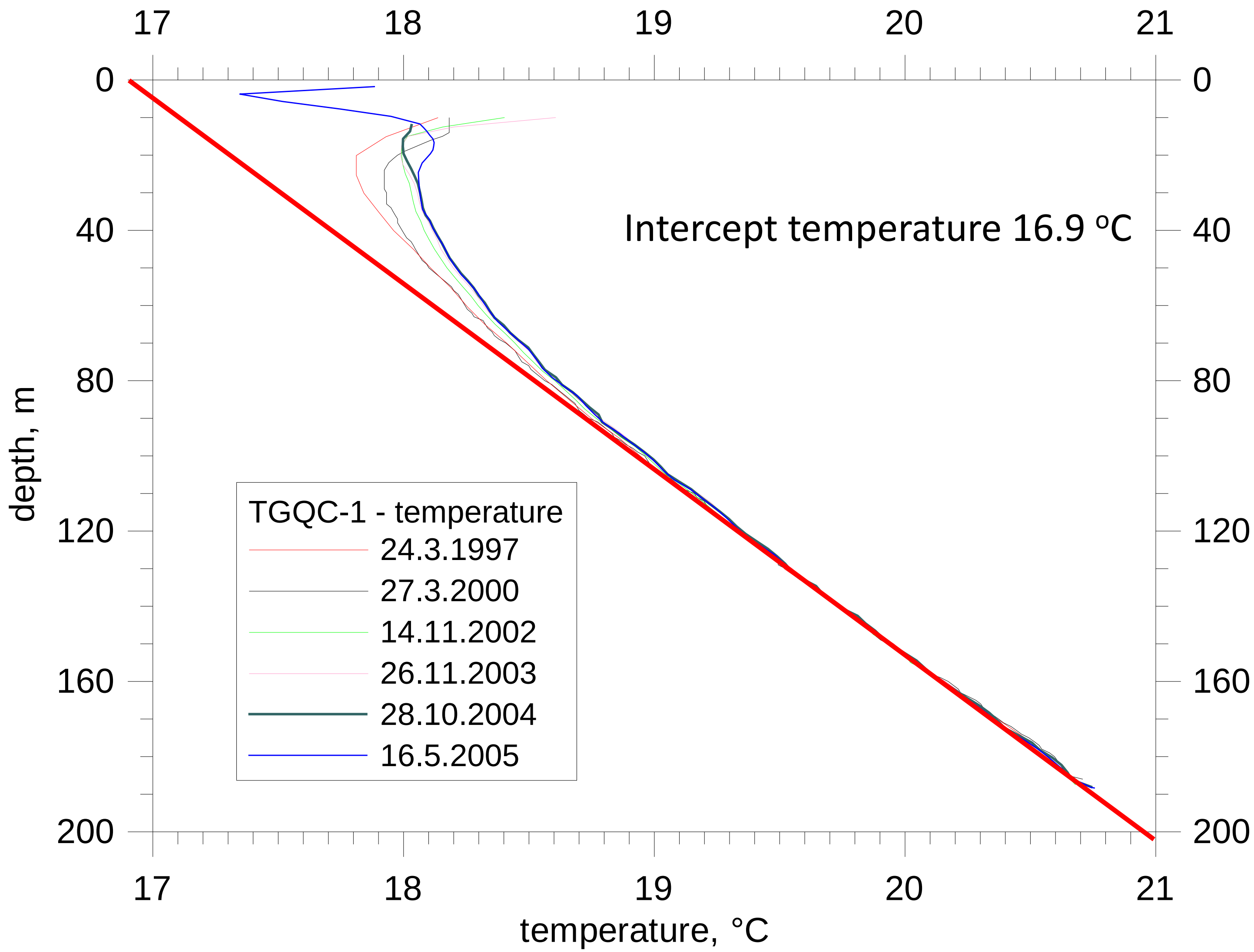
The shape says a lot about climate

Increase of air temperature

Decrease of air temperature



Temperature logs for different years



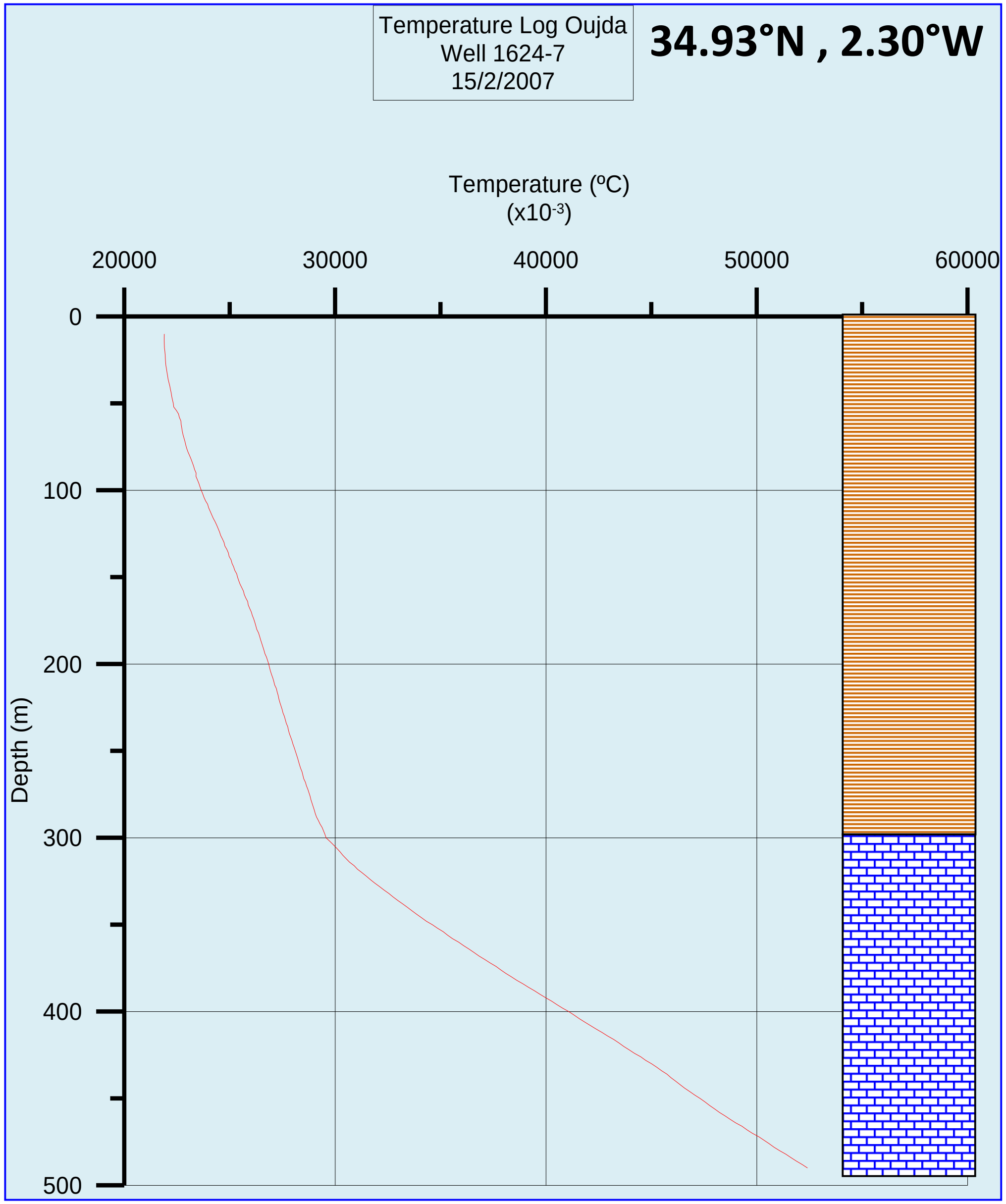
Examples (Morocco)

Examples (Morocco)

For the next examples try:

- To estimate the geothermal gradient
- Give plausible interpretion of the geological, geophysical and hydrodynamical situation

Examples (Morocco)



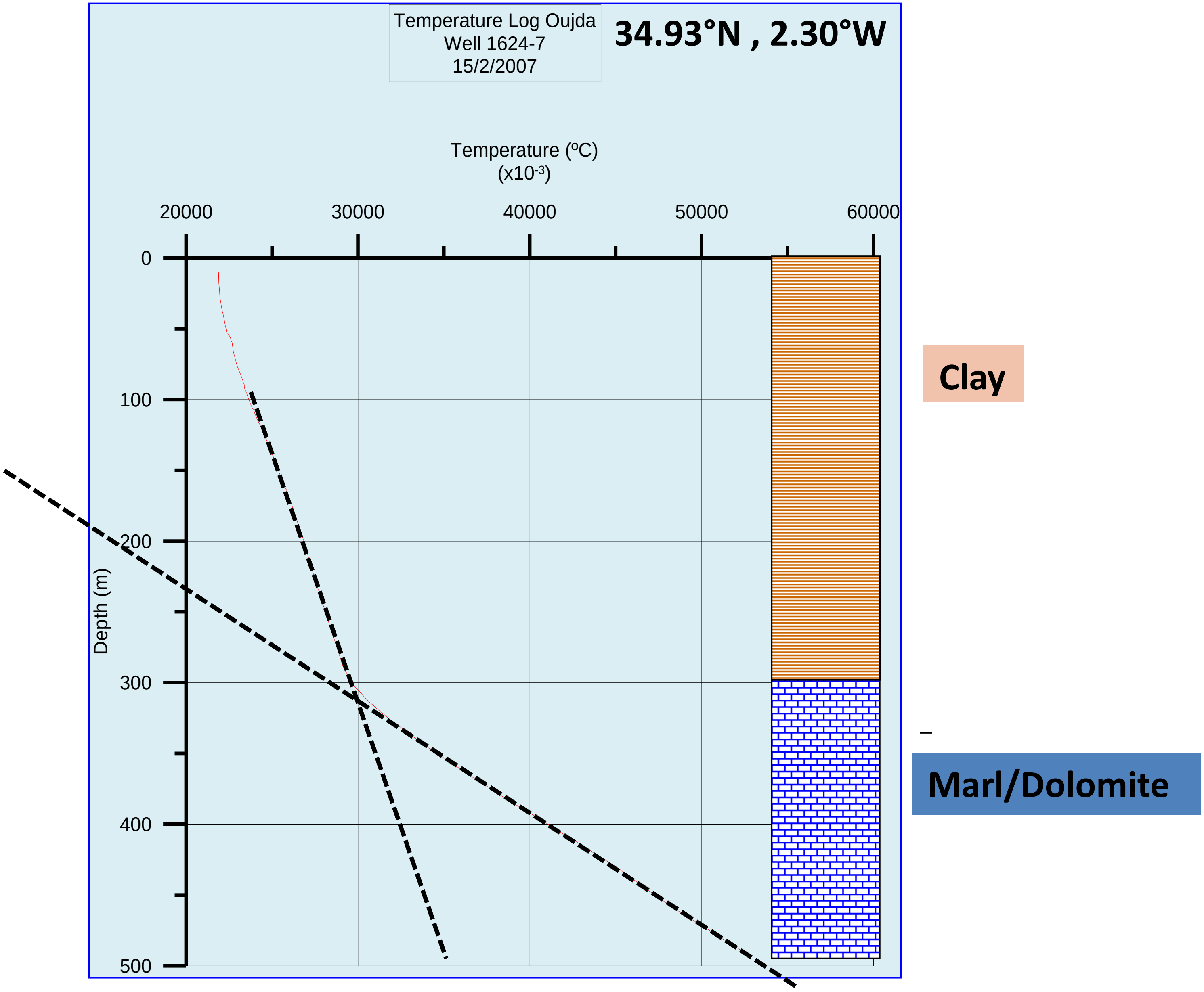
We can see 2 different gradients. What are their values?

What's happening at the surface?

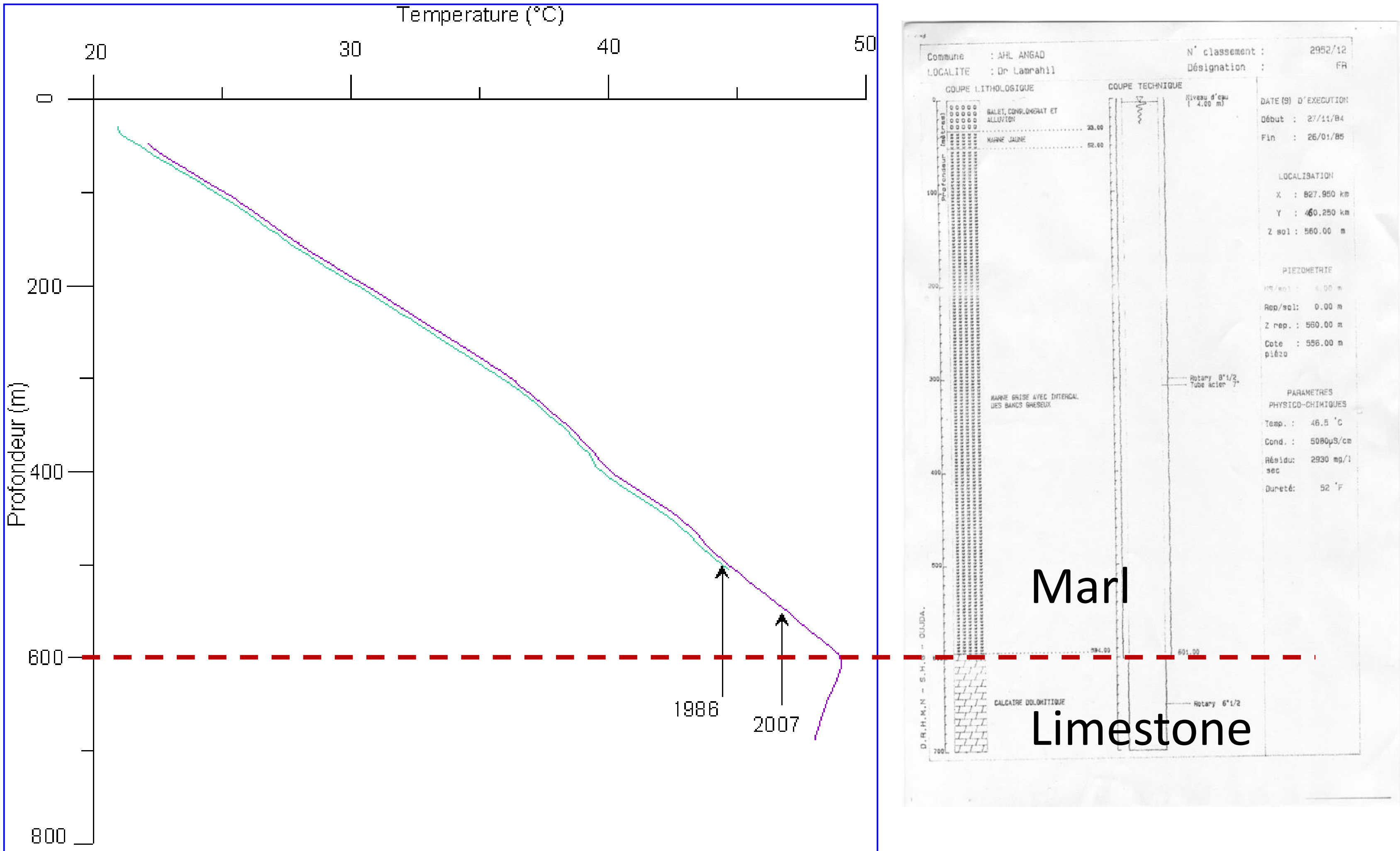
Clay

Marl/Dolomite

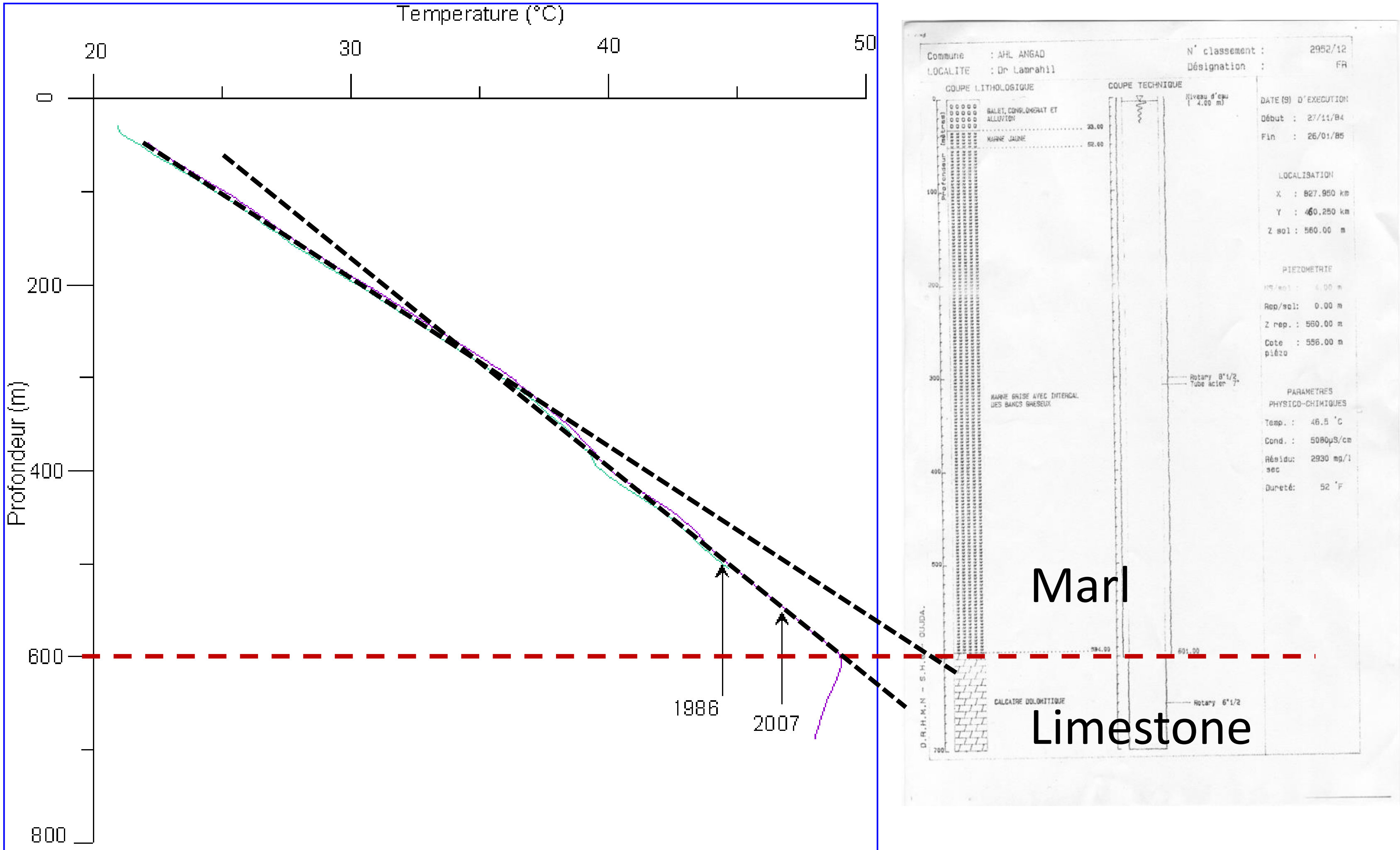
Examples (Morocco)



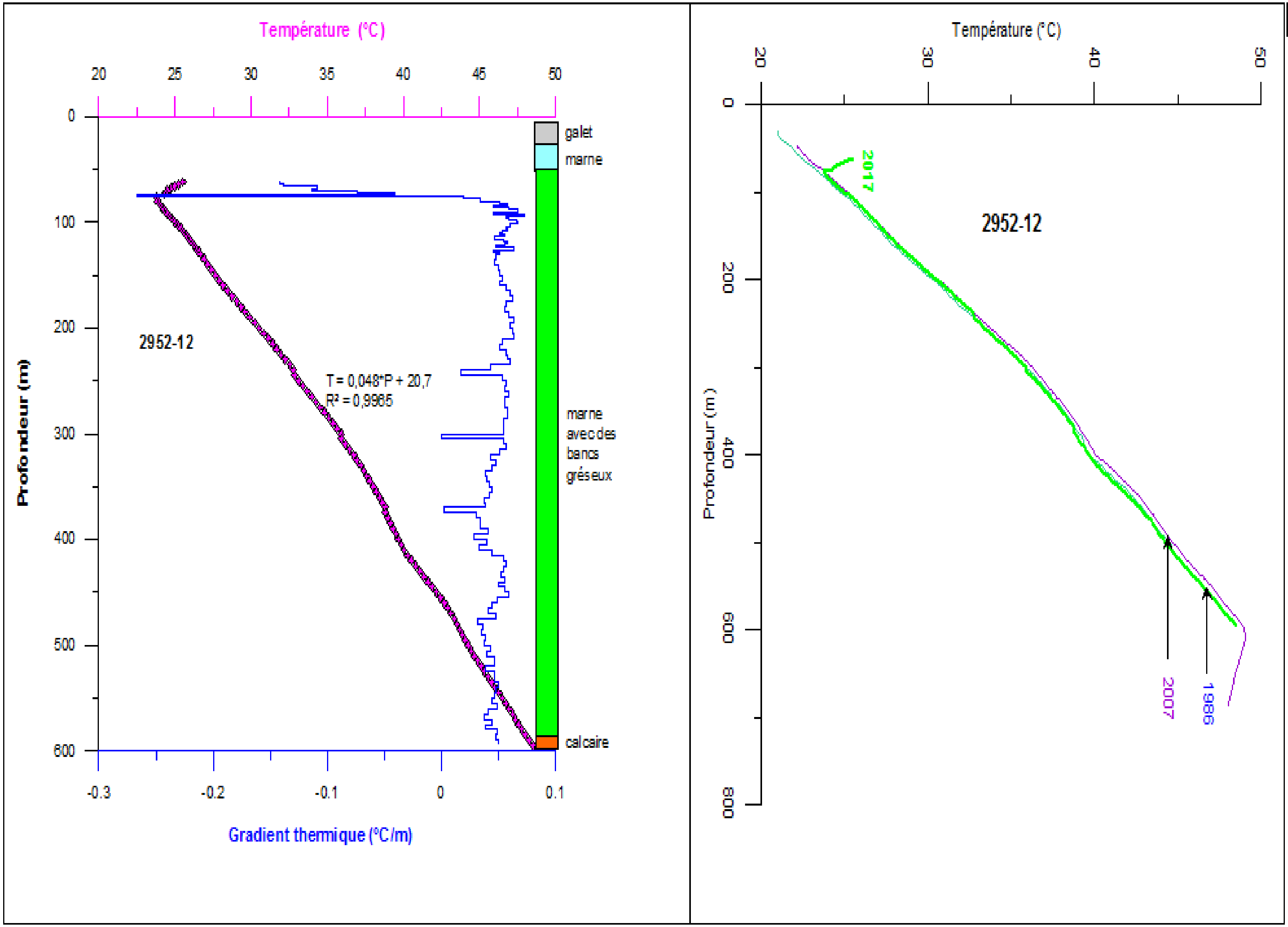
Examples (Morocco)



Examples (Morocco)



Examples (Morocco)



Examples (Morocco)

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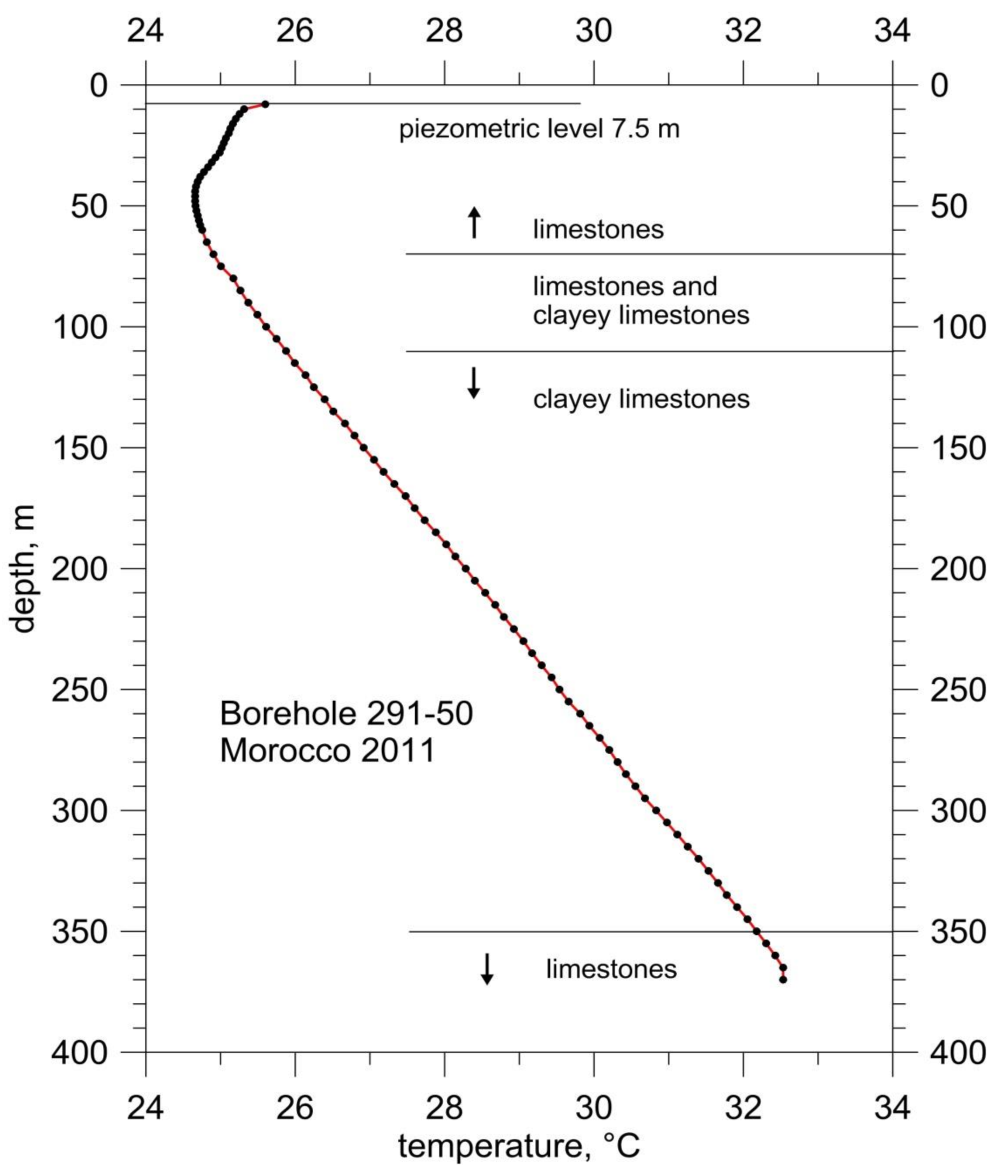
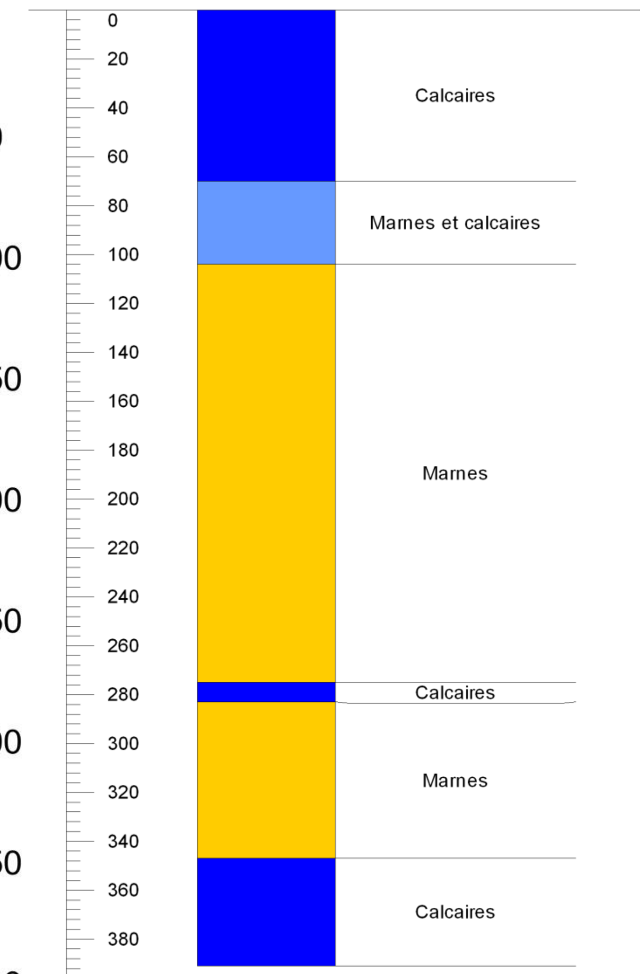
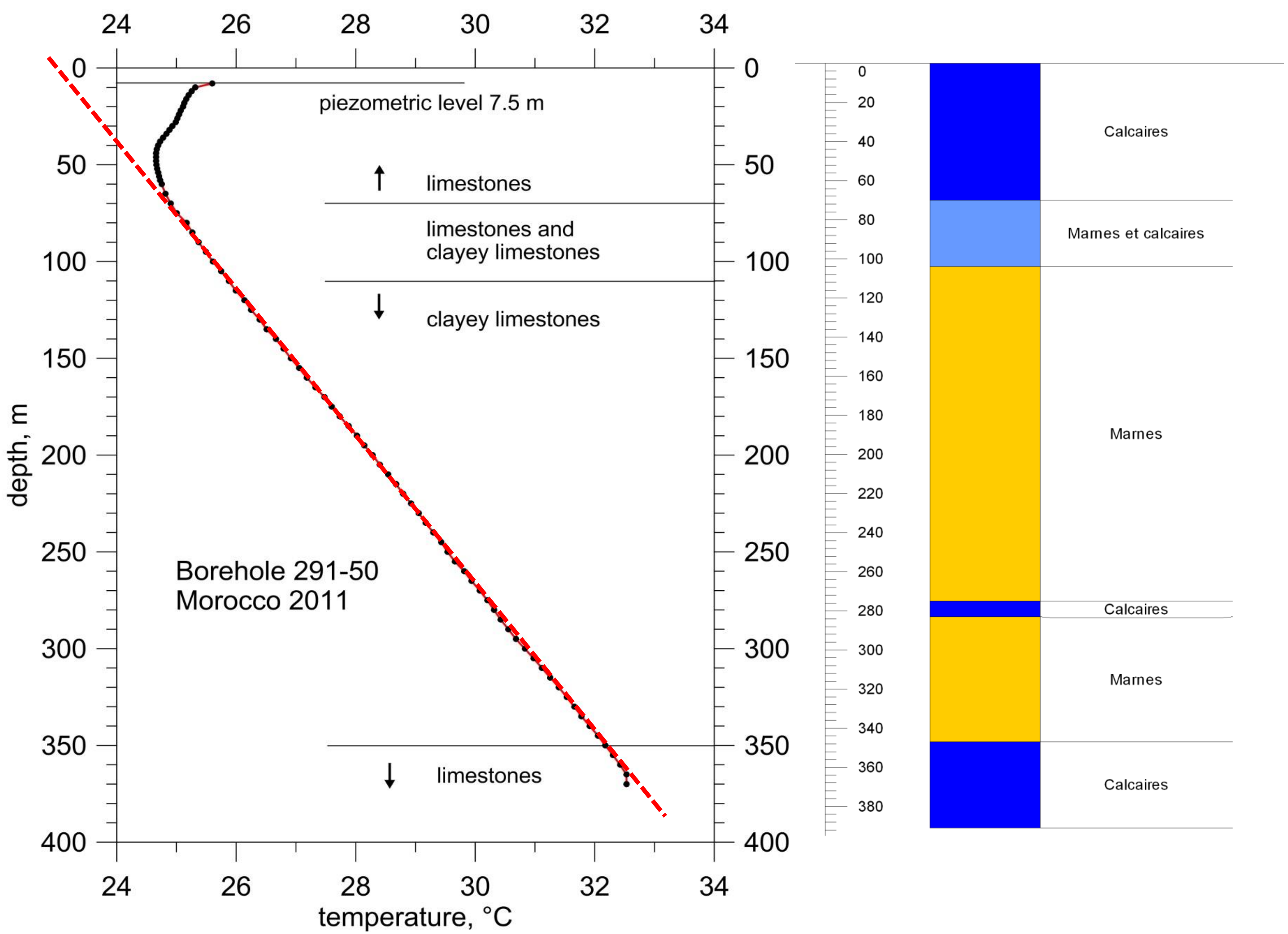
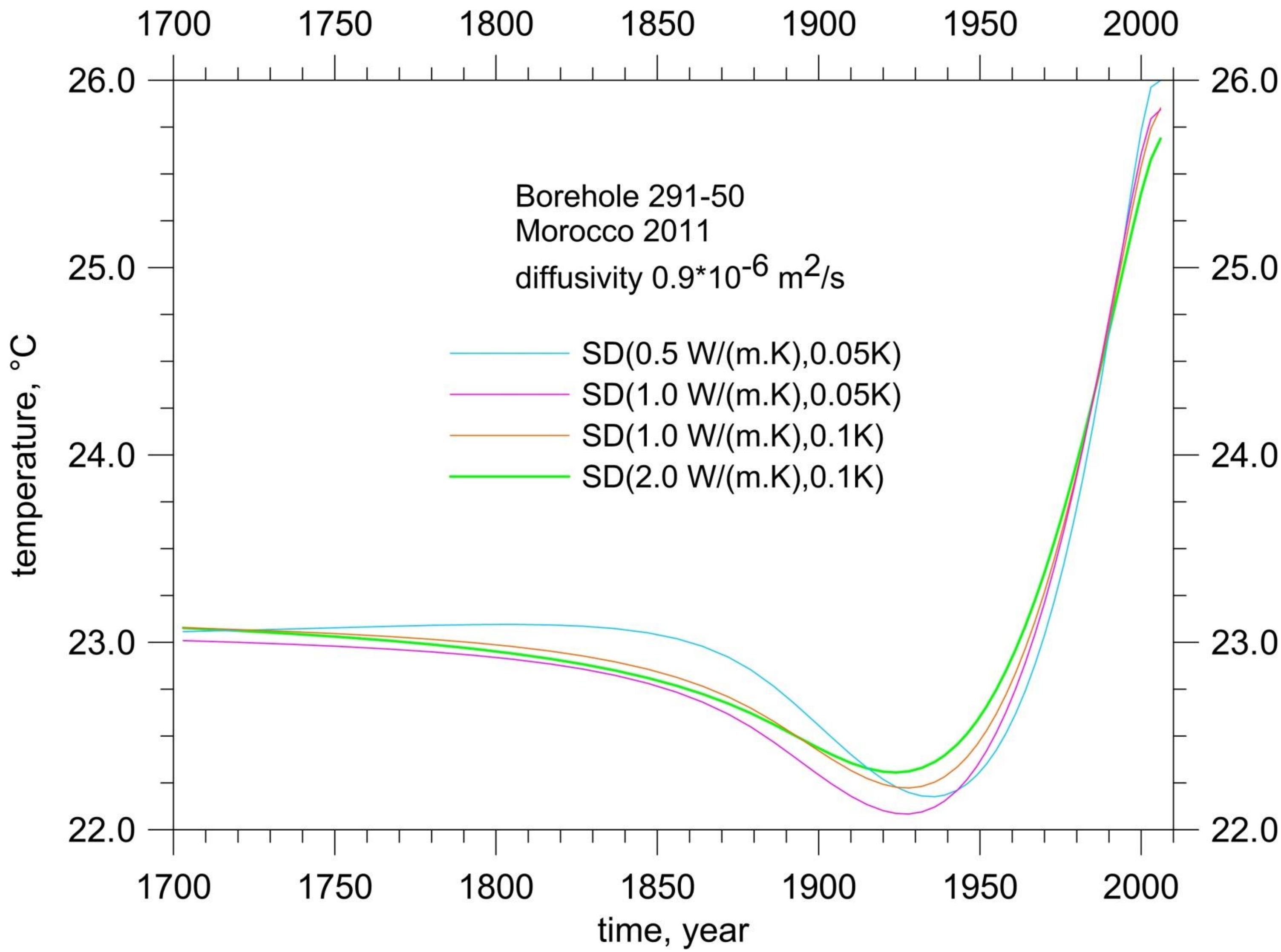




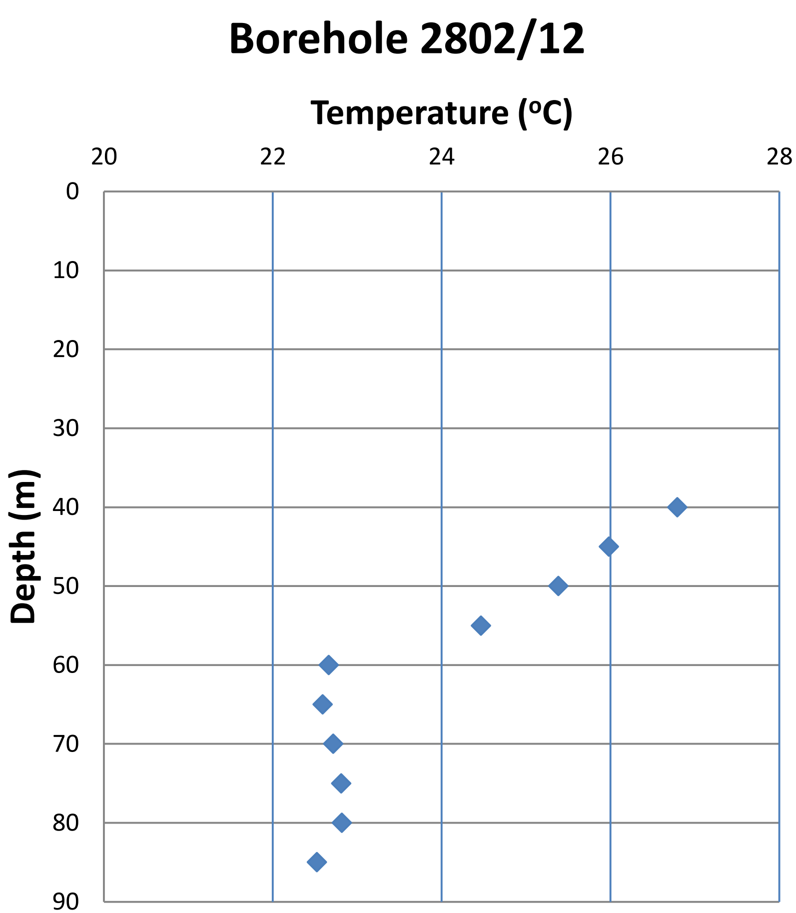
Examples (Morocco)

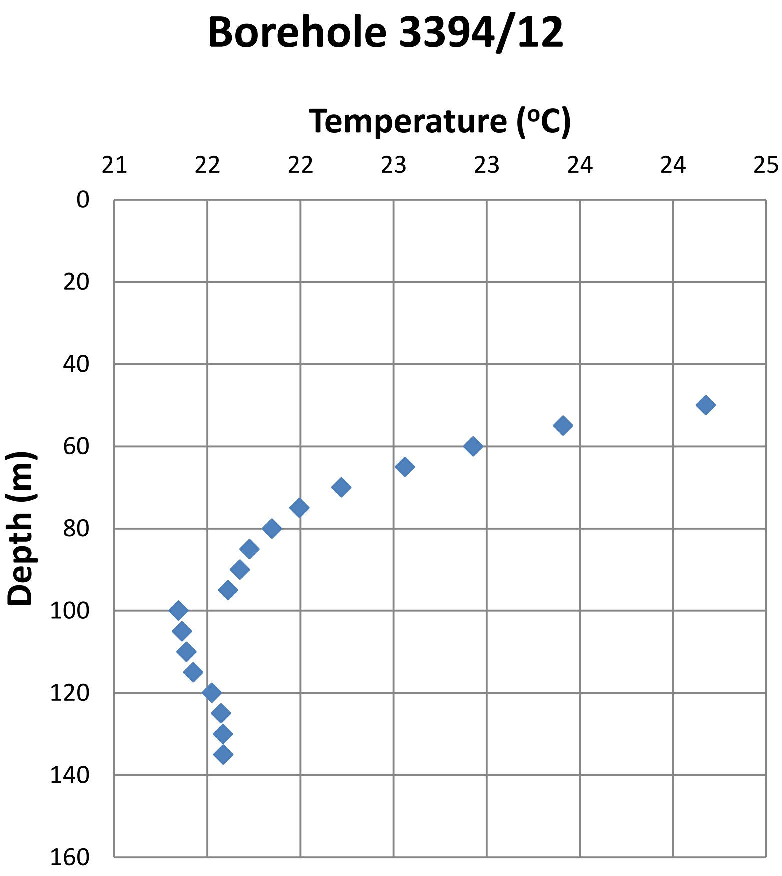


Examples (Morocco)

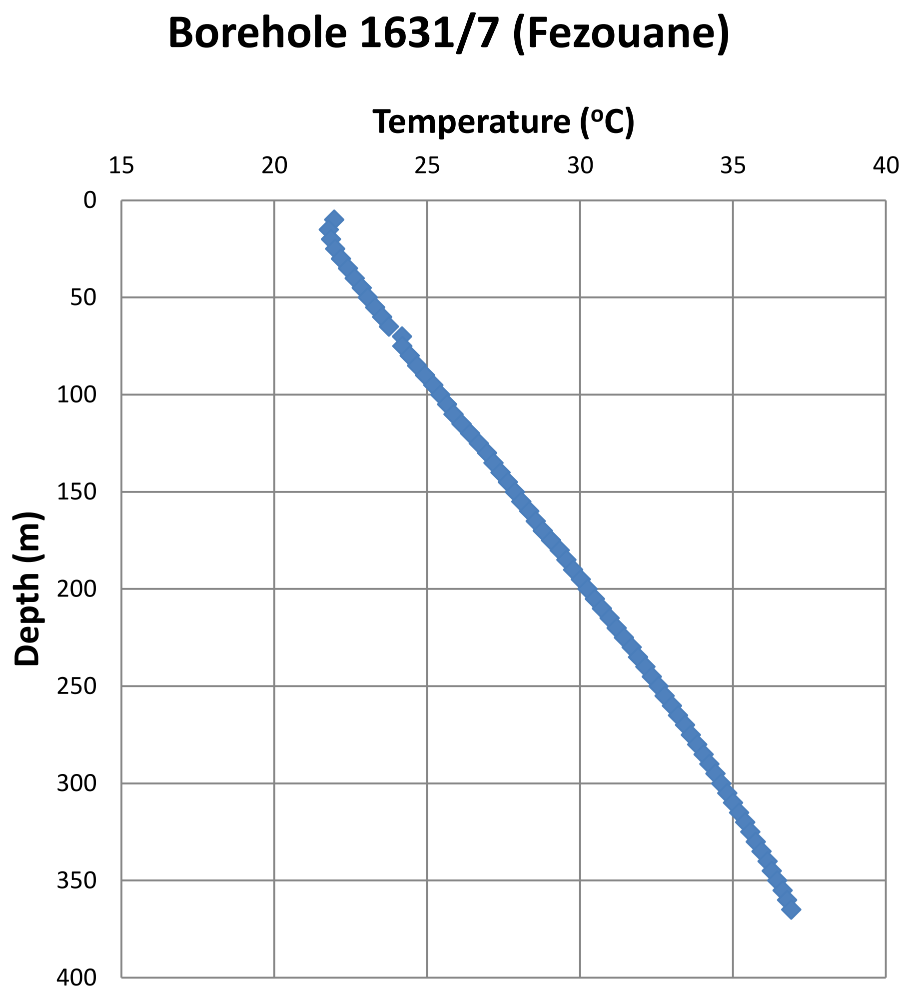


Examples (Morocco)

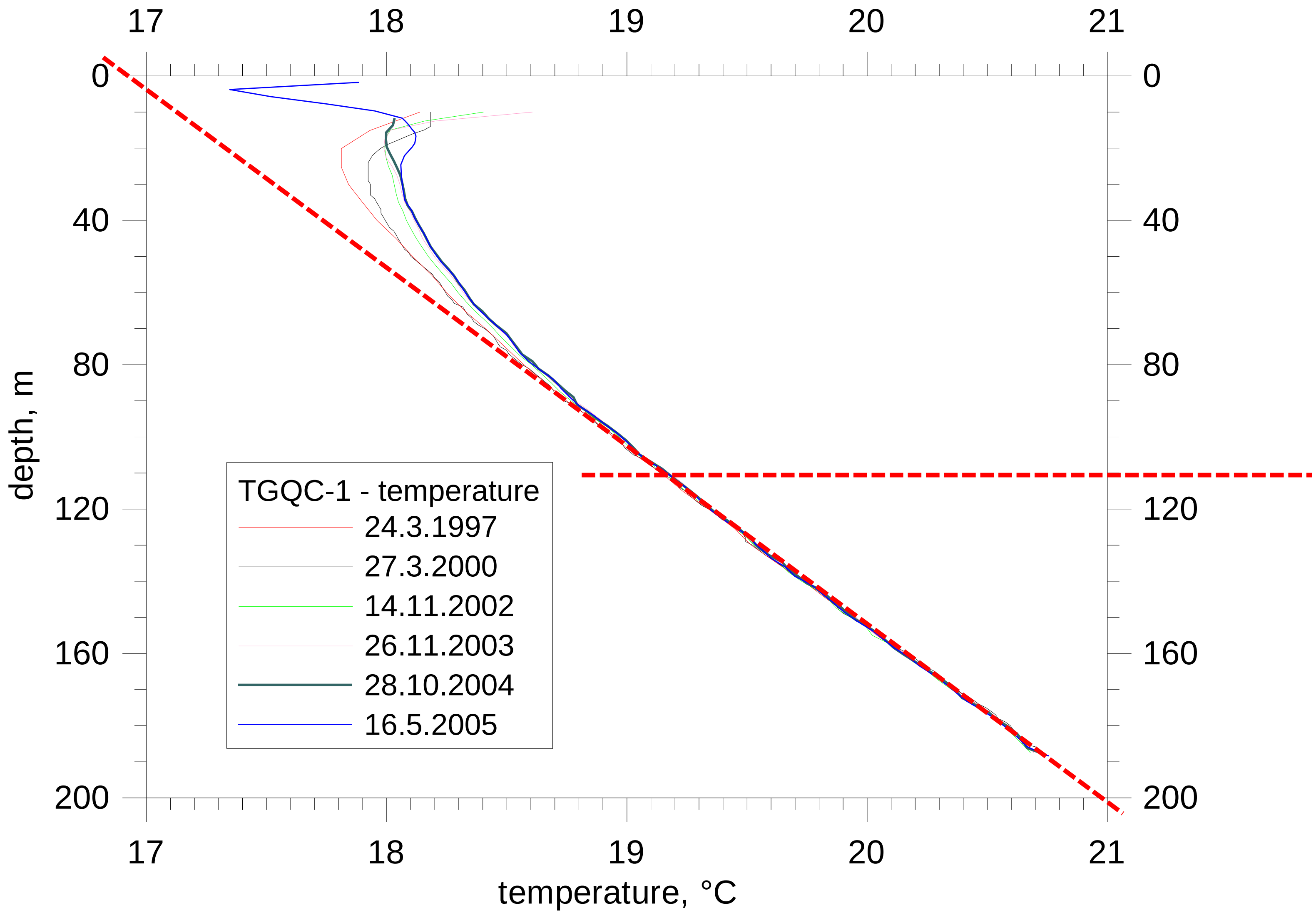




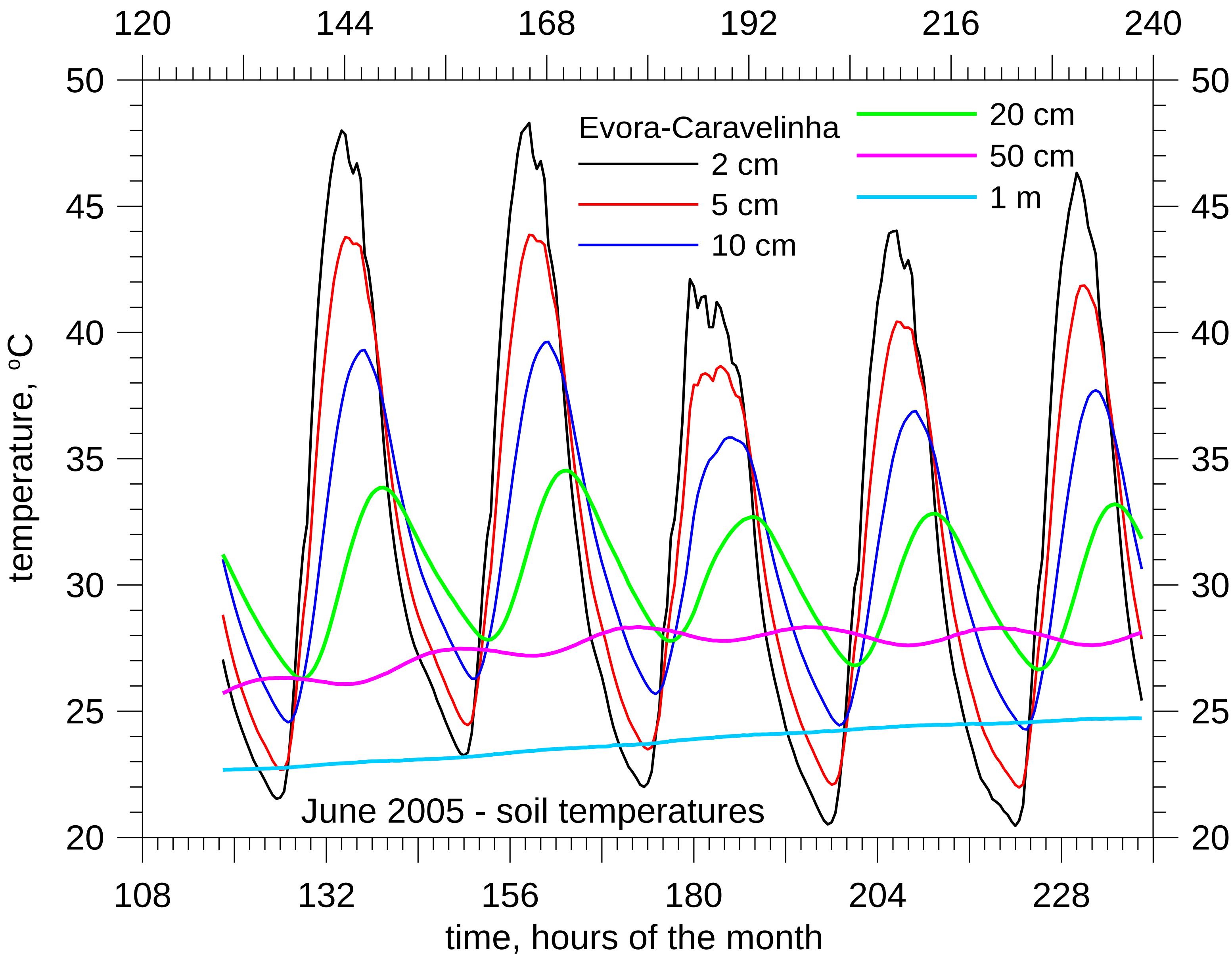
Examples (Morocco)



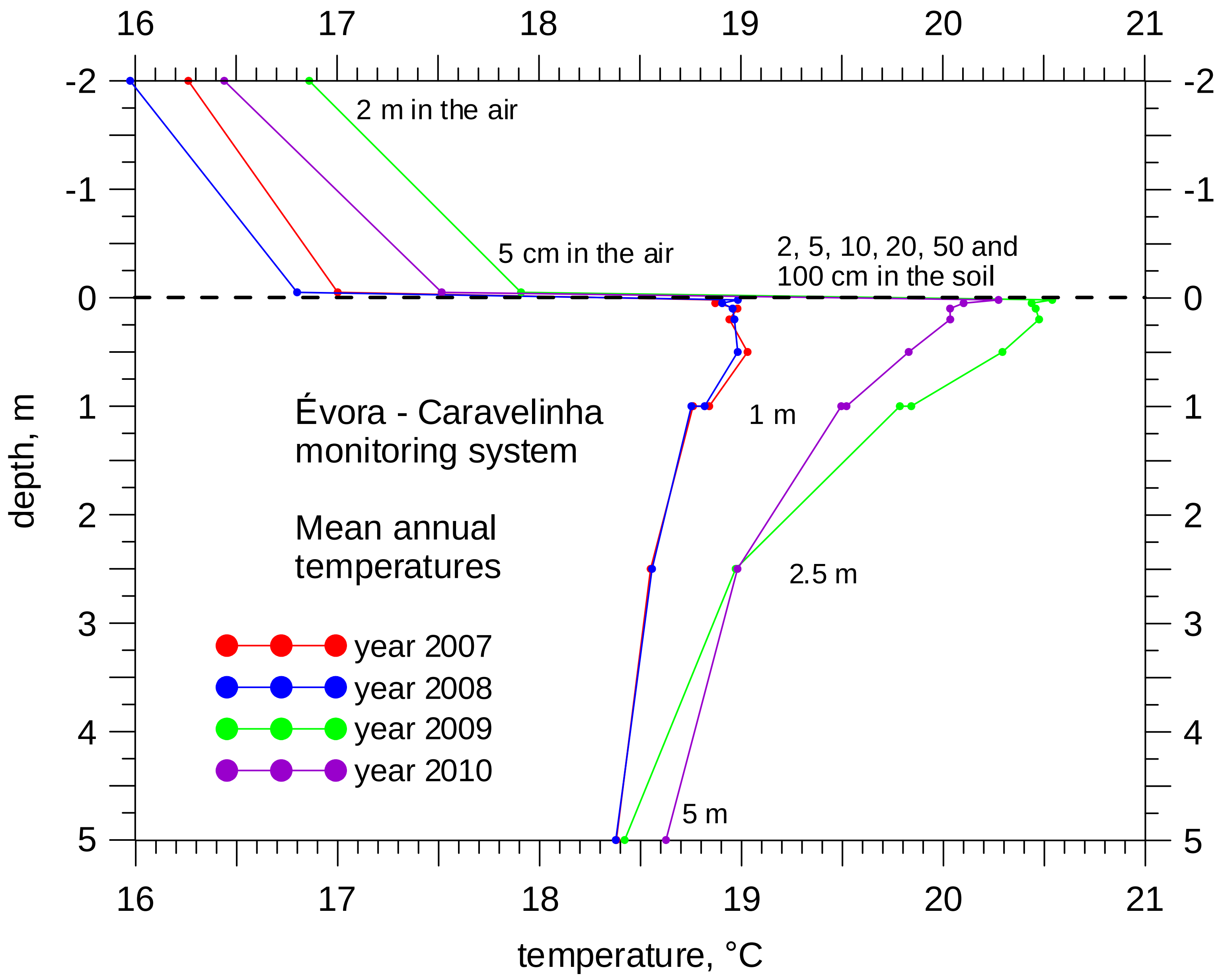
Temperature logs for different years



Temperature for different depths

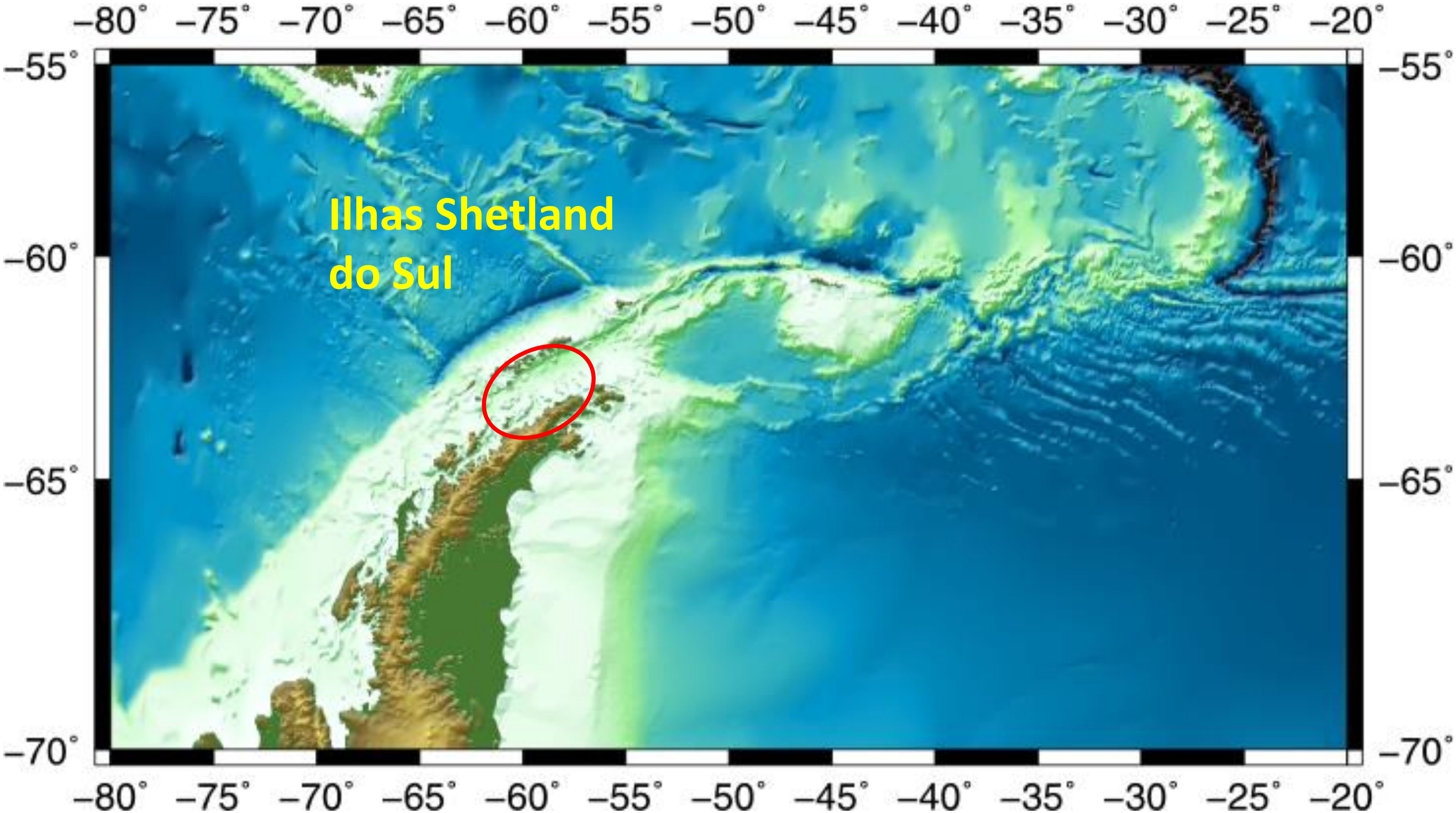


Temperature at ground's surface

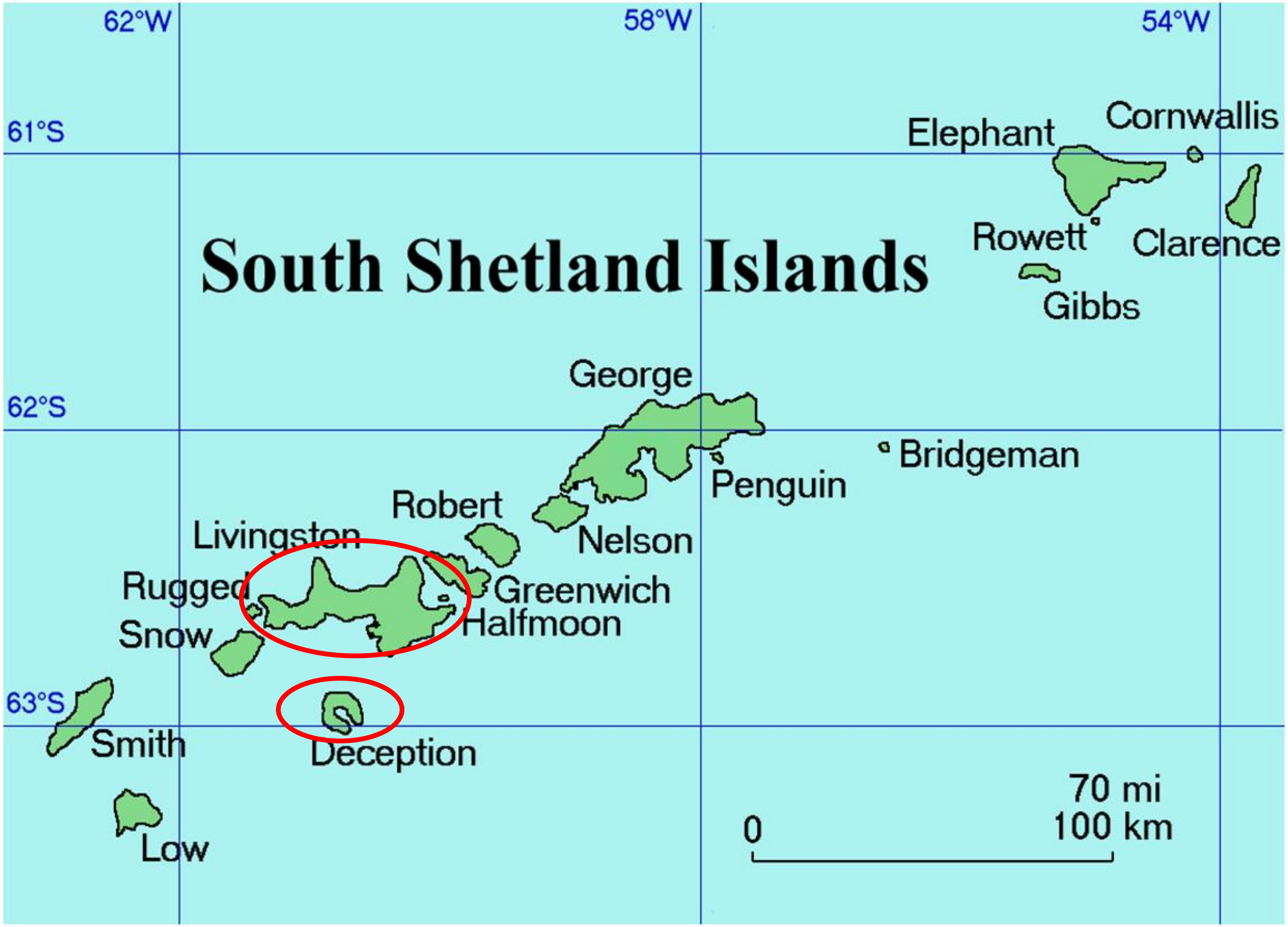


The Antarctica Continent

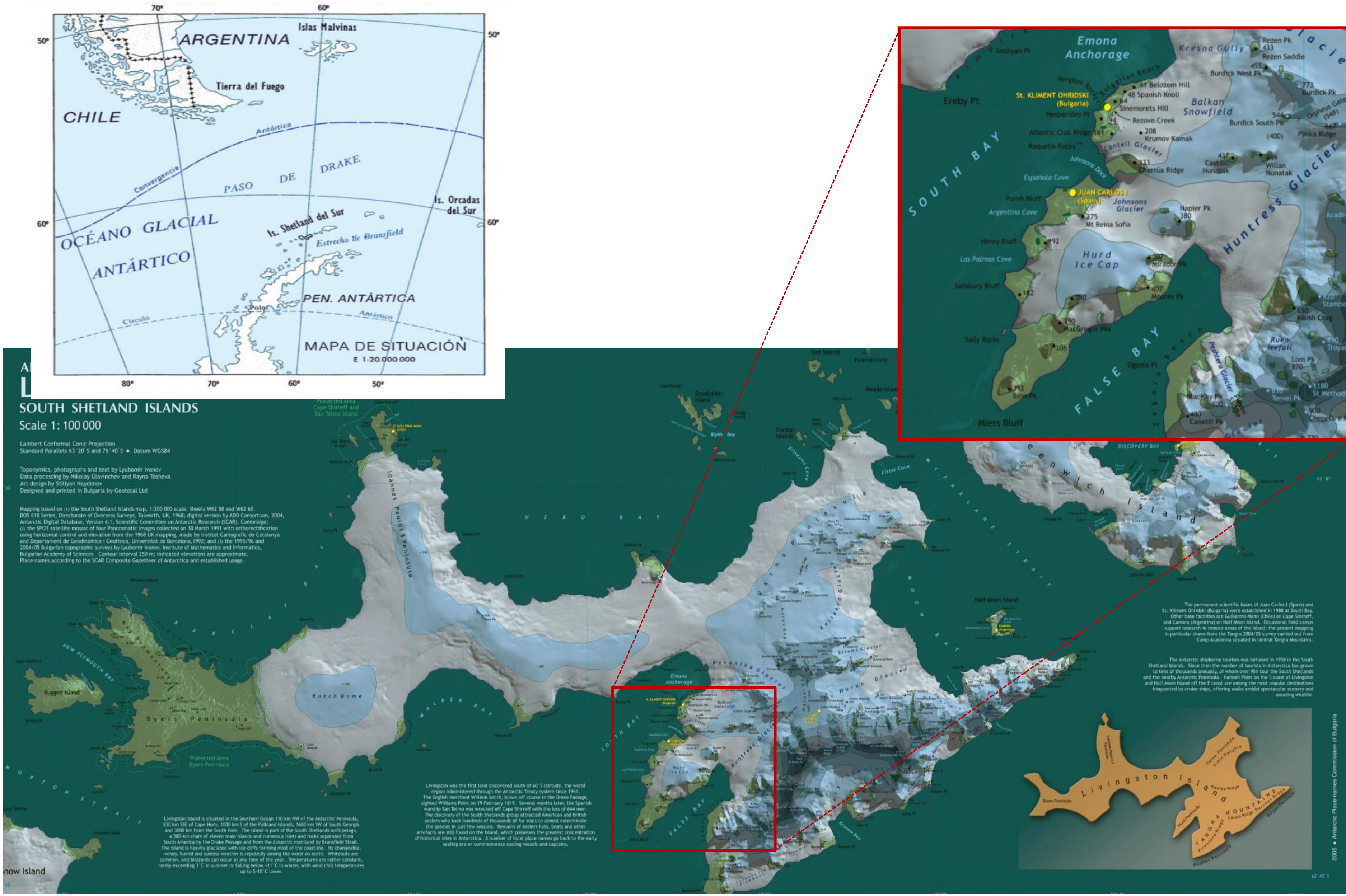
Antarctica Peninsula



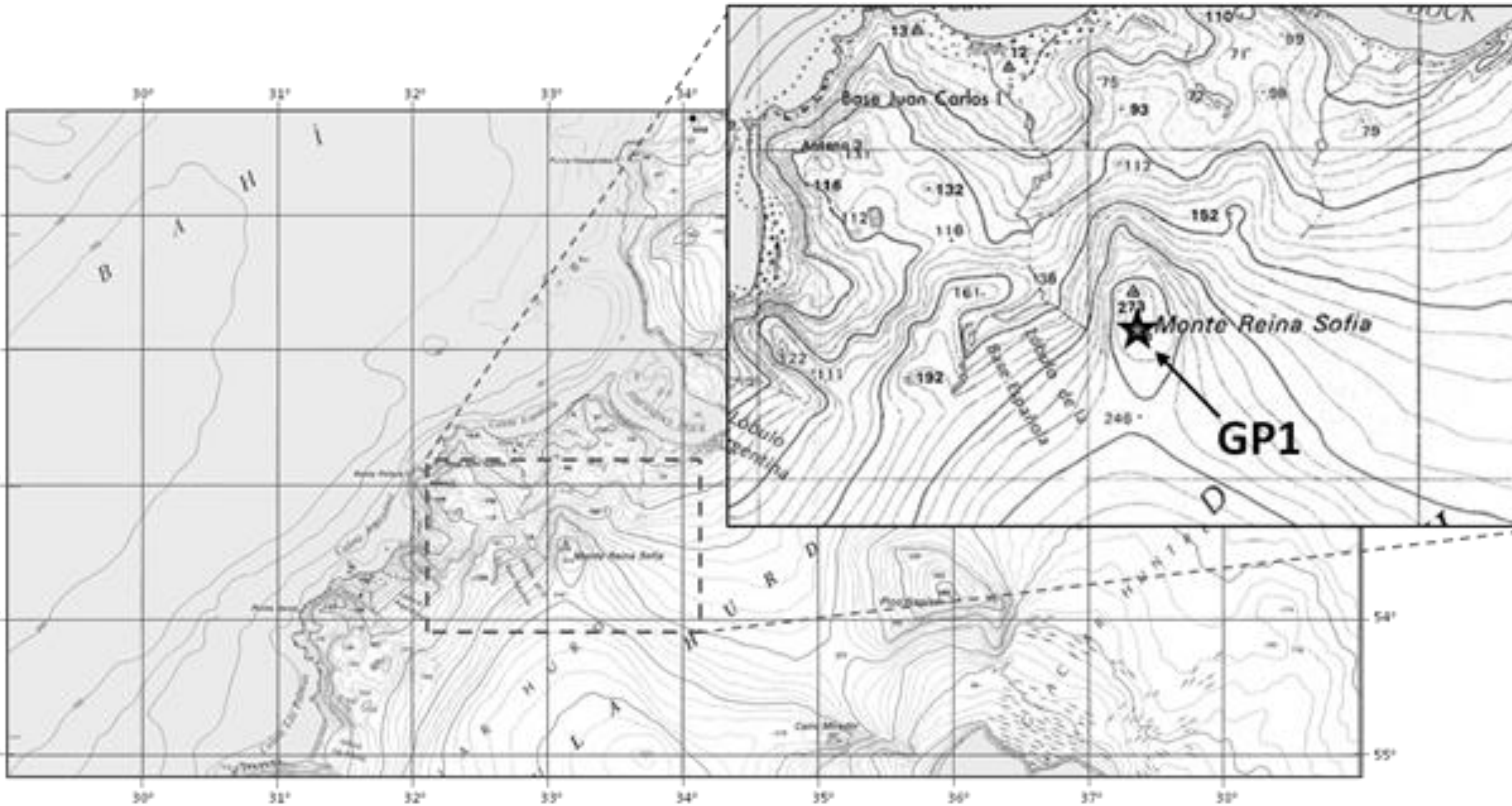
Livingston and Deception Islands



Hurd Peninsula in Livingston Island

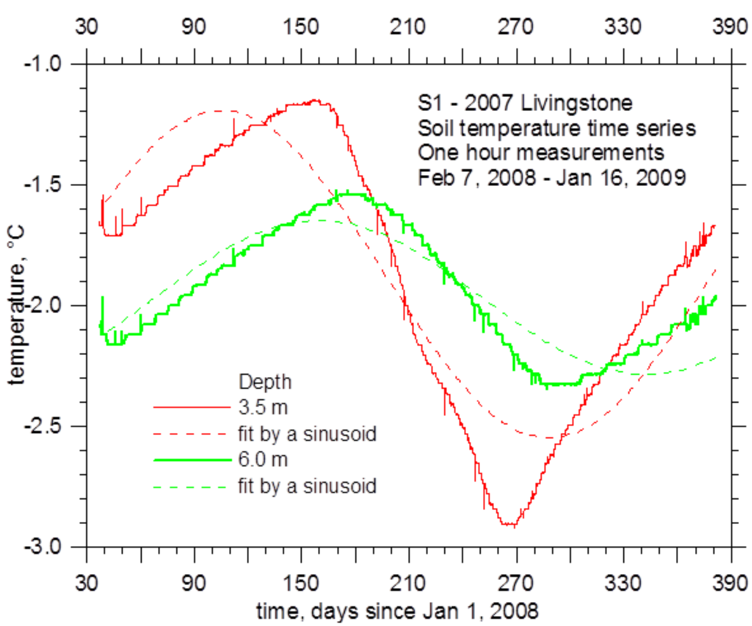


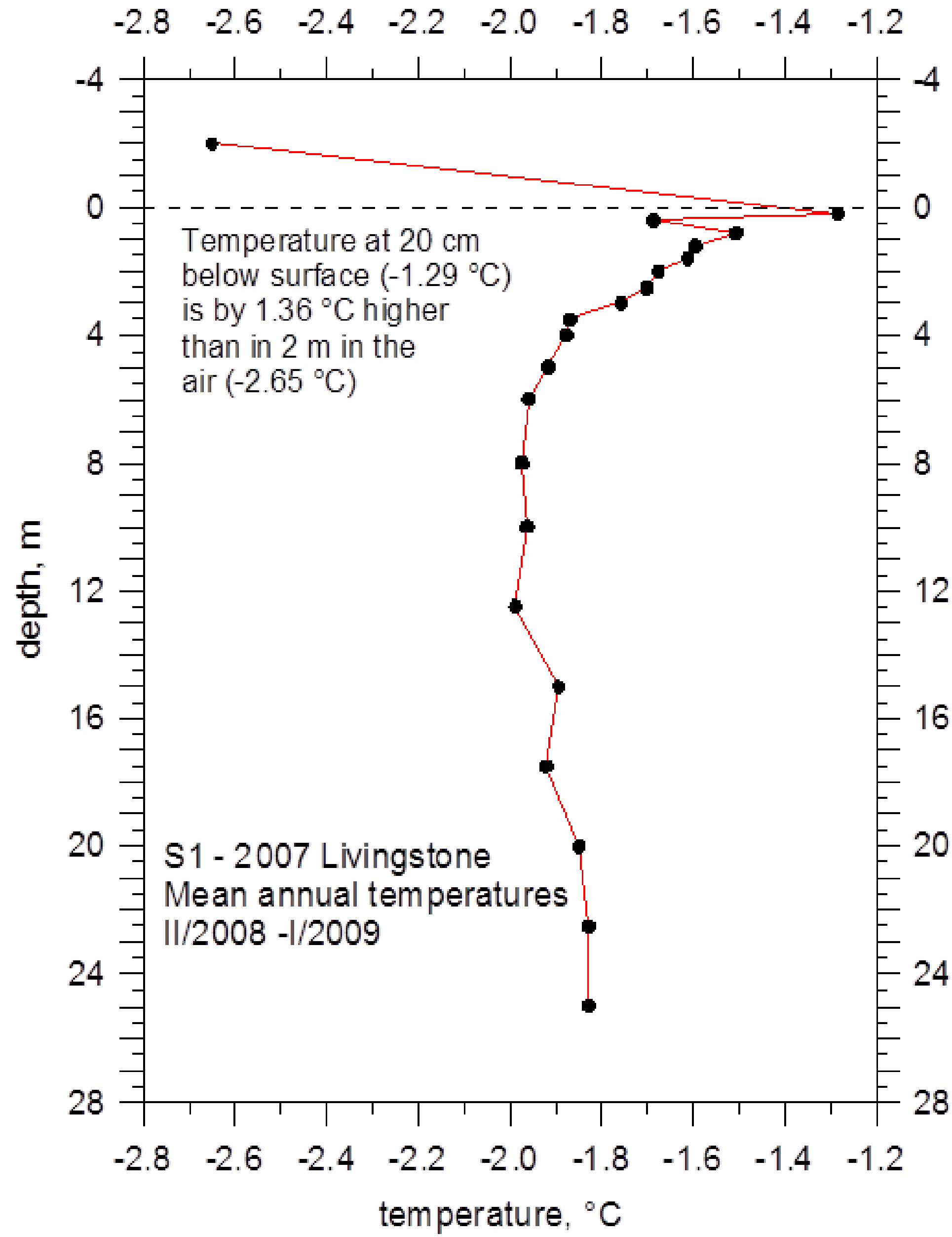
Hurd Peninsula in Livingston Island



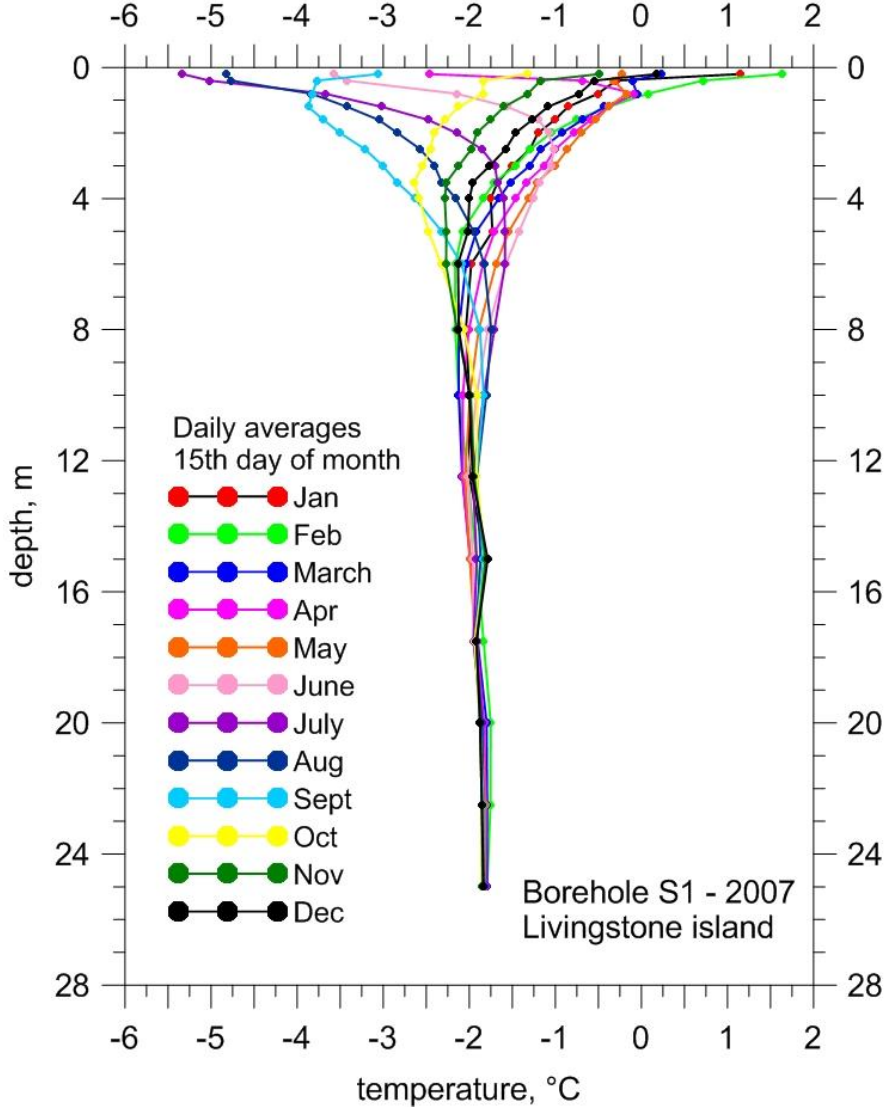
Hurd Peninsula in Livingston Island

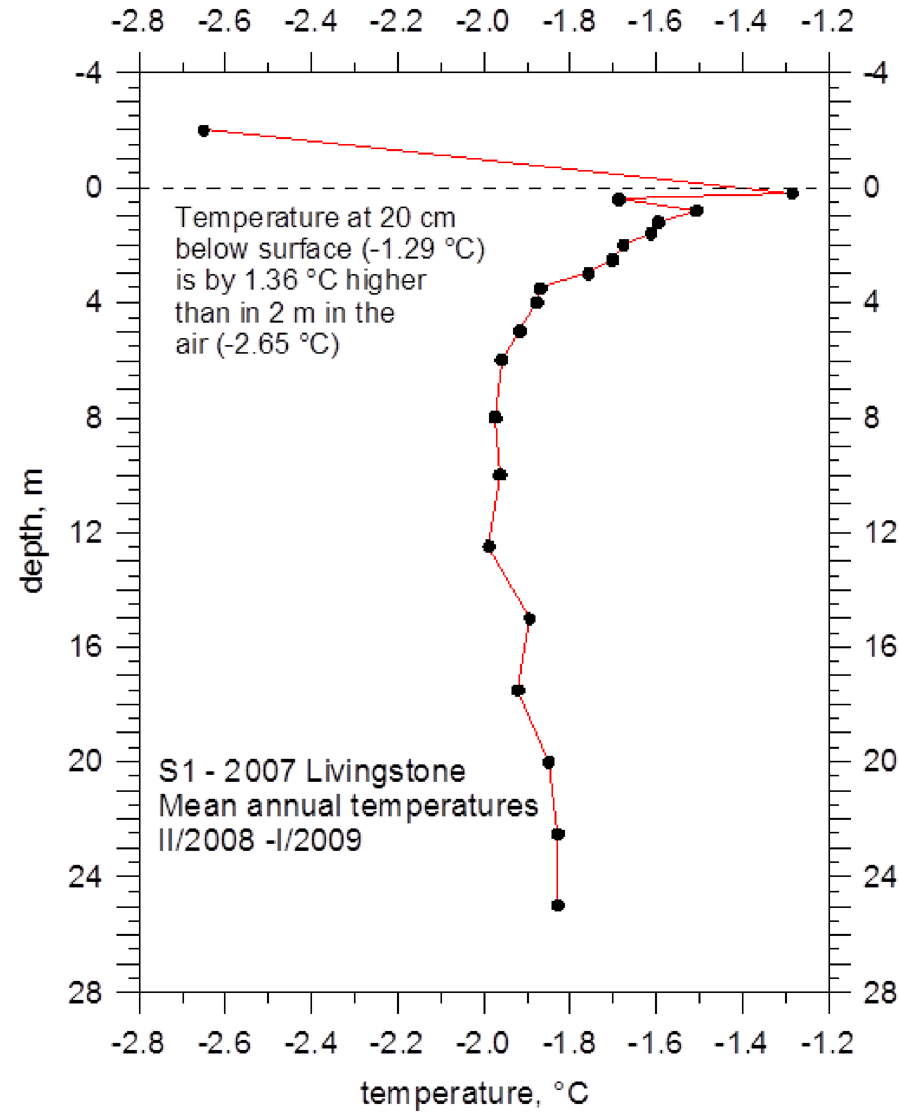
Hurd Peninsula in Livingston Island





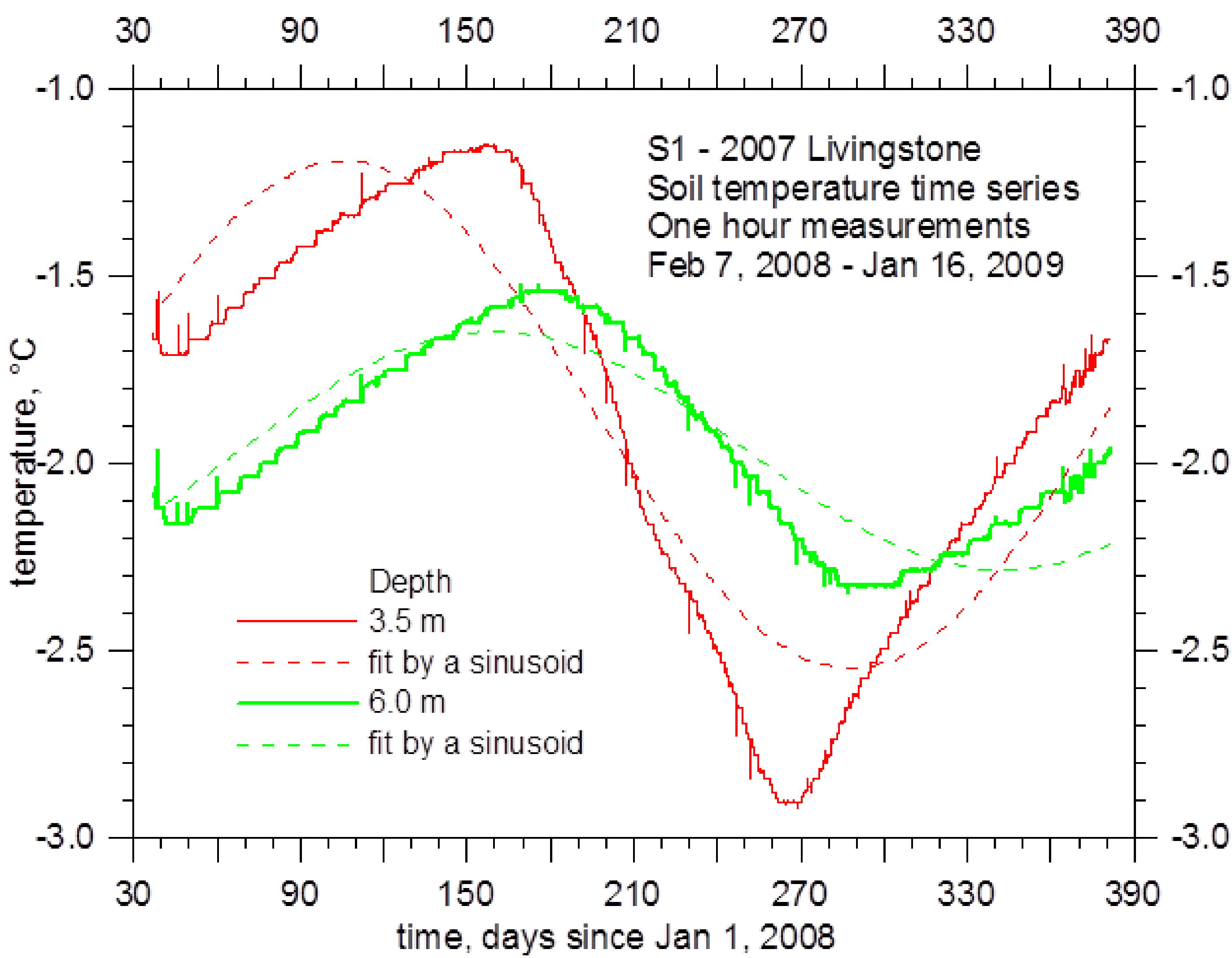
Hurd Peninsula in Livingston Island





Hurd Peninsula in Livingston Island

Why the real data does not fit the sinusoids?



Indirect temperature measurements

As a complement to direct methods to measure temperature, other indirect methods have been developed. However, in all of them the accuracy is much lower than in the direct methods. For heat flow density studies in the crust they serve as indicators of its thermal regime. They are:

- Groundwater Geochemistry
- Curie Depth
- Xenoliths
- Upper Mantle Resistivity

Groundwater Geochemistry

The solubility of many compounds in water increases with temperature. The temperature of a given porous formation can be estimated from the amount of dissolved material in its water. Relationships of this type (geothermometers) have been developed for several components or chemical elements present in the water filling the pores of different rocks. For silica (SiO_2) an empirical expression was developed by Swanberg and Morgan (1979):

$$T = [1315/(5.205 - \log_{10}(SiO_2))] \pm 0.5°C \quad (1)$$

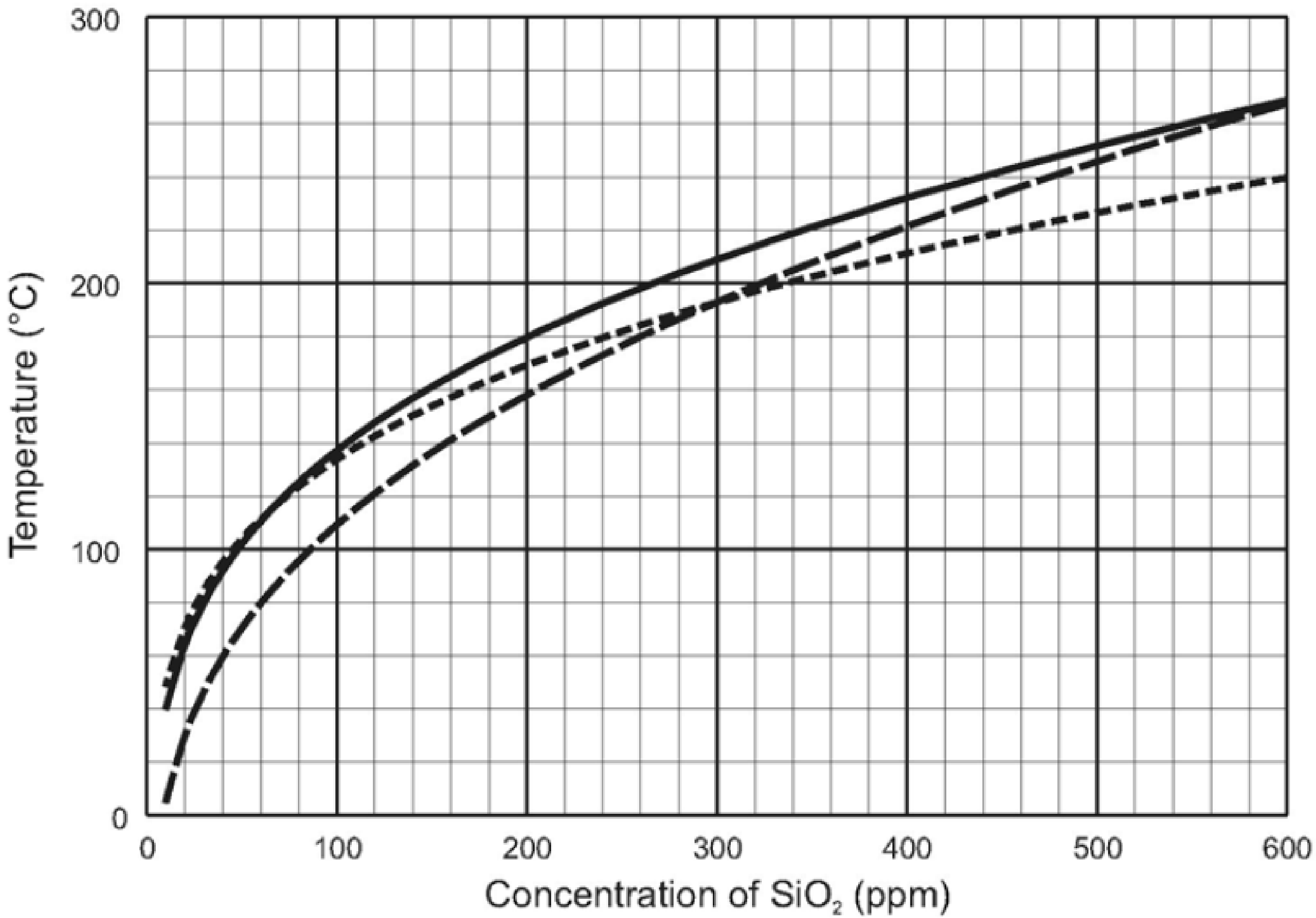
where T is the estimated water temperature (K) and SiO_2 is the amount of dissolved silica (ppm) in the formation water. The quoted precision is valid in the range 125-250 °C.

Other thermometers using silica where developed such as:

$$T = [1533.5/(5.768 - \log_{10}(SiO_2))] \pm 2.0\ °C \quad (2)$$

$$T = [1015.1/(4.655 - \log_{10}(SiO_2))] \pm 2.0\ °C \quad (3)$$

Groundwater Geochemistry



Relationships between concentration of dissolved silica in groundwater (ppm) and temperature of the water (°C). Solid line - Eq. (1); short-dashed line – Eq. (2); long-dashed line – Eq. (3).

Groundwater Geochemistry

Geothermometers for sodium (Na), potassium (K) and calcium (Ca),

$$T = [855.6/(0.8573 + \log10(Na/K))] \quad \pm\ 2.0\ °C \qquad (4)$$

where Na and K are concentrations in ppm and temperature is in kelvin.
Fournier and Truesdell (1973) suggested the following:

$$T = [846/(0.5964 + \log10(M_{Na}/M_K))] \qquad (5)$$

$$T = [1647.3/(2.24 + \log_{10}(M_{Na}/M_K) + \beta \cdot \log_{10}(\sqrt{M_{Ca}}/M_{Na}))] \qquad (6)$$

where M_x is the molar concentration of element X (moles per litre),
β is 4/3 for $(\sqrt{M_{Ca}}/M_{Na}) > 1$ and $T < 373.15$ K (100°C),
β is 1/3 for $(\sqrt{M_{Ca}}/M_{Na}) < 1$ or $T > 373.15$ K (100°C),
and T is in kelvin

Groundwater Geochemistry

Geothermometers are most useful in high-temperature geothermal reservoirs where direct temperature measurements are not possible or are difficult. As a matter of fact, they were developed for use in high-enthalpy geothermal reservoirs

Curie Depth

By definition, the ***Curie temperature*** is the point at which a mineral loses its ferromagnetic properties and the ***Curie depth*** is the depth at which crustal rocks reach the Curie temperature. The Curie depth can be used to constrain deep geothermal gradients.

- For pure magnetite the Curie temperature of 580 °C; however, inclusions of titanium can reduce that temperature to values as low as 300 °C.
- In andesites and alkali-basalts ferromagnetic minerals have Curie temperatures ranging from 100 to 300 °C.
- Intermediate to mafic rocks have Curie temperatures ranging from 300 to 450 °C.
- Fe-Co-Ni alloys have Curie temperatures ranging from 620 to 1100 °C .

Curie Depth

Several authors have used the theoretical possibility to determine Curie depths from total magnetic field intensity data. The area covered must be about 200 x 200 km, with a maximum grid spacing of 1 km. Curie depth estimates are based on a well known method developed by Spector and Grant.

The Curie depth does not necessarily define an isotherm. As a matter of fact, different rock types have different Curie temperatures, which means that the Curie depth can also correspond to a composition boundary.

Curie Depth

Seismic or gravity data may be used to infer if the Curie depth corresponds to an isotherm or to composition boundary. When the Curie depth coincides with an inferred velocity boundary or an inferred density boundary, it is probable it reflects a composition boundary. When the Curie depth does not coincide with a velocity boundary or density boundary, it is probable it represents the Curie temperature isotherm (about 580 °C in most continental areas).

The Curie depth-temperature pair then allows the to estimate the deep geothermal gradient in the study area.

Xenoliths

By definition a xenolith is a piece of country rock picked up by magma as it rises through the crust. The pressure-temperature stability conditions indicated by mineral assemblage in the a xenolith provide information on the depth and temperature of its origin point. Thus, xenoliths provide an independent estimate of temperature at depths down to hundreds of kilometres (e.g. O'Reilly and Griffm, 1985).

Upper Mantle Resistivity

The electrical resistivity of mantle rocks is strongly dependent on temperature. So, in principle, electrical resistivity can be used to infer temperature in the lower crust and upper mantle.

Some authors have identified a direct relationship between the 450 °C isotherm and the depth to an electrically conductive layer in the Canadian Cordillera and in other regions of the Earth.

Magnetotelluric methods are used to make the electrical resistivity measurements and can provide information about the depth of the 450 °C isotherm.

However, the depth resolution is poor.

Surface temperature (onshore)

When temperatures are measured in boreholes to estimate the geothermal gradient it is also important to know the temperature at the surface of the Earth; that is a way of constraining the geothermal gradient for stationary conditions.

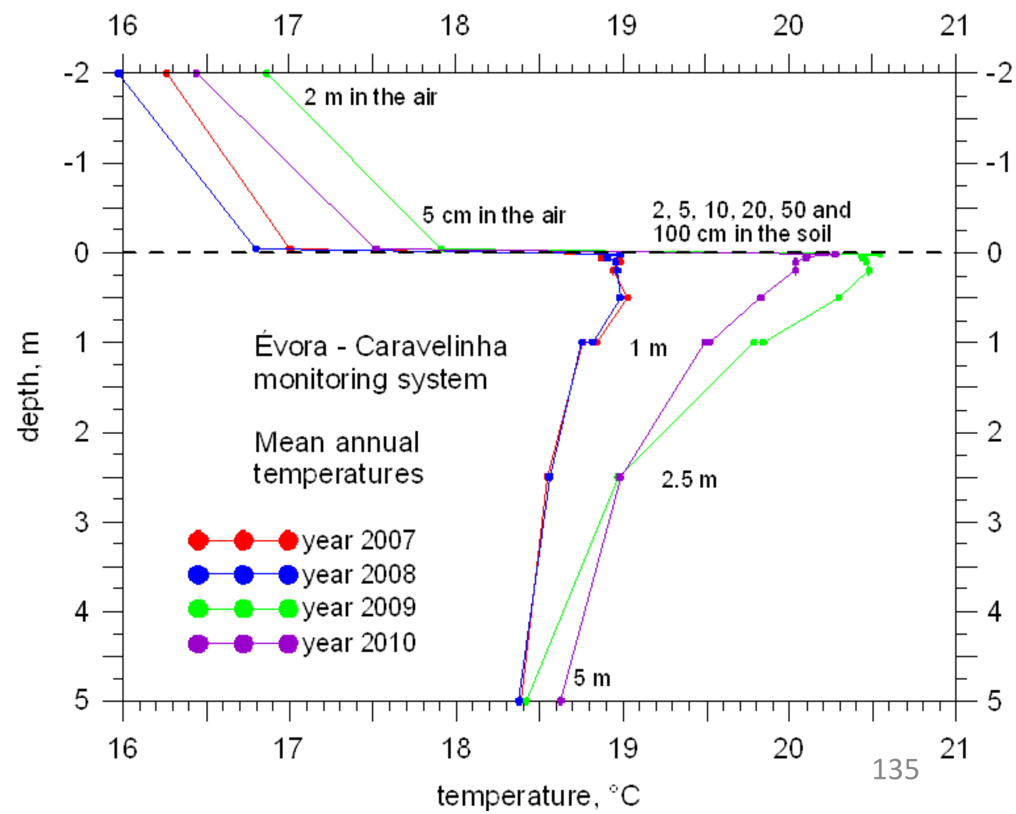
Onshore the temperature of the surface rocks generally exceeds the average air temperature by a few degrees due to surface albedo and interaction of the solar radiation with the ground.

The best way to estimate the average surface temperature (T_o) is to use local meteorological records and the equation:

$$T_0 = 3 + \frac{(T_{av.min} + T_{av.max})}{2}$$

Surface temperature (onshore)

$$T_0 = 3 + \frac{(T_{av.min} + T_{av.max})}{2}$$

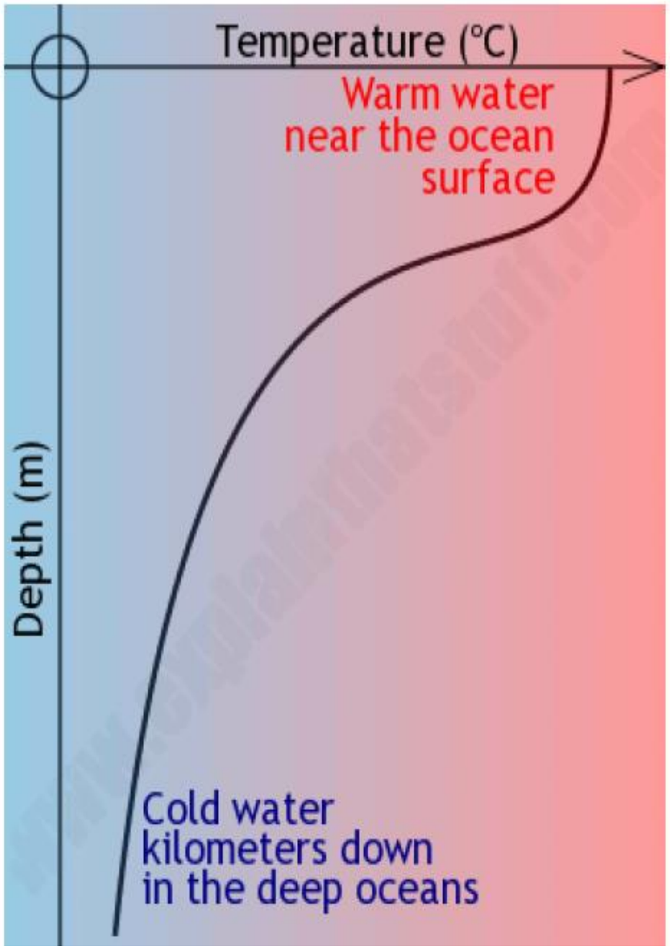


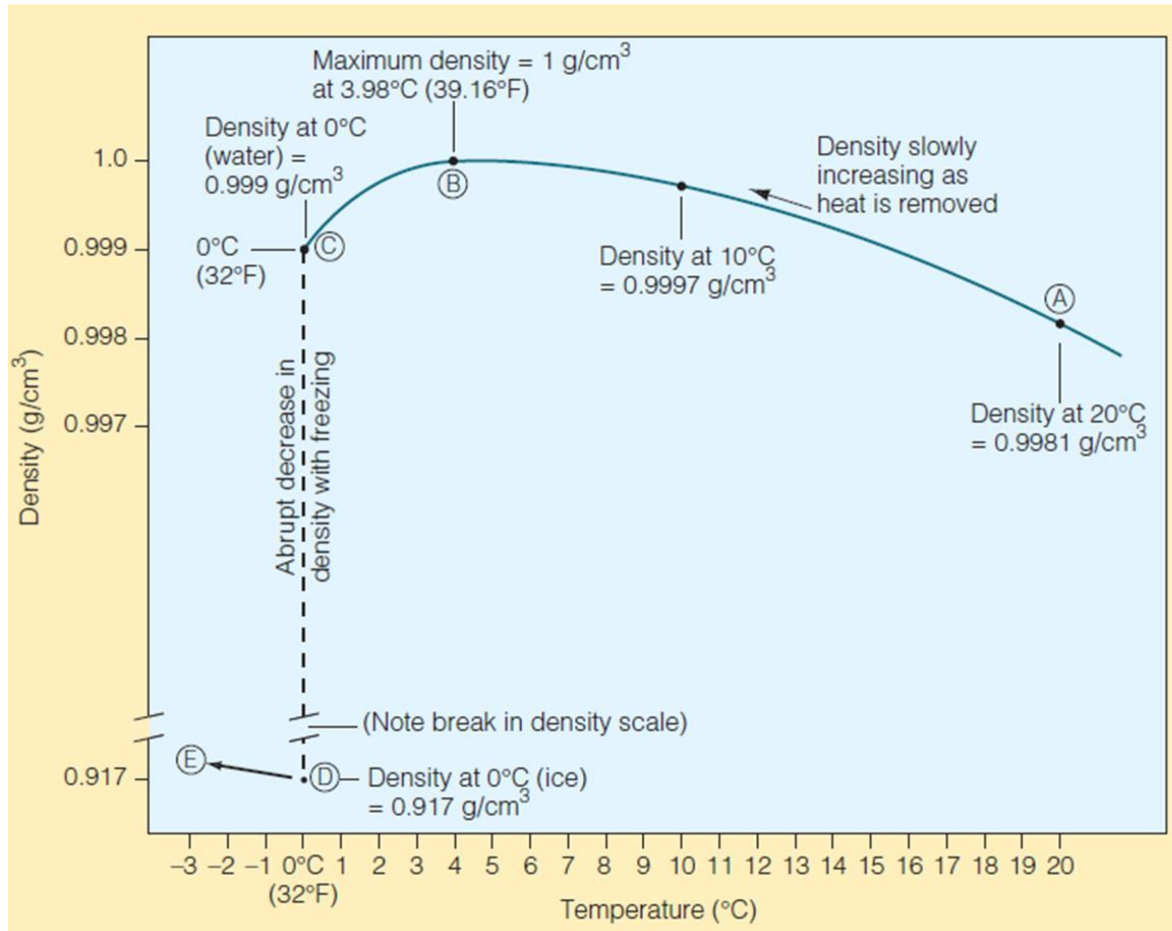
Surface temperature (offshore)

When speaking about "surface temperature" for offshore temperature logs people is referring to the top of the sediment column or the bottom of the water column.

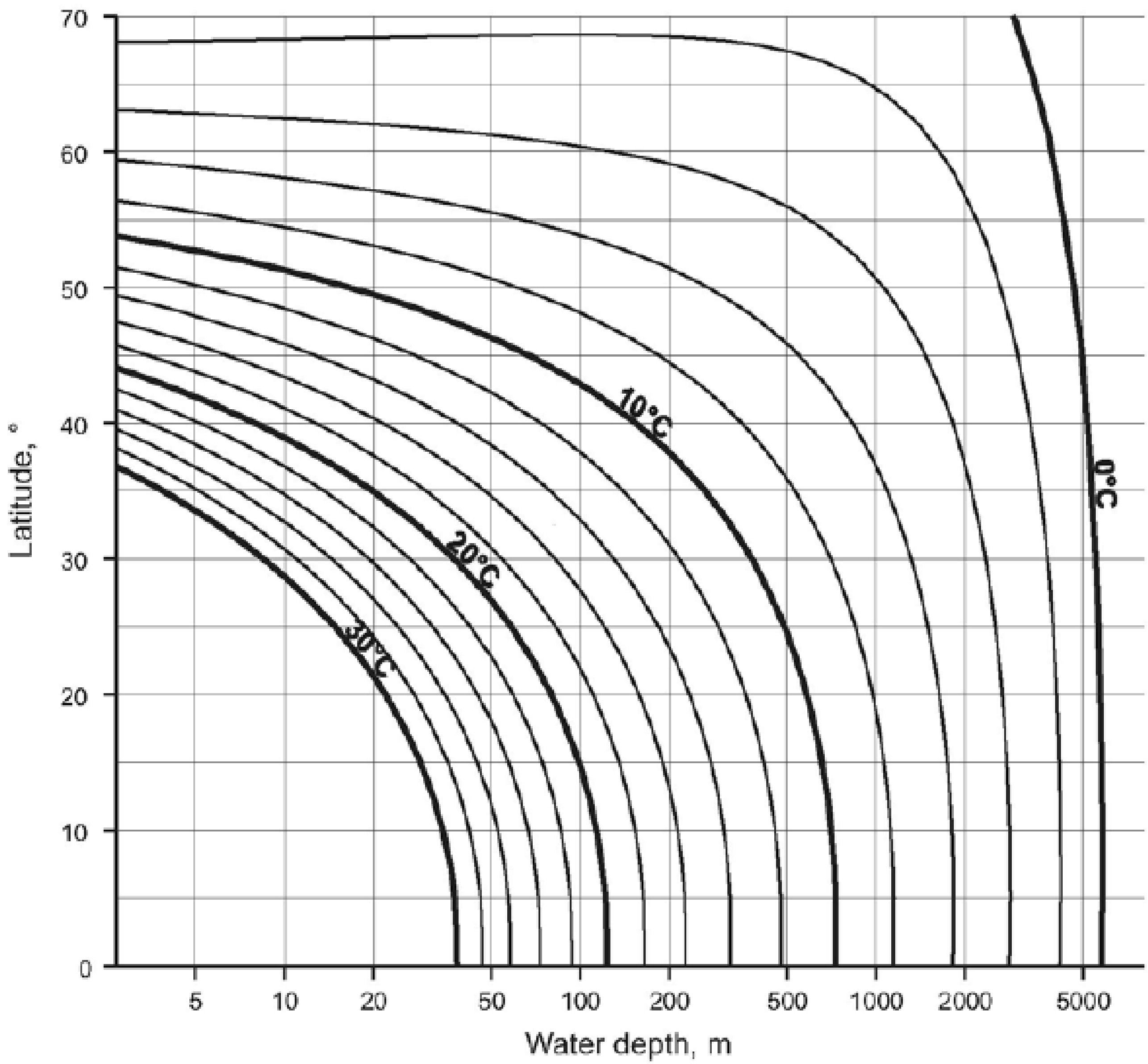
In this case the surface temperature is the bottom water temperature which is obviously different from the average sea surface temperature. Therefore, to calculate a geothermal gradient we must know the temperature at the bottom of the water column, in case we are looking at a conductive process of heat transfer. For the ocean there are graphs that give the temperature at their bottom in accordance with some ocean models.

Surface temperature (offshore)



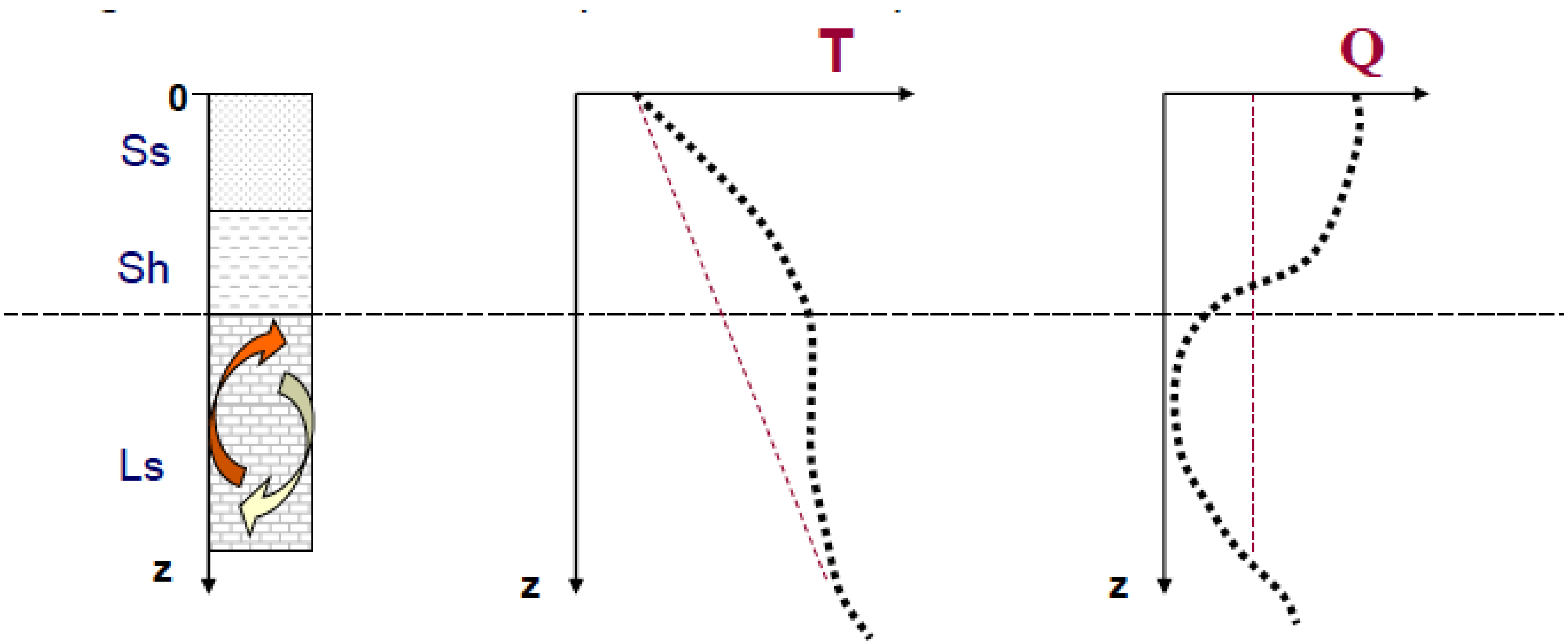


Surface temperature (offshore)



Bottom-water temperature (BWT) (°C) as a function of water depth (m) and latitude (°)

Fluid flow in boreholes

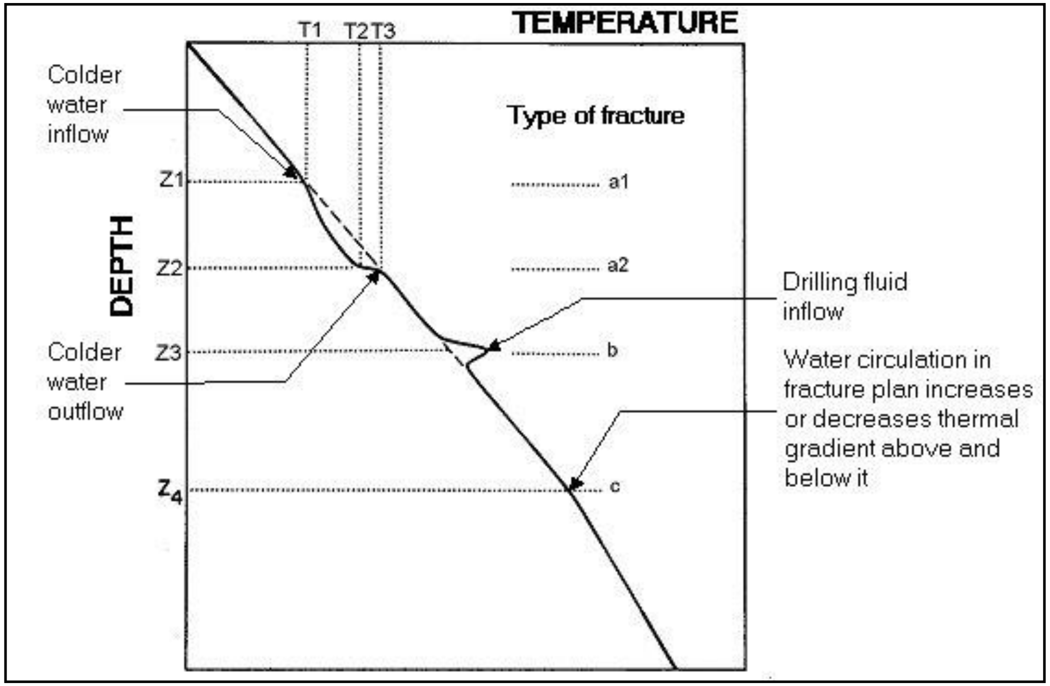


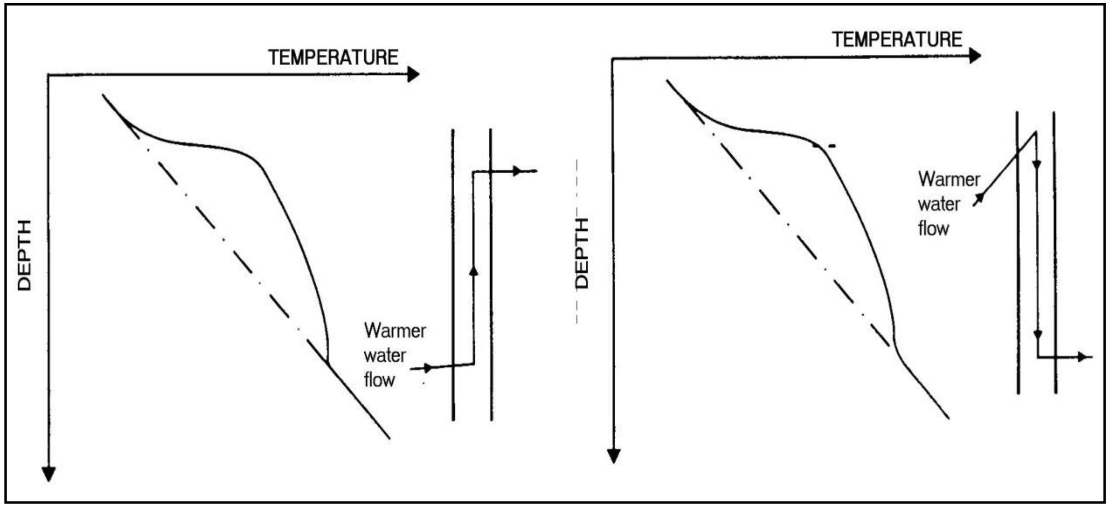
$$q_z = {}^c q_z + {}^a q_z = -K\frac{dT}{dz} + \rho C_p T V_z$$

Fluid velocity

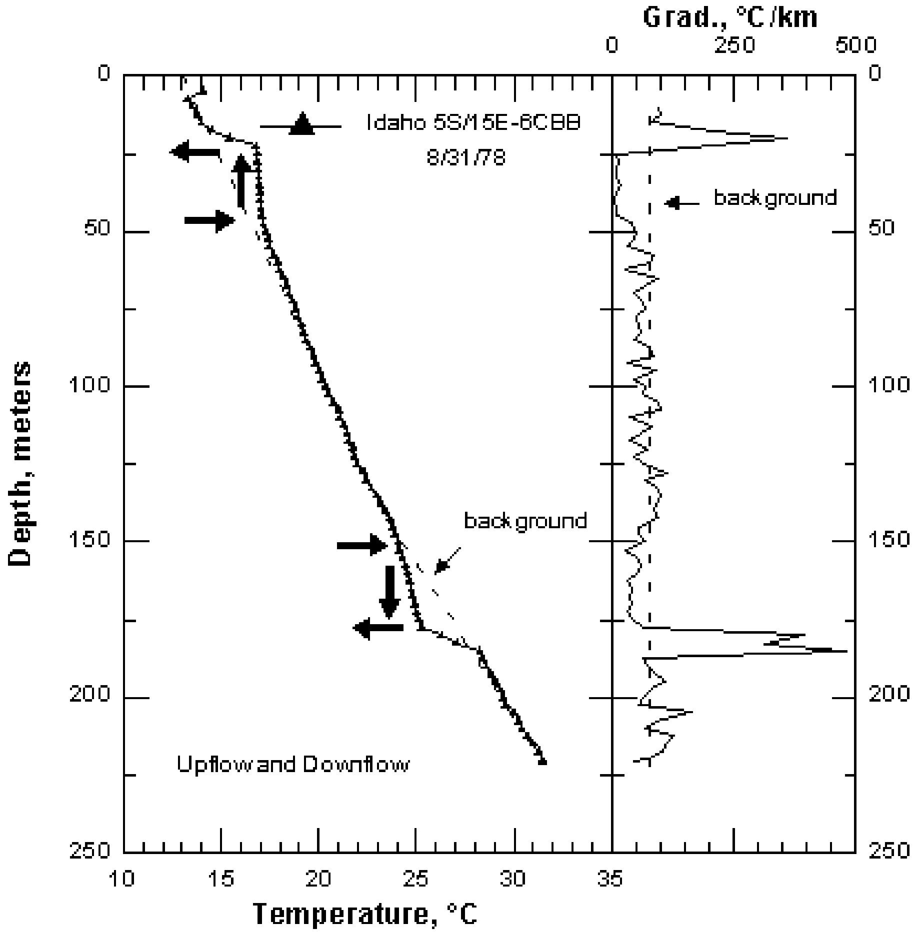
Fluid flow in boreholes

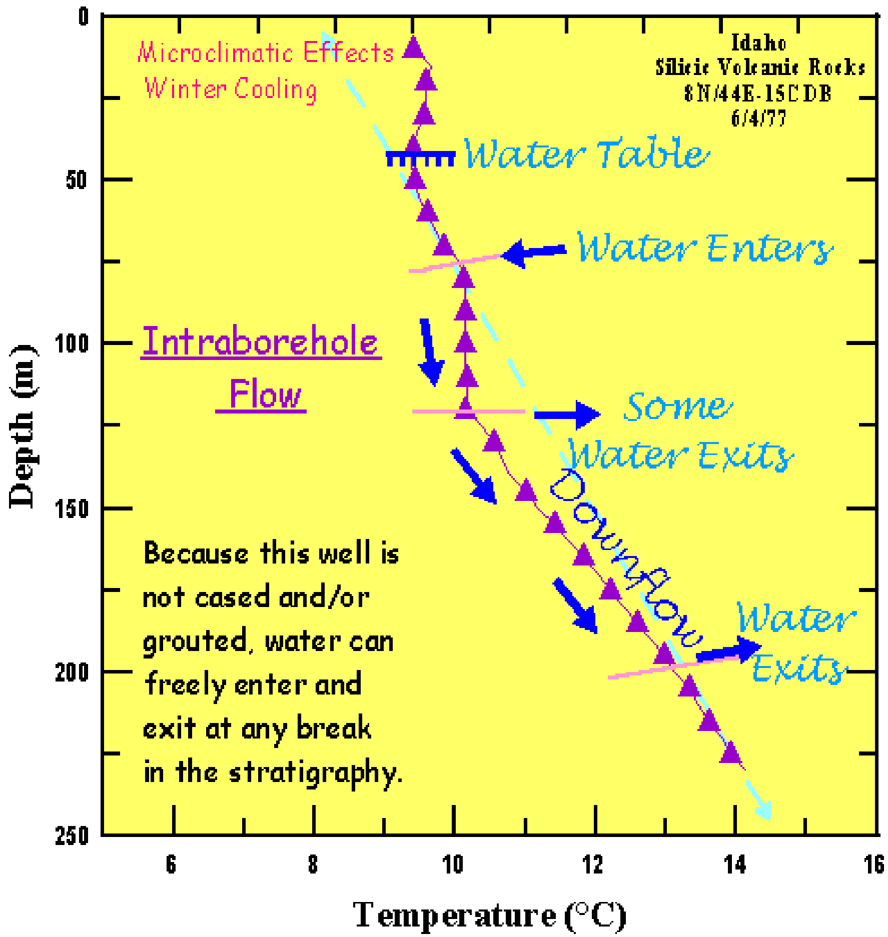
Temperature logs with evidence of fluid flow





Fluid flow in boreholes





Direct temperature measurements

Bottom-Hole Temperatures (BHT)

Drilling in the oil industry consist of (very short summary):

- Start to drill to, let’s say, 200 m using mud to cool the drill bit and transport the cuttings to the surface
- Stop drilling and start mud circulation for some time (let’s say 2 hours) to homogenize the mud column for well logging
- Pull out the drilling column
- Carry out several runs with different probes for measuring, e.g., gamma ray, electrical resistivity, spontaneous potential, density logs, etc.

Direct temperature measurements

Bottom-Hole Temperatures (BHT)

Drilling in the oil industry consist of (very short summary (cont.)):

- If we are lucky the operators of the geophysical probes will attach a maximum thermometer to each probe

- Each probe reaches the bottom of the borehole some time after the mud circulation has stopped and records the temperature (BHT); the time each probe reaches the bottom is also registered

- Well logging comes to an end and drilling restarts to a new depth (let's say 400 m) and the drilling, the mud circulation and the well logging processes repeat to the new depth

Direct temperature measurements

Bottom-Hole Temperatures (BHT)

So, for **each depth** (200 m, 400 m, and so on) we have:

- Several BHT (we must have at least 2)
- The time each probe reached the bottom of the borehole (200 m, 400 m, and so on) after circulation has stopped
- And the circulation time
- With this information it is possible to correct the BHT to obtain the virgin temperature of the geological formations crossed by the borehole
- Sometimes, though, the logging companies also carry out a continuous temperature log, which also has to be corrected for thermal disturbances during the drilling.

III
Thermal Conductivity

Reminder

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K\,\frac{dT}{dz} \qquad \alpha = \frac{K}{\rho \cdot c_p}$$

q_z is a vector as well as the temperature gradient dT/dz. To calculate (estimate) the heat flux we have to know thermal conductivity K and geothermal gradient dT/dz. Most of the times a q_z vertical is assumed. α is the thermal diffusivity.

Thermo-physical properties

$$\alpha = \frac{K}{\rho \cdot c_p}$$

Thermal conductivity is a measure of how easily heat is transmitted through a material by conduction (W/mK)

Thermal diffusivity is a measure of how easily a given material accumulates or looses heat. It controls the rate at which heat dissipates through that material (m^2/s)

The **specific heat** is the amount of thermal energy that has to be supplied to the unit mass of a given material to raise its temperature 1 K (or the amount of thermal energy that has to be withdrawn from the unit mass of a given material to decrease its temperature 1 K) (J/kgK)

Table 4.1. Thermal Conductivity, λ, of Some Common Minerals

Mineral	Temp. range (K)	λ ($W\,m^{-1}\,K^{-1}$)	Comments[Source]
Quartz	273.2–623.2	$4134 \times T^{-1.053}$	Parallel to *c*-axis[1]
	273.2–623.2	$820 \times T^{-0.861}$	Perpendicular to *c*-axis[1]
		7.69	Average[2]
Calcite	273.2–374.2	$264.5 \times T^{-0.727}$	Perpendicular to *c*-axis[1]
	273.2	5.51	Parallel to *c*-axis[1]
	332	5.16	Average[1]
		3.59	Average[2]
Dolomite		5.51	Average[2]
Orthoclase		2.32	Average[2]
Albite		2.37	Average[2]
Anorthite		1.68	Average[2]
Plagioclase		1.91	Average[2]
Sillimanite	333.2	2.60	Average[1]
Cordierite	320.8–398.1	$116.3 \times T^{-0.635}$	Average[1]
Salt	273–460	$4610 \times T^{-1.146}$	Average[1]
Topaz	314.2–419.8	$9946 \times T^{-1.094}$	Average[1]
Forsterite		5.12	Average[1]
Wollastonite	317.2–397.2	2.65	Average[1]
Zircon	318.8–411.6	4.03	Parallel to *c*-axis[1]
	318.7–414.2	4.14	Perpendicular to *c*-axis[1]
Tourmaline	398.2–723.2	$0.492 \times T^{0.297}$	Parallel to *c*-axis[1]
	393.2–729.2	$0.108 \times T^{0.556}$	Perpendicular to *c*-axis[1]

Sources: [1] Touloukian et al. (1970b), [2] Horai and Simmons (1969).

Table 4.2. Comparison of Published Compilations of Thermal Conductivities ($W\,m^{-1}\,K^{-1}$)

Lithology	Sources										
	1[a]	2	3	4	5	6	7	8	9	10	11
Sandstone	7.1	4.2 ± 1.4	3.1 ± 1.3		3.7 ± 1.2	2.8		3.7 ± 1.2			4.7 ± 2.8
Claystone	2.9				2.0						1.8
Mudstone	2.9							2.0 ± 0.4			1.9 ± 0.4
Shale	2.9	1.5 ± 0.5	1.4 ± 0.4		2.1 ± 0.4	1.4		2.1 ± 0.4			1.8 ± 1.2
Kaolinite									1.8 ± 0.3		
Glauconite									0.5 ± 0.2		
Siltstone	2.9	2.7 ± 0.9	3.2 ± 1.3		2.7 ± 0.2	2.7 ± 0.9		2.7 ± 0.2			
Limestone	3.1	2.9 ± 0.9	2.4 ± 0.9	2.21	2.8 ± 0.4		3.4 ± 3.0	2.8 ± 0.3			2.5 ± 0.6
Marl	3.2	2.1 ± 0.7	3.0 ± 1.1		2.7 ± 0.5						2.4 ± 0.5
Dolomite		5.0 ± 0.6	3.1 ± 1.4		4.7 ± 0.8		4.8 ± 1.5	4.7 ± 1.1			3.7 ± 1.8
Halite		5.5 ± 1.8	5.7 ± 1.0		5.4 ± 1.0			5.4 ± 0.3			5.9
Chert		4.2 ± 1.5	1.4 ± 0.5		1.4 ± 0.5						
Quartzite				6.0			5.0 ± 2.4	5.9 ± 0.8		3.5 ± 0.4	5.6 ± 1.9
Granite							3.4 ± 1.2			3.5 ± 0.4	2.8 ± 0.6
Basalt	1.8			1.7			1.7 ± 0.6			2.0 ± 0.2	1.5
Tuff				1.7 ± 0.3							
Conglomerate		2.4 ± 0.8	3.2 ± 1.8		2.1 ± 1.0						
Coal		0.3 ± 0.1	0.2 ± 0.2	0.2 ± 0.04	0.2 ± 0.1	0.3 ± 0.1					
Loose sand				2.44 ± 0.8							
Typical sediment			2.3 ± 2.0								

[a]Matrix conductivity values, only representing bulk conductivity when $\phi = 0$.

Sources: 1 = Beardsmore (1996), 2 = Majorowicz and Jessop (1981), 3 = Beach, Jones and Majorowicz (1987), 4 = Raznjevic (1976), 5 = Reiter and Jessop (1985), 6 = Taylor, Judge and Allen (1986), 7 = Roy et al. (1981), 8 = Reiter and Tovar (1982), 9 = Touloukian et al. (1970b), 10 = Drury (1986), 11 = Barker (1996).

Table 4.3. Thermal Conductivity of Common Pore Fluids

Fluid	Conductivity ($W\,m^{-1}\,K^{-1}$)	Source
Water		
Fresh	$-7.42 \times 10^{-6} \times T^2 + 5.99 \times 10^{-3} \times T - 0.522$	1
Saline	$-7.42 \times 10^{-6} \times T^2 + 5.99 \times 10^{-3} \times T - 0.5$	
Hydrocarbon		
Oil	0.1	1
Gas	$0.000143 \times T - 0.0159$	1
Air	0.023	2

Note: T = absolute temperature in the range 295–450 K.

Source: 1 = Touloukian et al. (1970a), 2 = Giancoli (1984).

Thermal conductivity and lithology

Thermal conductivity is closely related to lithology. Whenever possible, each lithology within the region of interest should be sampled for thermal conductivity measurements. Ideally, these should use a steady-state method on core samples, but drill cuttings and transient methods can also be used with reduced accuracy and precision.

The important parameter to be determined is thermal conductivity of the matrix of the rock.

Apart from temperature, thermal conductivity also varies with pressure, saturation, pore fluid, dominant mineral phase, and anisotropy of different rock types.

Thermal conductivity and compaction

Compaction reduces the amount of pore fluid in a rock and generally increases the bulk thermal conductivity.

Different lithologies compact at different rates.

In the absence of porosity logs, compaction models must be constructed for each lithology under investigation.

Once porosity is known, it can be combined with matrix thermal conductivity to determine the bulk thermal conductivity of each formation. This thermal conductivity must then be corrected for in situ temperature conditions.

Thermal conductivity measurement

Thermal conductivity is measured on samples from wells (cores or large cuttings or cell with small cuttings) or in samples from rock outcrops.

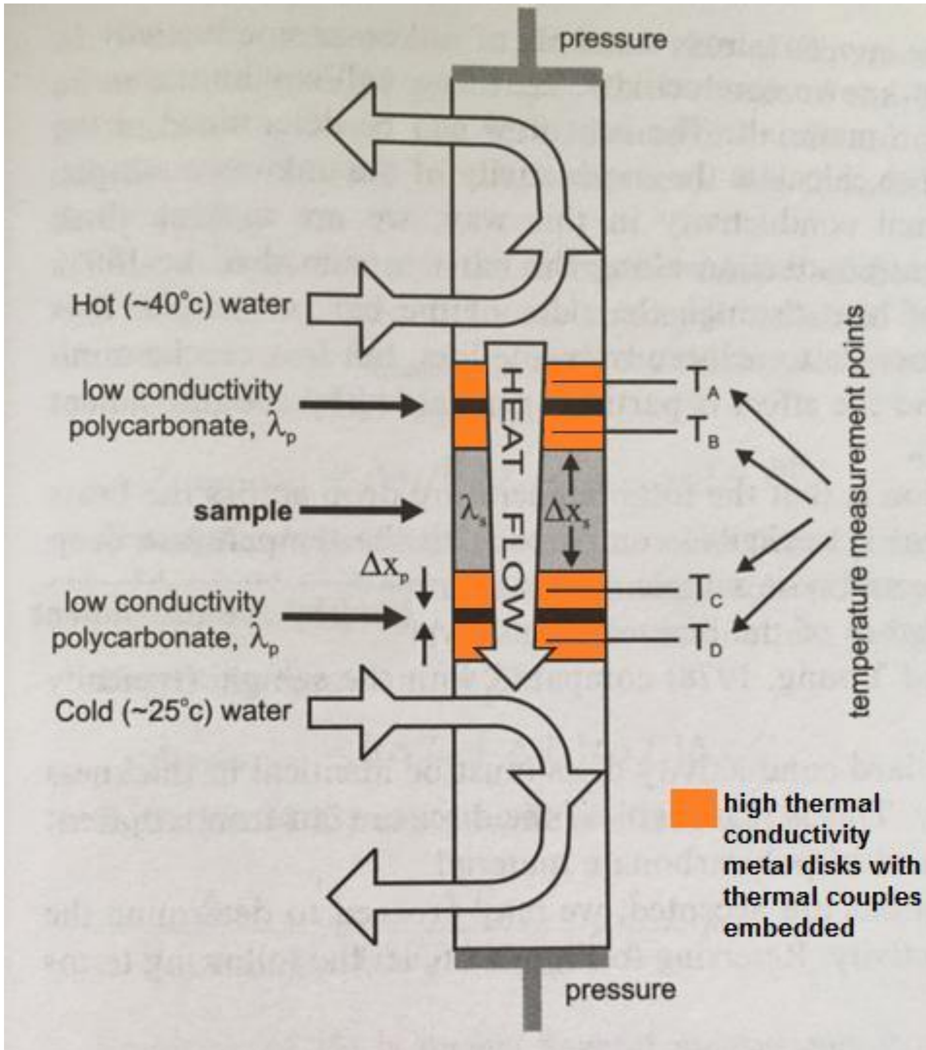
There are two main methods to measure thermal conductivity:

- “Divided Bar” measures K in steady state mode
- Transient methods, with a linear source of heat (needle within the sample) or a planar source of heat (hot plate on the sample surface) in transient state

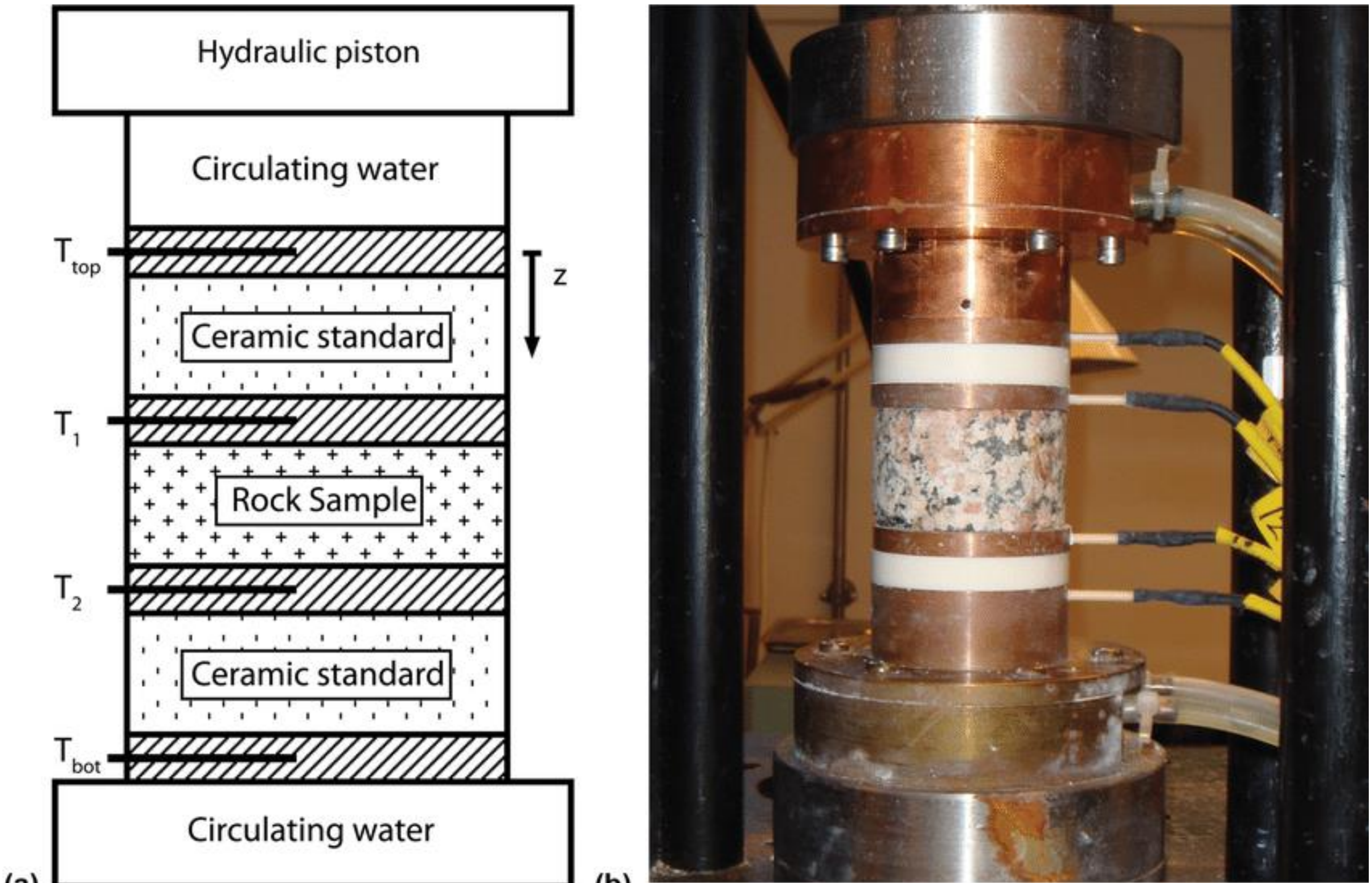
Steady-State Methods

Consolidated rocks

Divided-bar apparatus



Divided-bar apparatus



Divided-bar apparatus

Sample preparation: they should be cut as cylinders with the bases polished and saturated with water

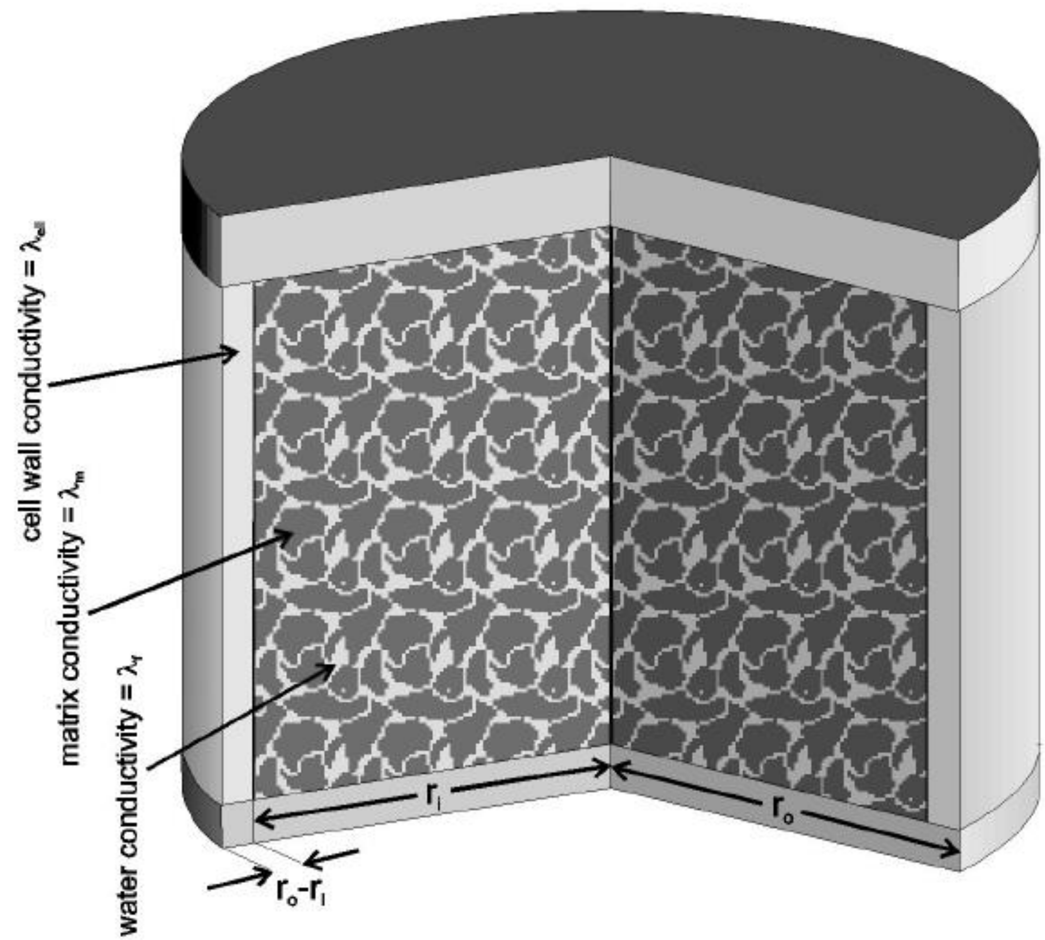
The heat flow density is constant along the pile

$$Q = K_s \cdot \frac{\Delta T_s}{\Delta x_s} = K_r \cdot \frac{\Delta T_r}{\Delta x_r}$$

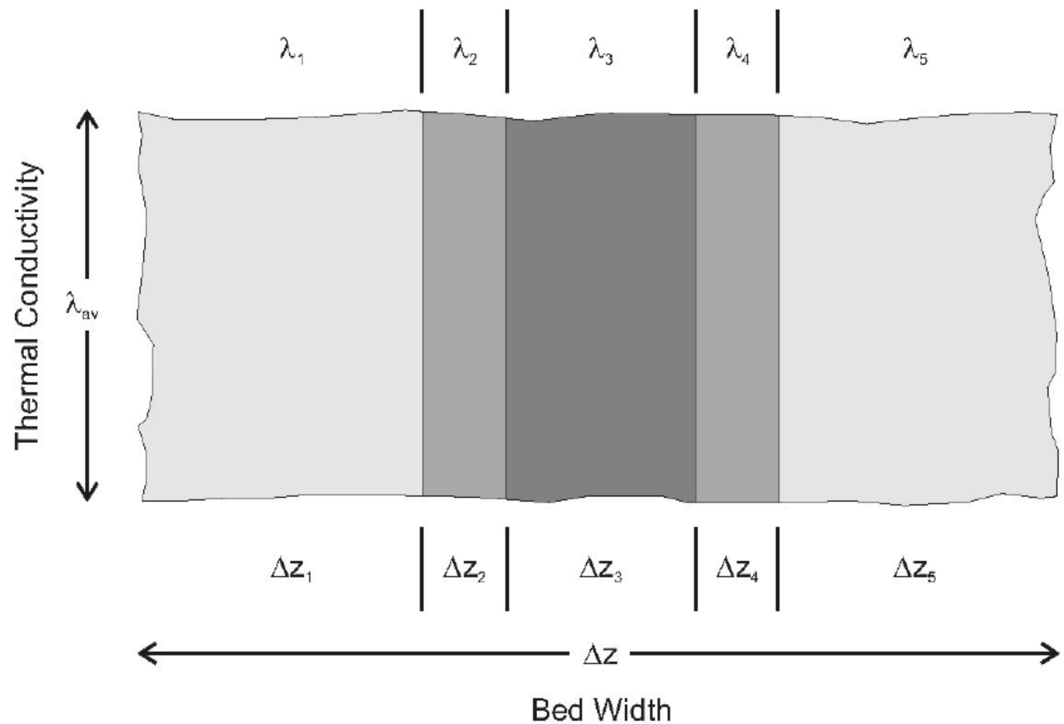
Solving for the thermal conductivity of the rock sample we get

$$K_r = K_s \cdot \frac{\Delta T_s}{\Delta x_s} \cdot \frac{\Delta x_r}{\Delta T_r}$$

Cell for cuttings or unconsolidated material



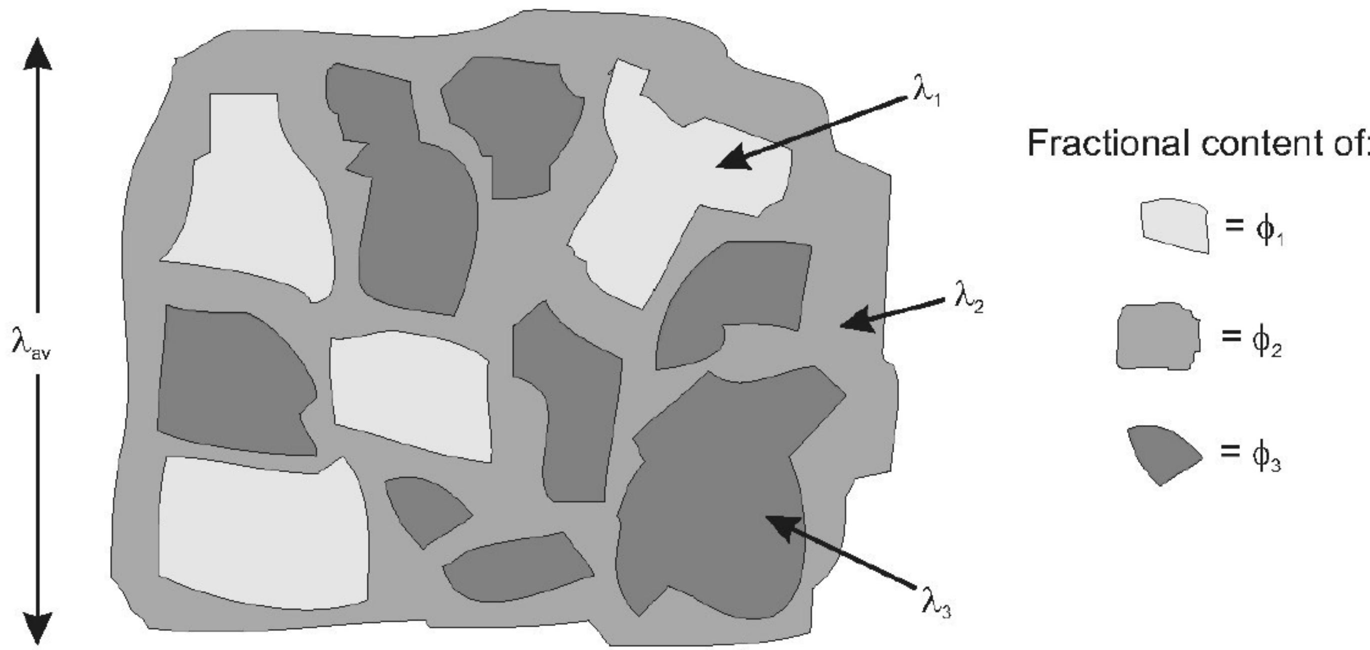
Cell for cuttings or unconsolidated material



Heat flow parallel to the bedding: apply the aritmetic mean

$$\lambda_{total} = \frac{\sum_{i=1}^{n} (\Delta z_i \cdot \lambda_i)}{\Delta z}$$

Cell for cuttings or unconsolidated material

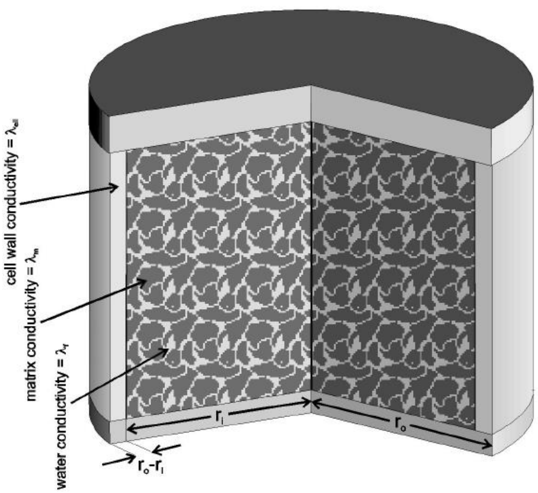


Square-root mean (Roy, Beck and Touloukian (1981))

$$\sqrt{\lambda_B} = \sum_{i=1}^{n} \emptyset_i \cdot \sqrt{\lambda_i}$$

Cell for cuttings or unconsolidated material

$$\lambda_{total} = \lambda_{\mathrm{ag}} \cdot \frac{\pi(\mathrm{r_i})^2}{\pi(\mathrm{r_0})^2} + \lambda_{\mathrm{cell}} \cdot \frac{\pi(r_0)^2 - \pi(r_i)^2}{\pi(\mathrm{r_0})^2}$$



$$\lambda_{\mathrm{ag}} = (\lambda_{total} - \lambda_{\mathrm{cell}})\left(\frac{r_0}{r_i}\right) + \lambda_{\mathrm{cell}}$$

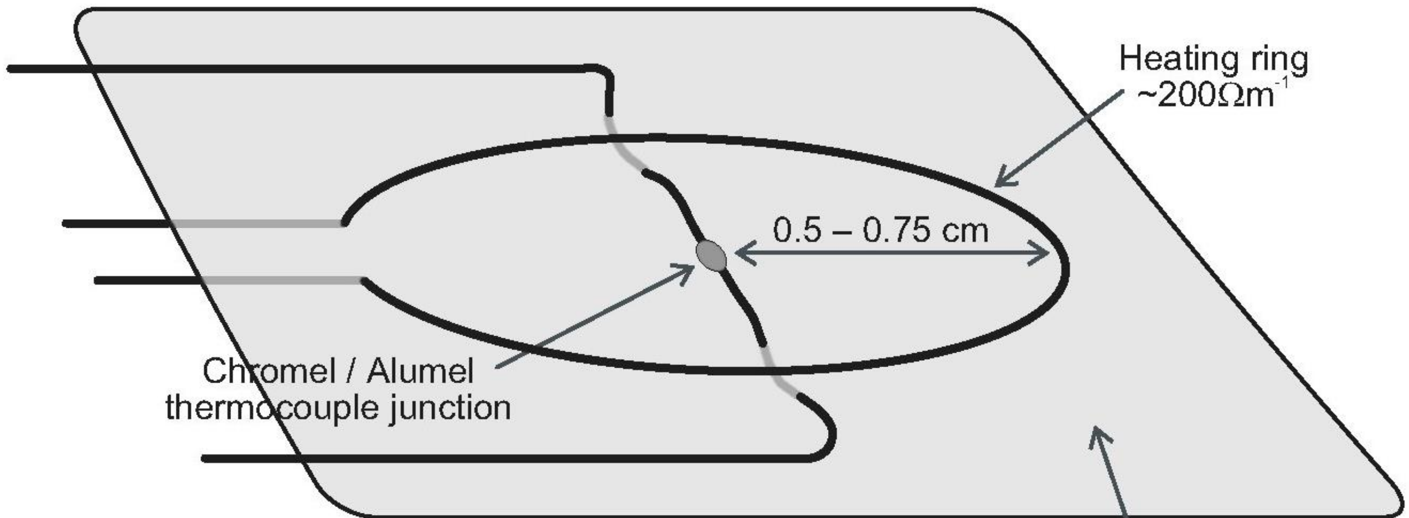
$$\lambda_{\mathrm{m}} = \left[\frac{\sqrt{\lambda_{\mathrm{ag}}} - \emptyset\sqrt{\lambda_{\mathrm{w}}}}{(1-\emptyset)}\right]^2$$

Transient methods

Unconsolidated rocks

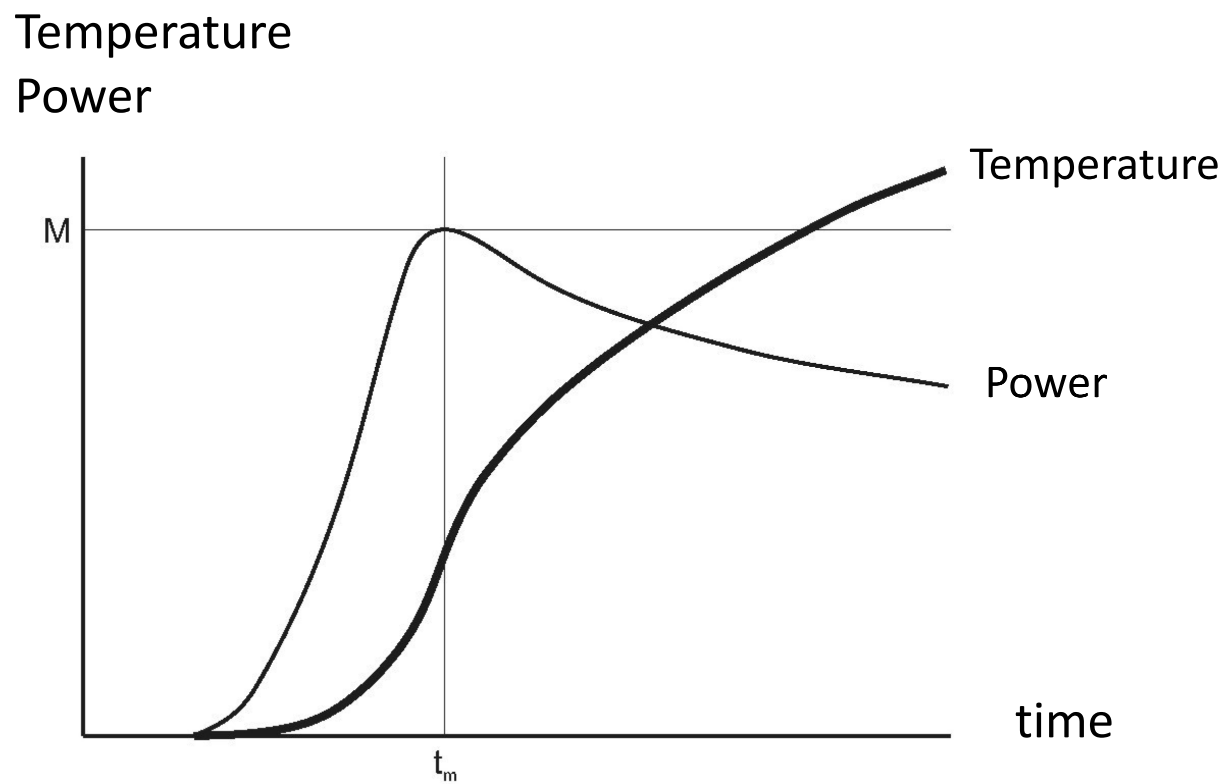
Thermal conductivity meter

Line-source methods



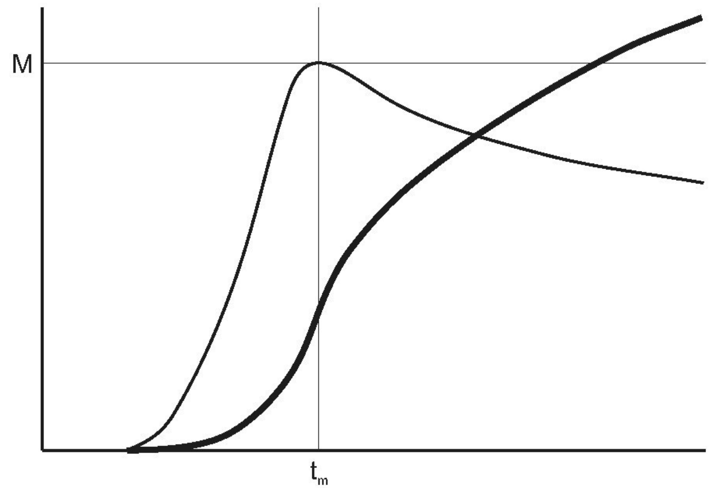
$$K = 0.0771 \cdot \frac{Q}{M \cdot t_m}$$

Line-source methods



Temperature vs time (thick line) and heating rate vs time (thin line) at a position offset from the line heat source

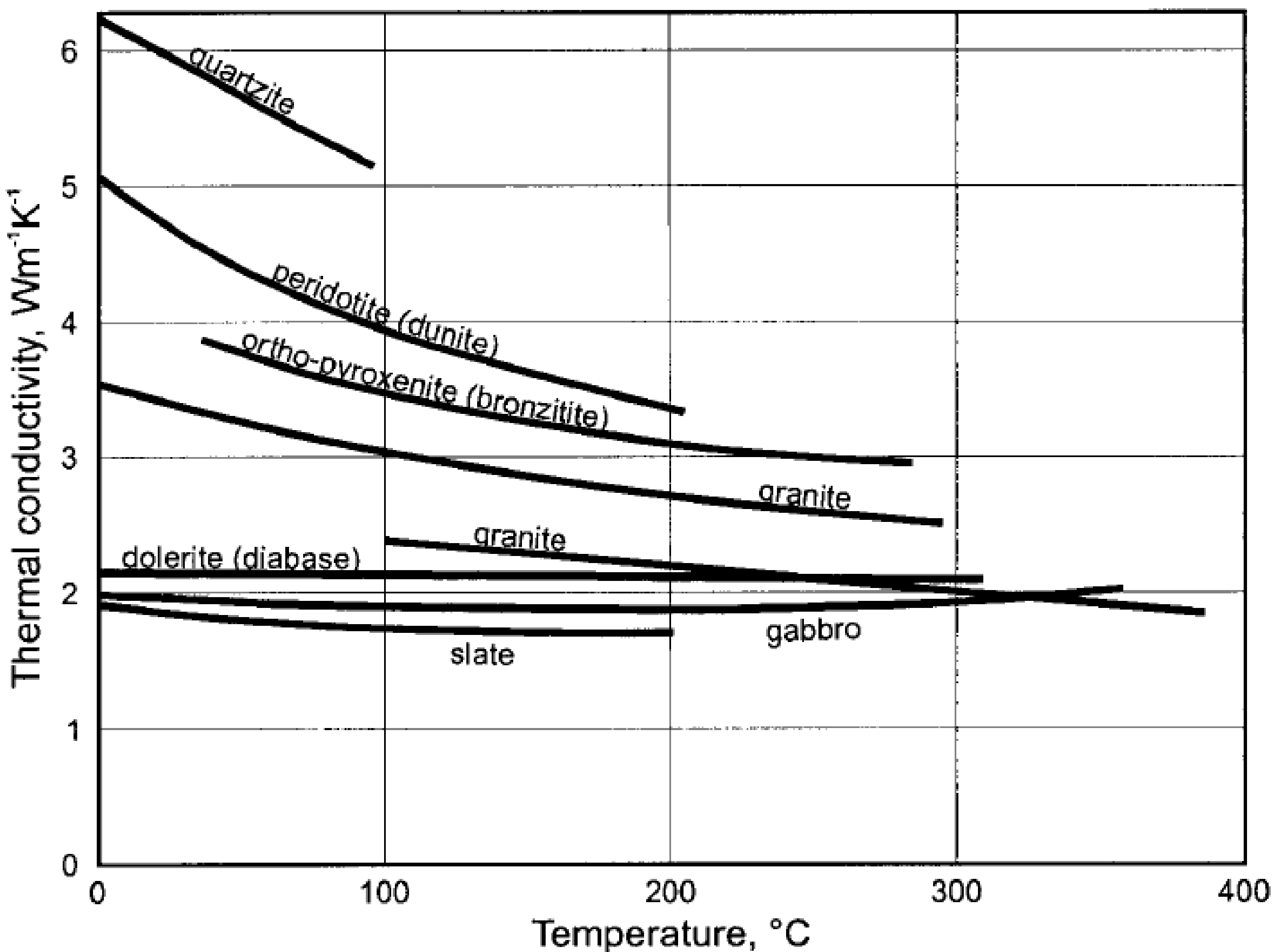
Line-source methods



$$K = 0.0771 \cdot \frac{Q}{M \cdot t_m}$$

where Q (Wm^{-1}) is the heat liberated per unit length of wire per unit time which is equal to $Q_T/2\pi r$ and Q_T (W) is the total amount of heat produced per unit time by a ring of radius r.

Thermal conductivity vs. temperature



(Birch and Clark, 1940).

Thermal conductivity vs. temperature

$$K(T) = [K(22)] \cdot \left(\frac{295}{T+273}\right)$$

Correia *et al.*, 1989

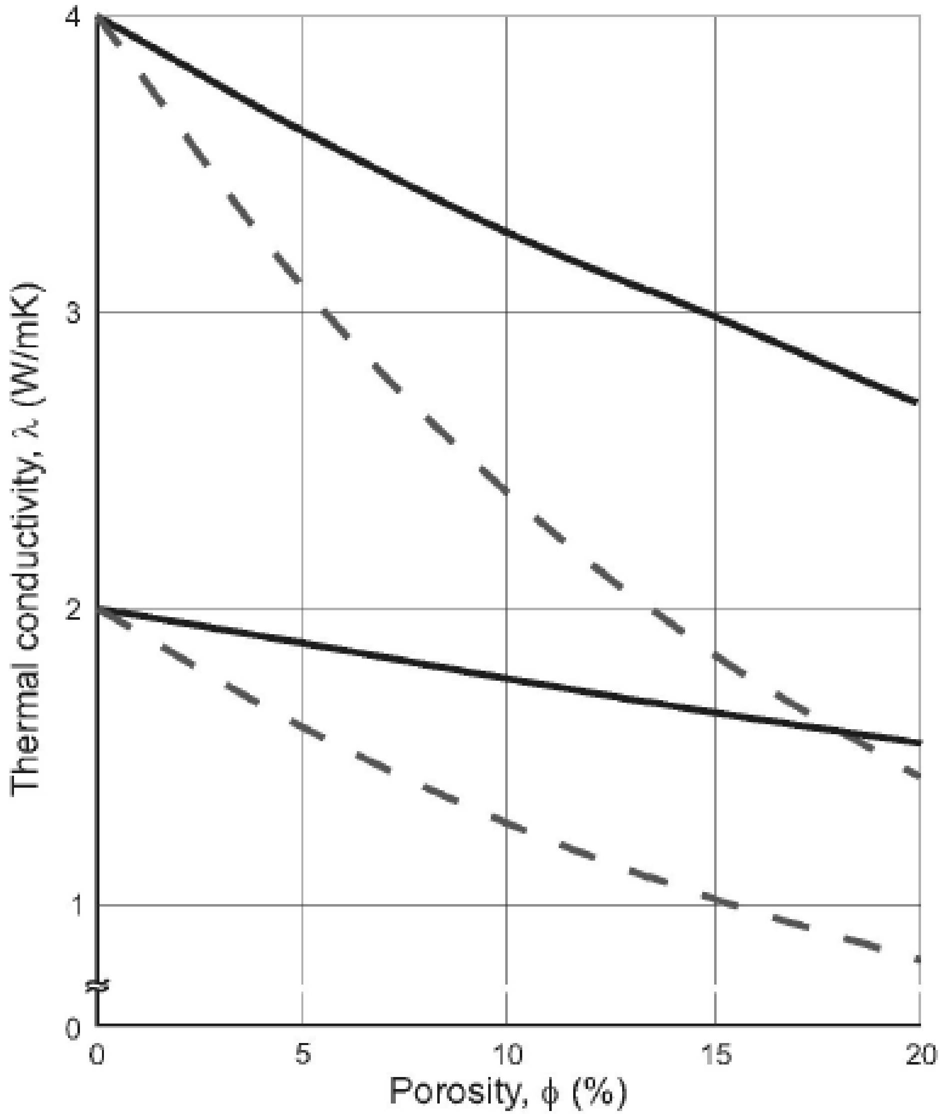
$$\lambda = A + \frac{B}{350 + T}$$

Zoth and Haenel, 1988

Table 8.13. Values for the constants A and B in eq. (8.33) for different rock types; data: [1988Zot].

rock type	T (°C)	A ($W\ m^{-1}\ K^{-1}$)	B ($W\ m^{-1}$)
(1) rock salt	-20 – 0	-2.11	2960
(2) limestones	0 – 500	0.13	1073
(3) metamorphic rocks	0 – 1200	0.75	705
(4) acid rocks	0 – 1400	0.64	807
(5) basic rocks	50 – 1100	1.18	474
(6) ultra-basic rocks	20 – 1400	0.73	1293
(7) rock types (2)-(5)	0 – 800	0.70	770

Thermal conductivity vs. porosity



Conductivity versus porosity for two rocks with λ = 2 W/mK and with λ = *4* W/mK. Solid line assumes water-filled pores, dashed line air-filled pores. Modified after Jessop (1990).

Thermal conductivity vs. porosity

There are at least three popular models describing the compaction of sediments with increasing burial.

That of Sclater and Christie (1980) who state that porosity decays exponentially with depth of burial, z:

$$\emptyset = \emptyset_0 \cdot e^{(-Az)}$$

where $\emptyset_0$ is the porosity of the sediments at the time of deposition and A is constant called compaction coefficient (A~2.7 x 10^{-4}).

That of Falvey and Middleton (1981) who proposed:

$$\frac{1}{\emptyset} = \frac{1}{\emptyset_0} + Bz$$

where B is a constant compaction coefficient (B~7.8688 x 10^{-4}).

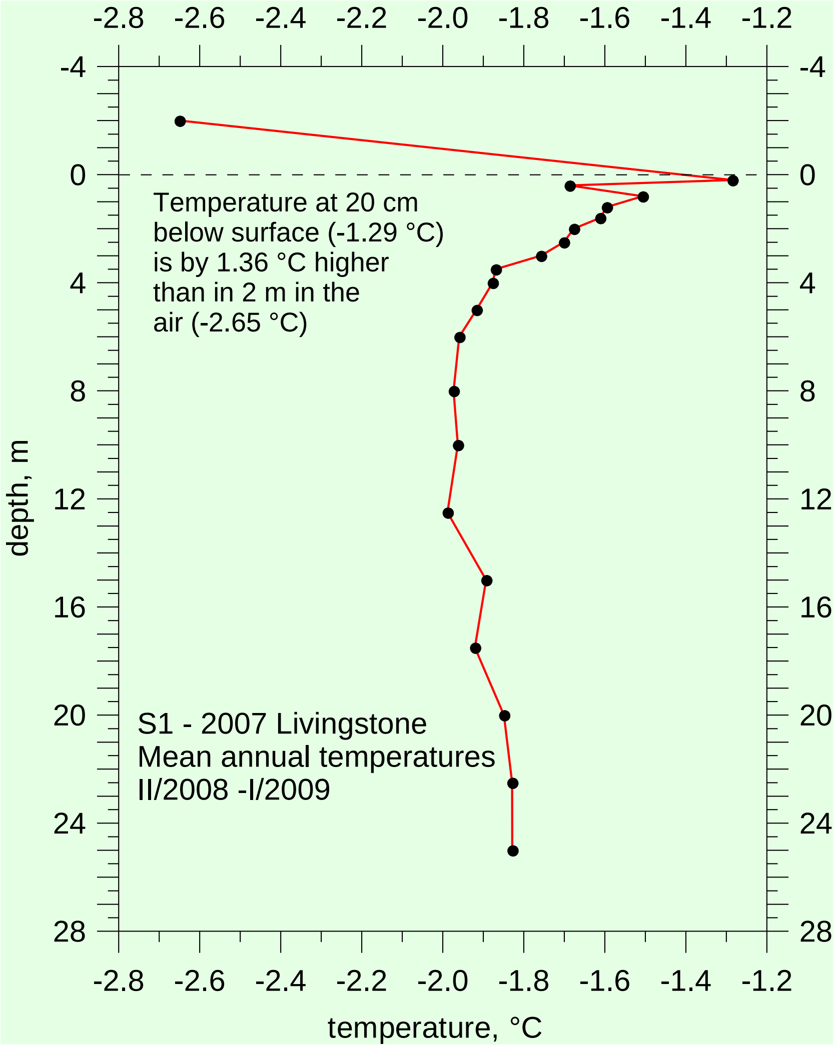
Thermal conductivity vs porosity

And that of Baldwin and Butler (1985) who suggested that while the Sclater and Christie (1980) model is best for sandstone, the compaction of shale and limestone is best explained using a power law model:

$$z = z_{max}(1 - \emptyset)^{C}$$

where z_{max} is the depth at which all fluid is expelled and *C* is a compaction constant (C~6.35).

Antarctica


-2.8 -2.6 -2.4 -2.2 -2.0 -1.8 -1.6 -1.4 -1.2
-4
0
4
8
12
16
20
24
28
depth, m
Temperature at 20 cm
below surface (-1.29 °C)
is by 1.36 °C higher
than in 2 m in the
air (-2.65 °C)
S1 - 2007 Livingstone
Mean annual temperatures
II/2008 -I/2009
temperature, °C

Reminder

$$\frac{\partial T}{\partial t} = \alpha \frac{\partial^2 T}{\partial z^2}$$

$$T(z = 0, t) = T_0 + A_0 \sin(\omega t - \varepsilon_0)$$

$$T(z, t) = T_0 + z \cdot \text{grad}\, T + A_0 \cdot e^{-z/d} \cdot \sin(\omega t - z/d - \varepsilon_0)$$

where grad T is the geothermal gradient and $d = \sqrt{2\alpha/\omega}$

$$\alpha = \frac{\omega \cdot (z_2 - z_1)^2}{2 \cdot \left(ln(A_1/A_2)\right)^2}$$

Periodic temperature change at the Earth's surface

The heat conduction equation can be integrated to give the temperature distribution as a function of time (t) and depth (z)

$$\frac{\partial T}{\partial t} = \alpha \cdot \frac{\partial^2 T}{\partial z^2}$$

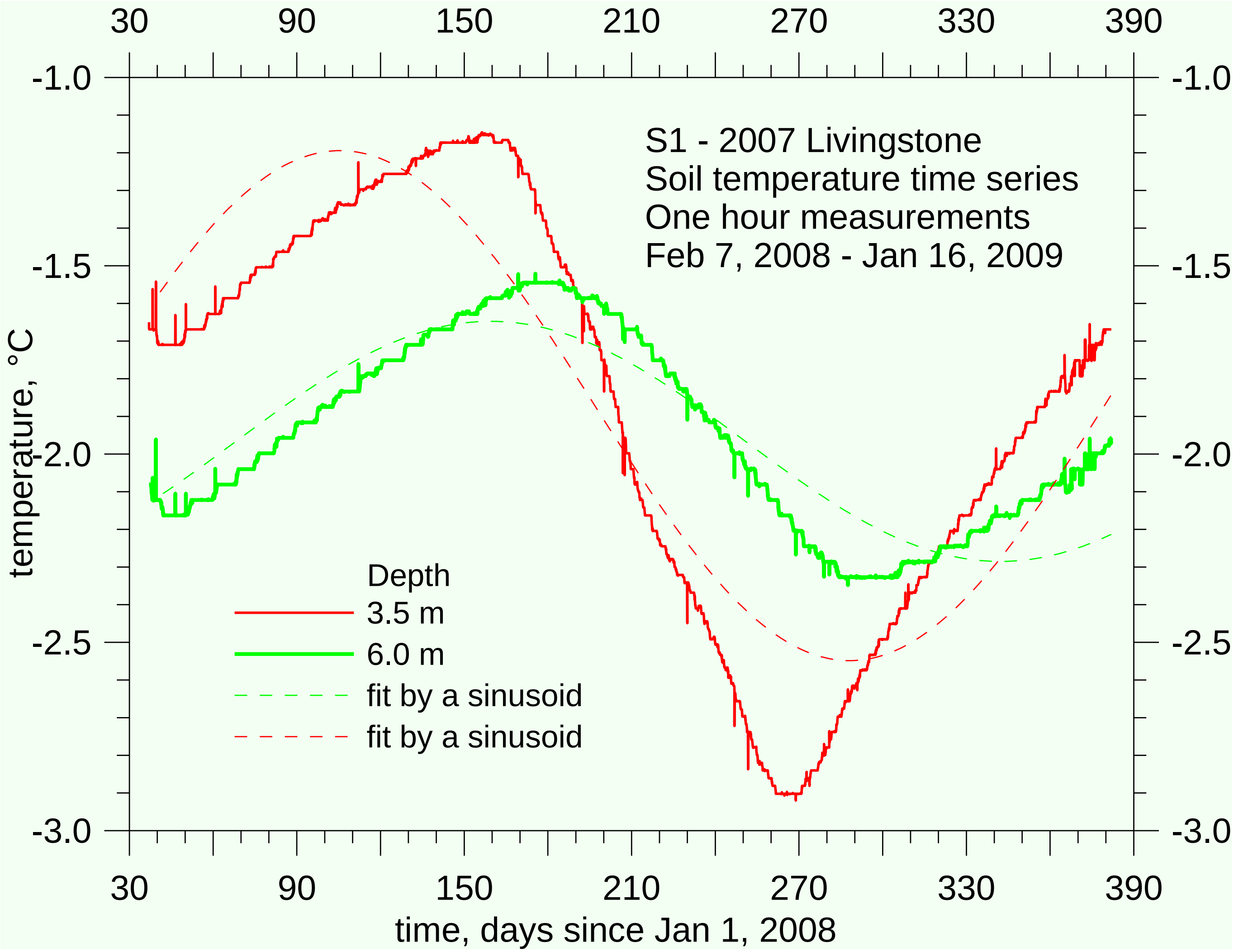
$$T(0,t) = T_0 \cdot e^{i\omega t}$$

$$T(z,t) \rightarrow 0 \quad \text{for} \quad z \rightarrow \infty$$

$$T(z,t) = T_0 \cdot \exp\left(-\sqrt{\frac{\omega \cdot \rho \cdot c_P}{2K}} \cdot z\right) \cdot \exp\left[i\left(\omega \cdot t - \sqrt{\frac{\omega \cdot \rho \cdot c_P}{2K}} \cdot z\right)\right]$$

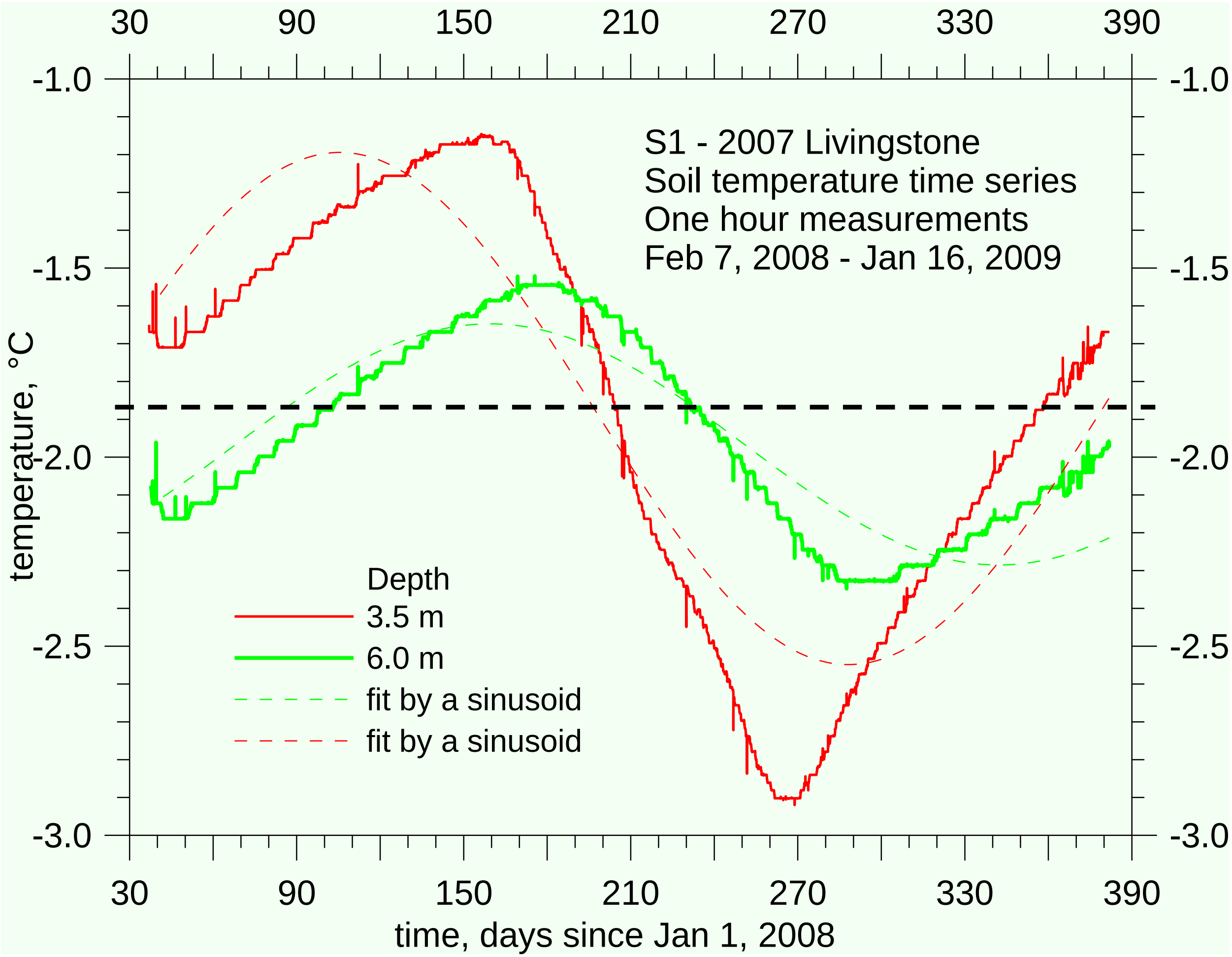
$$L = \sqrt{\frac{2K}{\omega \cdot \rho \cdot c_P}} \qquad \Phi = \sqrt{\frac{\omega \cdot \rho \cdot c_P}{2K}}$$

Antarctica



30
90
150
210
270
330
390
-1.0
-1.5
-2.0
-2.5
-3.0
S1 - 2007 Livingstone
Soil temperature time series
One hour measurements
Feb 7, 2008 - Jan 16, 2009
Depth
3.5 m
6.0 m
fit by a sinusoid
fit by a sinusoid
temperature, °C
time, days since Jan 1, 2008

Antarctica



$$\frac{\partial T}{\partial t} = \alpha \frac{\partial^2 T}{\partial z^2}$$

$$\alpha = \frac{\omega \cdot (z_2 - z_1)^2}{2 \cdot \left(ln(A_1/A_2)\right)^2}$$

Antarctica

A few results

- The thermal conductivity in Antarctica samples is similar to those measured in other regions of theEarth.

- The thermal diffusivity measured in Antarctic cores collected in the borehole GP1 range between 1.09×10^{-6} and 1.58×10^{-6} m^2/s.

- The thermal diffusivity estimated using temperatures measured in the borehole is 2.2×10^{-6} m^2/s.

IV
Heat Production

Reminder

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K \frac{dT}{dz}$$

Fourier equation

$$\frac{\partial T}{\partial t} = \alpha \cdot \frac{\partial^2 T}{\partial z^2}$$

Heat conduction equation

$$T = T_0 + \frac{Q_0}{K} \cdot z - \frac{A'}{2K} \cdot z^2$$

Heat generation

Radioactivity in the Earth

Heat is generated in rocks principally through the radioactive decay of unstable isotopes that release energy in the form of alpha (α) and beta (β) particles, gamma radiation (ϒ). However, the surrounding rocks absorb the kinetic energy carried by those particles and radiation, thus generating heat.

The rate of radiogenic heat generation within rocks is related to the abundance of isotopes in them and determines the rate of heat production. Approximately 98% of geothermal radiogenic heat arises from the decay of uranium (U^{238}), thorium (Th^{232}) and potassium (K^{40}). The energy released by the decay of the uranium is considerably greater than the other two.

Heat production

To determine the heat production of a rock it is necessary to know its uranium, thorium and potassium concentrations which are generally obtained using a gamma-ray spectrometer. Heat production can be calculated using the equation

$$A = \rho \cdot (0.097 \cdot C_U + 0.026 \cdot C_{Th} + 0.036 \cdot C_K)$$

where A is in $\mu W/m^3$, ρ is the density of the rock (g/cm^3), and C_U, C_{Th}, and C_K are the concentrations of U and Th (in ppm) and K (in %), respectively.

Heat production

	Concentration [p.p.m. by weight]			Heat production [10^{-11} W kg^{-1}]			
Rock type	U	Th	K	U	Th	K	Total
Granite	4.6	18	33,000	43.8	46.1	11.5	101
Alkali basalt	0.75	2.5	12,000	7.1	6.4	4.2	18
Tholeiitic basalt	0.11	0.4	1,500	1.05	1.02	0.52	2.6
Peridotite, dunite	0.006	0.02	100	0.057	0.051	0.035	0.14
Chondrites	0.015	0.045	900	0.143	0.115	0.313	0.57
Continental crust	1.2	4.5	15,500	11.4	11.5	5.4	28
Mantle	0.025	0.087	70	0.238	0.223	0.024	0.49

Estimates of radioactive heat production in selected rock types, based on heat production rates (from Rybach, 1976, 1988) and isotopic concentrations.

Heat production

Rock Type	No. of Samples	A (μW m^{-3})	V_p (km s^{-1})
Acidic rocks			
Rhyolite	5	2.80 ± 0.28	5.57 ± 0.12
Granite	16	2.82 ± 1.03	5.85 ± 0.16
Granodiorite	11	2.45 ± 1.29	5.88 ± 0.22
		1.0	
Tonalite	8	1.48 ± 0.54	5.89 ± 0.20
Diorite	7	0.88 ± 0.30	6.24 ± 0.25
Basic rocks			
Gabbro	6	0.11 ± 0.13	6.19 ± 0.82
		0.03	
Amphibolite	6	0.37 ± 0.19	6.30 ± 0.30
Ultrabasic rocks			
Hornblendite	3	0.12 ± 0.15	6.91 ± 0.78
Pyroxinite	4	0.06 ± 0.04	7.20 ± 0.33
Peridotite	4	0.01 ± 0.01	7.82 ± 0.12
Serpentinite	6	0.01 ± 0.02	6.49 ± 0.64
Miscellaneous crustal rocks			
Various	23	1.91 ± 1.40	5.78 ± 0.53
Granulite		*0.13*	
Bulk Earth		*0.014*	
Carbonaceous chondrite		*0.01*	
Ordinary chondrite		*0.015*	

Note: Uncertainty in A and V_p = one standard deviation.

Sources: Cermák et al. (1990) and Brown and Mussett (1993).

Heat Generation (A), P wave velocity (V_P) by rock type.

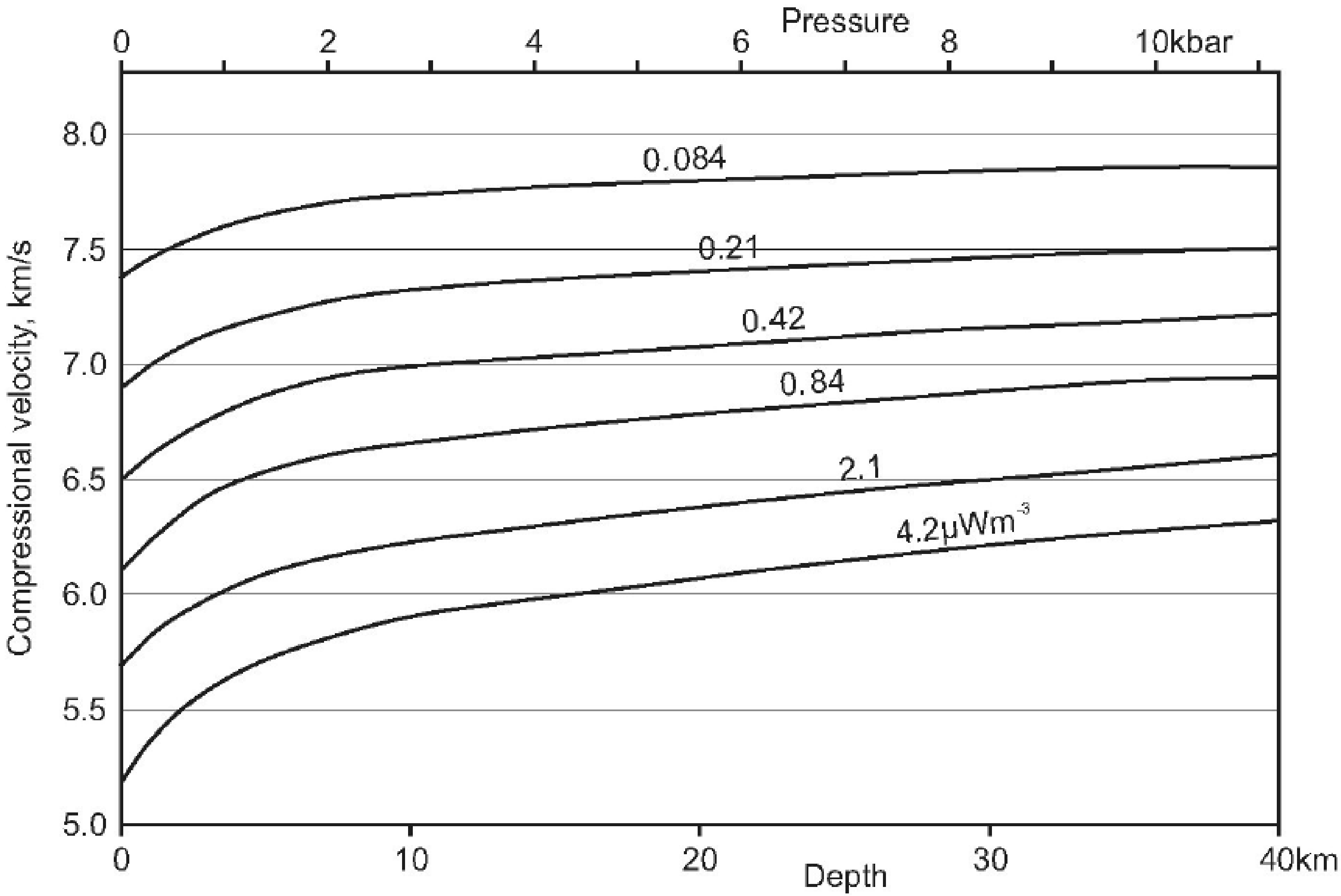
Heat production

Because many times there no samples to determine the concentrations of U, Th and K, several authors have suggested a relationship between heat production and velocity of P waves (Rybach and Bunterbath, 1984). One of those relationships is

$$ln(A) = B - 2.17 \cdot V_P$$

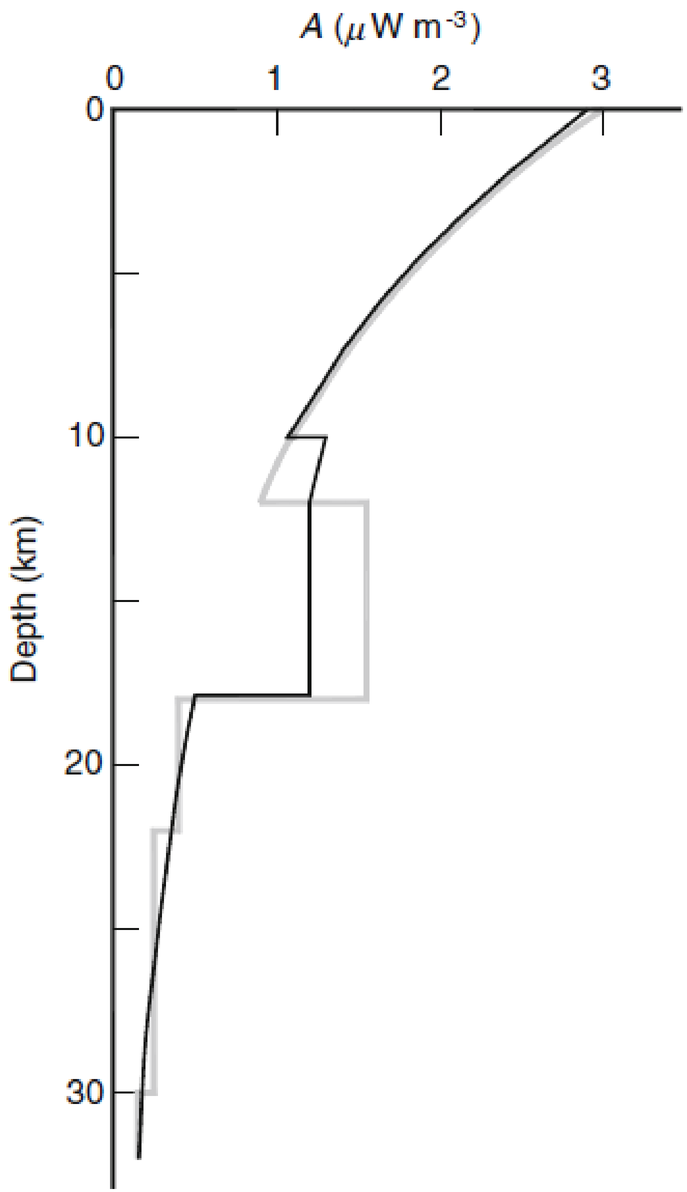
where A in $\mu W/m^3$ is the equivalent heat generation for compressional seismic velocity Vp (km/s), and B is equal to 12.6 for Precambrian formations and 13.7 for Phanerozoic formations which reflects variations in crustal evolution and seismic complexity. Some authors have developed other relationships; other authors consider these relationships not realistic.

Heat production



Correlation between depth/pressure, seismic velocity and heat production. Modified after Rybach (1979)

Heat production



Radiogenic heat production as a function of depth in the Variscan crust based on a compositional model (grey line) and estimated radiogenic heat production using relationship between heat production and velocity of P waves (black line).

$$ln(A) = B - 2.17 \cdot V_P$$

Heat production

Experimental results on volcanic rocks show that the radioelement concentration and, consequently, the radiogenic heat are related to magmatic differentiation processes, thus implying a decrease with depth within the crust. The most widely accepted model is the exponential model by Lachenbruch (1970):

$$A(z) = A_0 e^{-\frac{z}{D}}$$

where A_o (in μW m^{-3}) is the radiogenic heat production at the surface and D (in km) is the rate of heat decrease.

Heat production

Factor D, which ranges from 5 to 15 km, derives from the linear relationship

$$q_0 = q_m + A_0 \cdot D$$

where q_o is the heat flowing out from the Earth's surface and q_m is a constant component of heat flow from the mantle. It is widely accepted that the surface radiogenic heat in the continental areas is of the order of 3 μW m^{-3} as radioactivity measurements of igneous and metamorphic rocks restrict the range of radiogenic heat production to 2.5 – 3.5 μW m^{-3}.

Heat flow density

Reminder

$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K\,\frac{dT}{dz}$$

Fourier equation

$$\frac{\partial T}{\partial t} = \alpha \cdot \frac{\partial^2 T}{\partial z^2}$$

Heat conduction equation

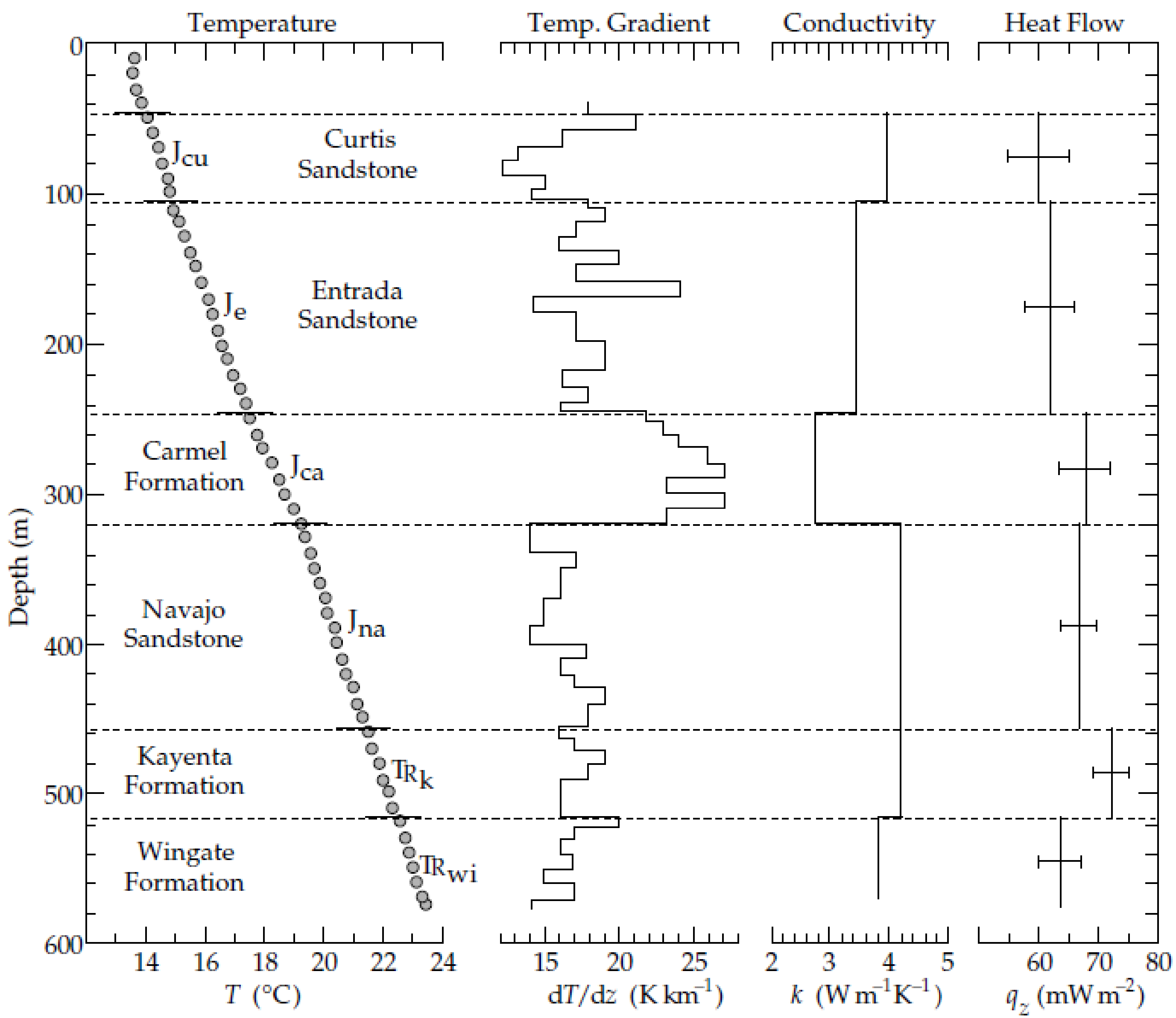
$$T = T_0 + \frac{Q_0}{K} \cdot z - \frac{A'}{2K} \cdot z^2$$

Heat flow density calculation the product method

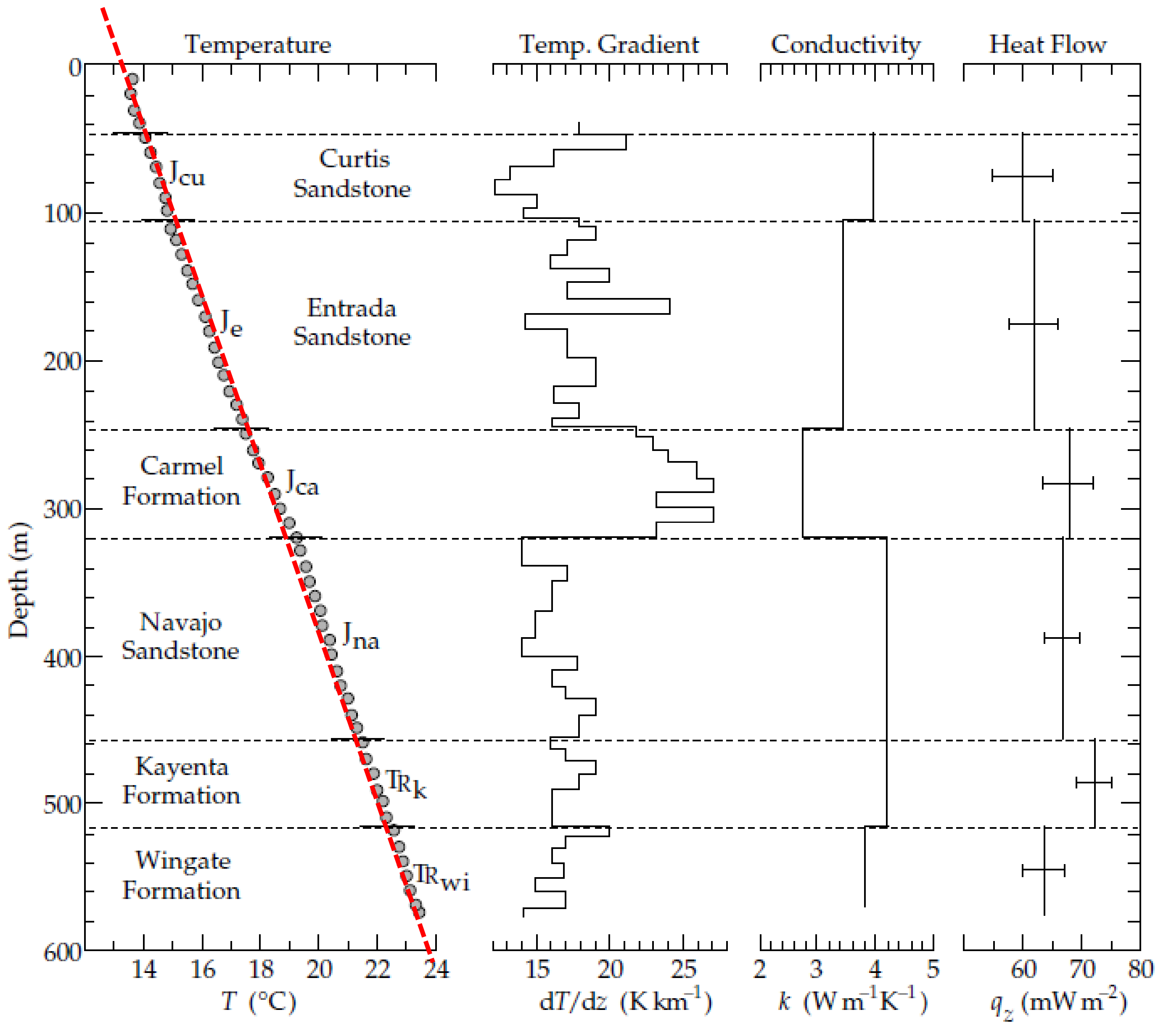
$$q_z = -\frac{\Delta Q}{A \cdot \Delta t} = -K\,\frac{dT}{dz}$$

To calculate the HFD only linear portions of the temperature log (or, equivalently, pure condutive portions of the temperature log) should be used to multiply by the average thermal conductivity for the same depth interval

The product method for HFD calculation



The product method for HFD calculation



Heat flow density calculation the Bullard plot method

A more coherent and demonstrable calculation of heat flow is based on the concept of *thermal resistance.* Thermal resistance (R) is defined as the integral of the reciprocal of thermal conductivity over the depth range z:

$$R = \int \frac{1}{\lambda} dz \qquad R = \sum_{i=1}^{n} \left(\frac{\Delta z_i}{\lambda_i} \right)$$

In SI units it is expressed in meters squared kelvin per watt (m^2KW^{-1}).

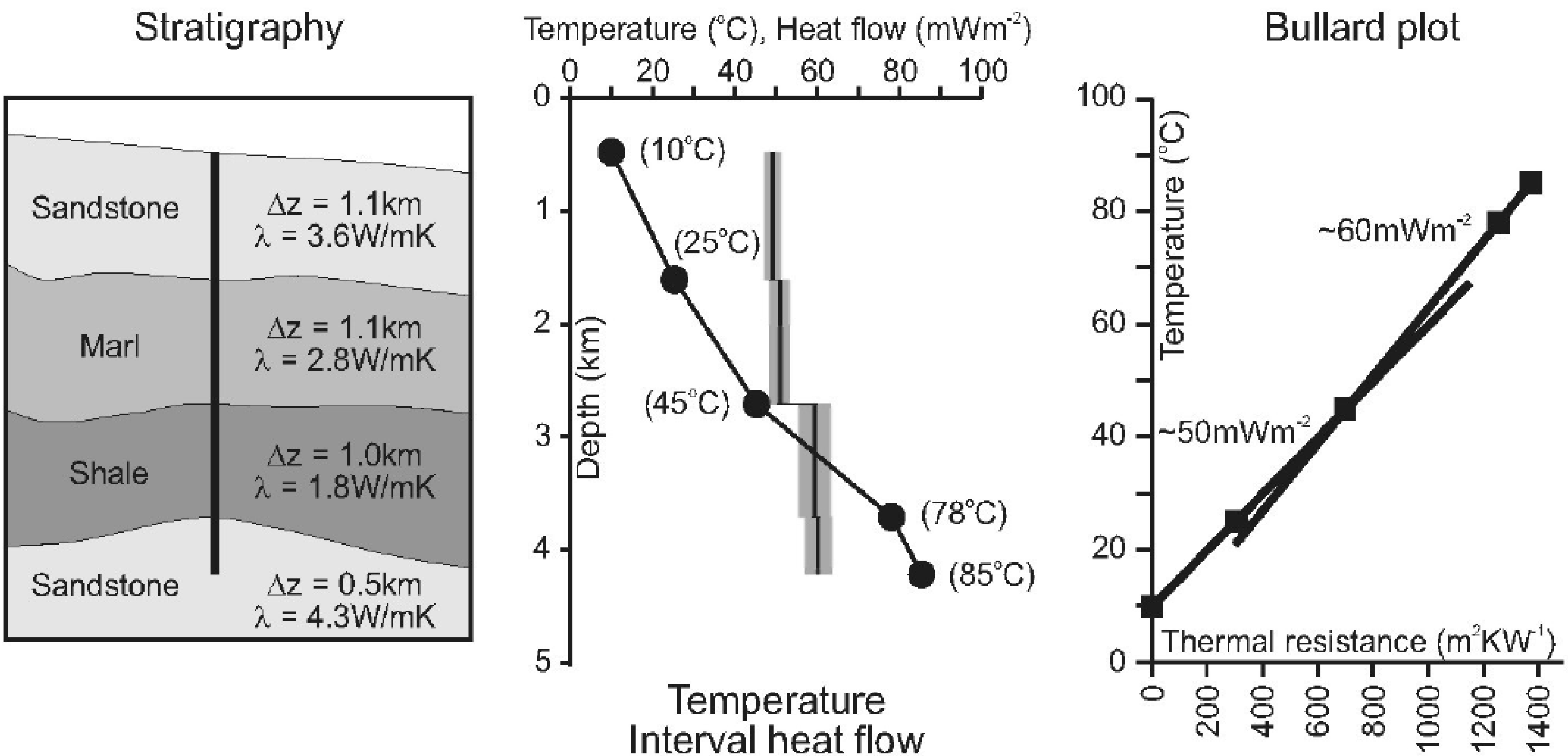
Heat flow density calculation the Bullard plot method

Temperature measured at a shallow level is used to estimate, by extrapolation, the temperature at depth and the heat flow. Assuming a pure conduction (no fluid circulation) the most common equation is (Bullard, 1939):

$$T_z = T_0 + q_0 \sum_{i=1}^{N} (\Delta_{zi} / k_i)$$

where T_z = T at depth *z,* T_0 = average surface *temperature,* q_0 i the heat flow density at the surface (constant), Δ_{zi} is the thickness of the i^{th} unit, k_i is the thermal conductivity of the i^{th} unit. A graph of T_z vs. (Δ_{zi}/k_i) is known as a Bullard plot.

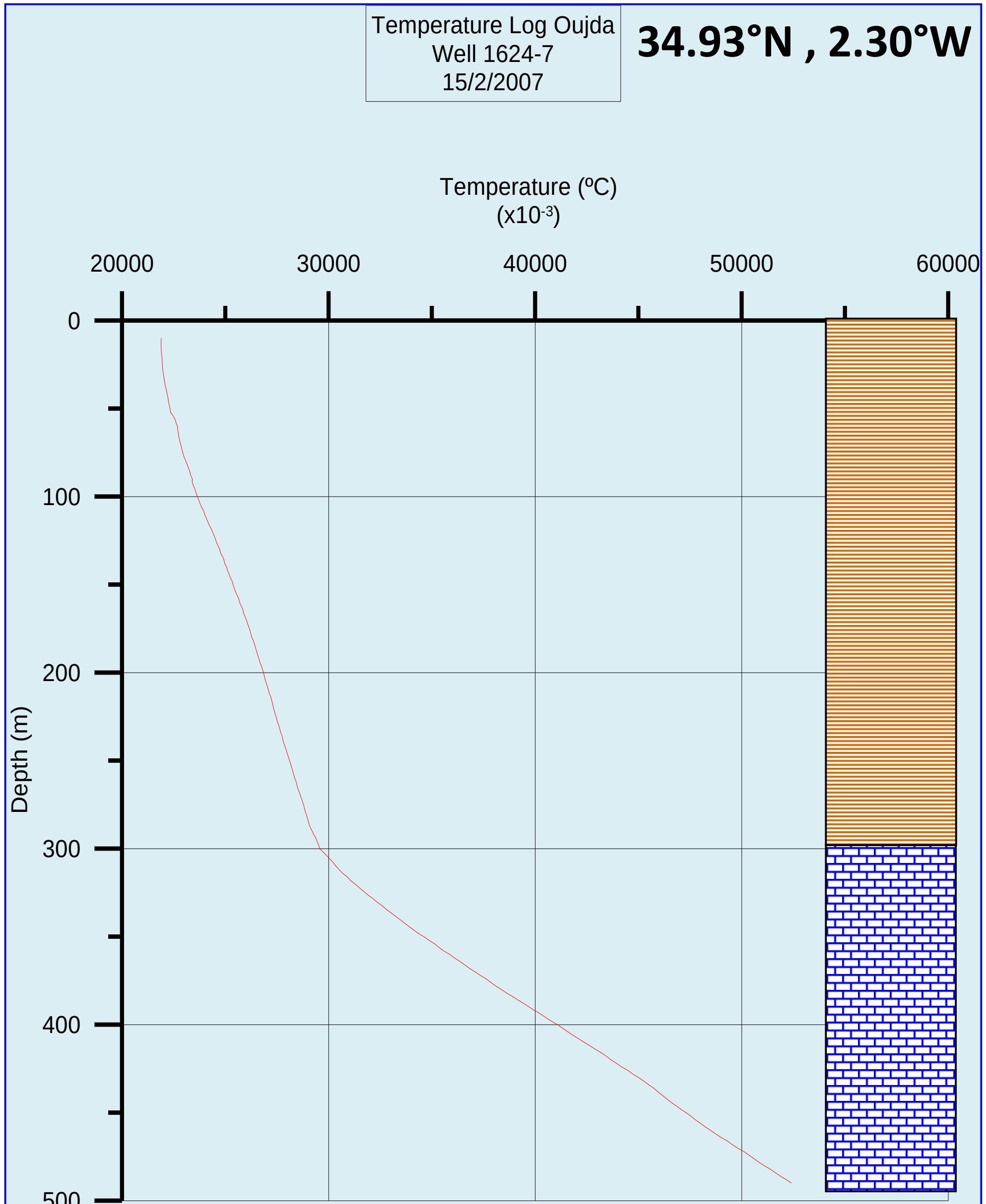
Product method / Bullard plot method



54.7 mW/m^2

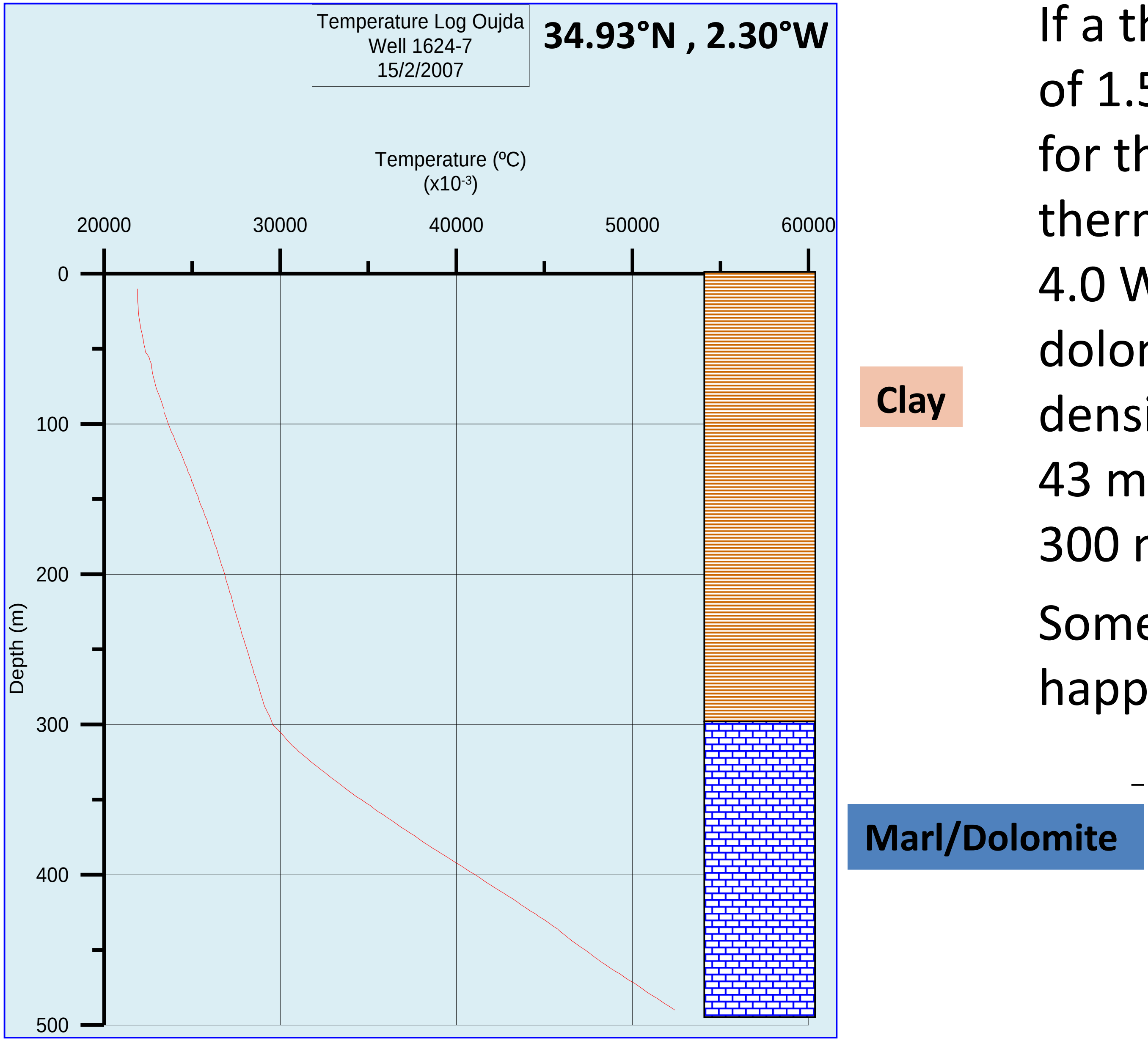
Q1= 49.1 mW/m^2
Q2= 50.9 mW/m^2
Q3= 59.4 mW/m^2
Q4= 60.2 mW/m^2

Morocco again



If the clay thermal conductivity were 1.5 W/mK and the Marl/dolomite 4.0 W/mK what would be the heat flow density in each formations? Give a plausible explanation. What would be the temperature at 1000 m depth?

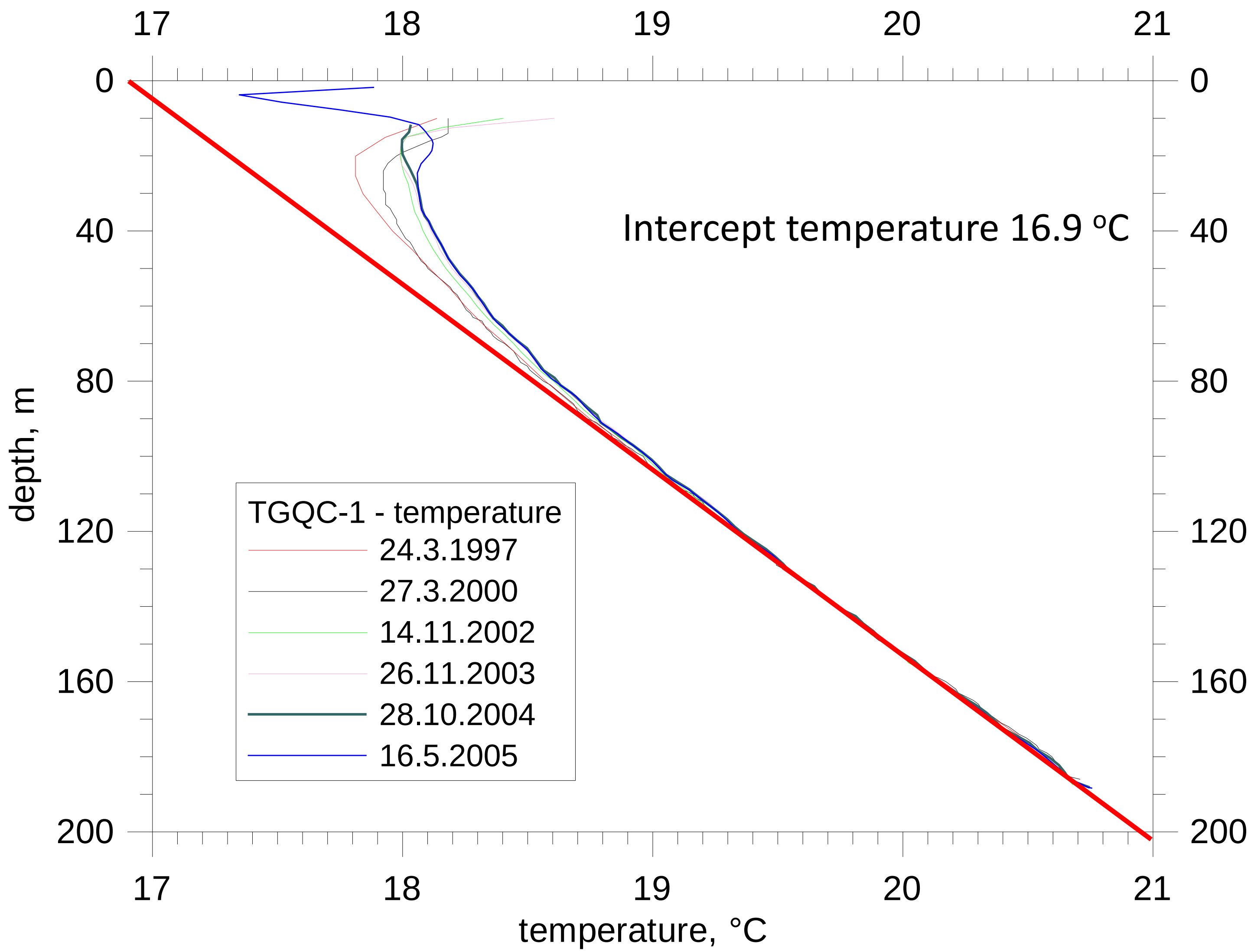
Morocco again



If a thermal conductivity of 1.5 W/mK is assumed for the clay and a thermal conductivity of 4.0 W/mK for the dolomite the heat flow density above 300 m is 43 mW/m^2 while below 300 m is 510 mW/m^2.

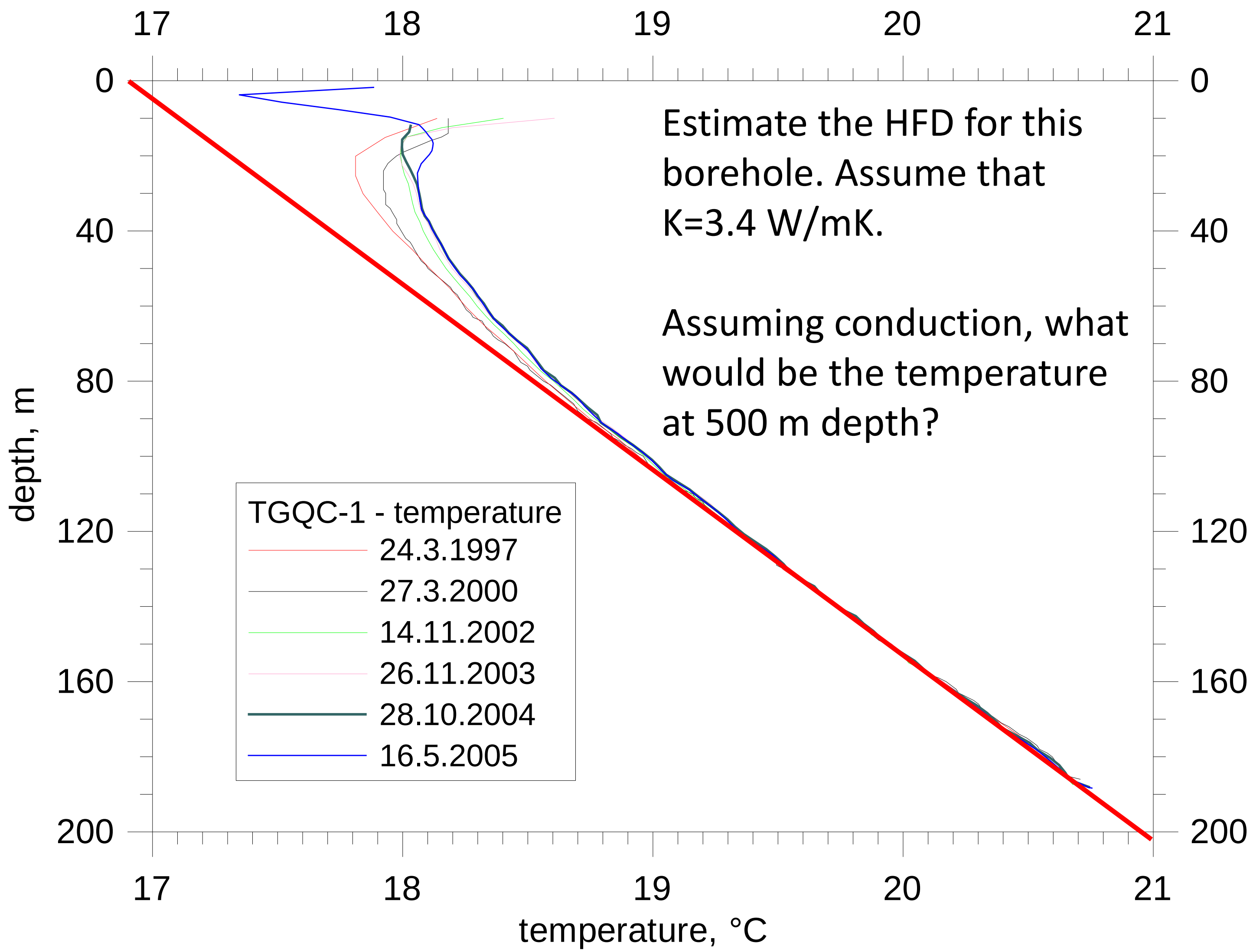
Something hot is happening here!

Temperature logs for different years



Temperature logs for different years

Intercept temperature 16.9 ^{o}C

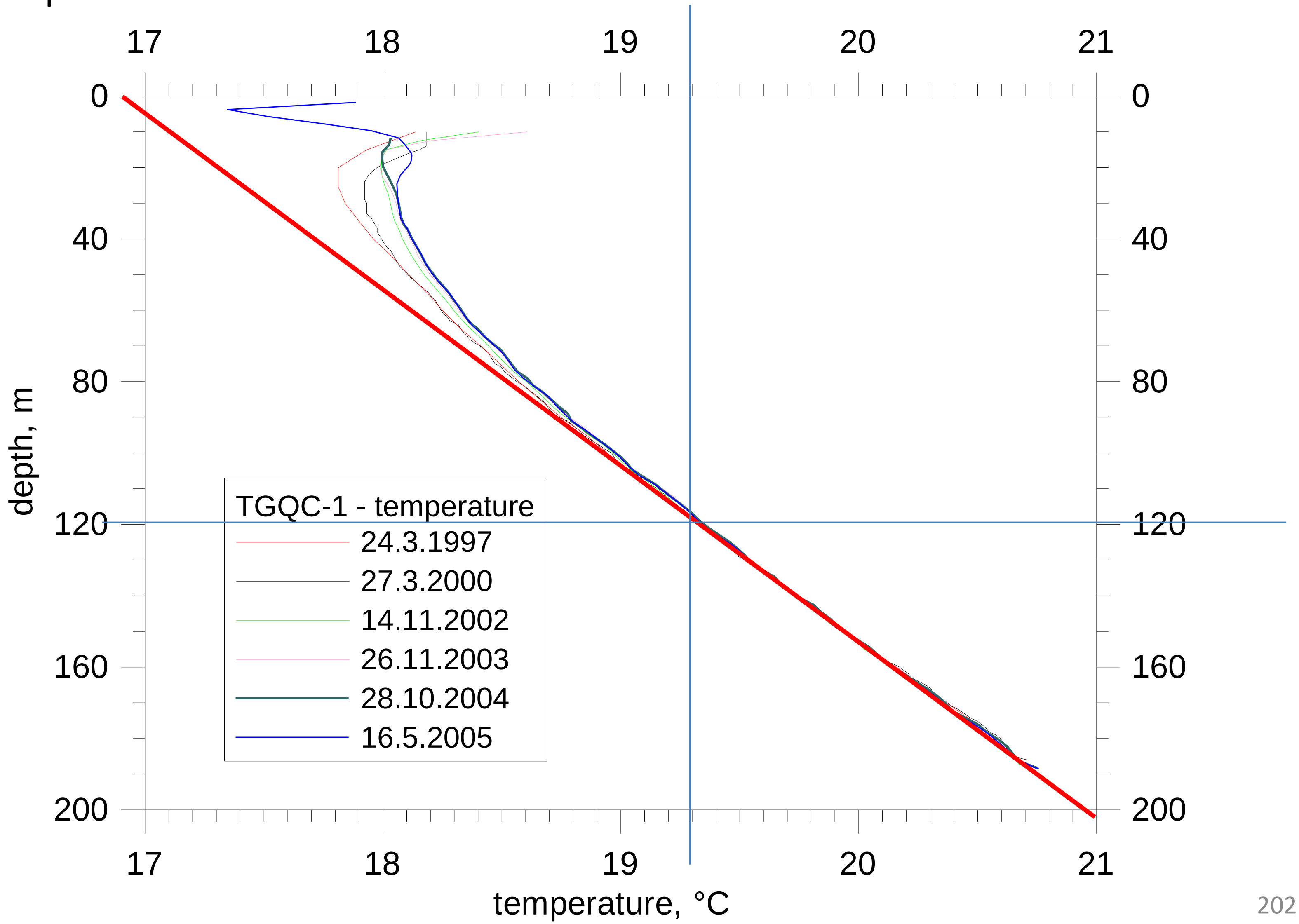


Estimate the HFD for this borehole. Assume that K=3.4 W/mK.

Assuming conduction, what would be the temperature at 500 m depth?

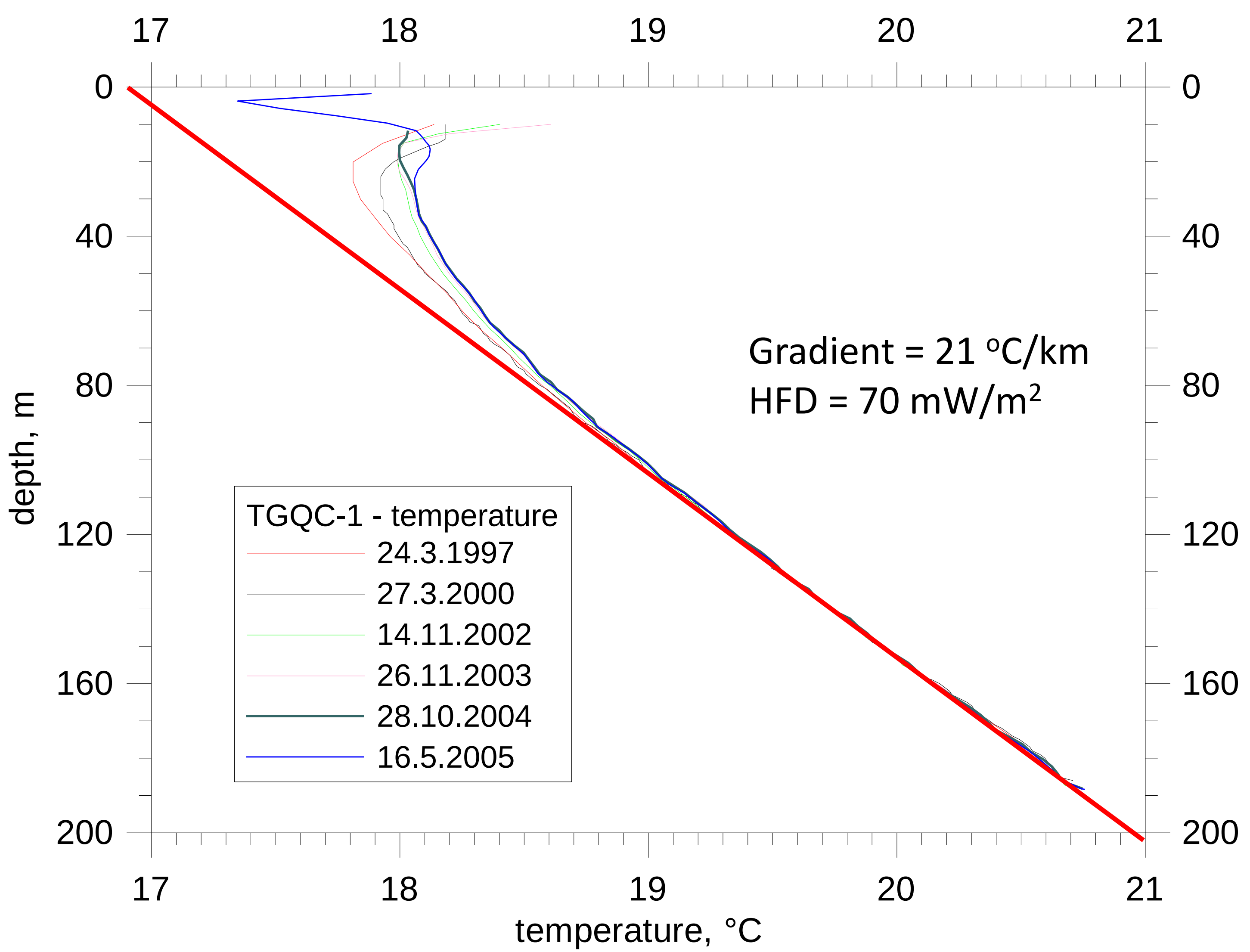
Temperature logs for different years

Intercept temperature 16.9 ^oC



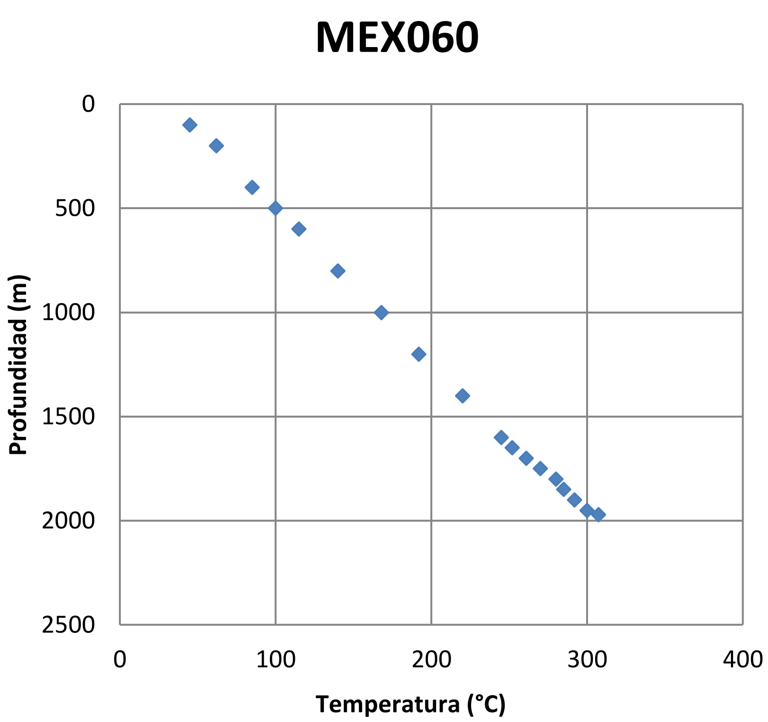
Temperature logs for different years

Intercept temperature 16.9 °C

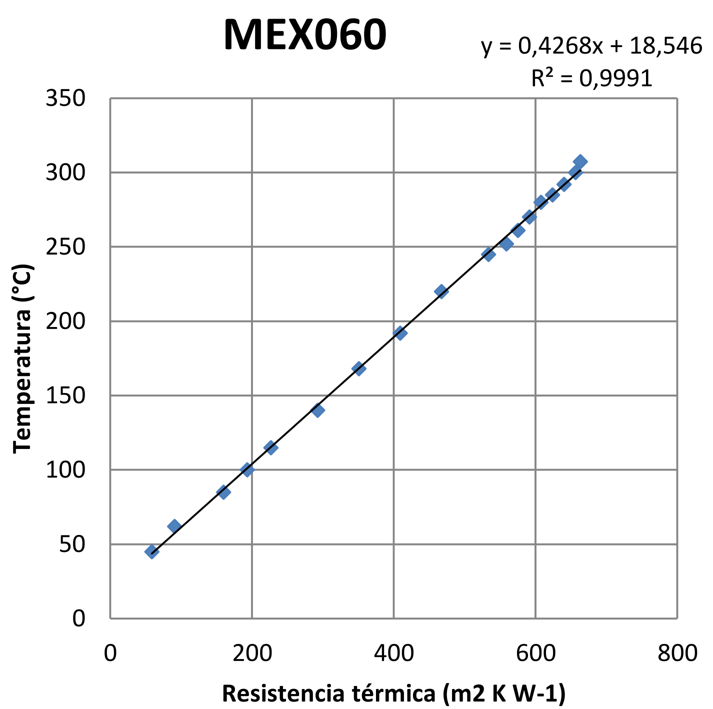


Mexico - Bullard plot method

Temperature Log



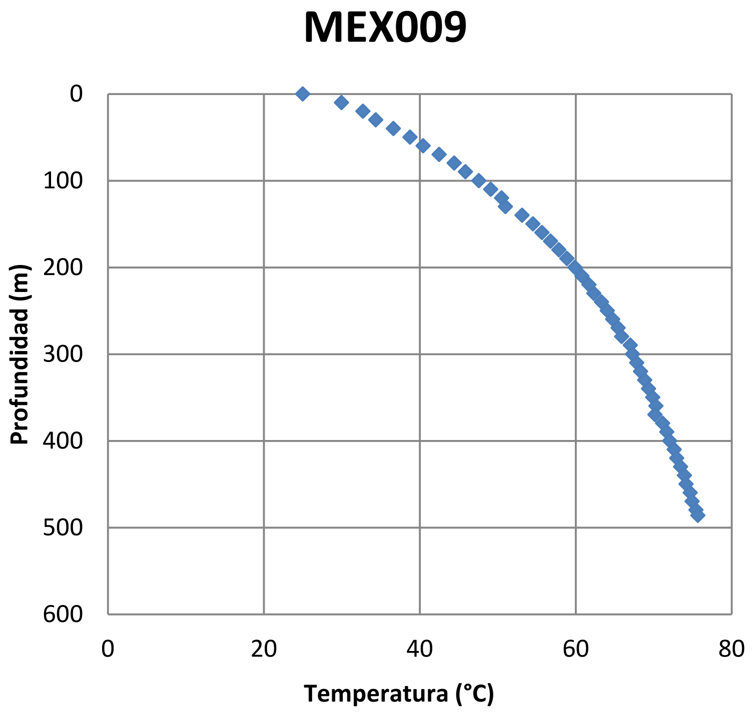
Bullard plot



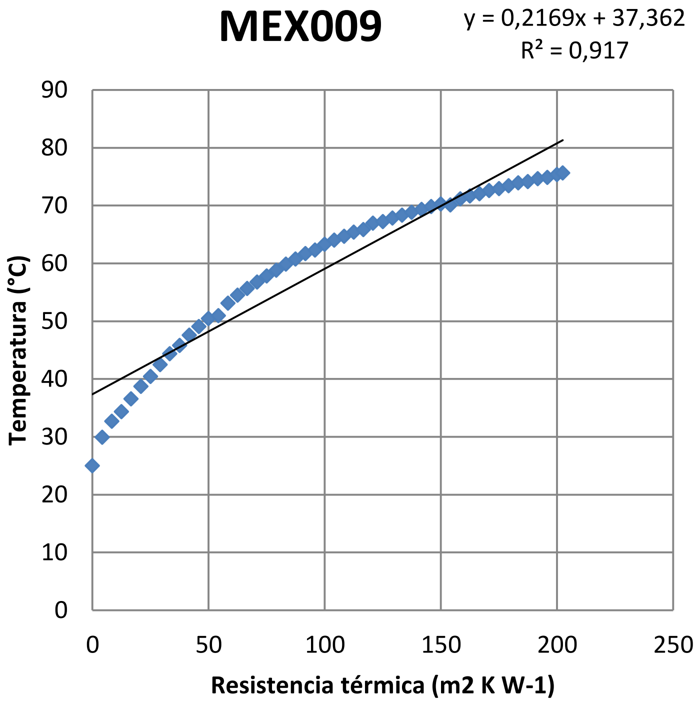
From the graphs estimate the geothermal gradient, the heat flow density, and the average thermal conductivity.

Mexico - Bullard plot method

Temperature Log

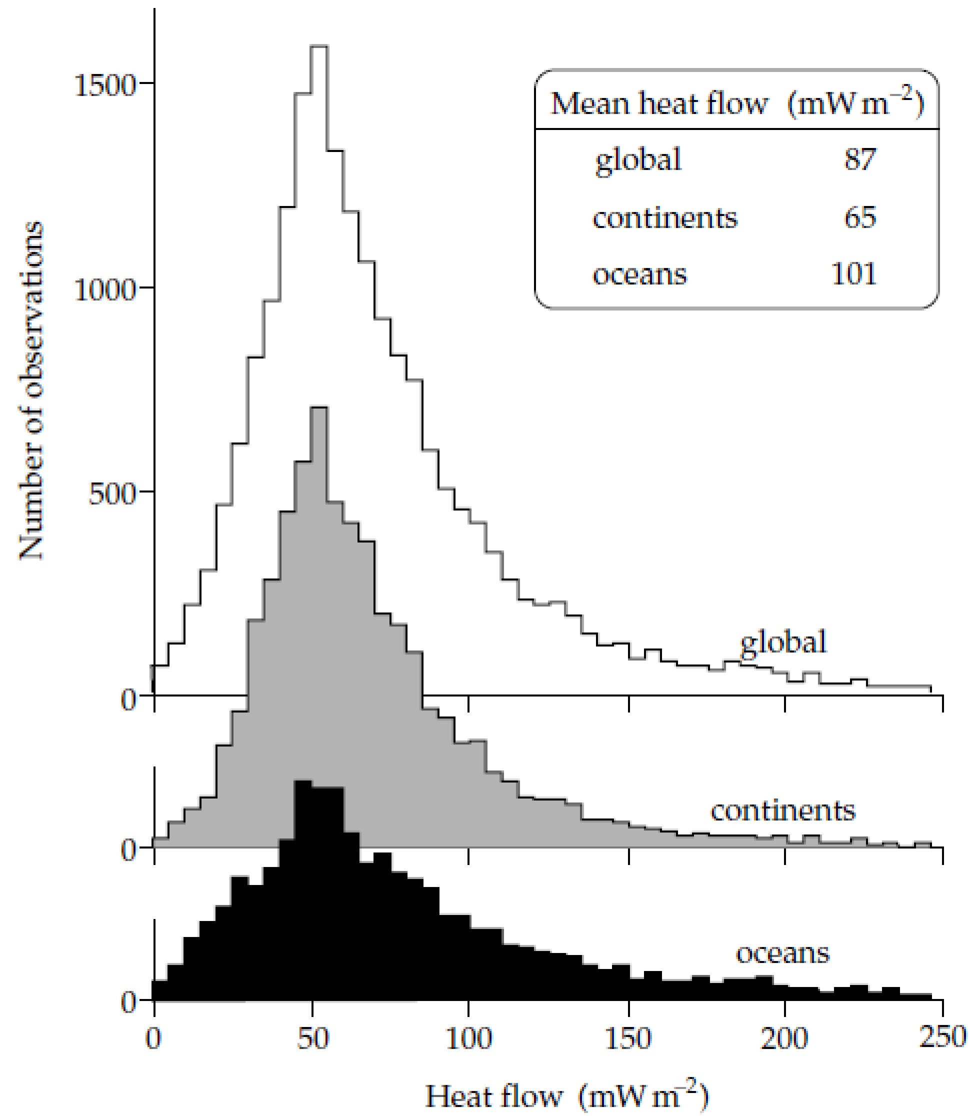


Bullard plot

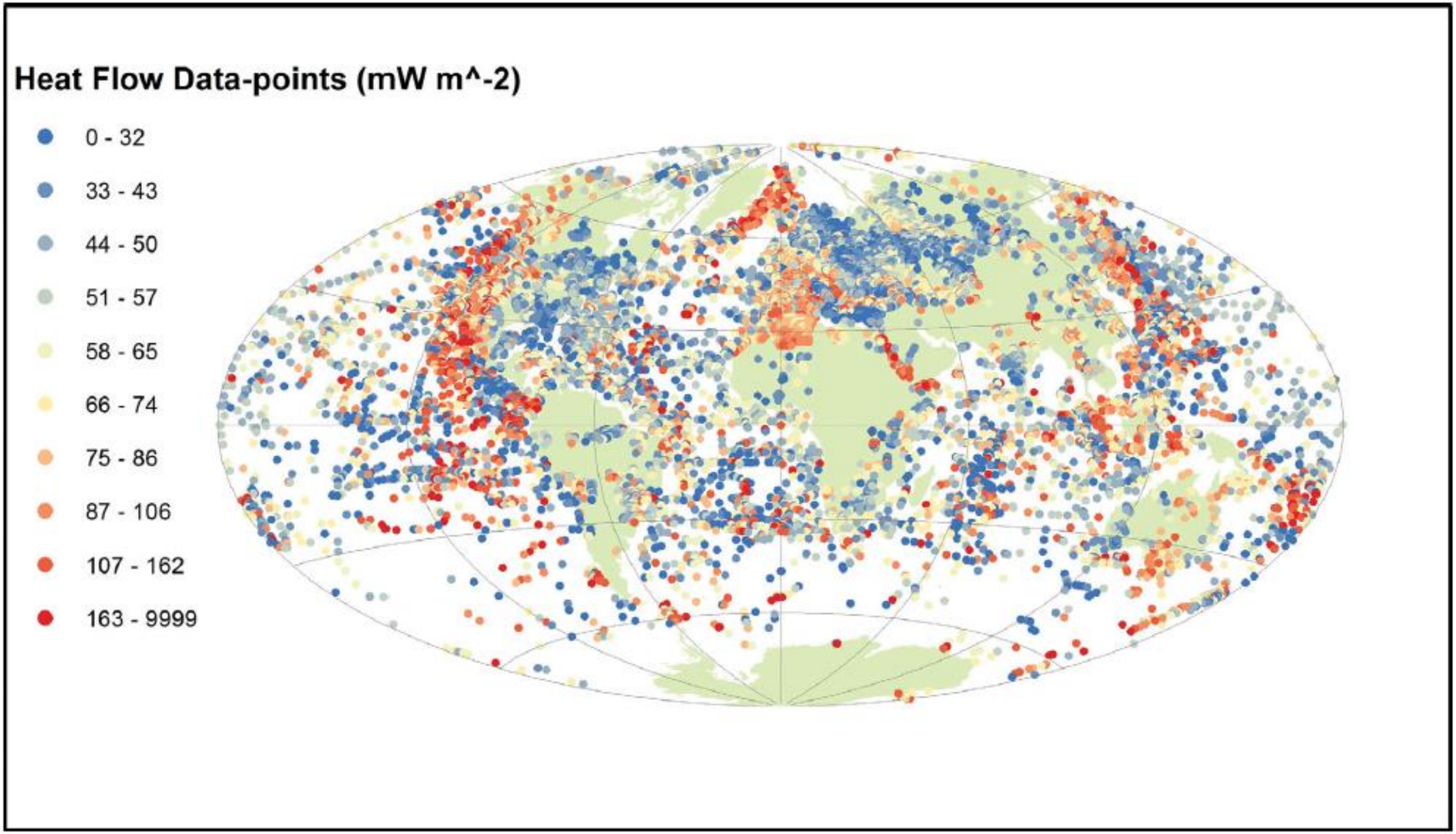


Global heat flow density

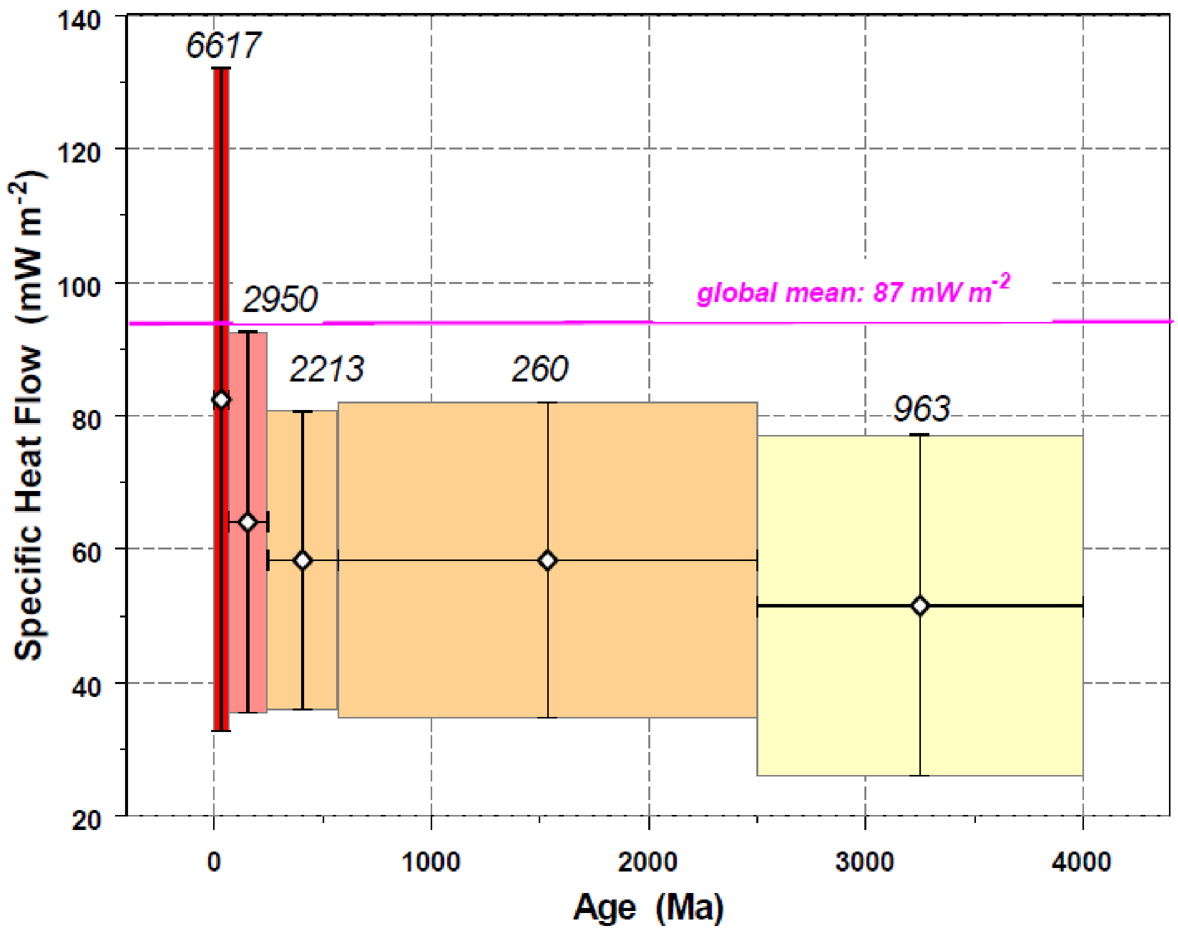
Histograms of continental, oceanic and global heat-flow. Values after Pollack et al. (1993).



Global heat flow density



Global heat flow density

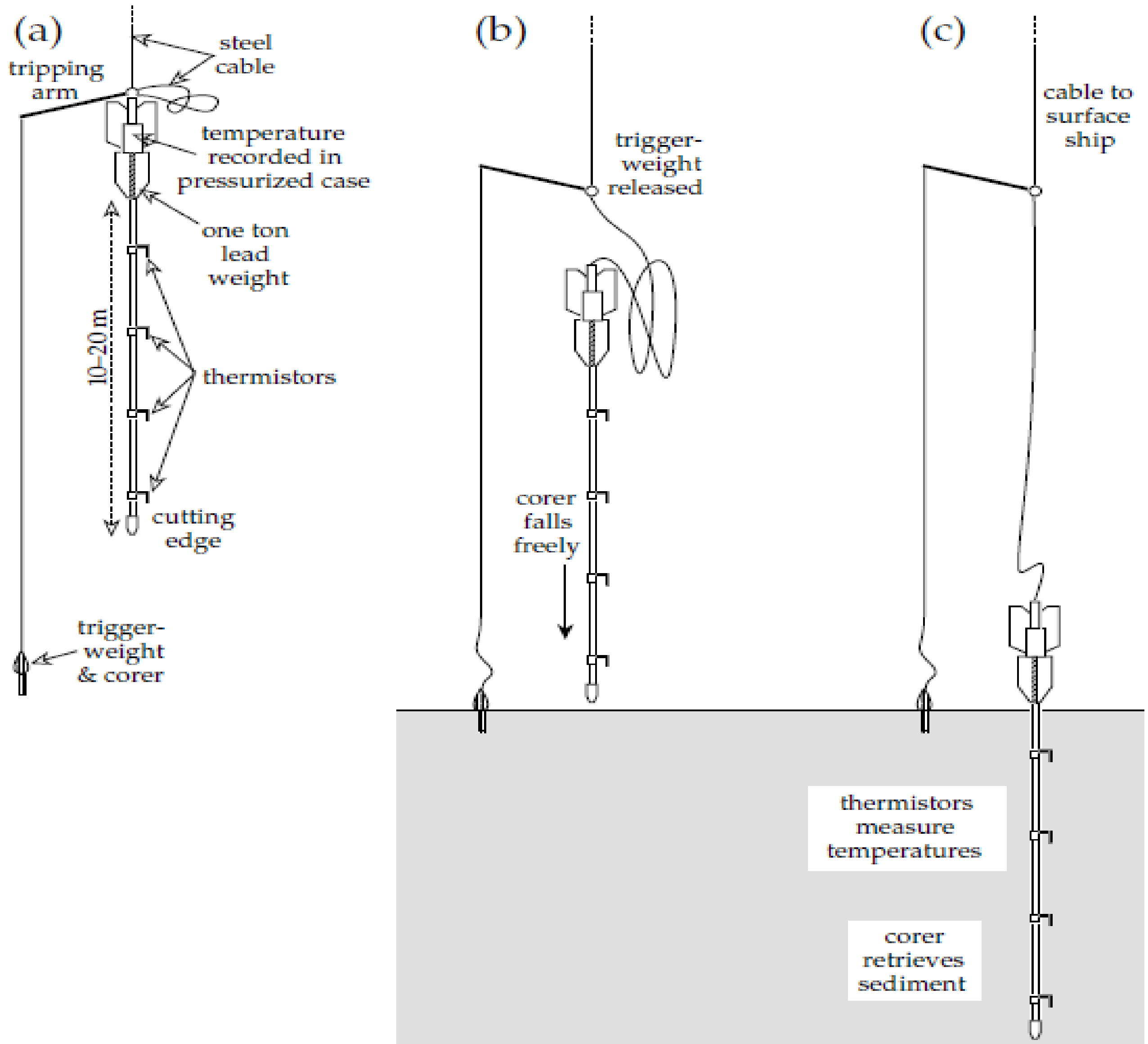


Studies of the heat flow density based on observations at 20201 sites worldwide reveal there is a decrease of the heat flow density with age: heat flow density is lower in old stable platforms than in young, tectonically active crust, on average by a factor of 1 ½.

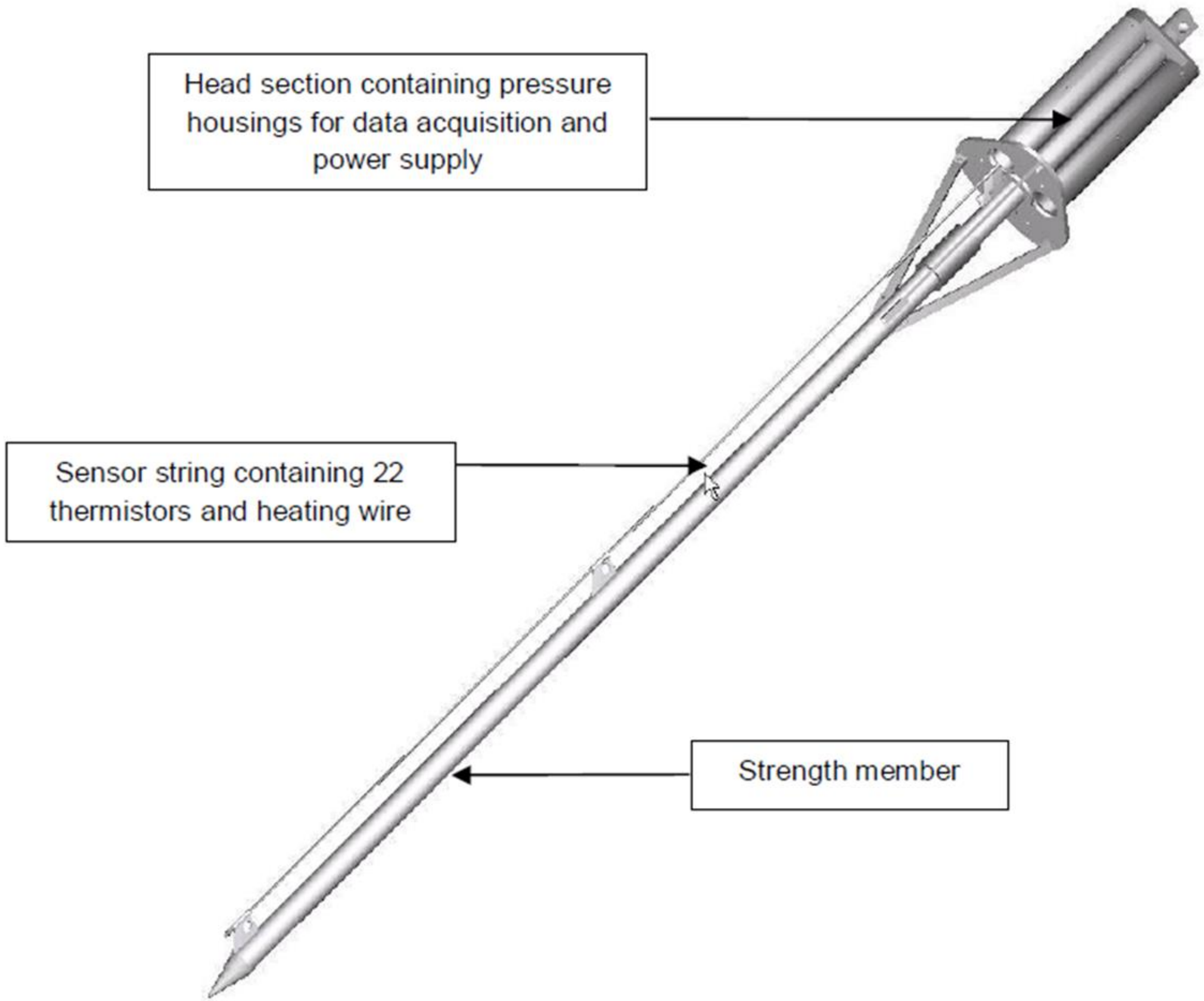
As a consequence, the mean heat flow density is larger in the generally young oceans (101 mWm^{-2}) than on the continents (67 mWm^{-2}).

Heat flow density in the ocean

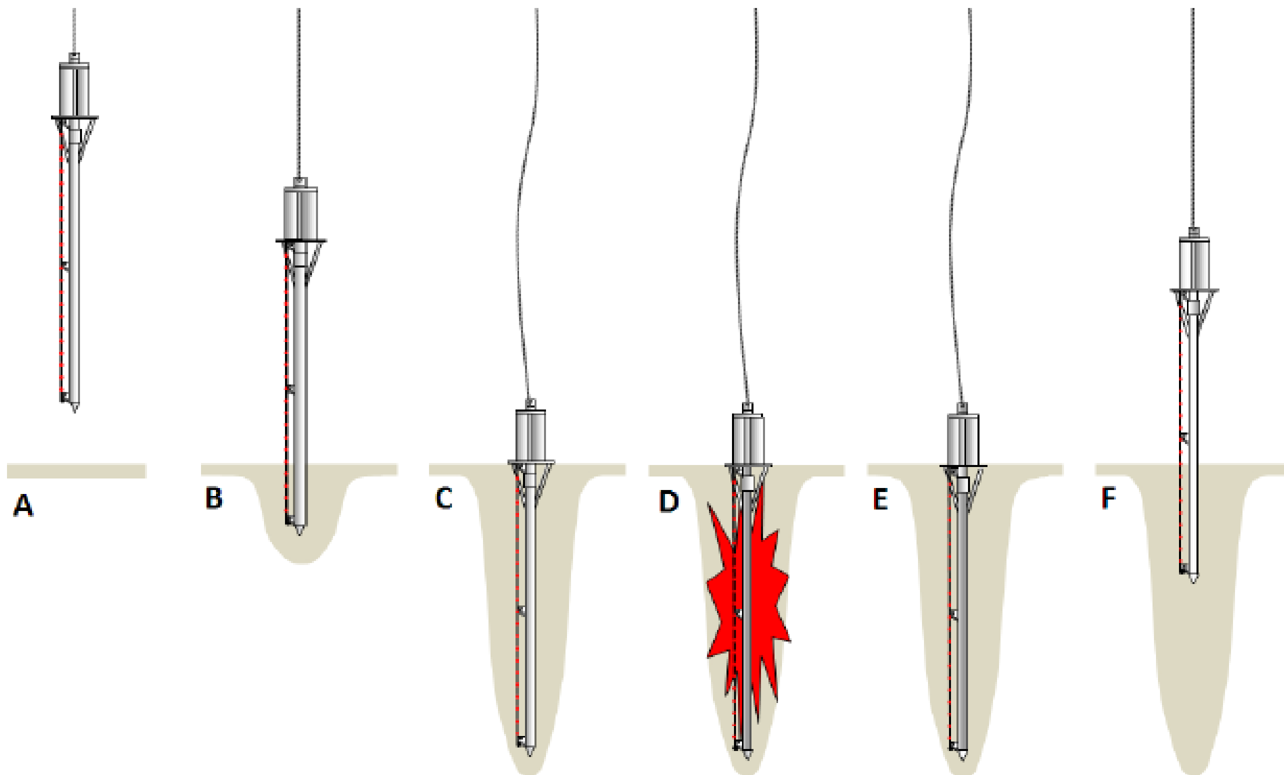
Heat flow density in the ocean



Heat flow density in the ocean



Heat flow density in the ocean



A: lowering to seabed

B: penetrating into seabed

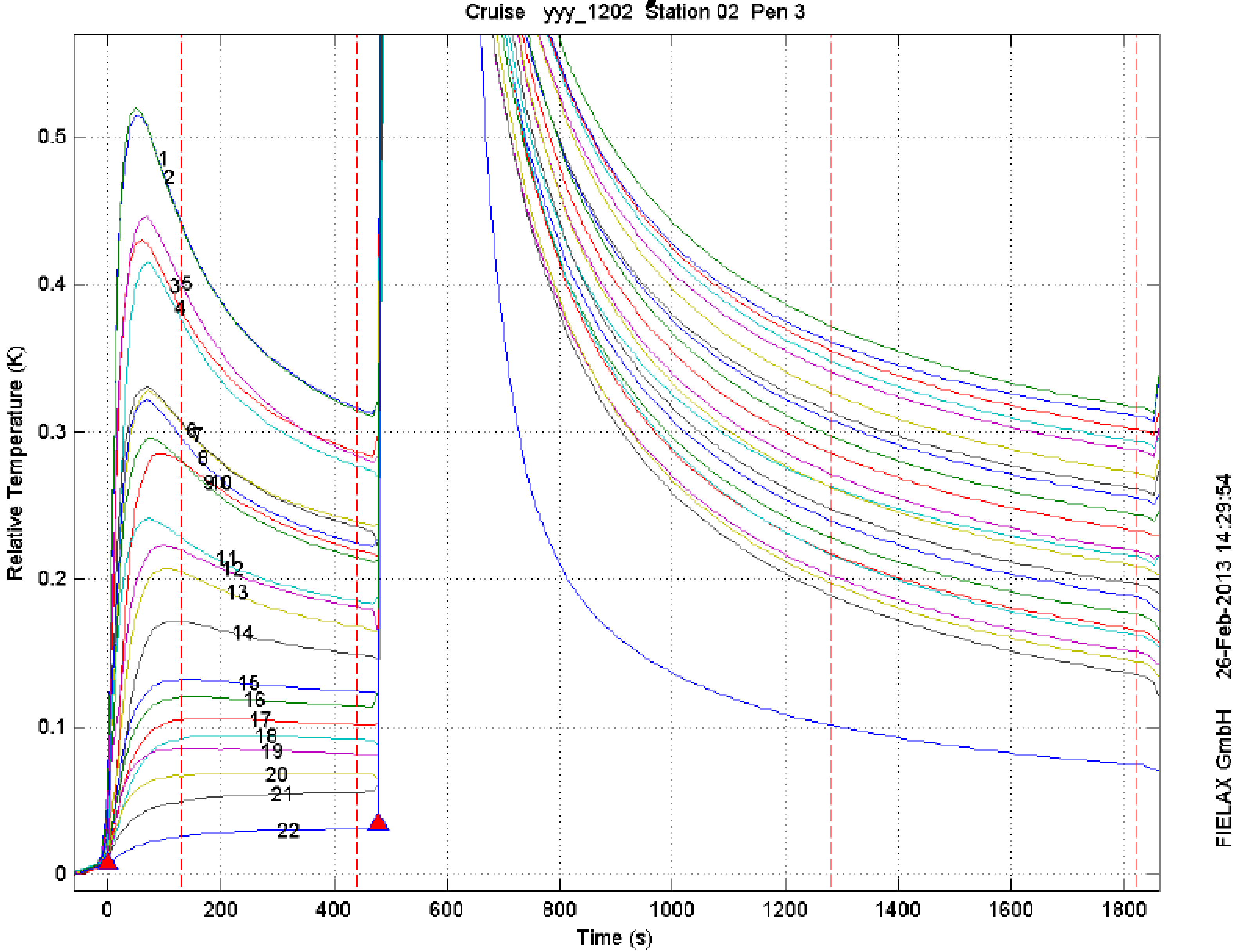
C: measuring thermal decay of frictional heat (approx.. 7-12 minutes)

D: heat pulse (approx.. 20 seconds)

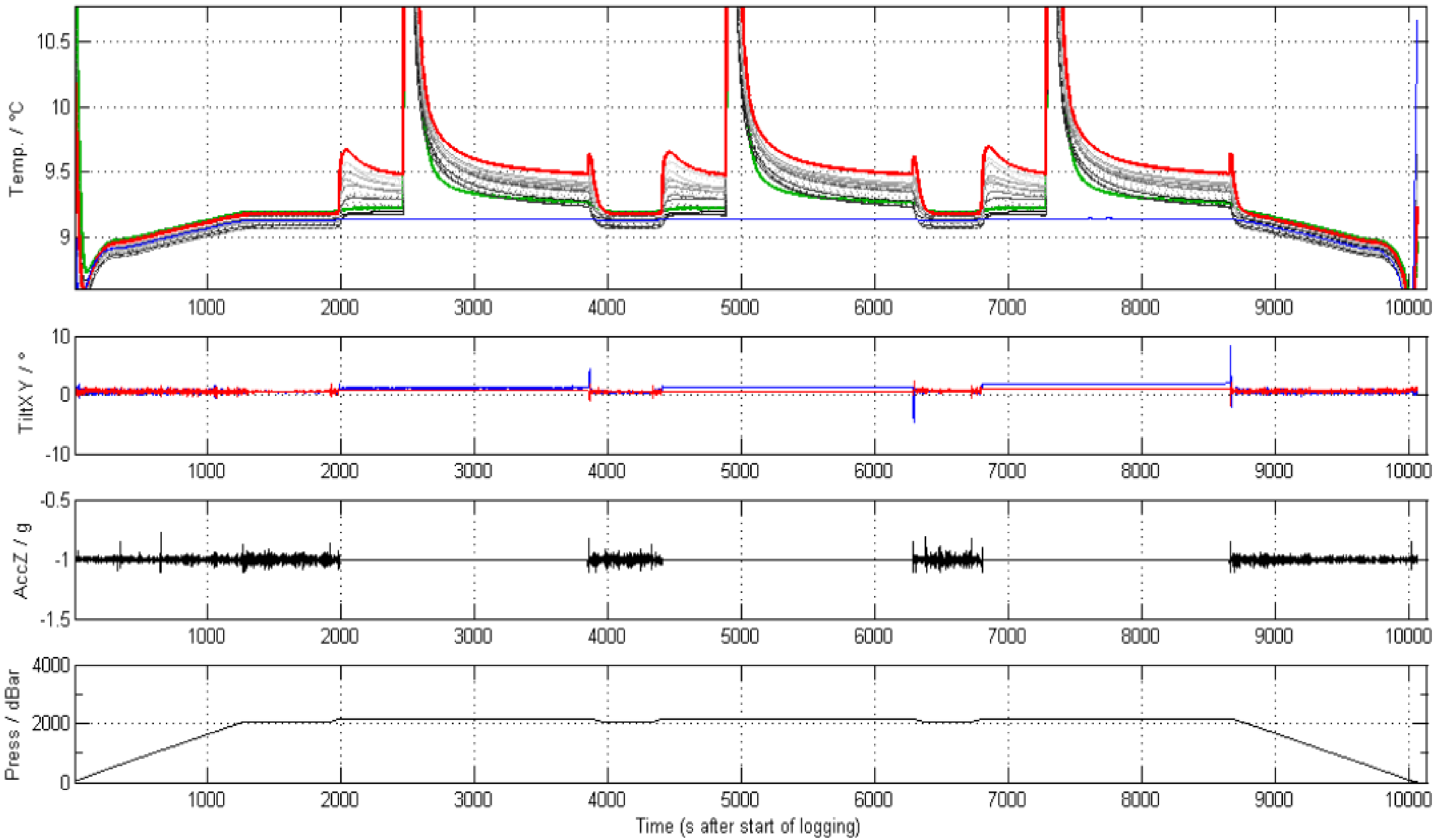
E: measuring thermal decay of heat pulse (approx.. 15-20 minutes)

F: pullout and retrieve to surface

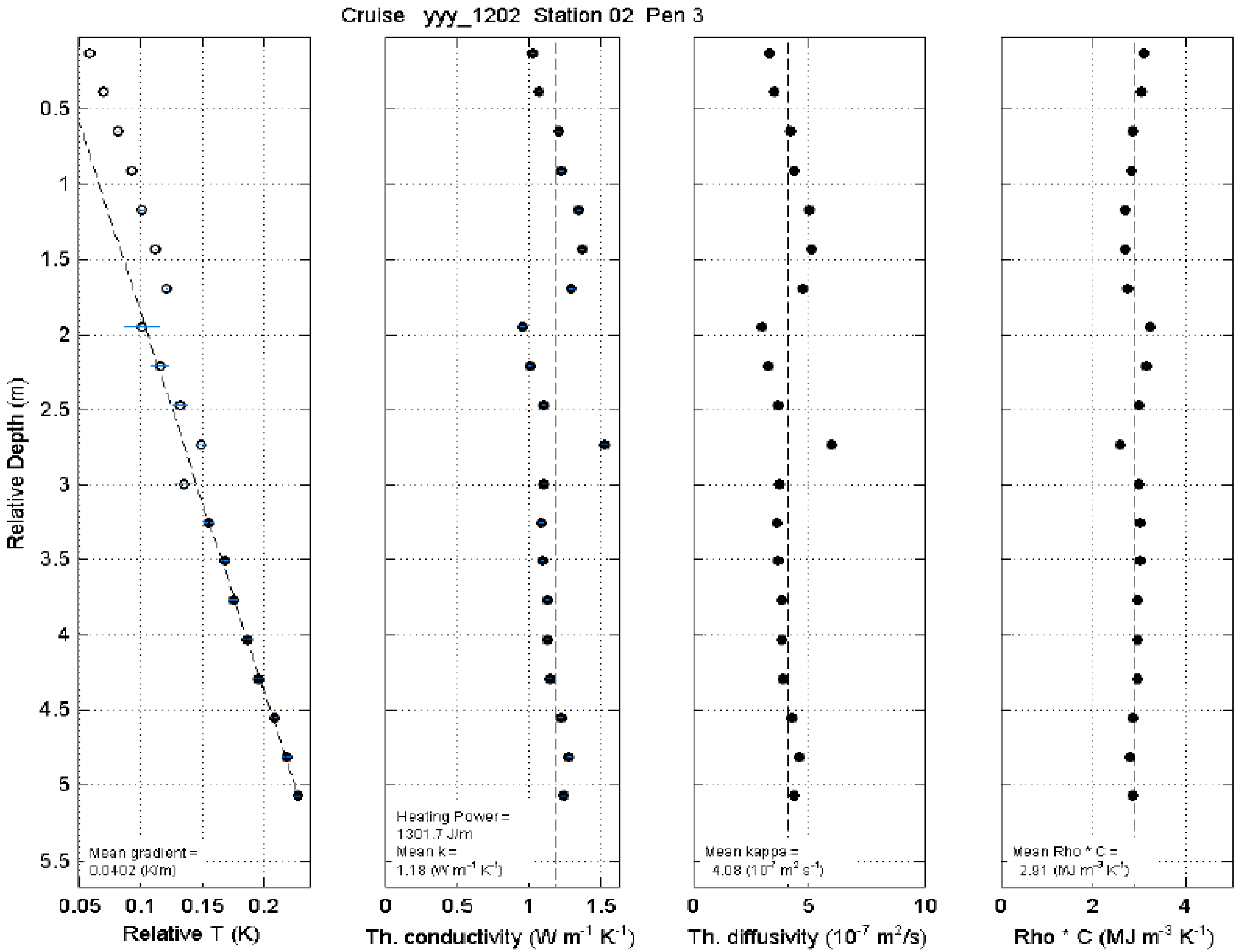
Heat flow density in the ocean



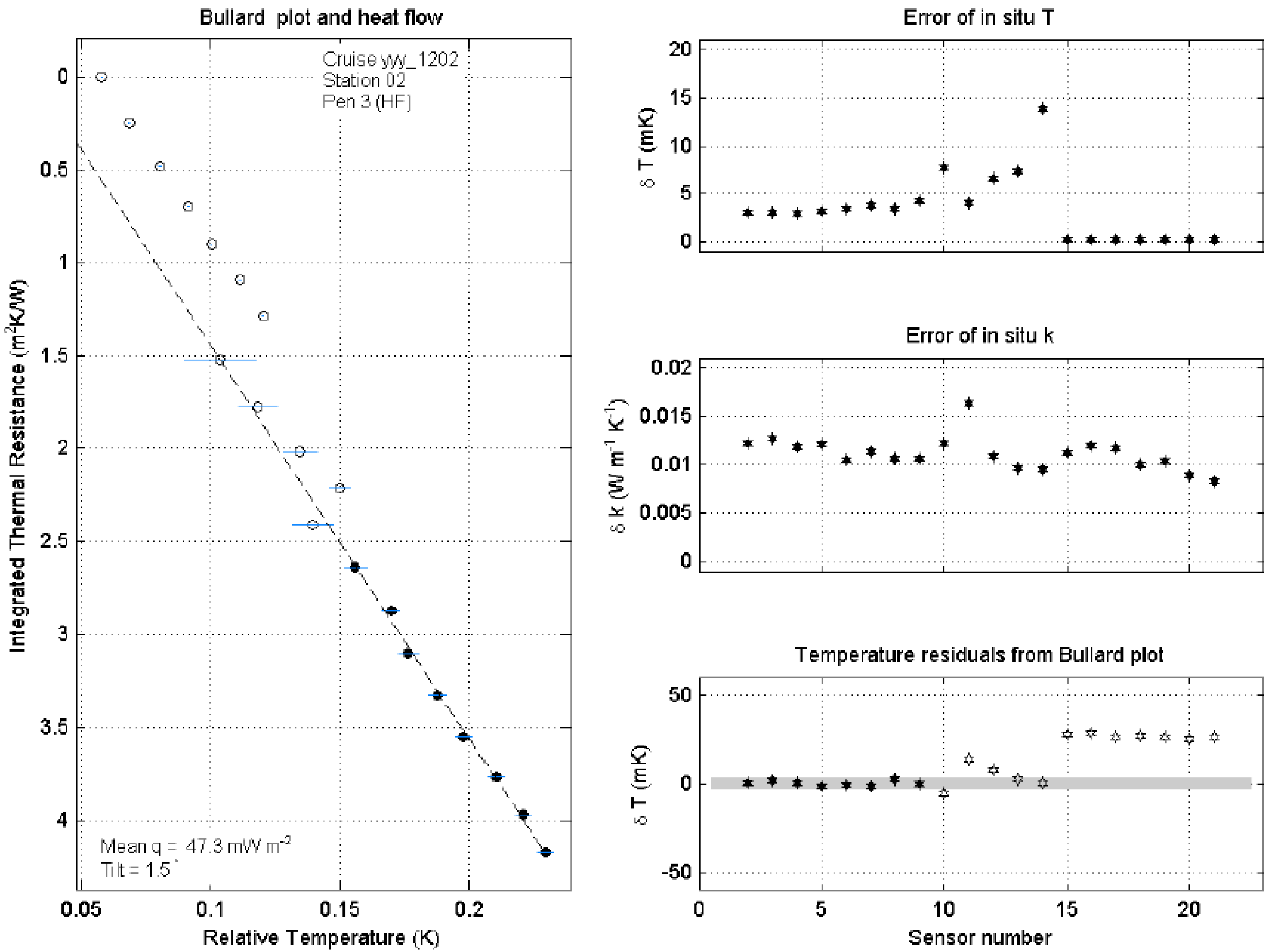
Heat flow density in the ocean



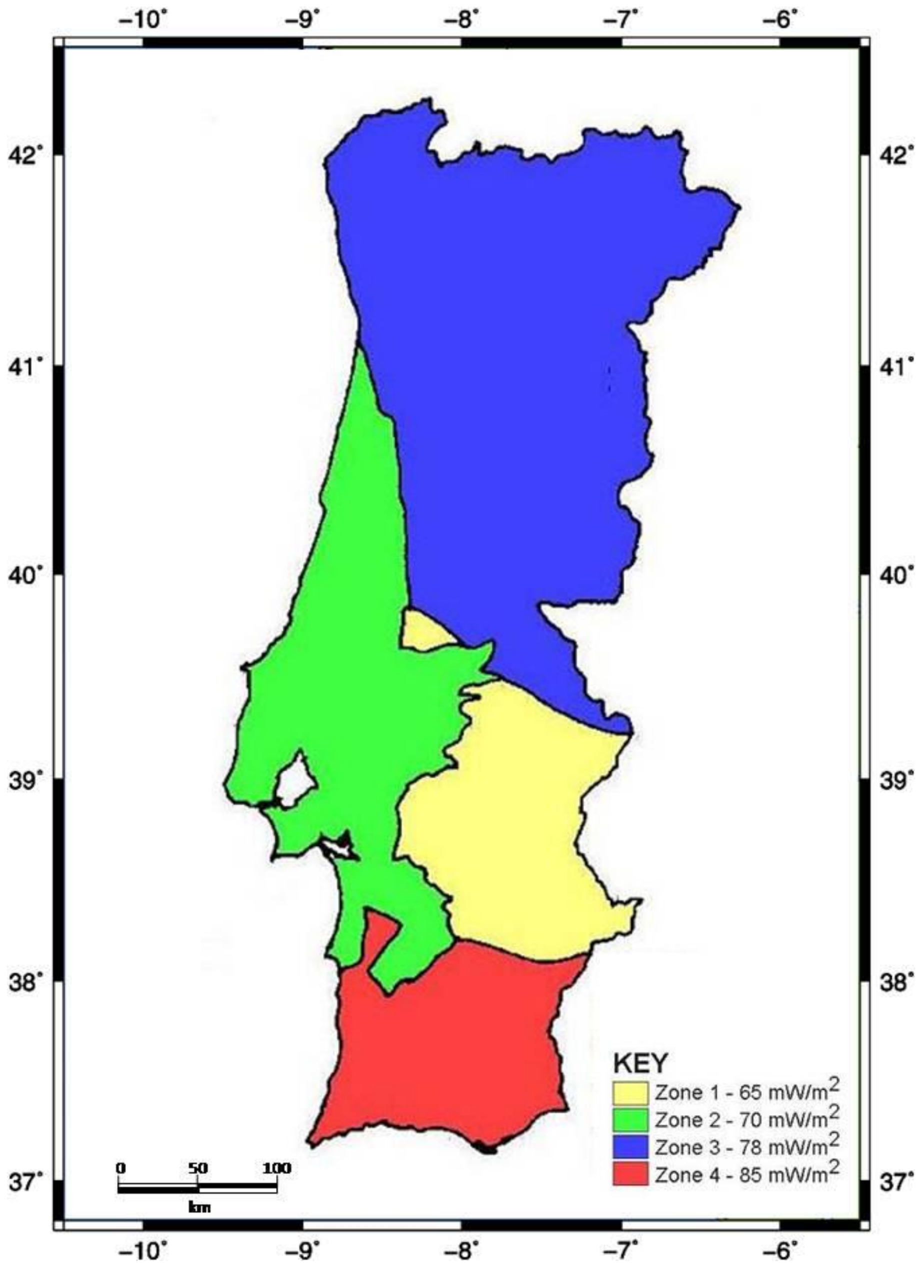
Heat flow density in the ocean

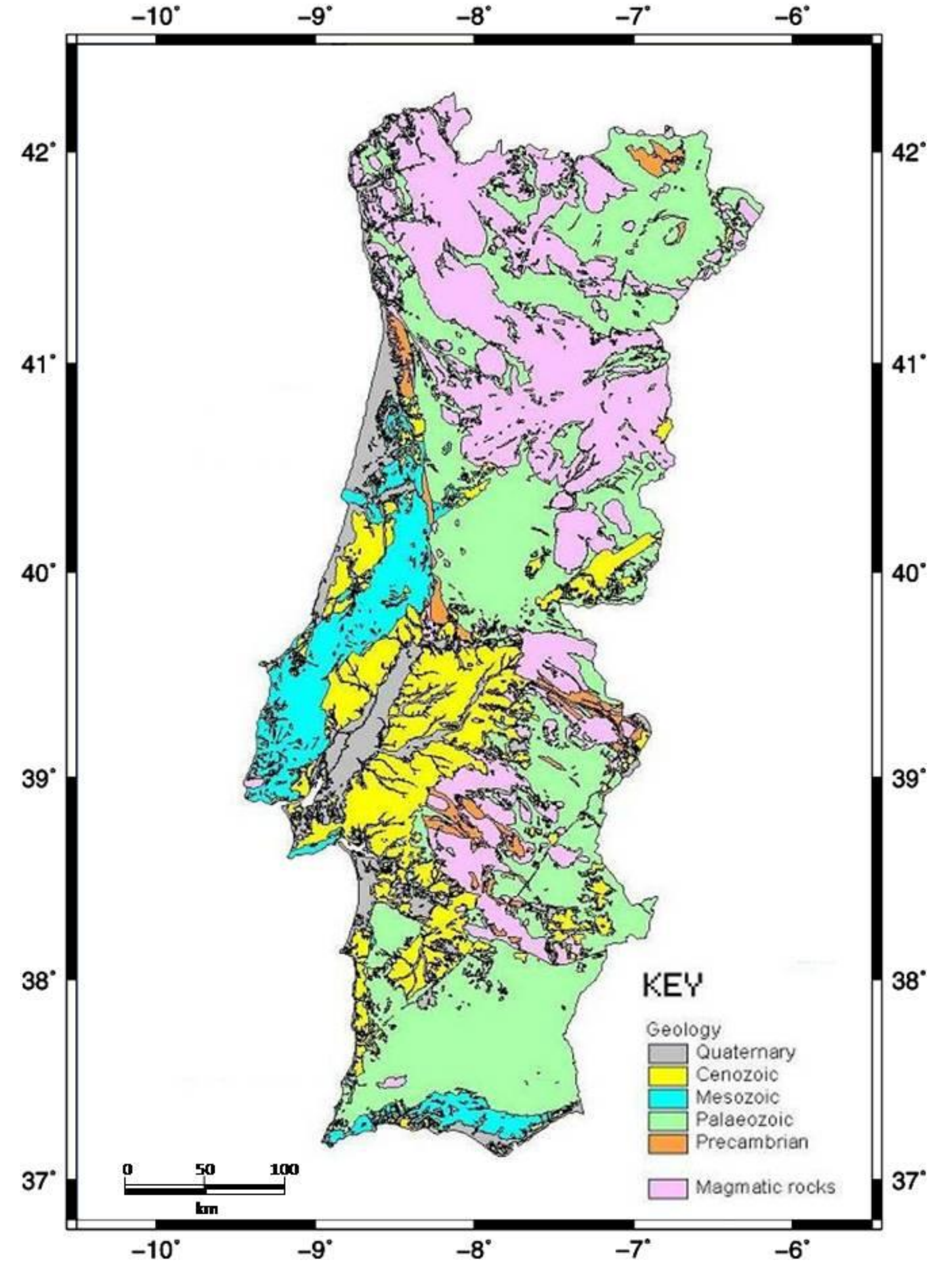


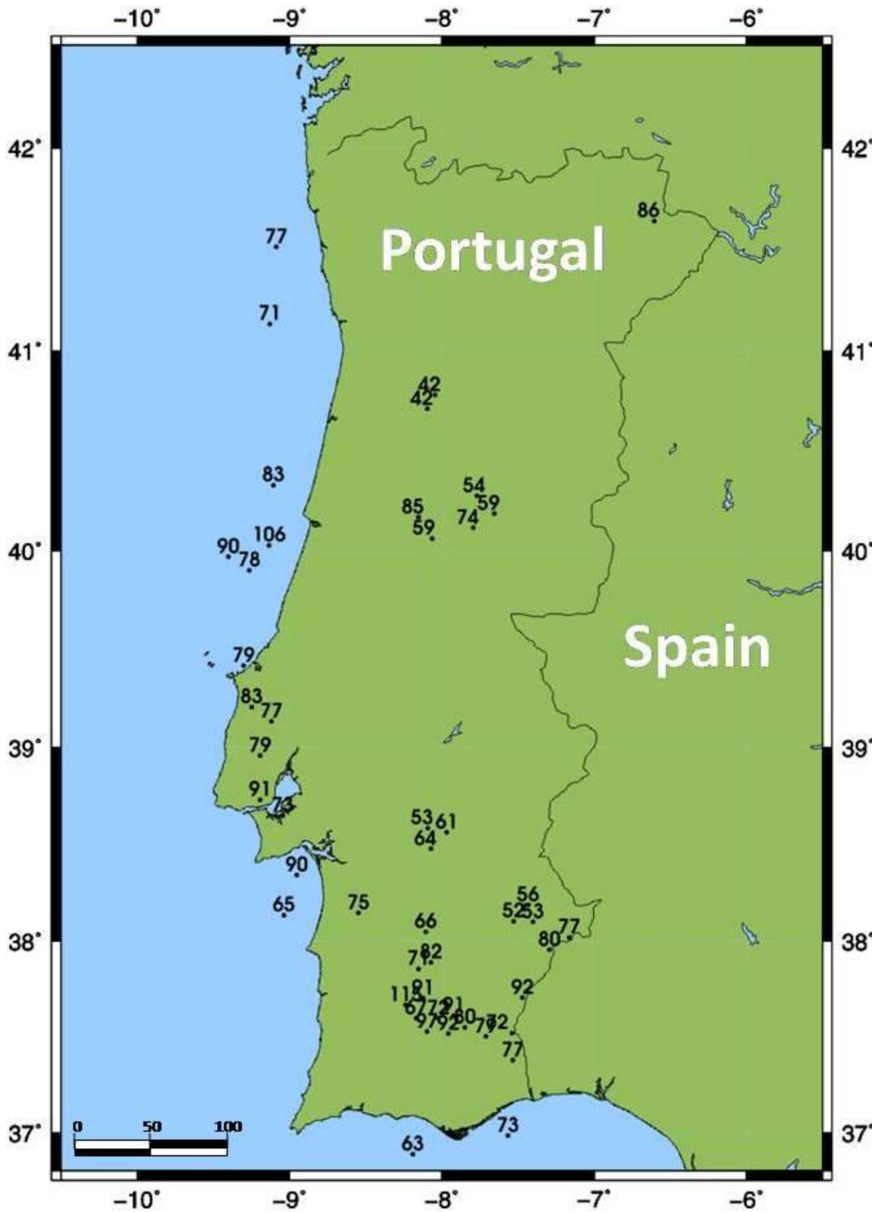
Heat flow density in the ocean



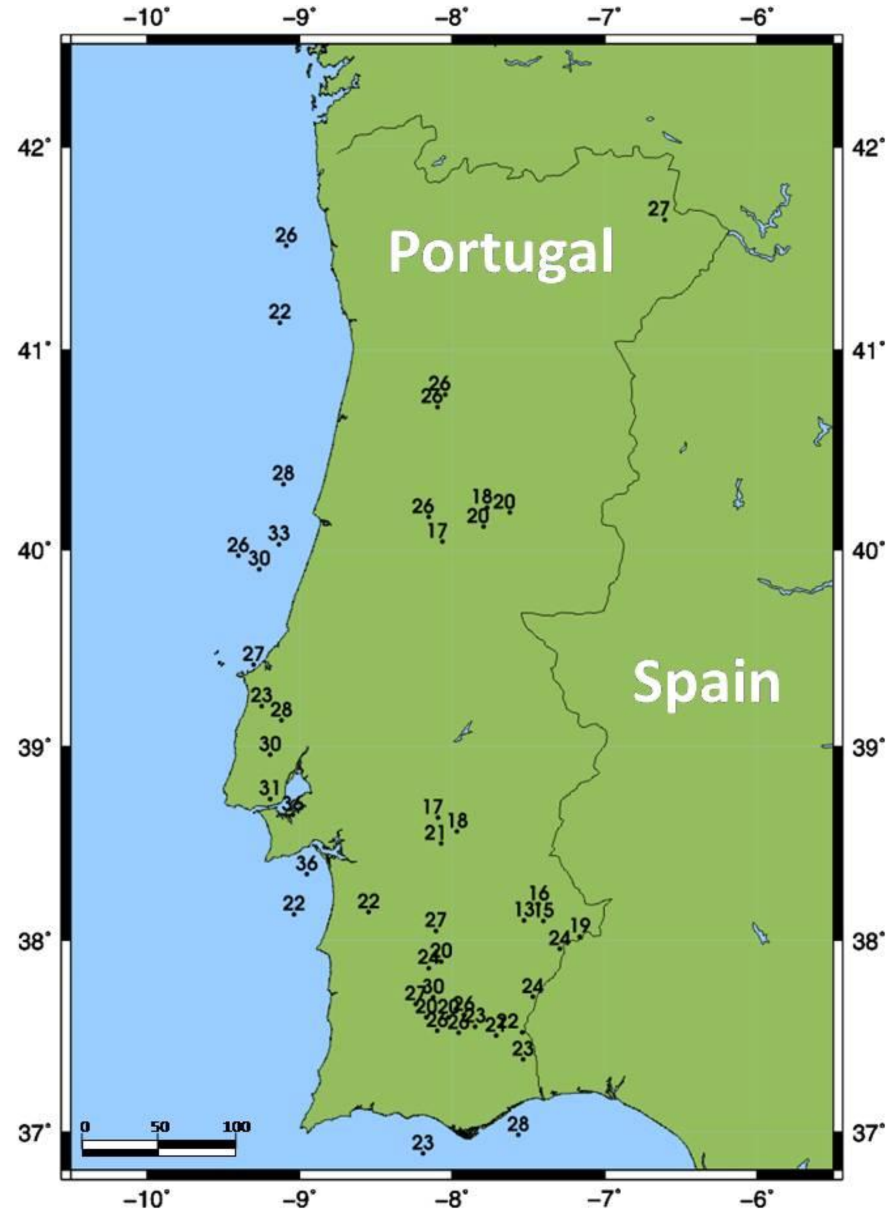
VI
Geothermal Resources


KEY
Zone 1 - 65 mW/m2
Zone 2 - 70 mW/m2
Zone 3 - 78 mW/m2
Zone 4 - 85 mW/m2
0 50 100
km


KEY
Geology
Quaternary
Cenozoic
Mesozoic
Palaeozoic
Precambrian
Magmatic rocks
0 50 100
km



HFD map



Gradient map

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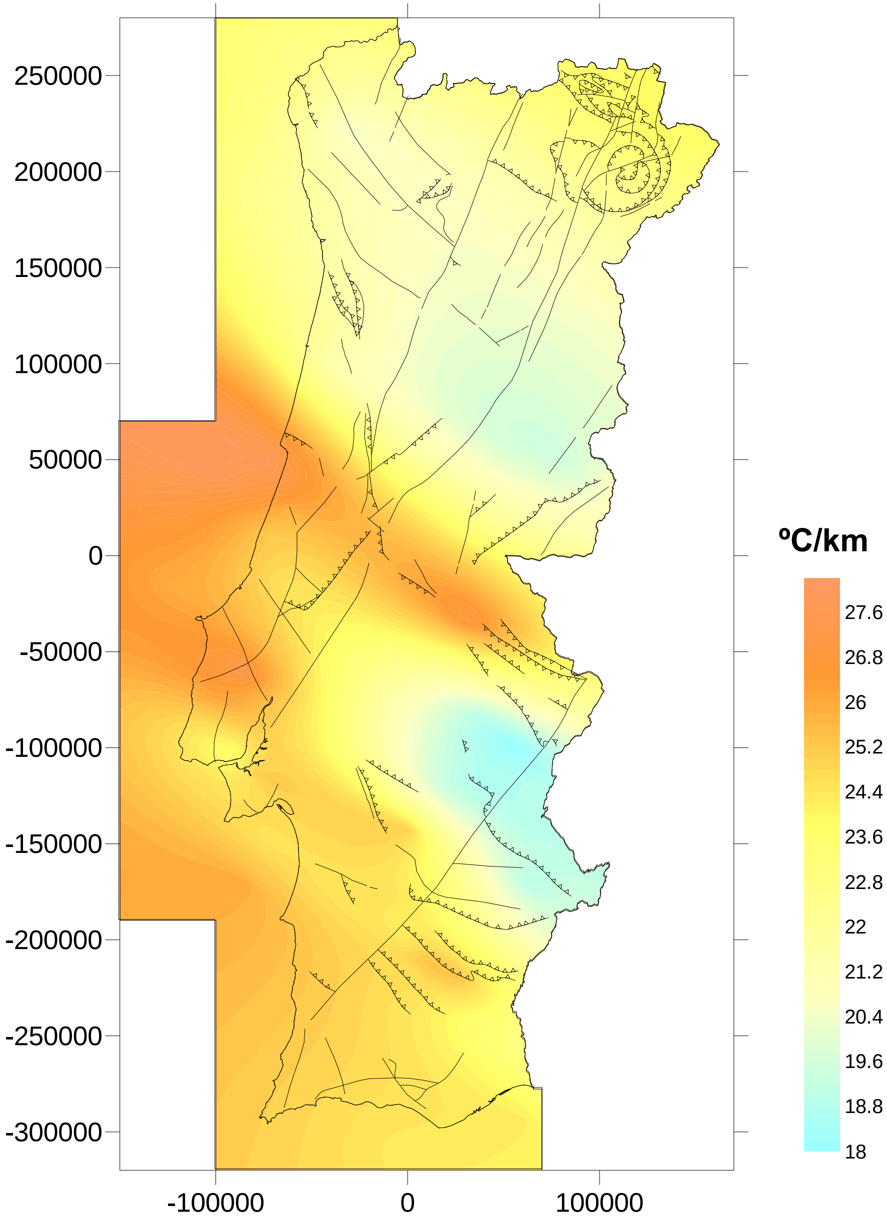
Laboratório de Geologia e Minas

LNEG

CARTA DO GRADIENTE GEOTÉRMICO DE PORTUGAL

2014



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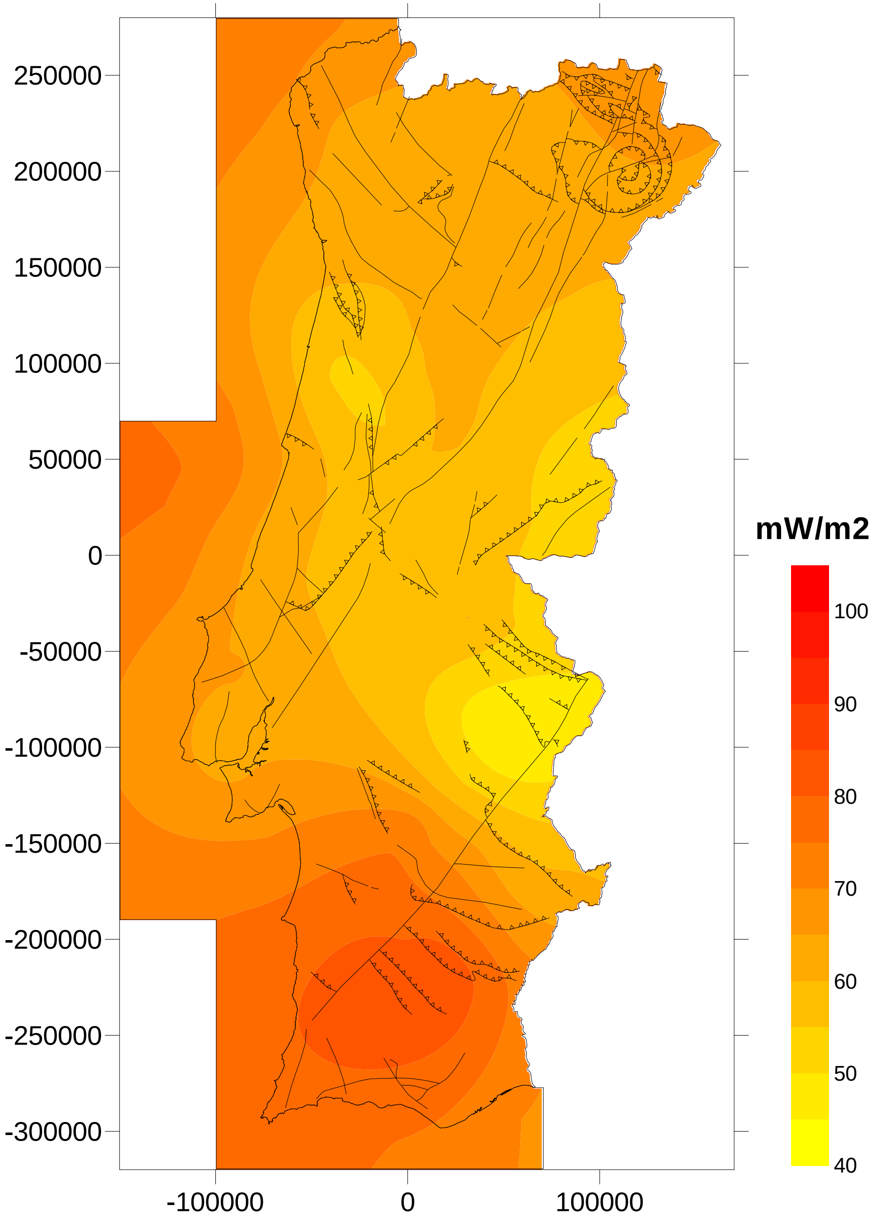
LNEG

CARTA DA DENSIDADE DE FLUXO DE CALOR À SUPERFÍCIE EM PORTUGAL

2014



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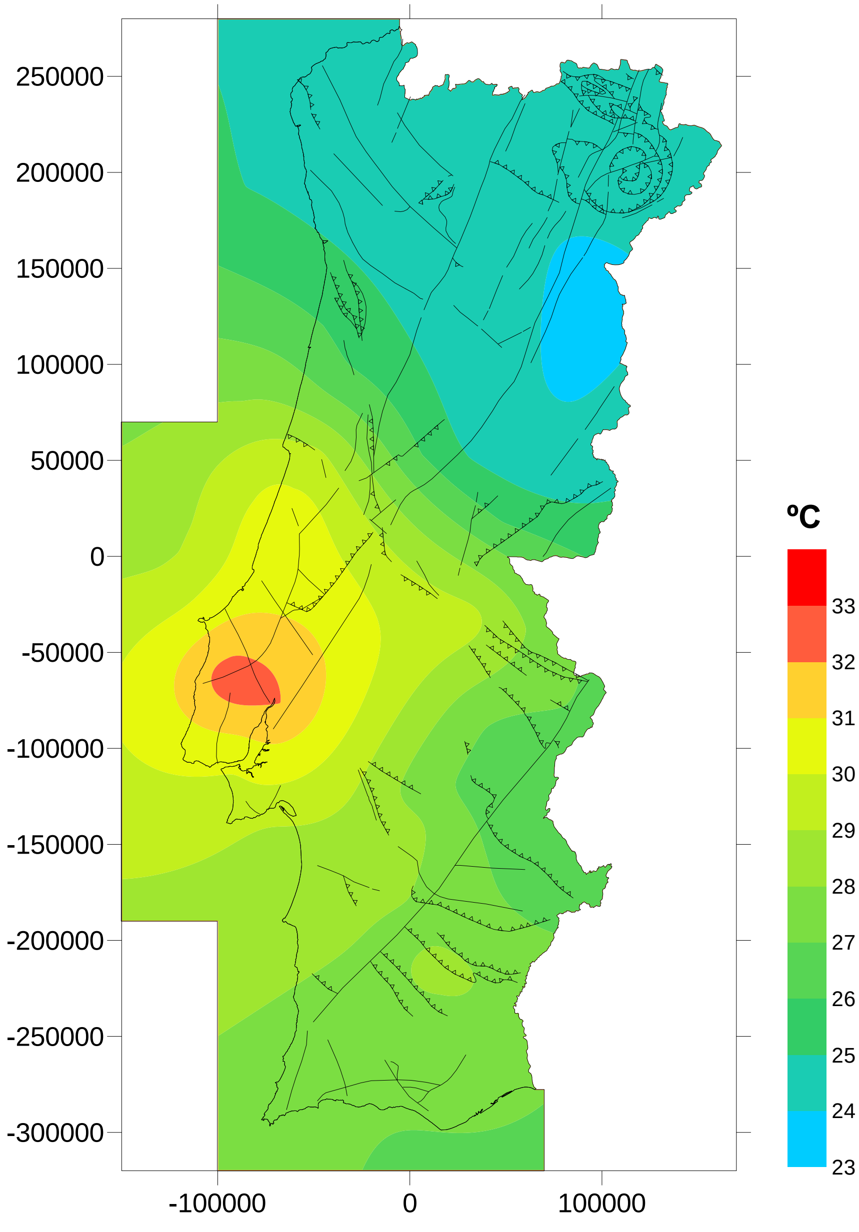
CARTA DA TEMPERATURA A 500 m DE PROFUNDIDADE

2014



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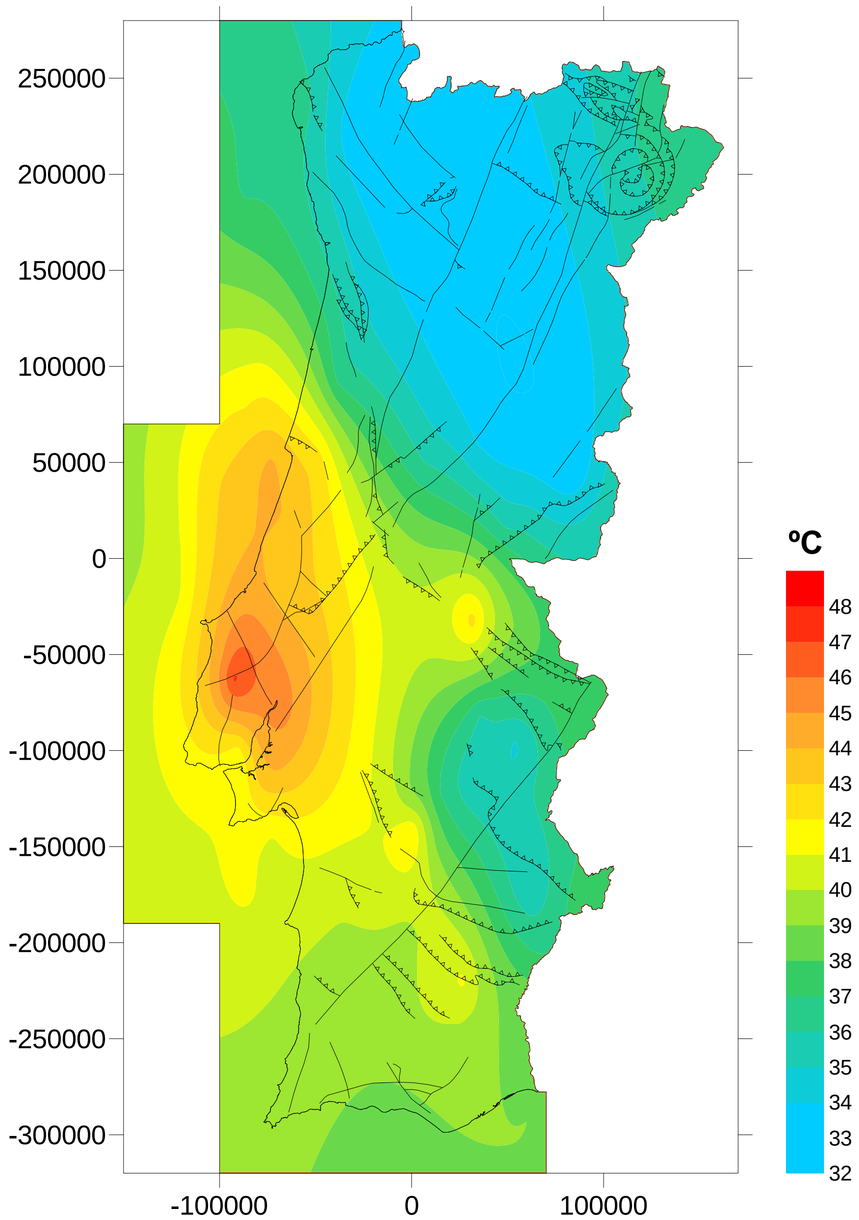
CARTA DA TEMPERATURA A 1000 m DE PROFUNDIDADE

2014



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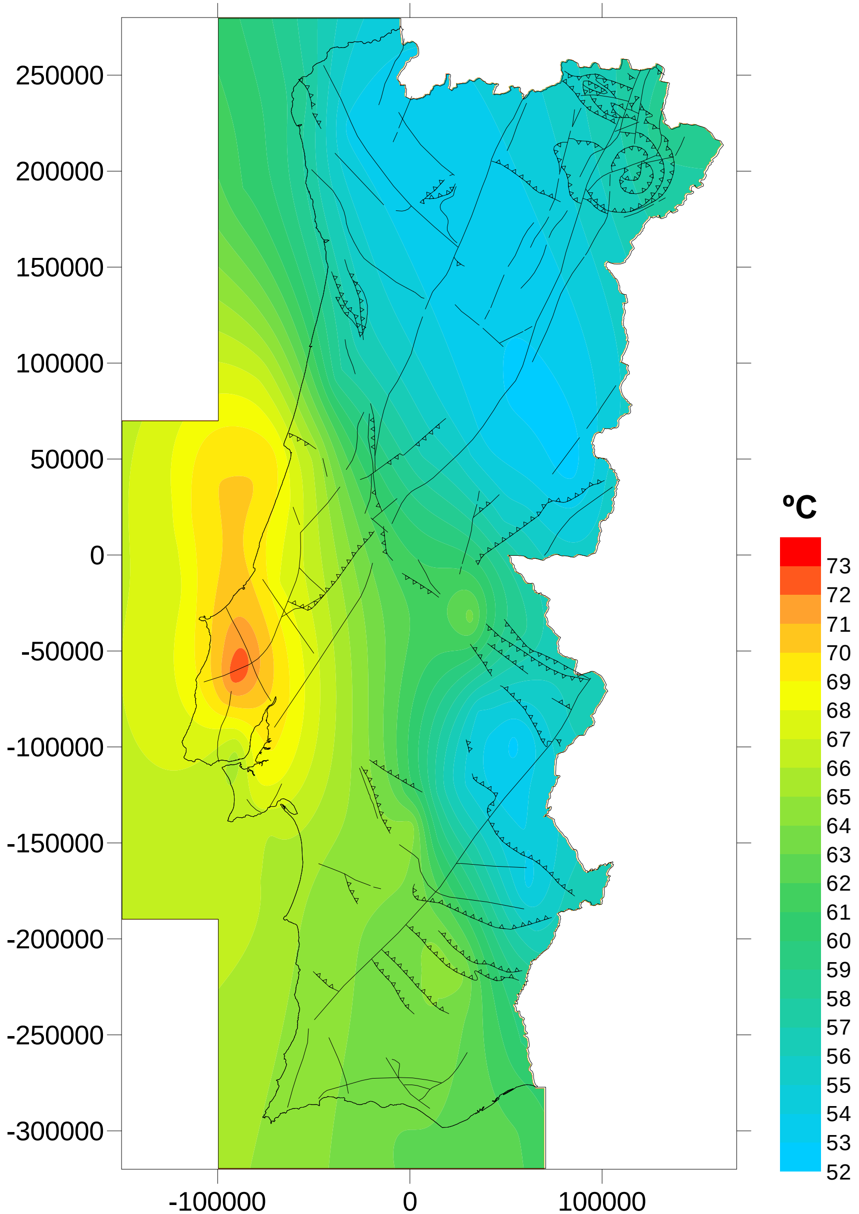
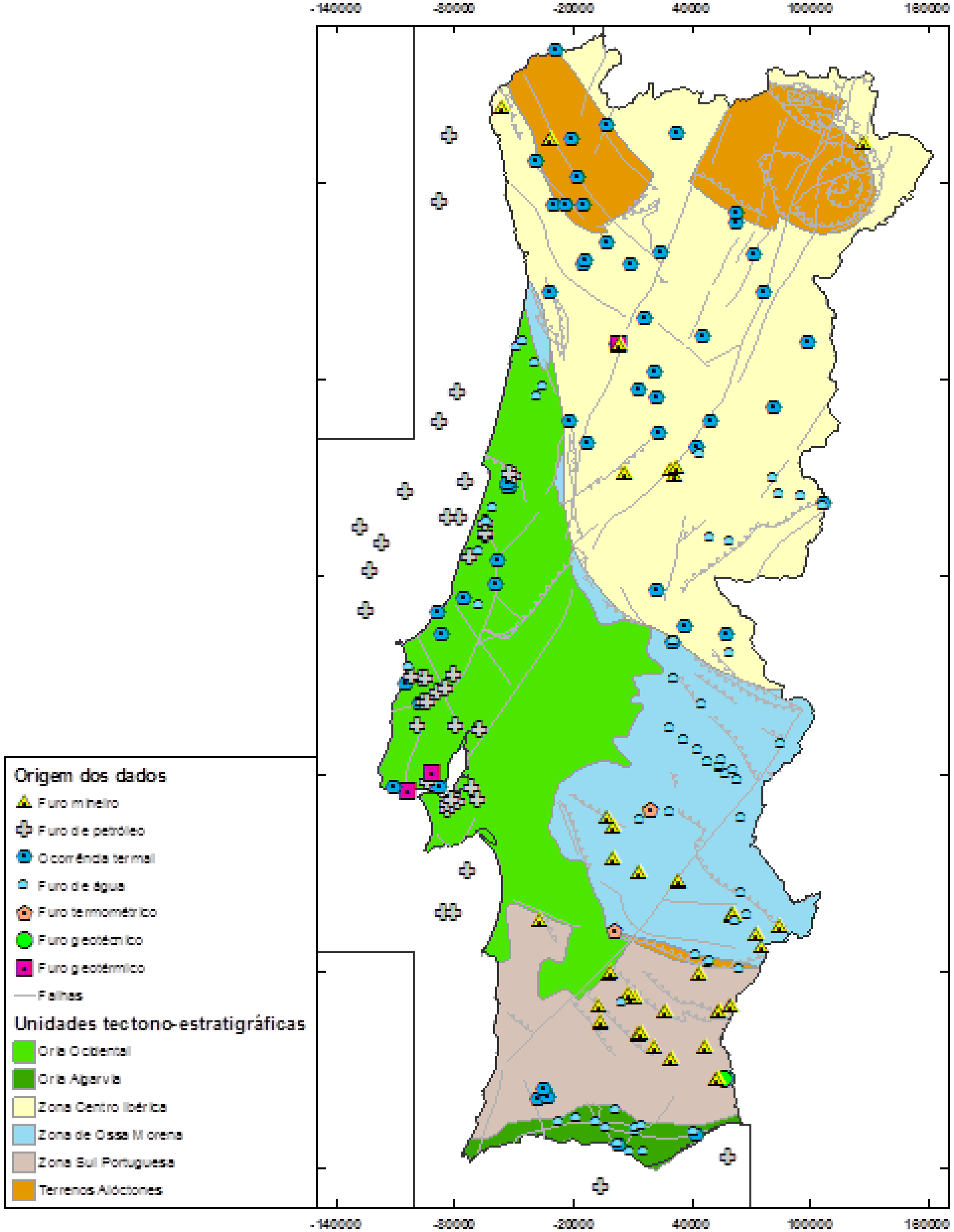
CARTA DA TEMPERATURA A 2000 m DE PROFUNDIDADE

2014




-140000
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40000
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160000
200000
100000
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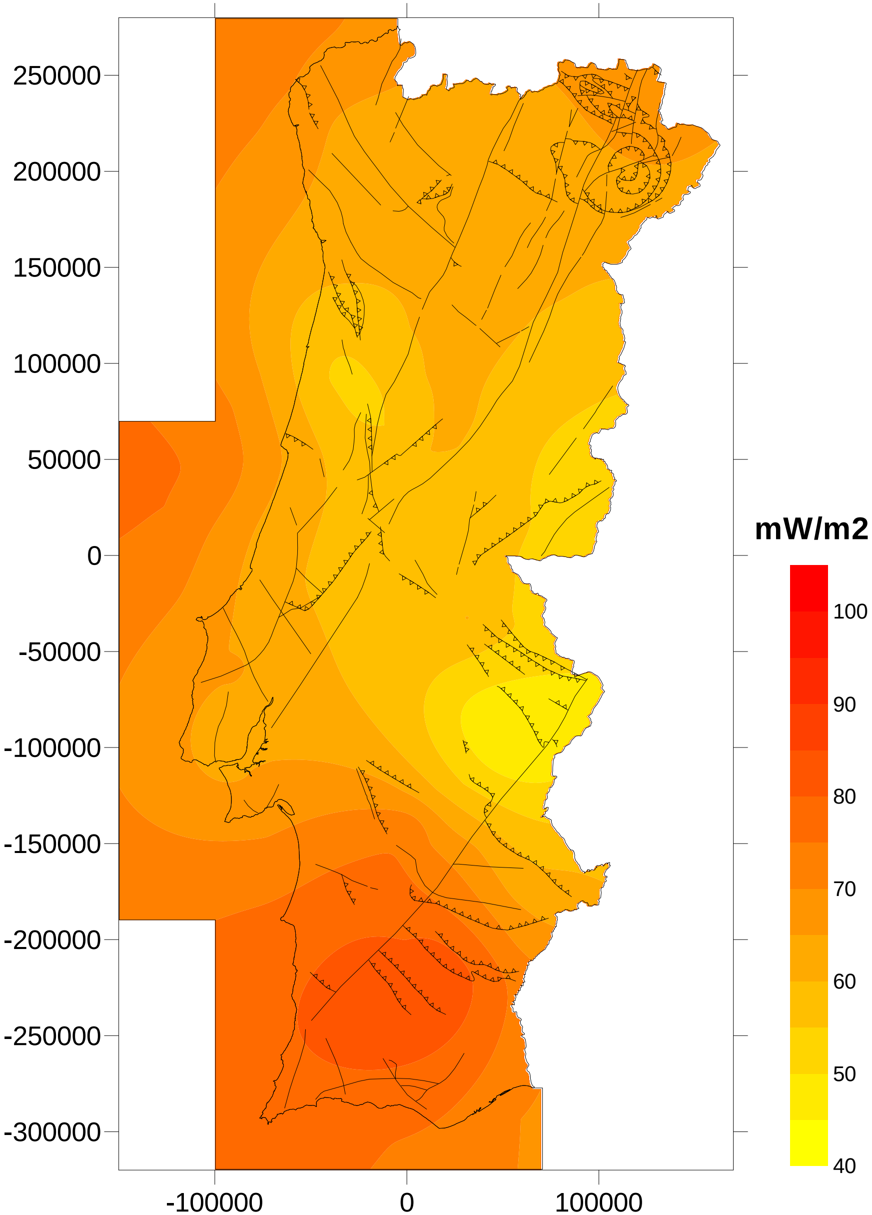
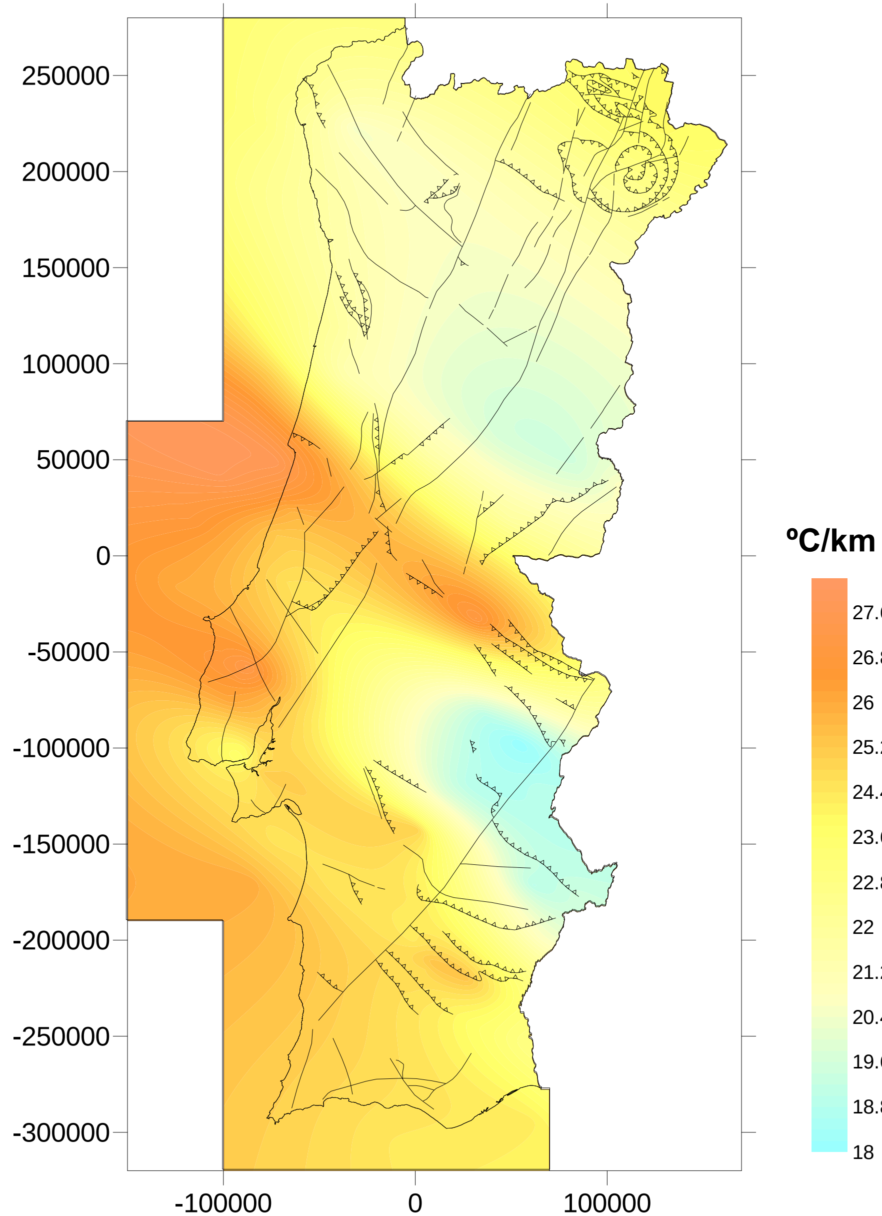
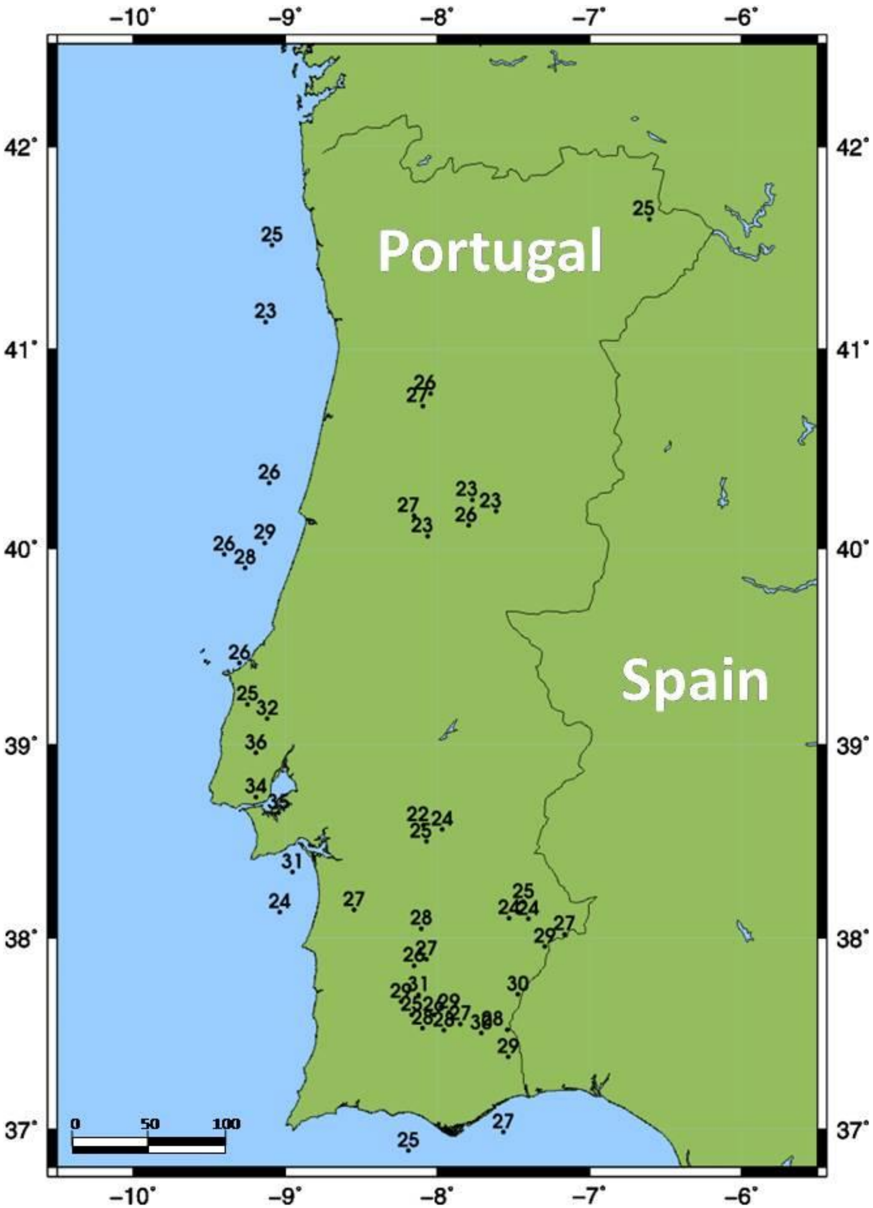
-200000
-300000
Origem dos dados
Furo mineiro
Furo de petróleo
Ocorrência termal
Furo de água
Furo termométrico
Furo geotécnico
Furo geotérmico
Falhas
Unidades tectono-estratigráficas
Orla Ocidental
Orla Algarvia
Zona Centro Ibérica
Zona de Ossa Morena
Zona Sul Portuguesa
Terrenos Alóctones



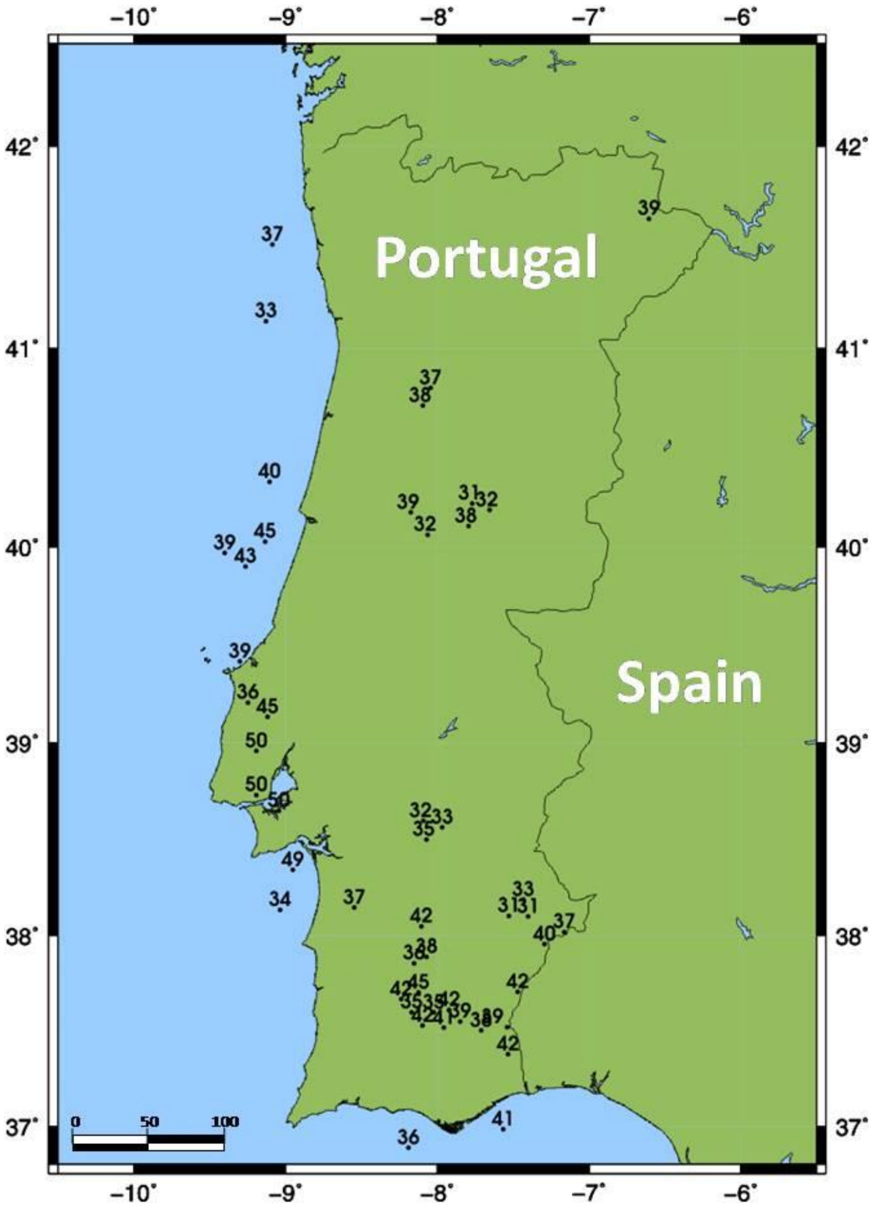
HFD map



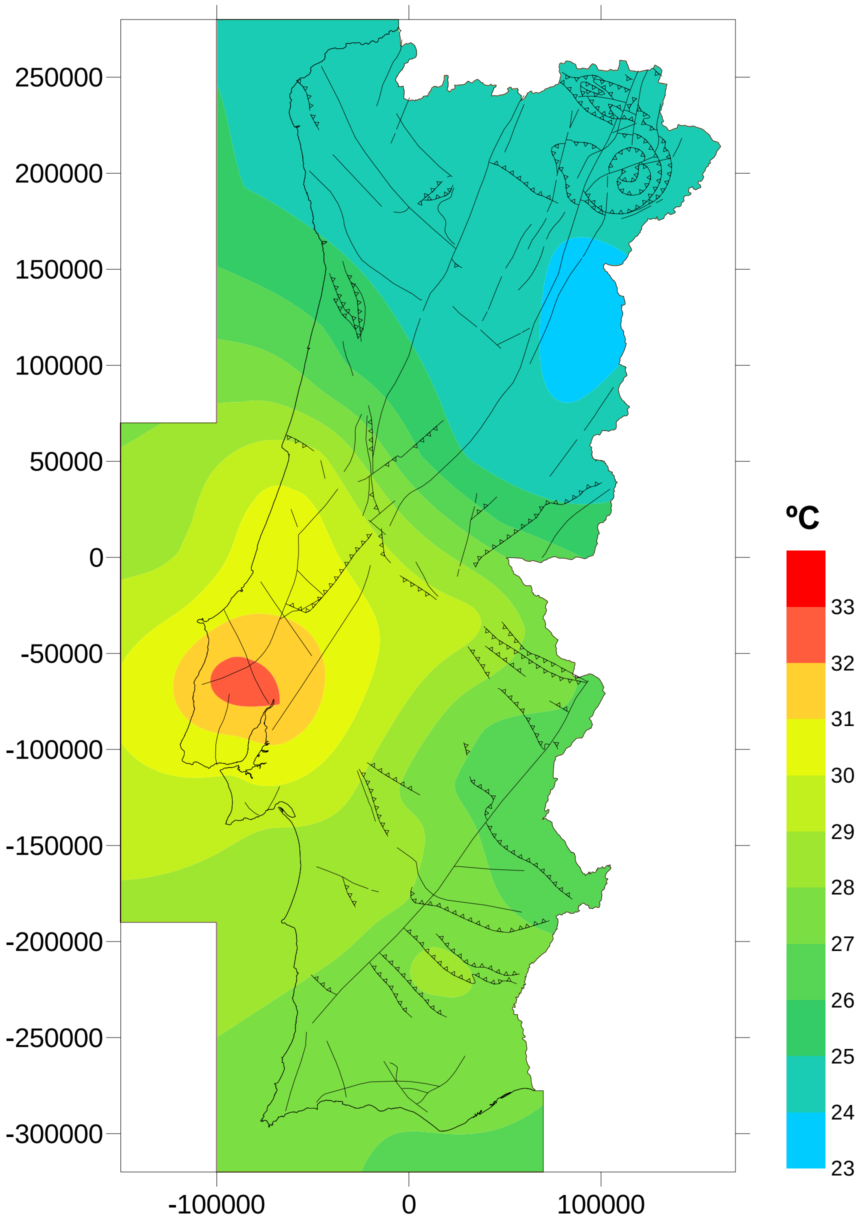
Gradient map



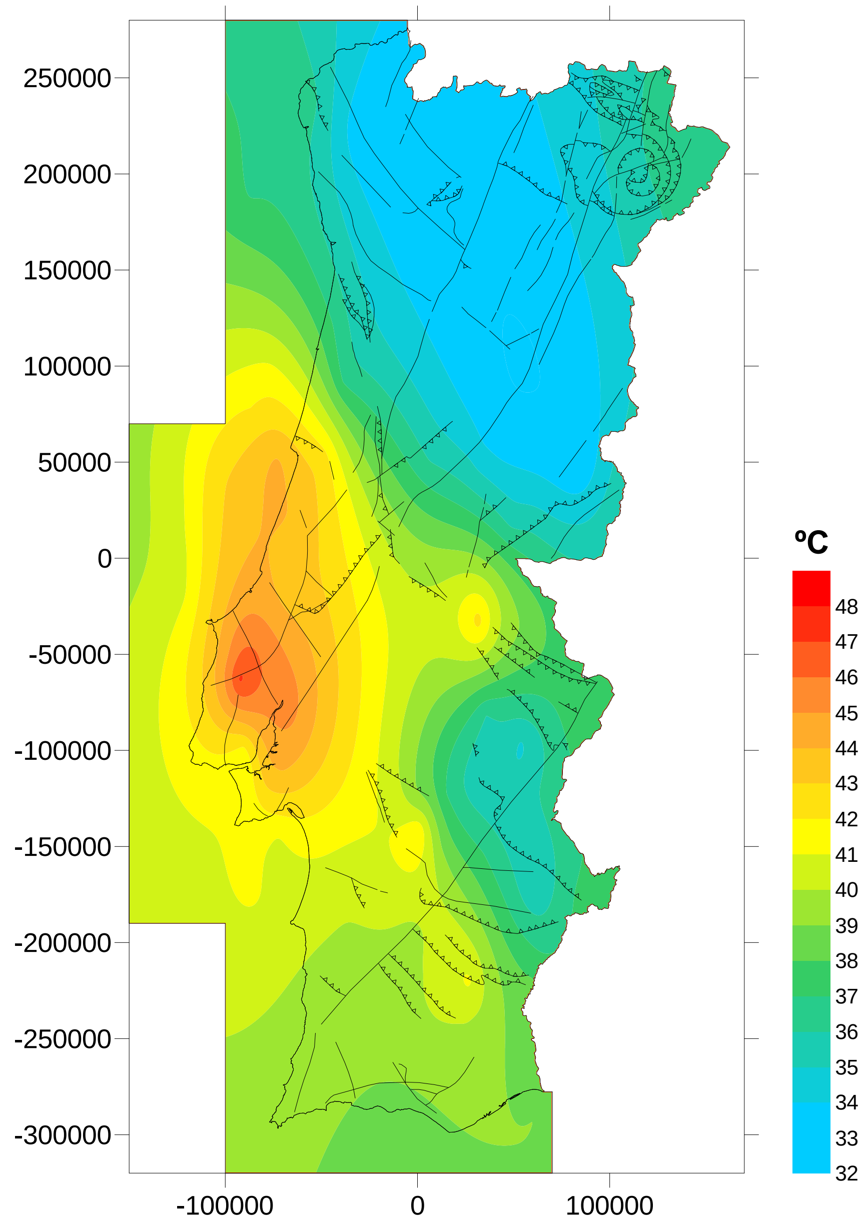
Temperature at 500 m depth



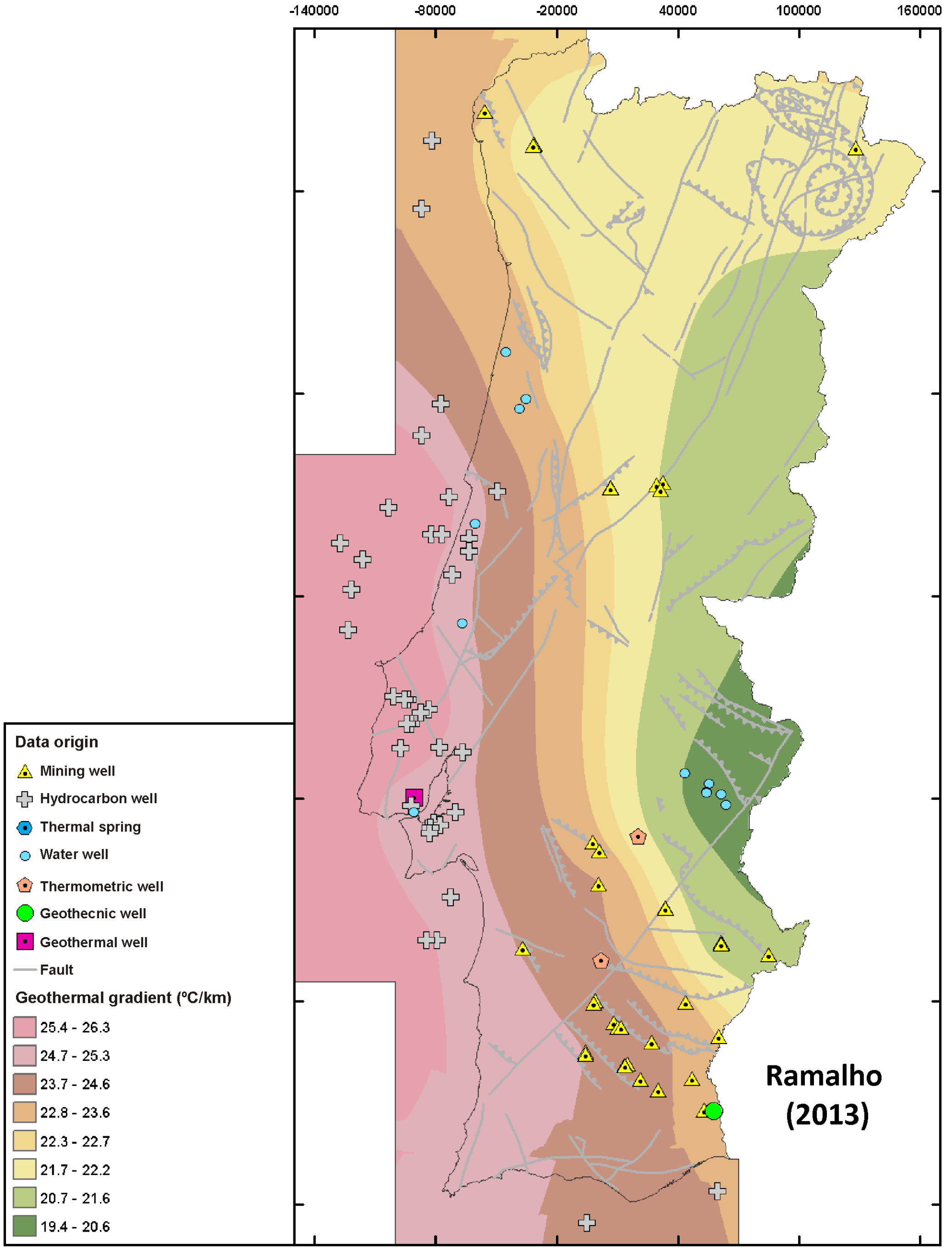
Temperature at 1000 m depth



Temperature at 500 m depth



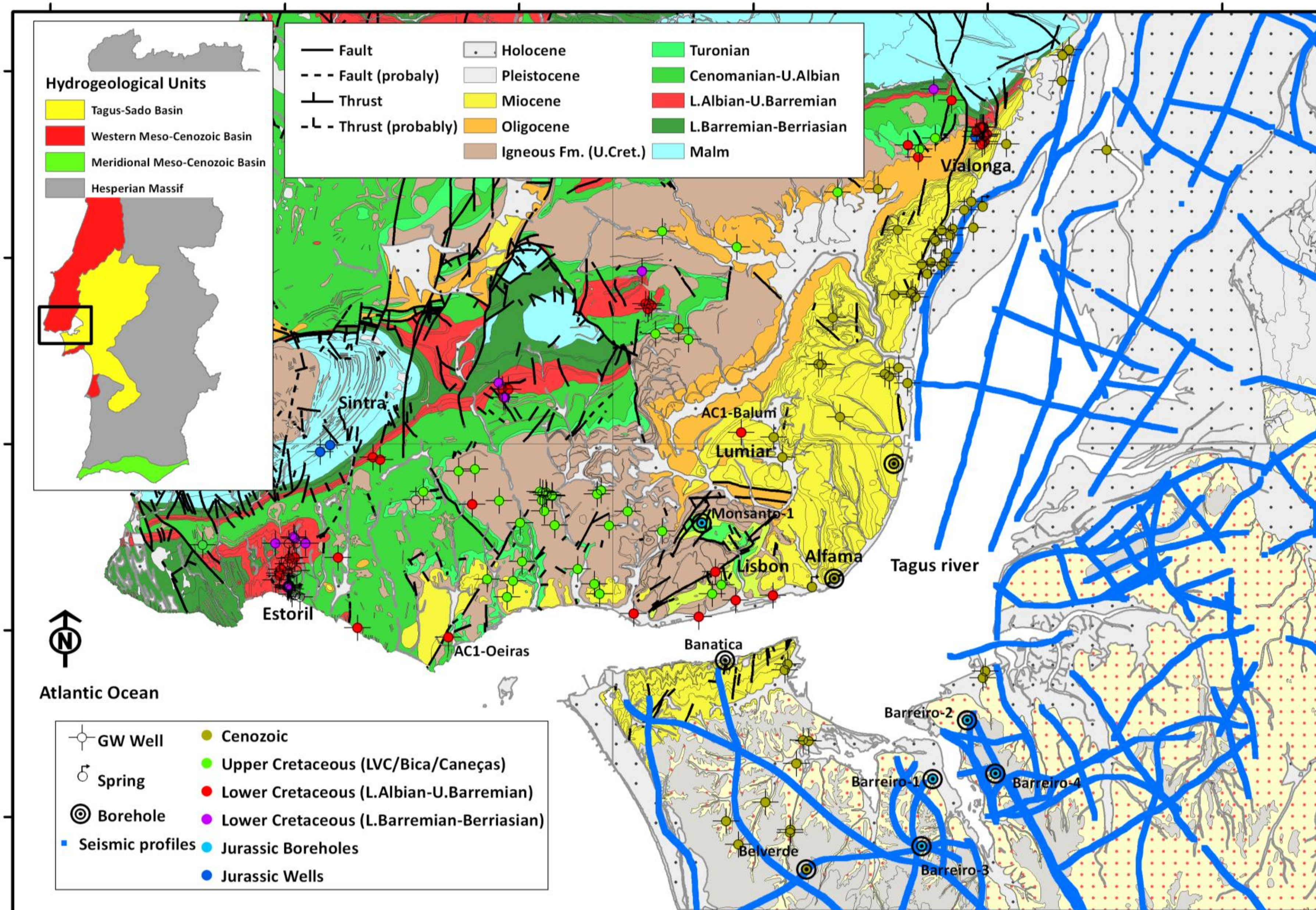
Temperature at 1000 m depth



Geothermal Resources Evaluation

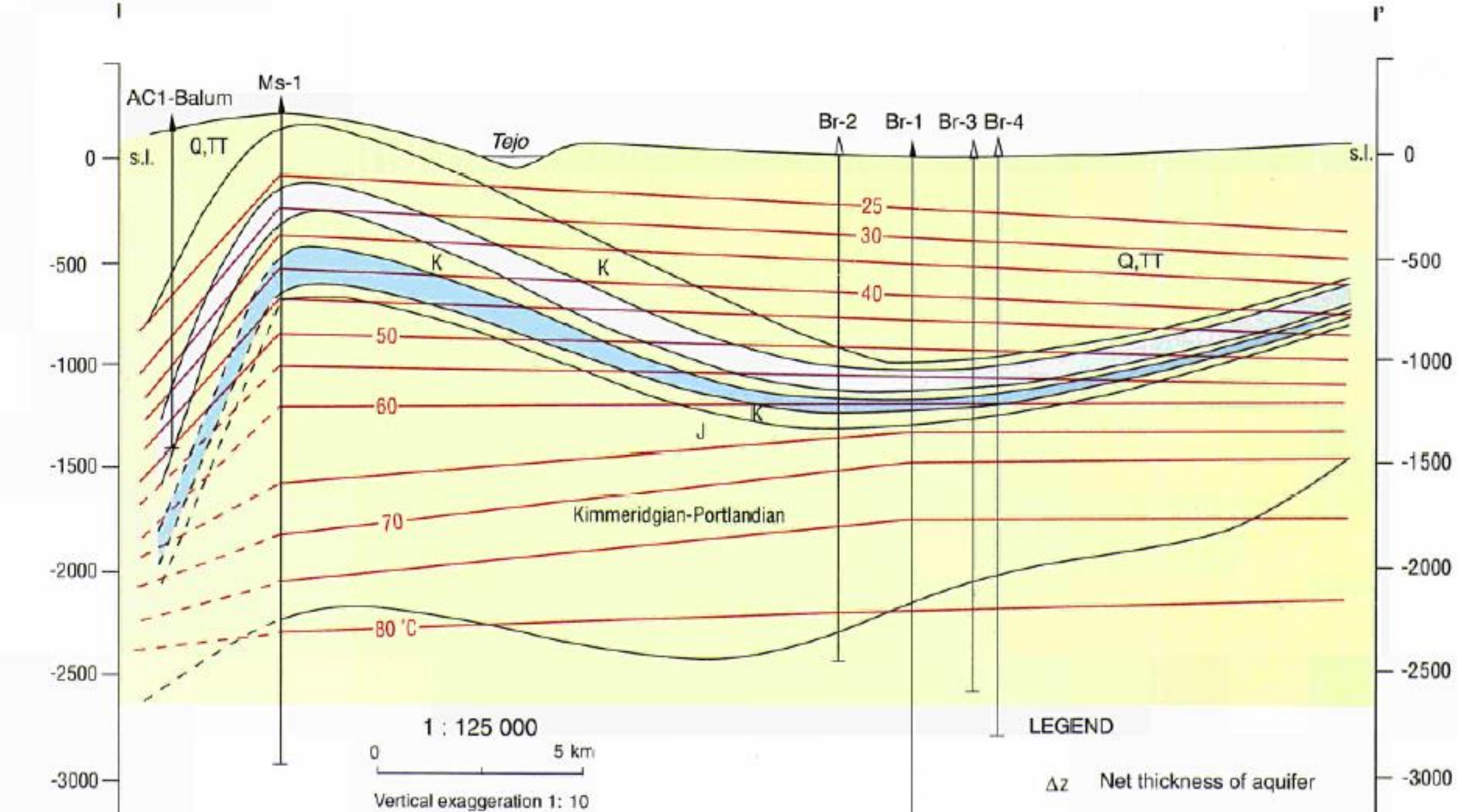
Lisbon region

Geothermal Resources Evaluation



Geothermal Resources Evaluation

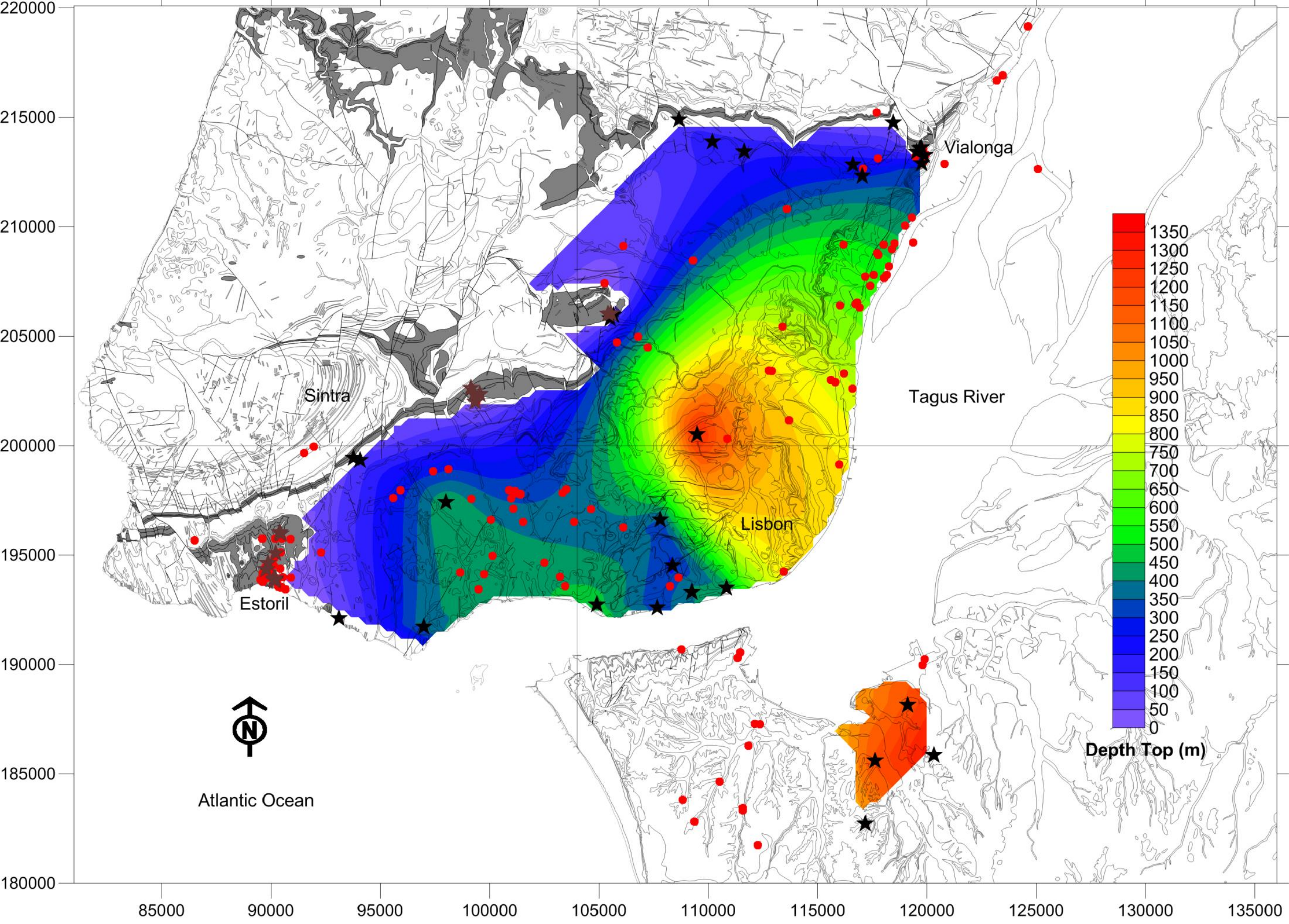
TEJO BASIN
Cross section



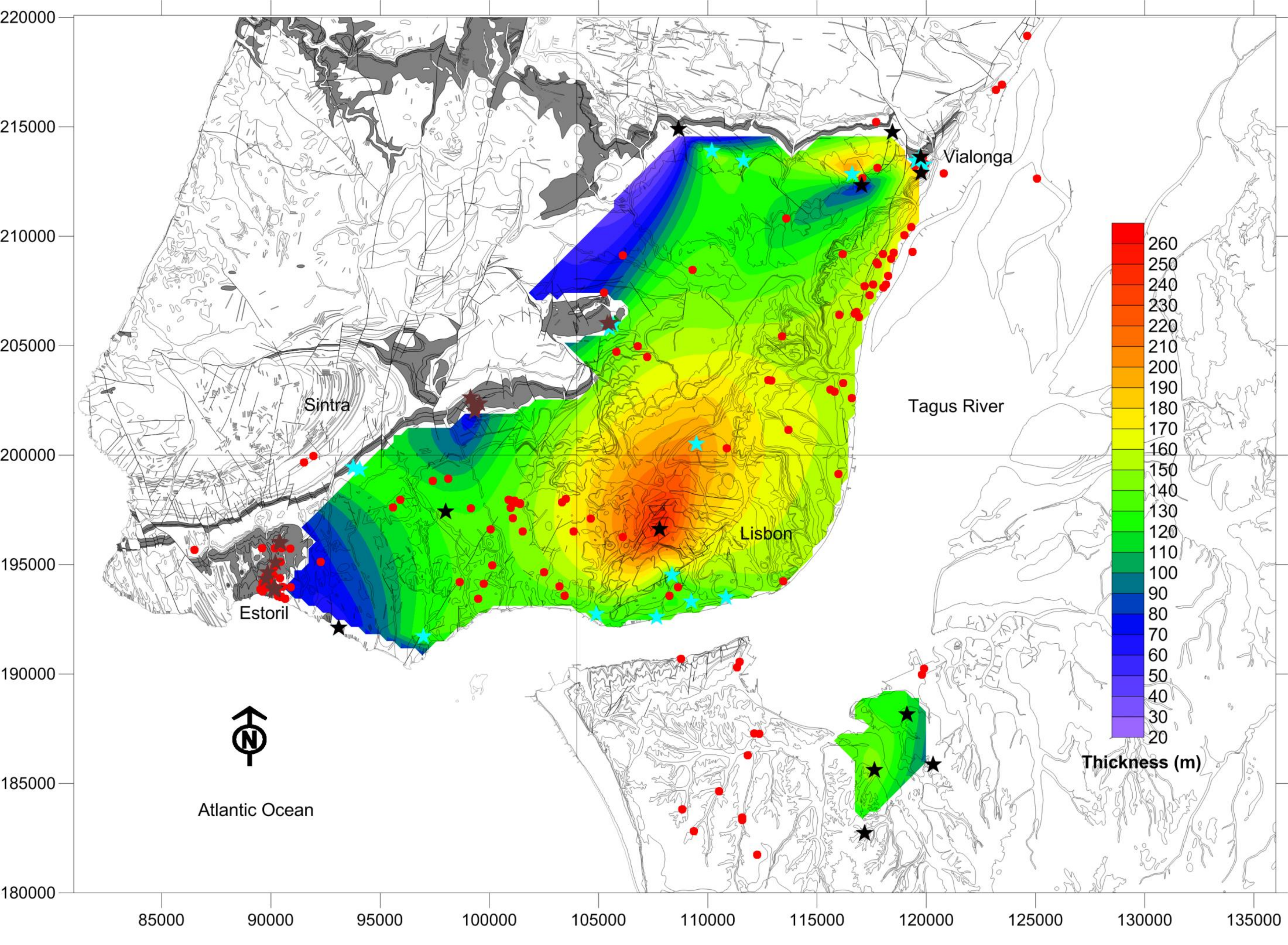
Q - Quaternary TT - Tertiary	Temperatures reaching 32 °C Very low thermal aquifer potentia	
K - Cretaceous	Aptian-Albian aquifer	Well head temperature of 50 °C
J - Jurassic	Temperatures that may reach 75°C at approximately 2400 m depth	

Aquifer	Δz m	P %		Δz m	P %
Aptian - Albian	135	15		125	15
Valanginian	202	15		75	15

Note: The values in the table refer to the aquifer properties at the point immediately above the table



220000
215000
210000
205000
200000
195000
190000
185000
180000
85000
90000
95000
100000
105000
110000
115000
120000
125000
130000
135000
Vialonga
Sintra
Tagus River
Lisbon
Estoril
Atlantic Ocean
N
1350
1300
1250
1200
1150
1100
1050
1000
950
900
850
800
750
700
650
600
550
500
450
400
350
300
250
200
150
100
50
0
Depth Top (m)



Vialonga
Sintra
Tagus River
Lisbon
Estoril
Atlantic Ocean
N
Thickness (m)
260
250
240
230
220
210
200
190
180
170
160
150
140
130
120
110
100
90
80
70
60
50
40
30
20
220000
215000
210000
205000
200000
195000
190000
185000
180000
85000
90000
95000
100000
105000
110000
115000
120000
125000
130000
135000

Heat in place

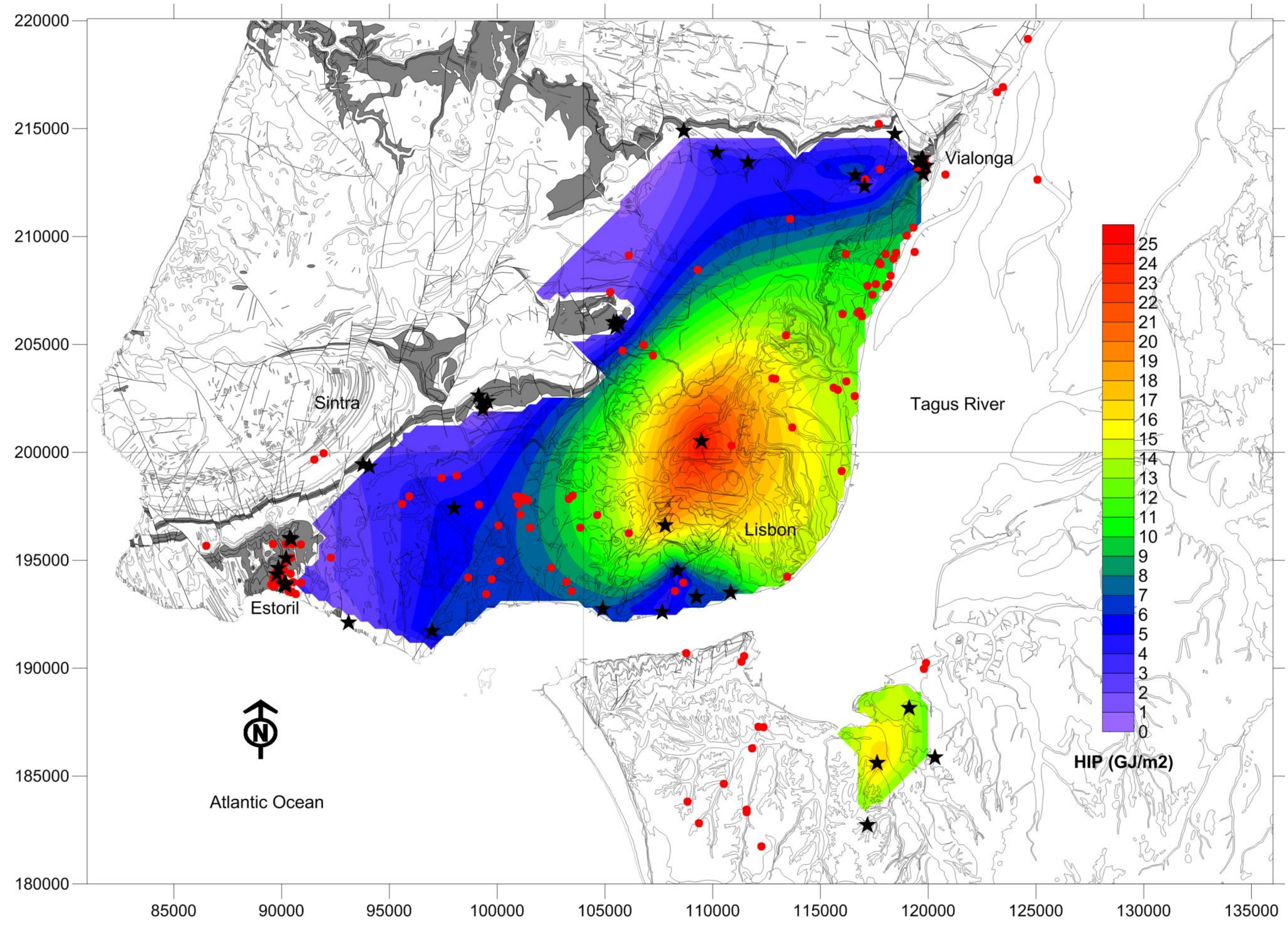
$$HIP = \left(\gamma_w \cdot \phi + \gamma_r \cdot \left(1 - \phi\right)\right) \cdot \mathrm{A} \cdot h \cdot (\mathrm{T_r} - \mathrm{T_0})$$

γ_w 4180 J/kg·^{0}C and γ_r value of 3740 kJ/m^3·^{0}C

Extractable Heat

$$EH = HIP \cdot R_g$$

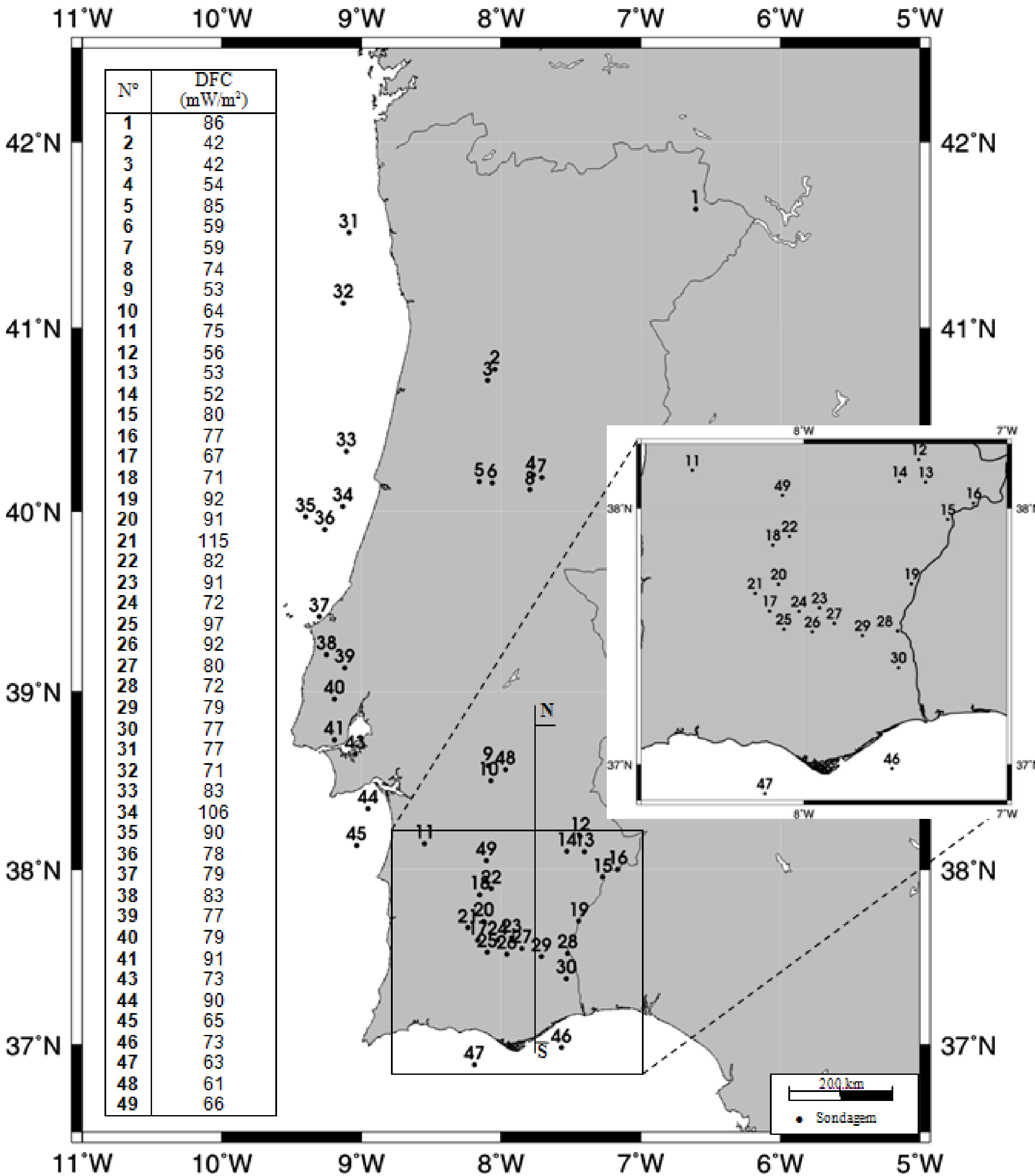
R_g can vary between 0.05 to 0.1



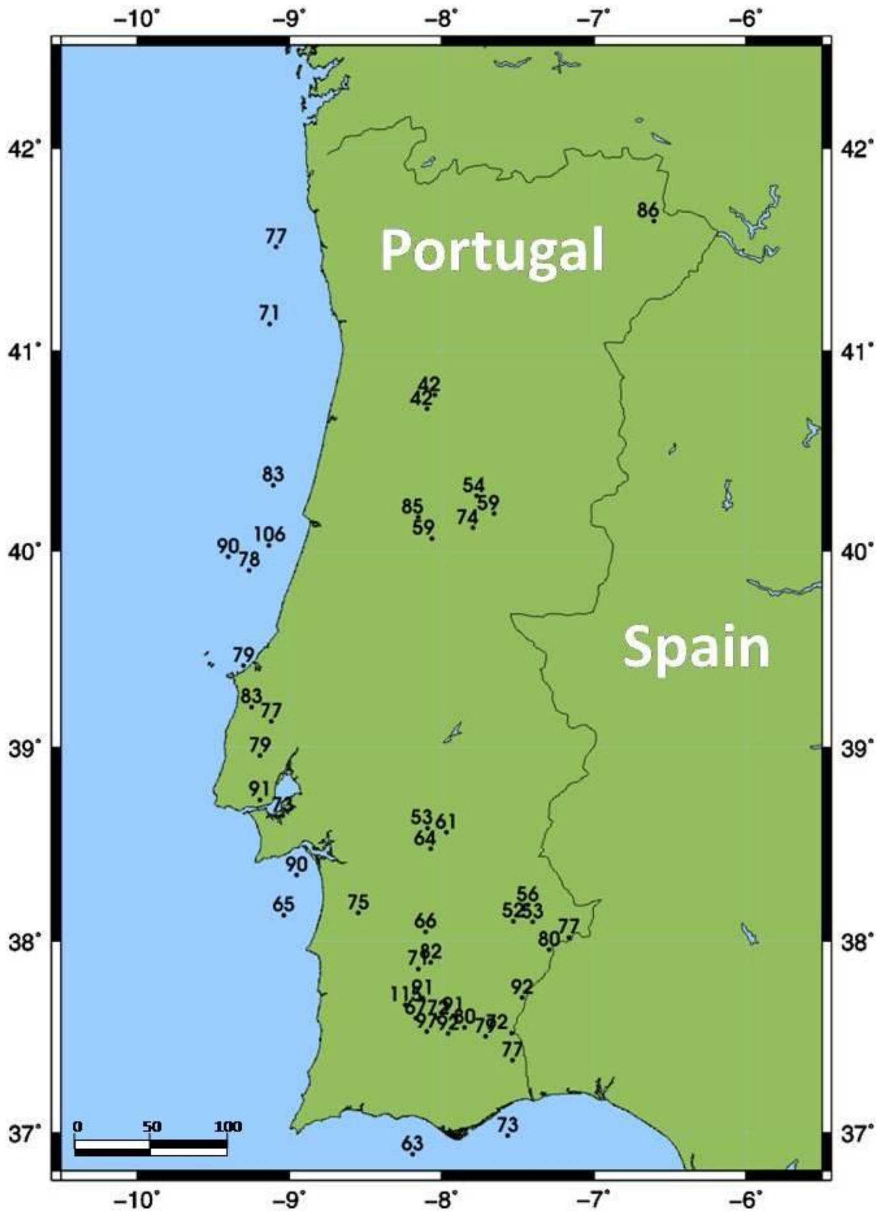
Heat in place (HIP) in GJ/m^2

Conclusions

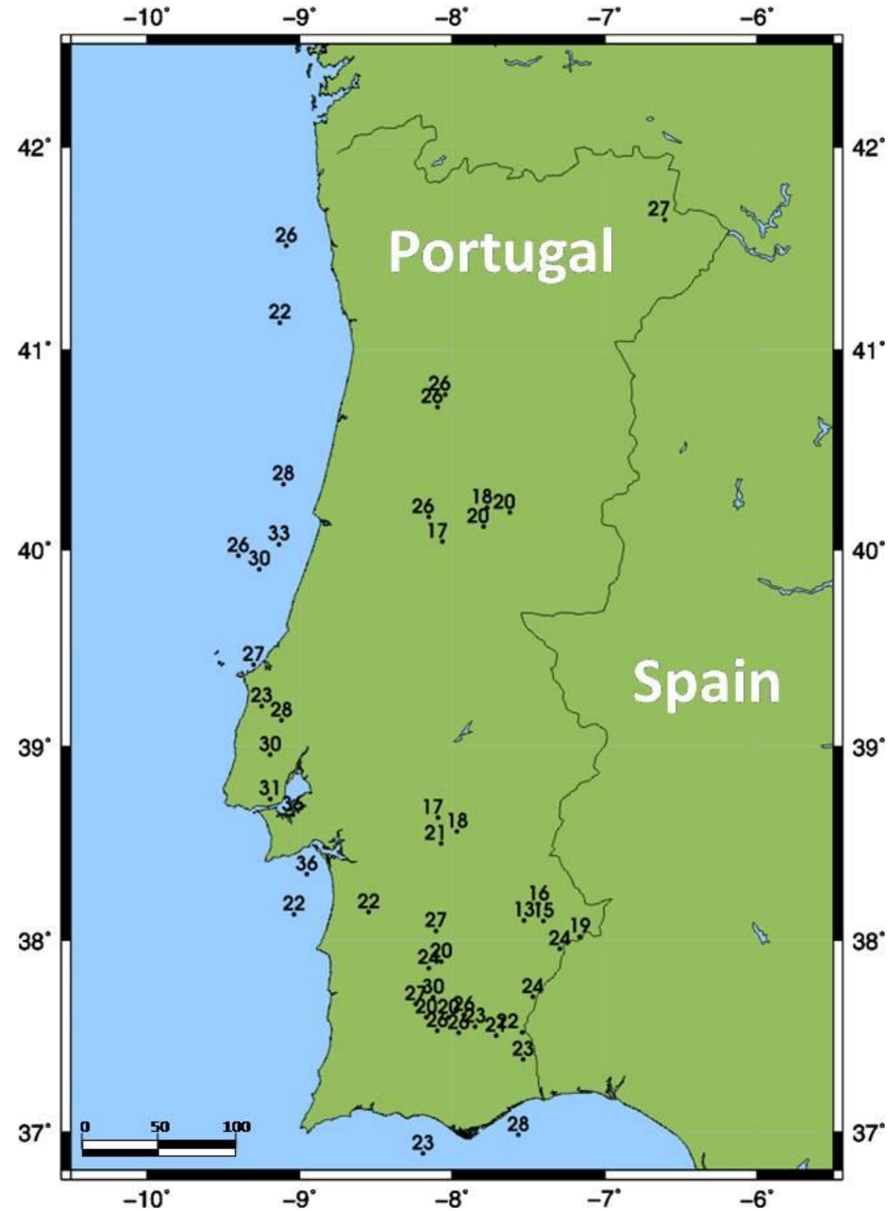
- **The Heat in Place in the Lisbon area is about 3.8 GJ/m^2.**
- **Assuming a recovery factor of 0.15, an Extractable Heat of 0.6 GJ/m^2 was obtained.**
- **This value is significantly lower than other deep sedimentary formations already exploited for district heating, such as the Paris Basin (7 GJ/m^2) or the Upper Rhine Graben (15-30 GJ/m^2).**



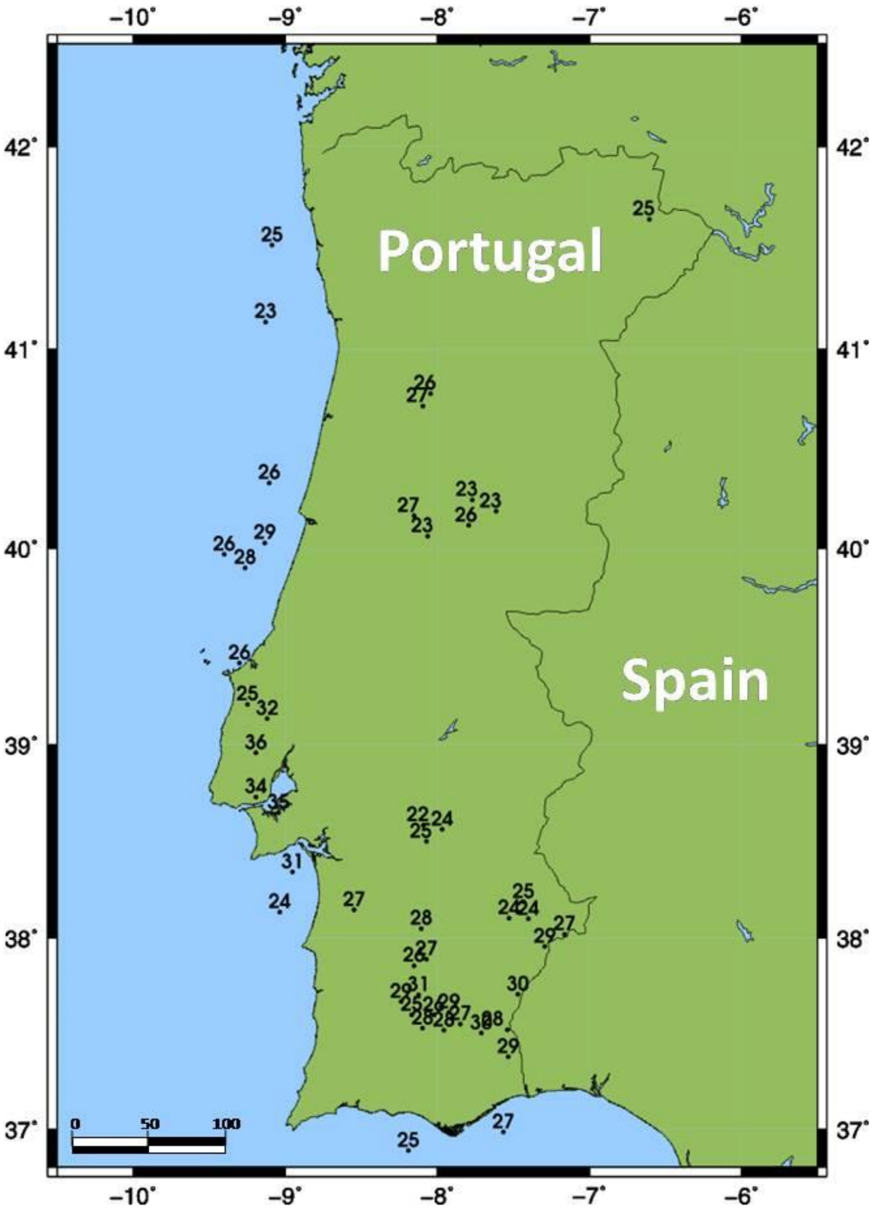
N°	DFC (mW/m²)
1	86
2	42
3	42
4	54
5	85
6	59
7	59
8	74
9	53
10	64
11	75
12	56
13	53
14	52
15	80
16	77
17	67
18	71
19	92
20	91
21	115
22	82
23	91
24	72
25	97
26	92
27	80
28	72
29	79
30	77
31	77
32	71
33	83
34	106
35	90
36	78
37	79
38	83
39	77
40	79
41	91
43	73
44	90
45	65
46	73
47	63
48	61
49	66



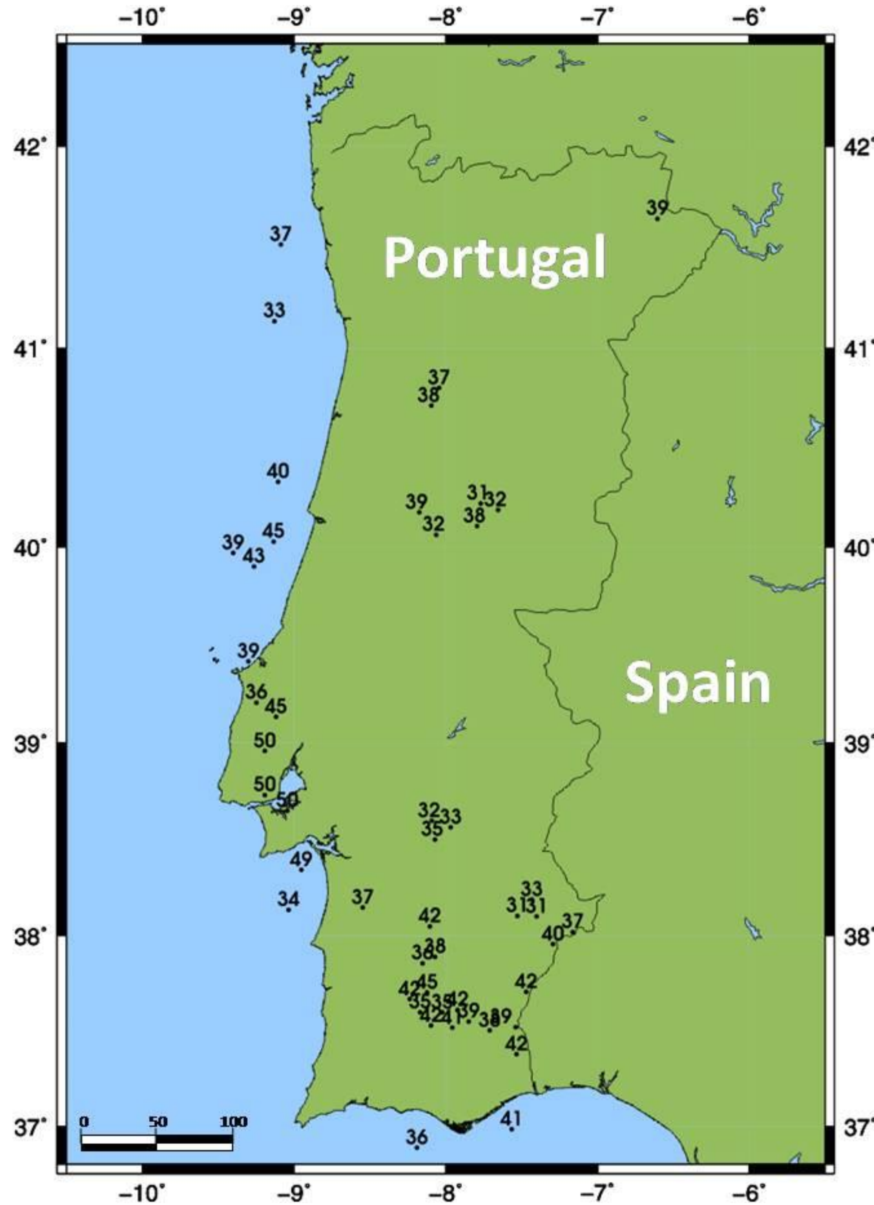
HFD map



Gradient map



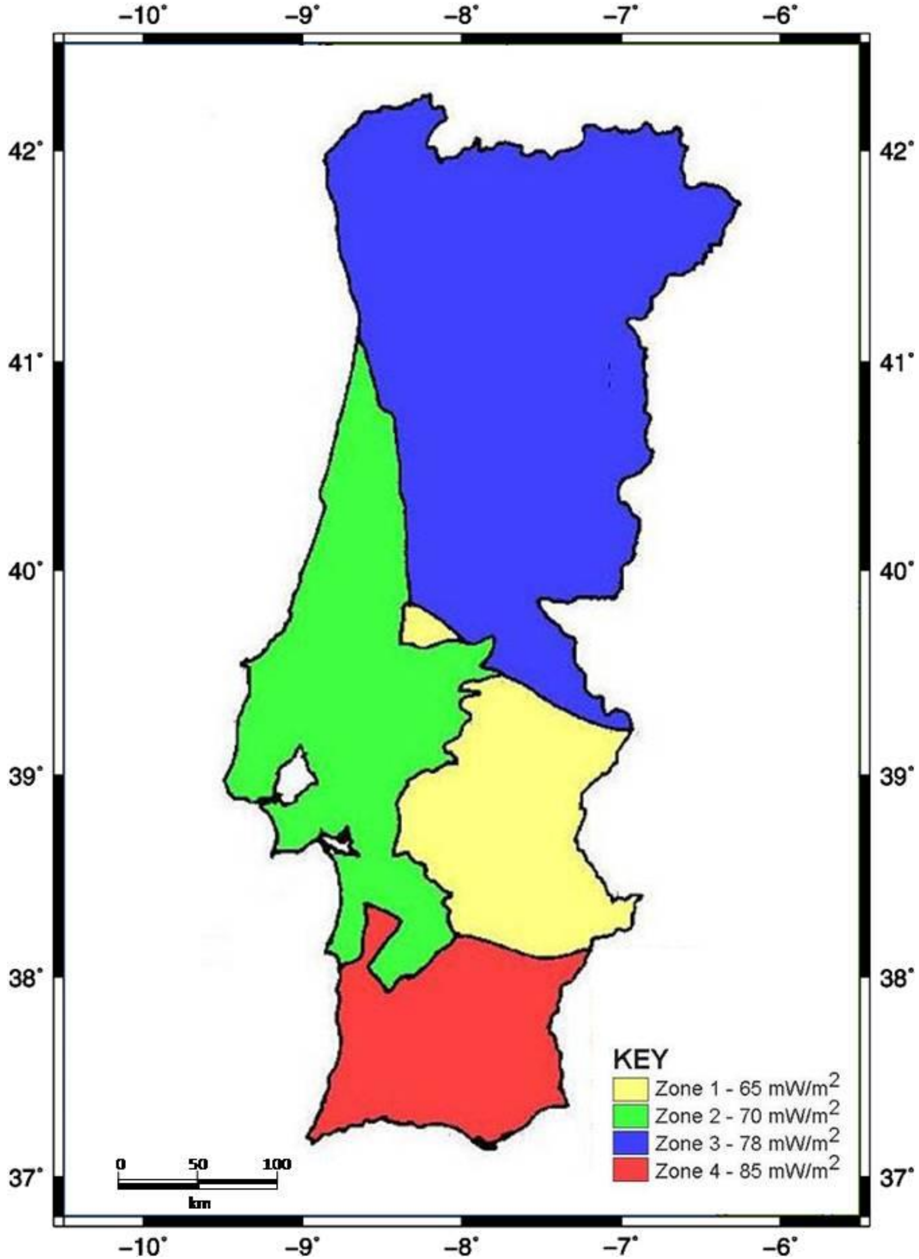
Temperature at 500 m depth

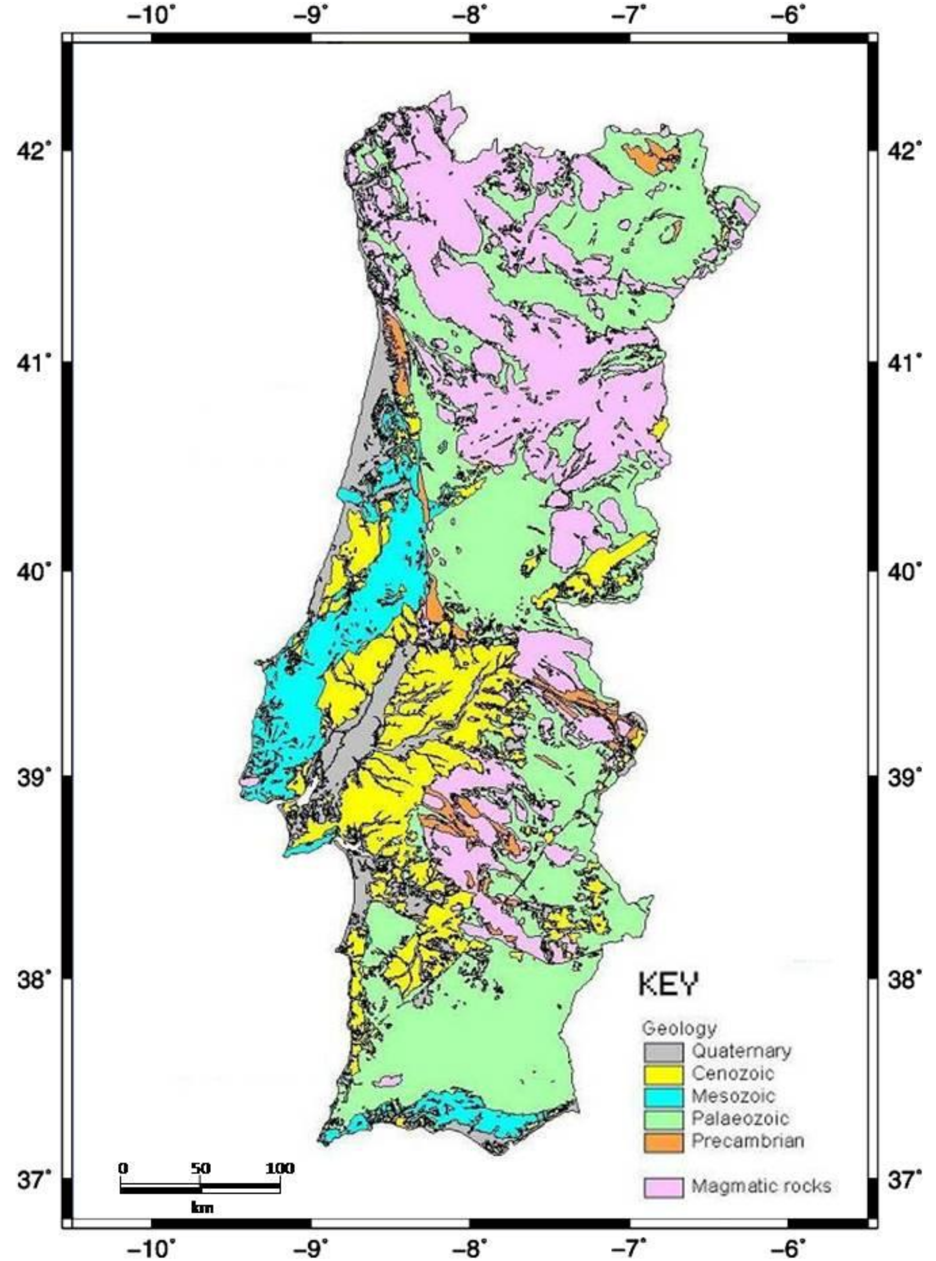


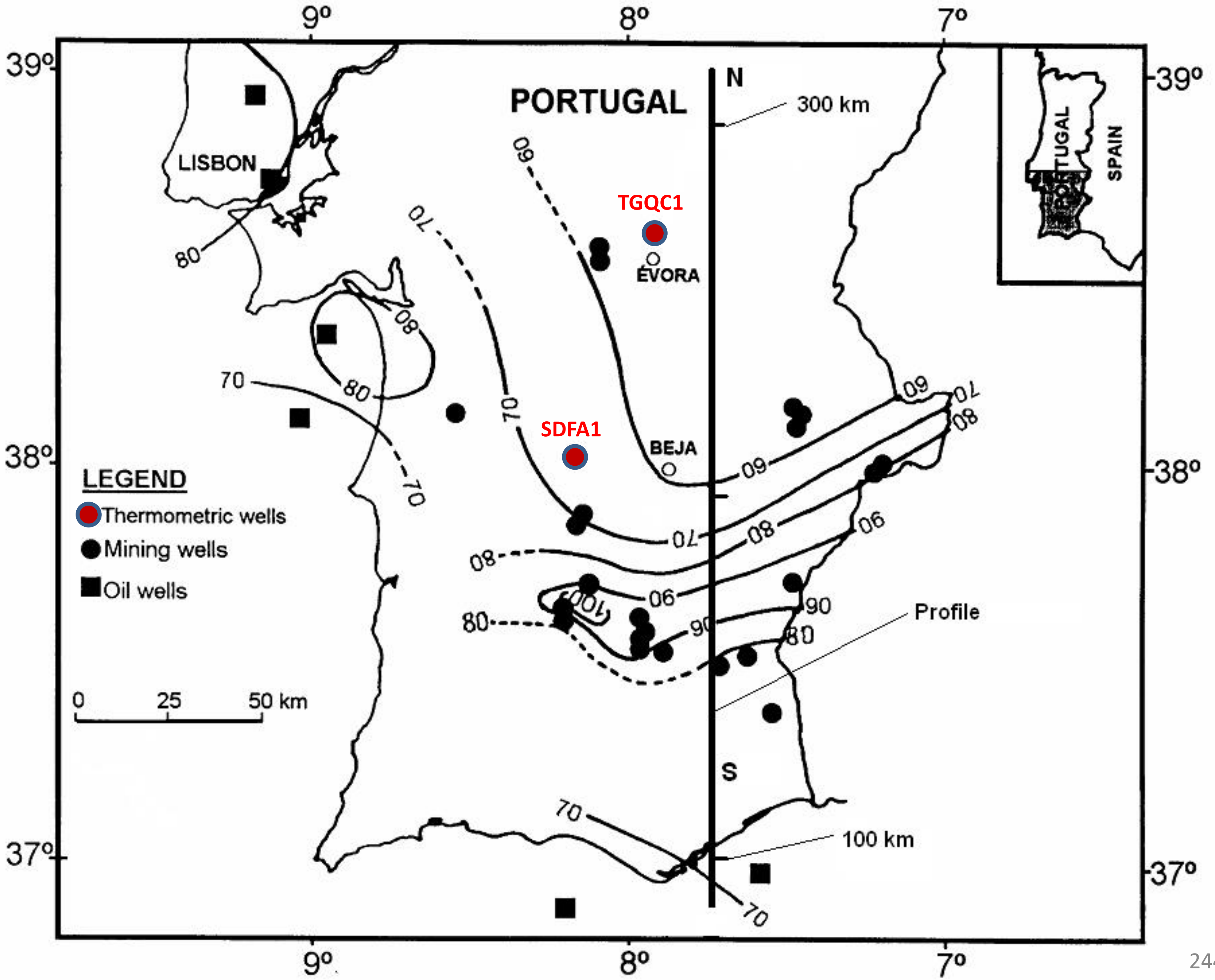
Temperature at 1000 m depth

So, what can we do with this information?

- Study the thermal regime of the crust and mantle
- Use heat flux and geothermometers
- Construct maps of geothermal energy potential
- Study climatic change
- Study rheology of the crust and mantle


KEY
Zone 1 - 65 mW/m2
Zone 2 - 70 mW/m2
Zone 3 - 78 mW/m2
Zone 4 - 85 mW/m2
0 50 100
km


KEY
Geology
Quaternary
Cenozoic
Mesozoic
Palaeozoic
Precambrian
Magmatic rocks
0 50 100
km

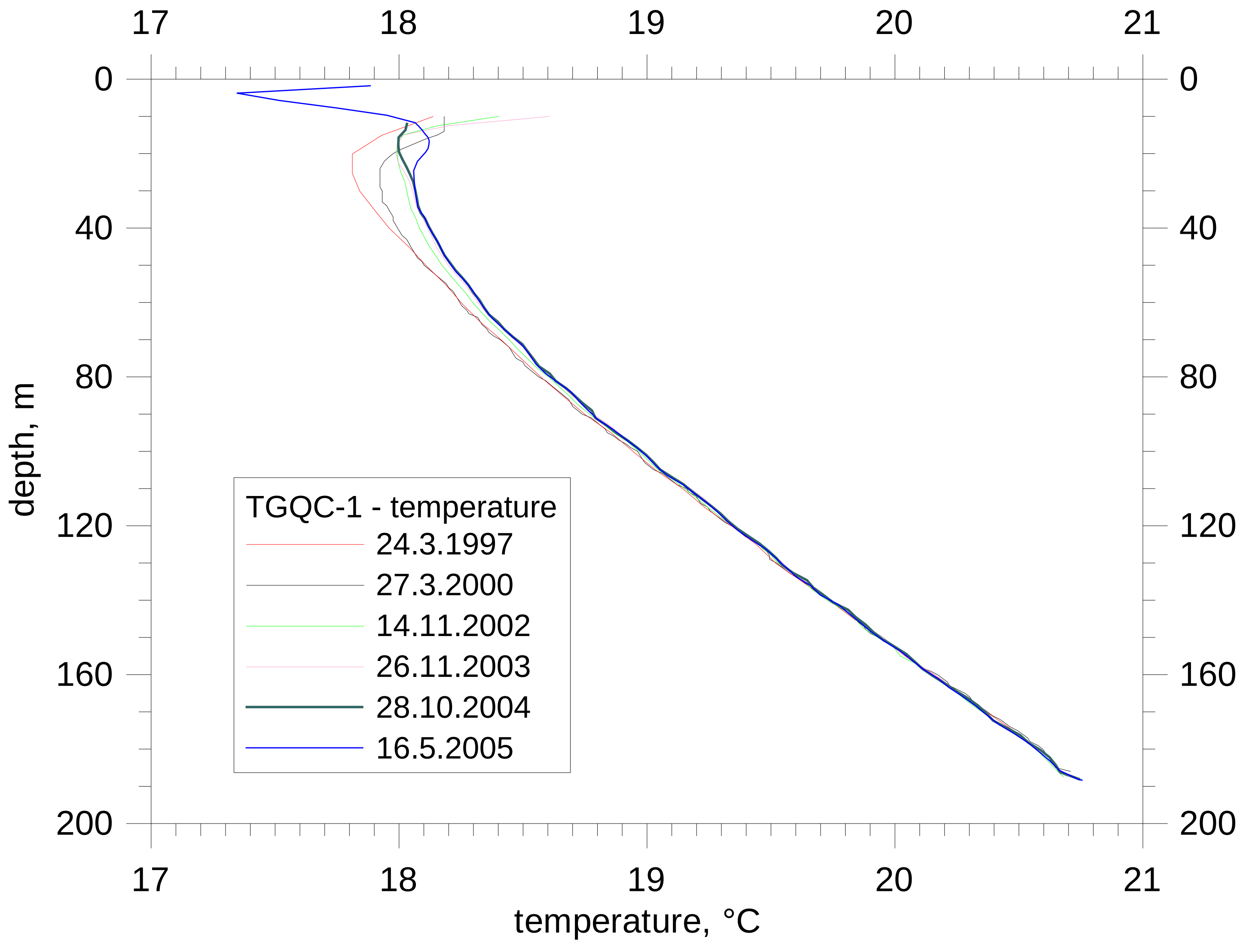

PORTUGAL
N
300 km
LISBON
TGQC1
ÉVORA
SDFA1
BEJA
LEGEND
Thermometric wells
Mining wells
Oil wells
0 25 50 km
Profile
S
100 km
SPAIN
9º
8º
7º
39º
38º
37º

Geothermal Climate Change Observatory in the TGQC-1 well

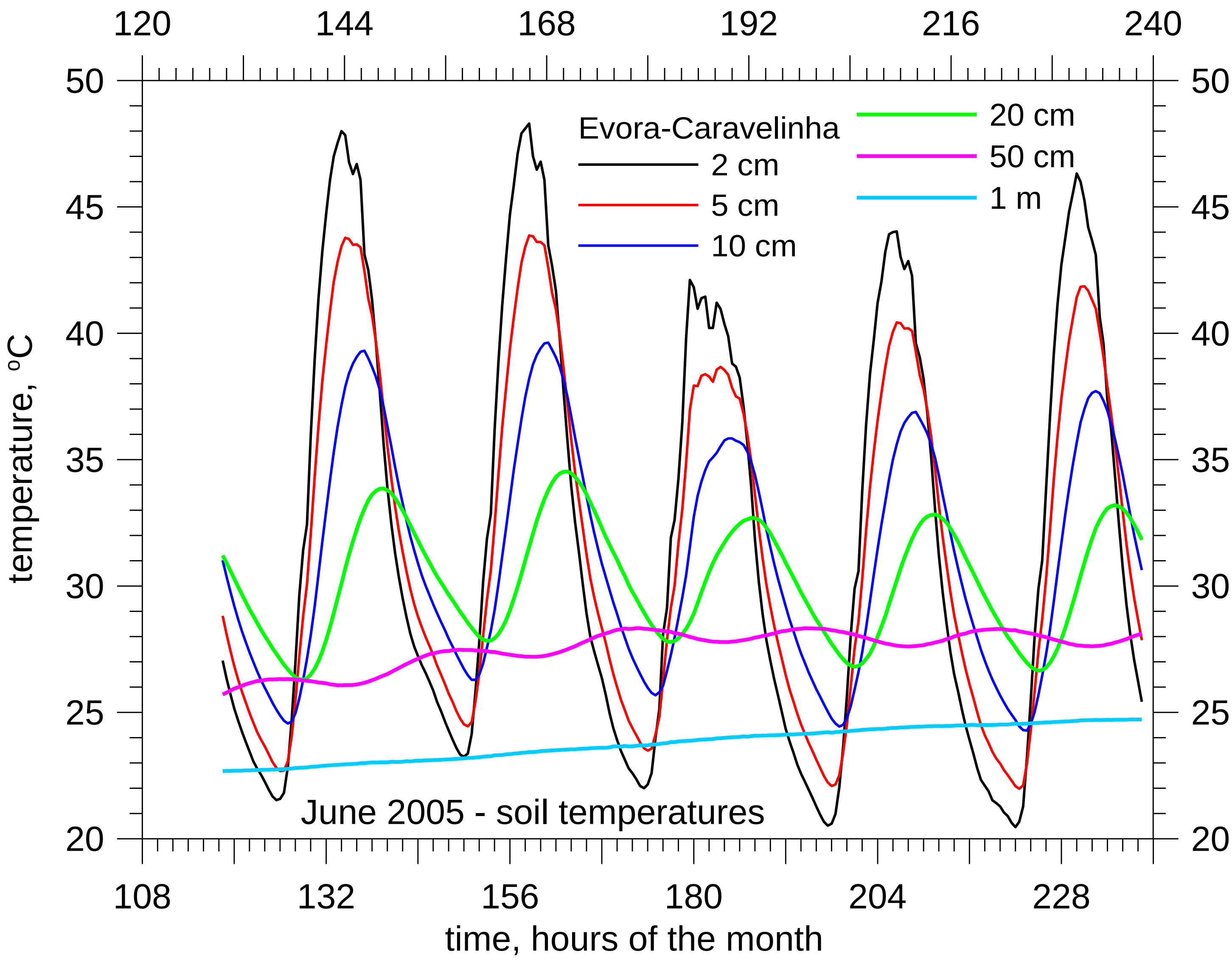
Depth (m) of sensors in the borehole:

0.02 0.05 0.10
0.20 0.50 1.0
2.0 5.0 10.0
20.0 30.0 e 40.0

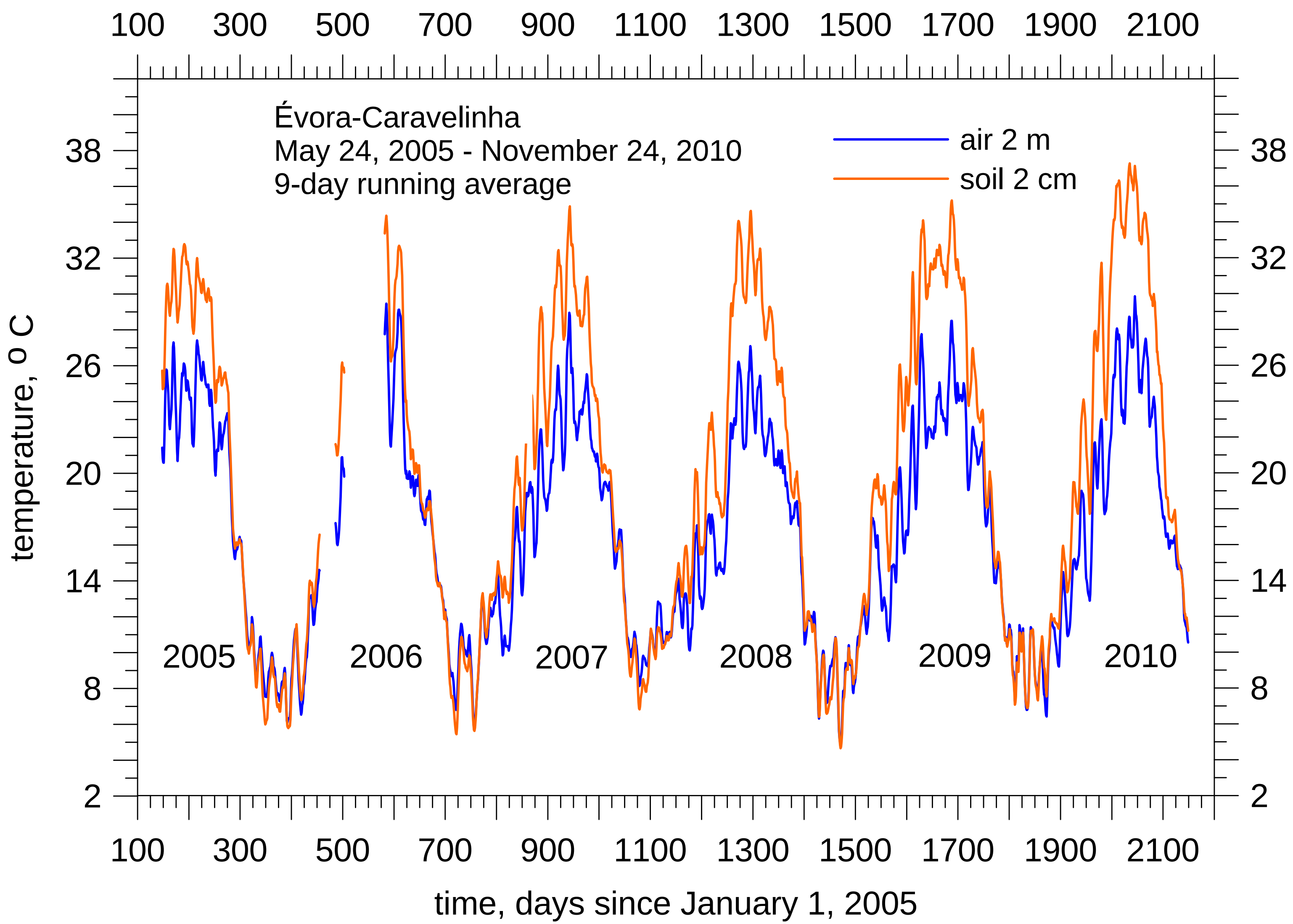
Temperature logs for different years



Temperature time series I



Temperature time series II



Conclusions

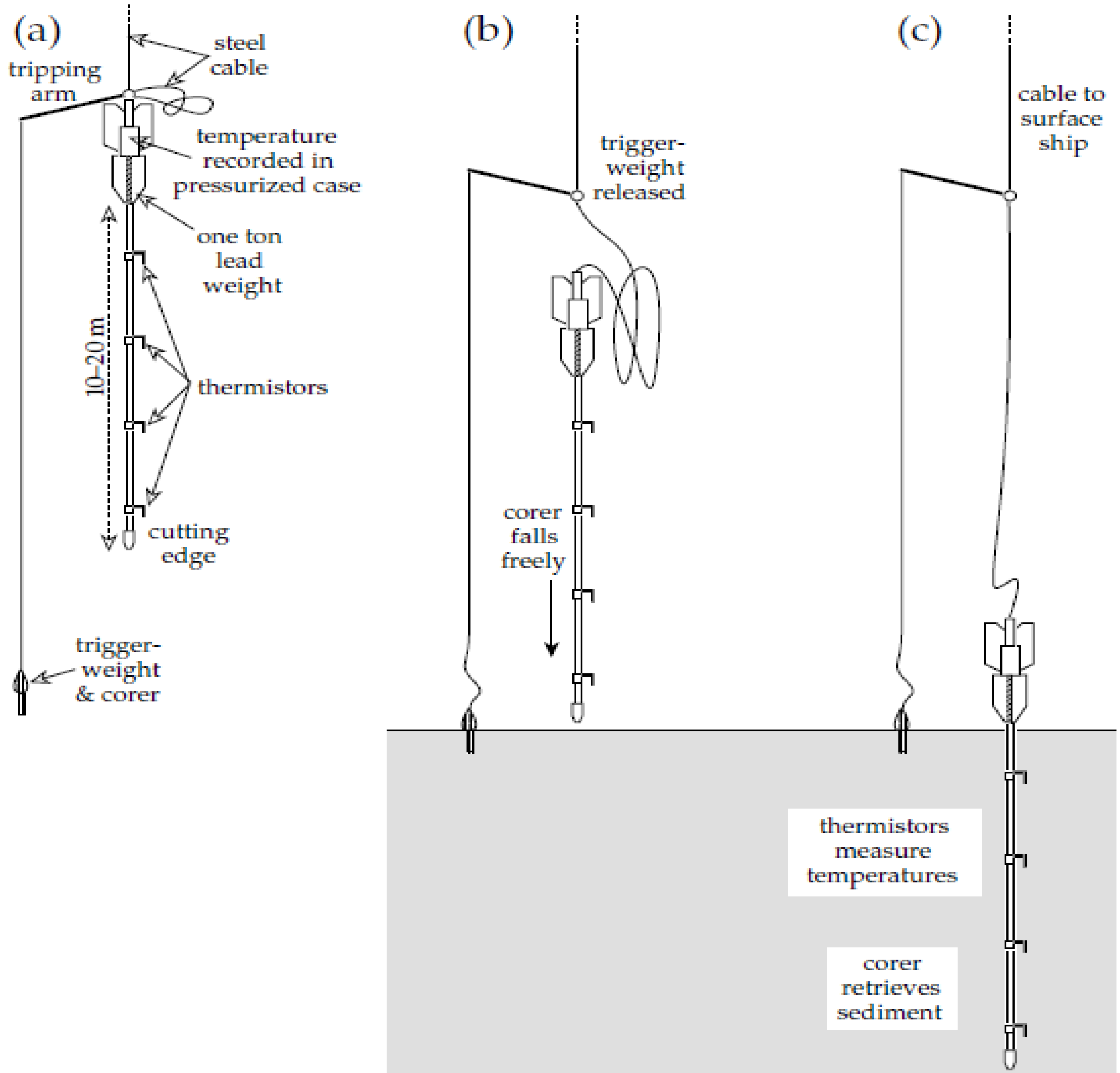
- **Temperature logs can be a useful tool to infer the past climate change.**

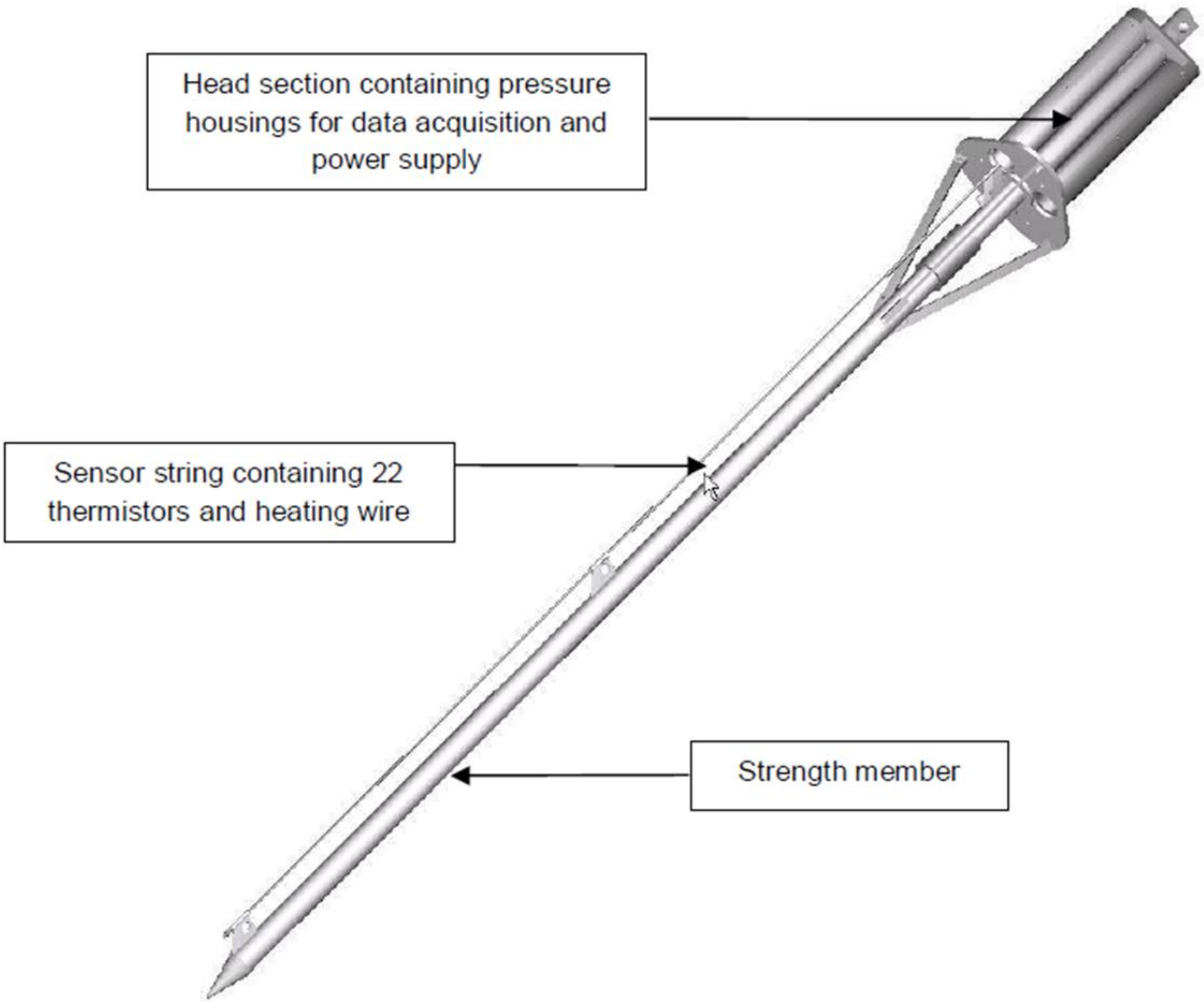
- **The GST reconstruction for the TGQC-1 well indicates a warming trend.**

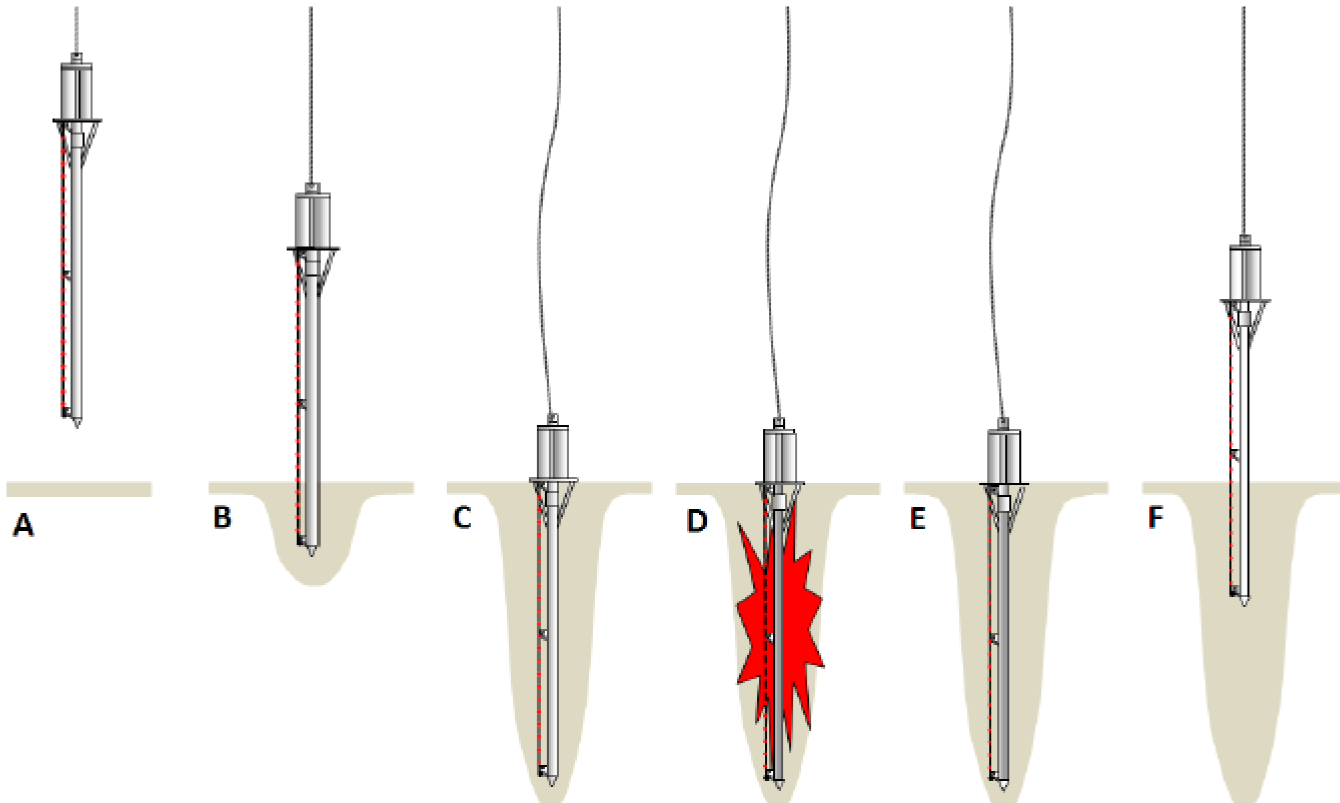
- **The warming trend appears to accelerate in the last 20 to 25 years.**

- **The installation of a geothermal climate change laboratory in the TGQC-1 well hopefully will allow understanding air-ground coupling and so constrain future GST inversions.**

VII
Heat Flow Density in the Sea


(a)
steel cable
tripping arm
temperature recorded in pressurized case
one ton lead weight
10–20 m
thermistors
cutting edge
trigger-weight & corer
(b)
trigger-weight released
corer falls freely
(c)
cable to surface ship
thermistors measure temperatures
corer retrieves sediment


Head section containing pressure housings for data acquisition and power supply
Sensor string containing 22 thermistors and heating wire
Strength member



A: lowering to seabed

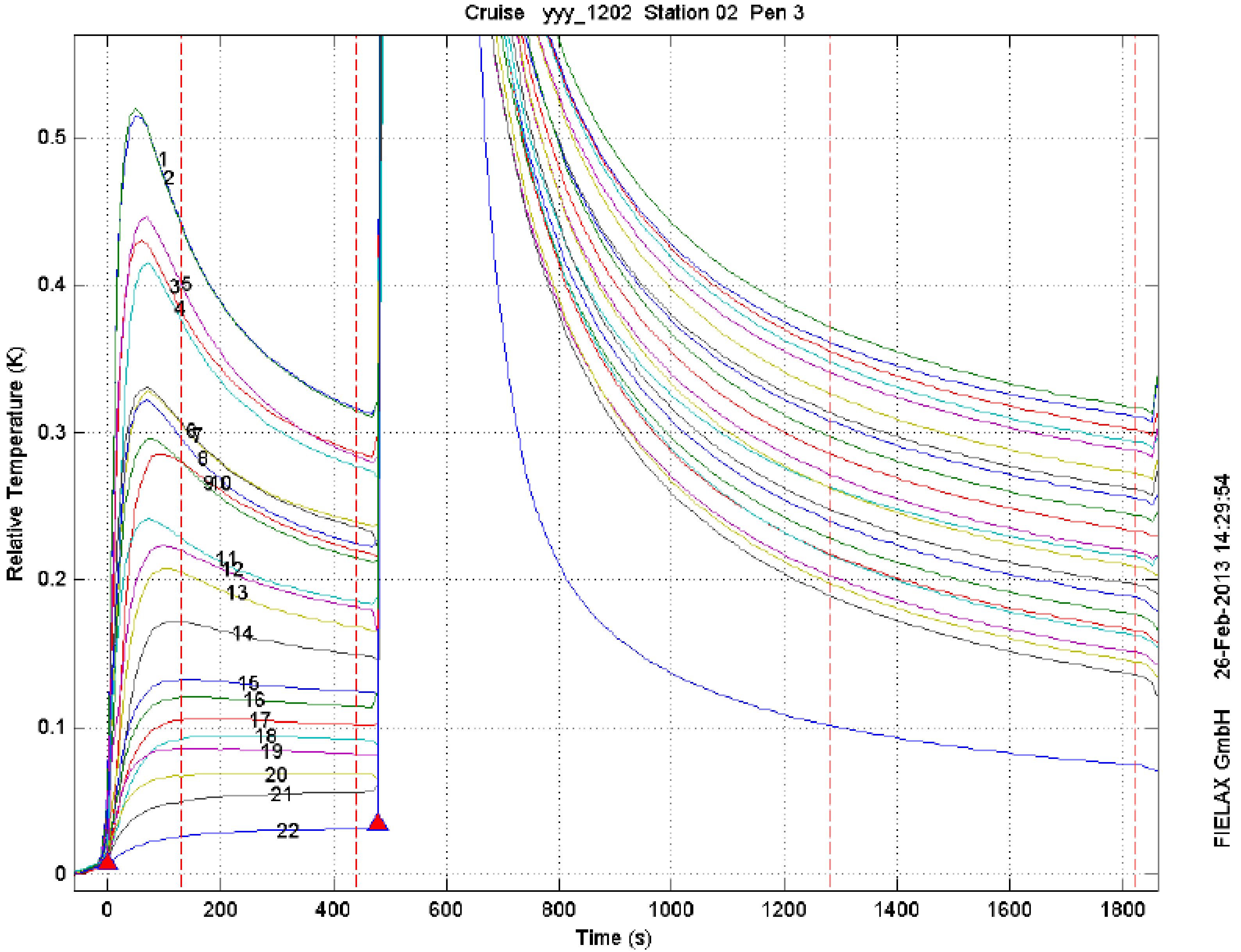
B: penetrating into seabed

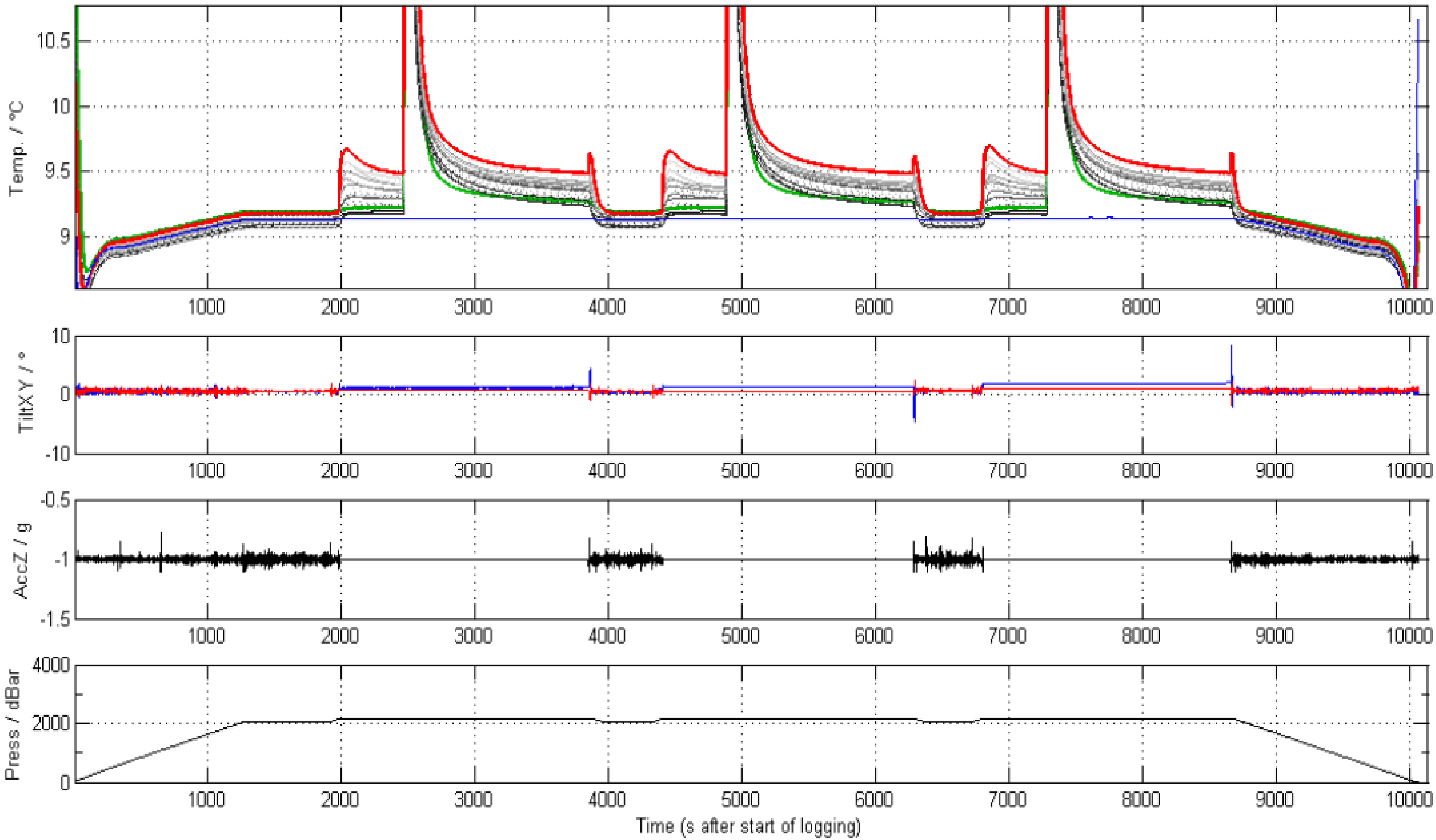
C: measuring thermal decay of frictional heat (approx.. 7-12 minutes)

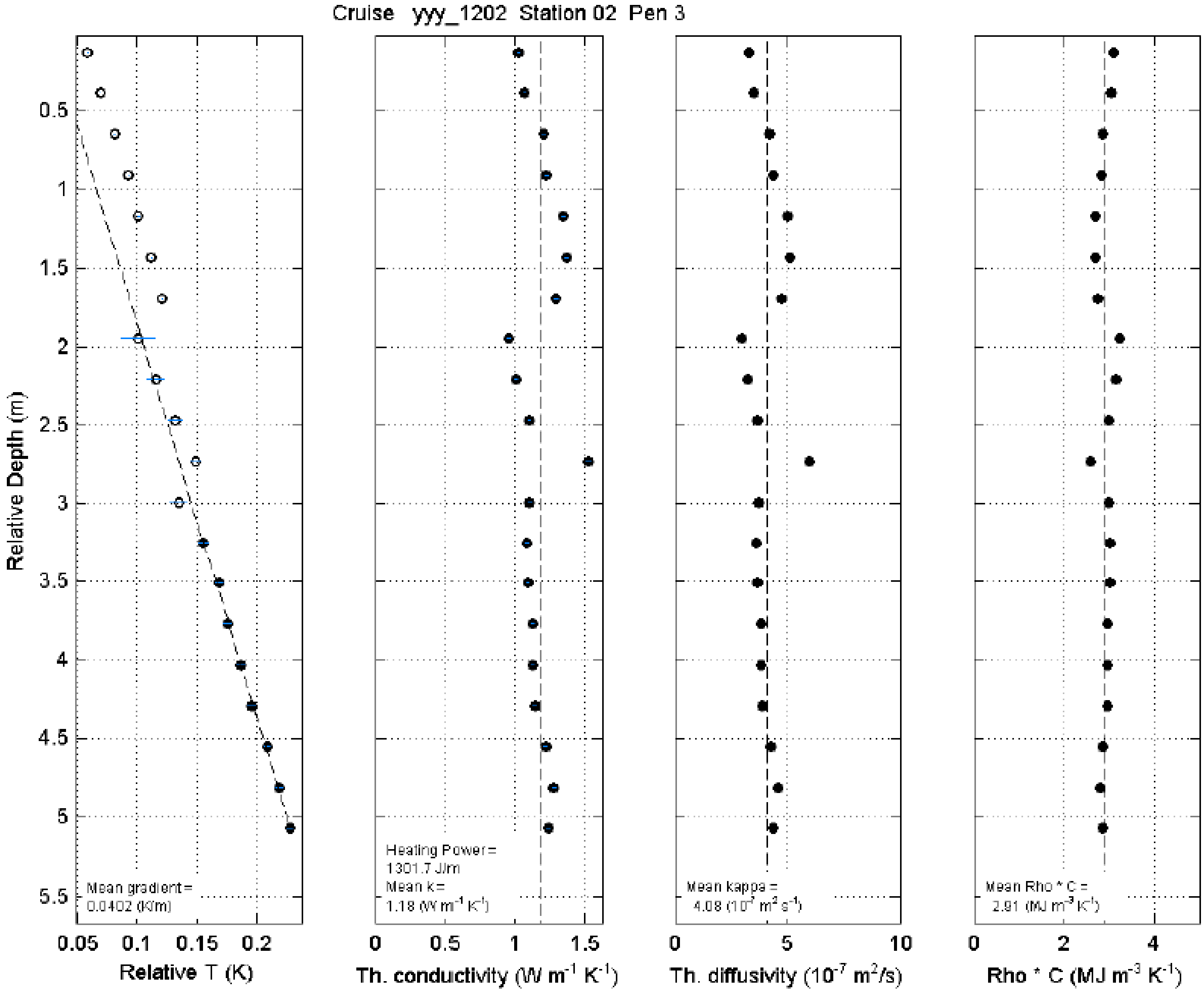
D: heat pulse (approx.. 20 seconds)

E: measuring thermal decay of heat pulse (approx.. 15-20 minutes)

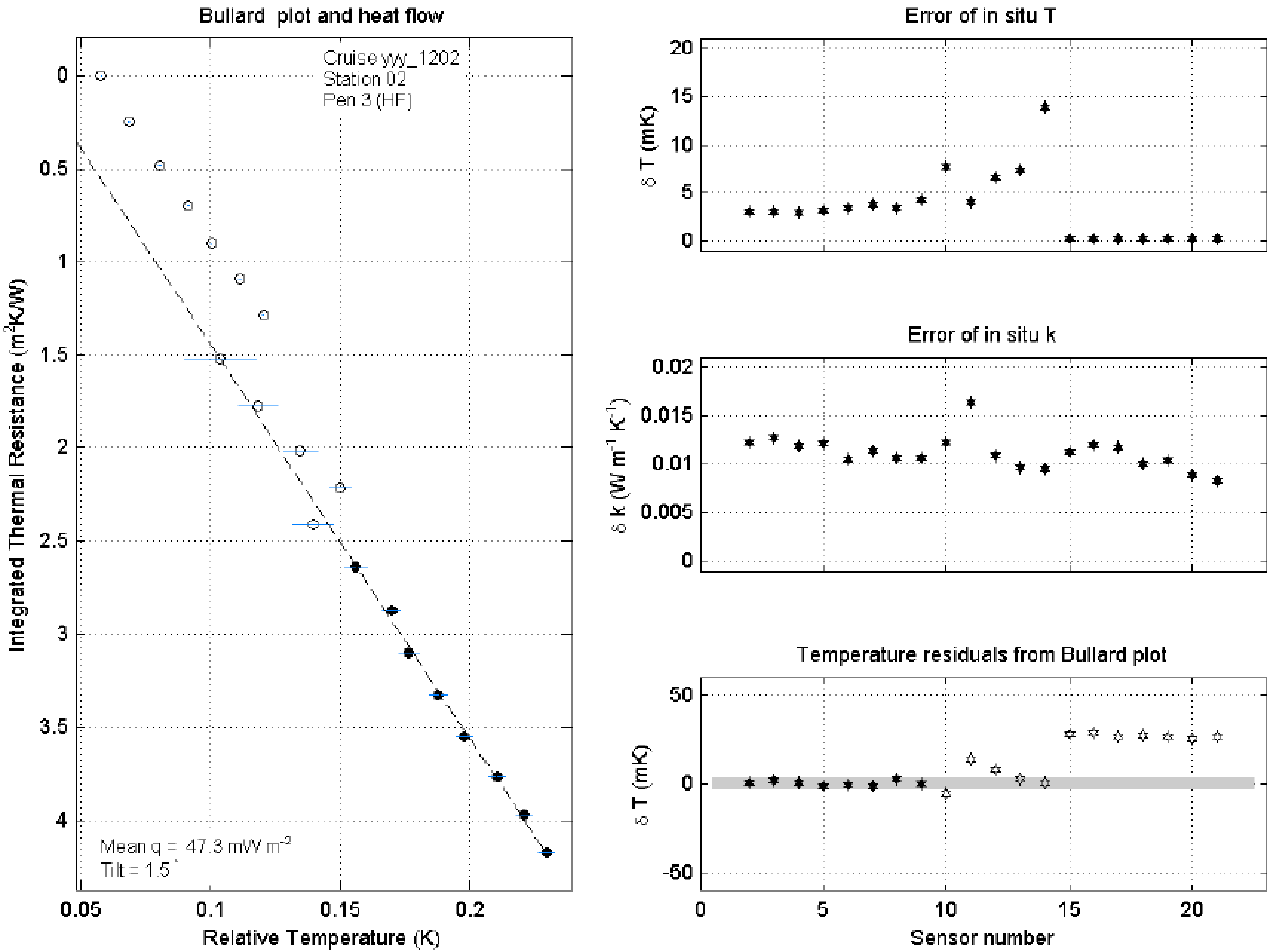
F: pullout and retrieve to surface


Cruise yyy_1202 Station 02 Pen 3
Relative Temperature (K)
Time (s)
0
0.1
0.2
0.3
0.4
0.5
0
200
400
600
800
1000
1200
1400
1600
1800
1
2
35
4
67
8
910
11
12
13
14
15
16
17
18
19
20
21
22
26-Feb-2013 14:29:54
FIELAX GmbH


Temp. / °C
10.5
10
9.5
9
TiltX Y / °
10
0
-10
AccZ / g
-0.5
-1
-1.5
Press / dBar
4000
2000
0
1000 2000 3000 4000 5000 6000 7000 8000 9000 10000
Time (s after start of logging)



Cruise yyy_1202 Station 02 Pen 3
Relative Depth (m)
Relative T (K)
Mean gradient = 0.0402 (K/m)
Th. conductivity (W m-1 K-1)
Heating Power = 1301.7 J/m
Mean k = 1.18 (W m-1 K-1)
Th. diffusivity (10-7 m2/s)
Mean kappa = 4.08 (10-7 m2 s-1)
Rho * C (MJ m-3 K-1)
Mean Rho * C = 2.91 (MJ m-3 K-1)

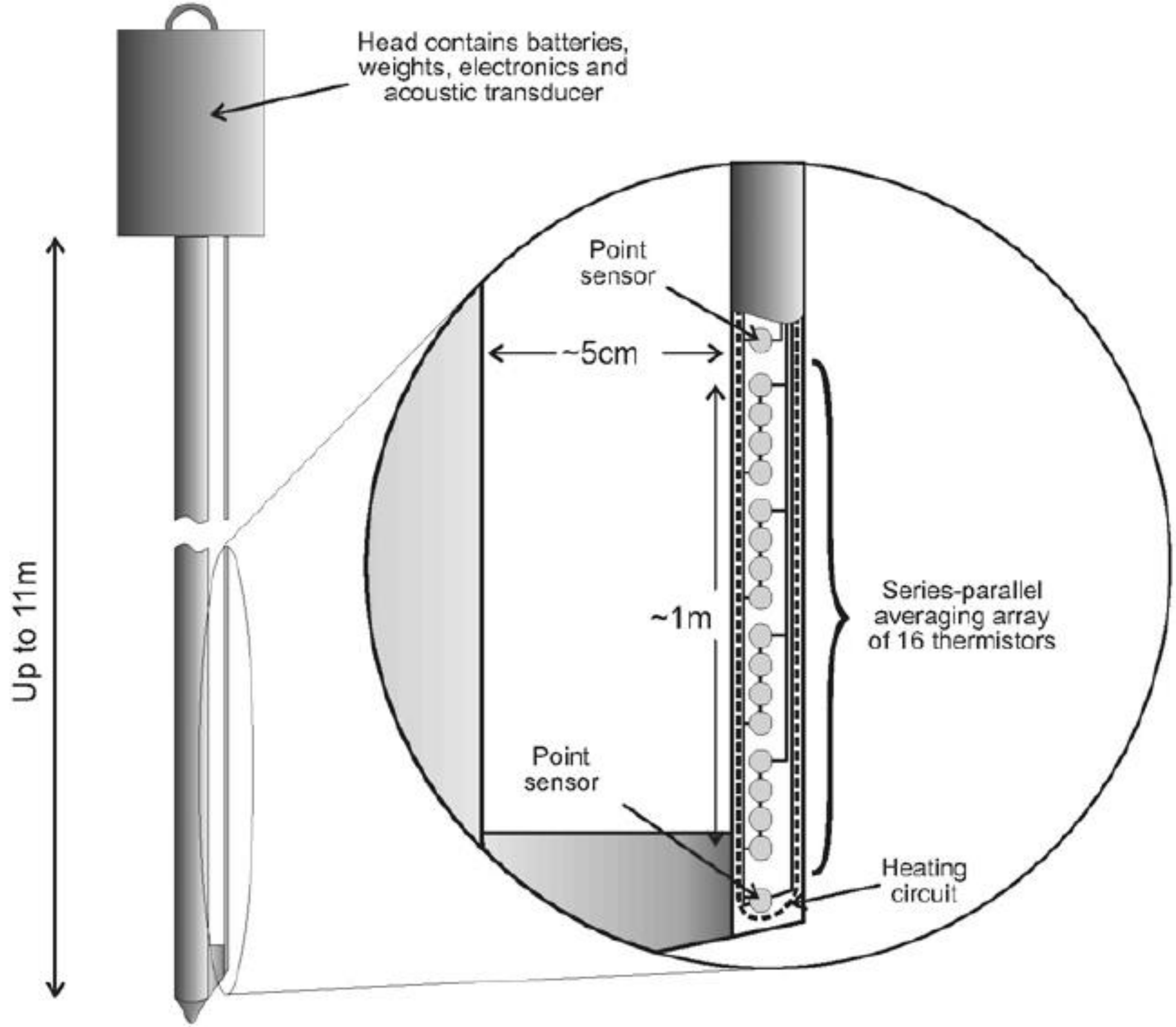


Bullard plot and heat flow
Cruise yyy_1202
Station 02
Pen 3 (HF)
Integrated Thermal Resistance (m2K/W)
Relative Temperature (K)
Mean q = 47.3 mW m-2
Tilt = 1.5°
Error of in situ T
δ T (mK)
Error of in situ k
δ k (W m-1 K-1)
Temperature residuals from Bullard plot
δ T (mK)
Sensor number

The approach that was talking here was to identify areas that show good prospects for high heat flow and then to use the heat flow probe to measure the temperatura and the seabed thermal gradient

La localización de las estaciones se definirá durante el crucero dependiendo de la información batimétrica y de cobertura de sedimentos que se obtenga mediante Multibeam, ecosonda y TOPAS.

Respecto a la localización de las estaciones, se especifica que las estaciones donde se realizarán las mediciones de flujo de calor terrestre con la sonda "FIELAX GmbH Heat Flow Probe" se seleccionarán con base en los datos del perfilador TOPAS, ya que es indispensable contar con una cubierta de sedimentos de espesor mayor a 10 m para que la sonda penetre y realice la medición del gradiente de temperatura.



A Lister “violin bow” heat flow probe