Seismic exploration of planets

Philippe Lognonné

Presentation’s outline

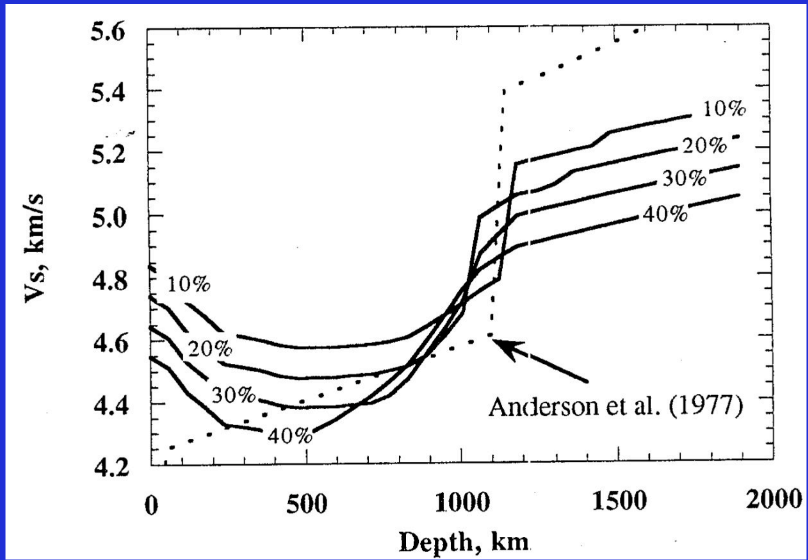
- What is geophysical exploration and sounding?
  - Focuse on seismic sounding
- The example of the Moon
- Mars present and future geophysical exploration
- Other planets and toward remote sensing

In situ geophysical exploration

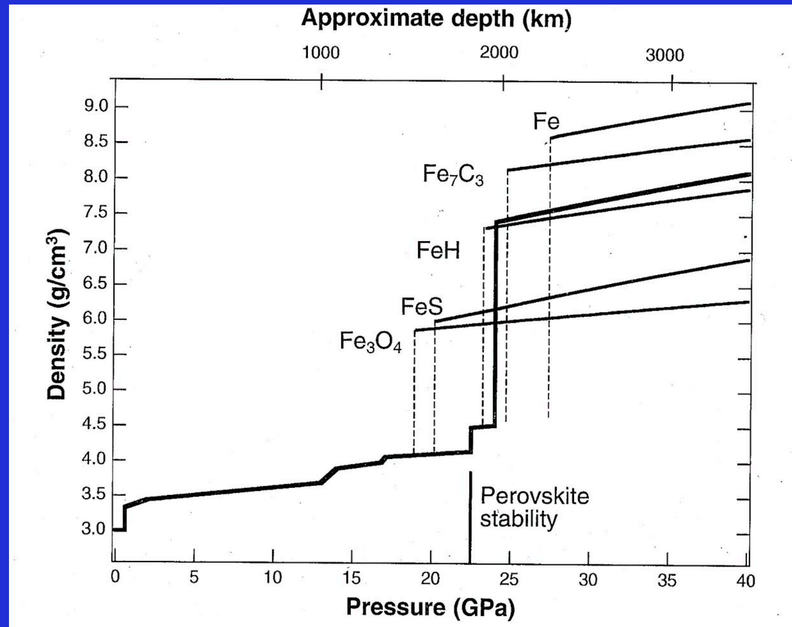
- Goal of in situ geophysical exploration is to determine the Internal structure of a planet
- Internal structure is
  - Therrmodynamical state (pressure and temperature)
  - Mineralogy
- The approach is based
  - On geophysical methods determining the profile with depth (or the 3D models, for the earth case) of geophysical parameters such as
    - Seismic velocities
    - Shear modulus
    - Density
    - Electrical conductivity
    - For subsurface, permittivity
  - On laboratory and theoretical studies determining the dependence of these geophysical parameters with respect to temperature and mineralogy
  - On mineralogical and geochemical analysis constraining directly or indirectly the mineralogy with depth

- A geophysical field must penetrate in the planet, must be reflected/transmitted and then recorded
  - Magnetic sounding (electrical conductivity)
  - Seismic sounding (seismic velocities, seismic attenuation)
  - Electro-magnetic sounding (permittivity, electrical conductivity)
- A geophysical signal must be produced by the planet with amplitude depending on its properties with depth
  - Gravity (density)
  - Heat flux (temperature, radioactivity, thermal conductivity)
- An external force deform the planet with a response depending on its properties with depth and the shape (or deformations) of the planet is recorded
  - Tidal deformation, Precession, nutation, etc (density and elastic modulus)
  - Plate flexure (density, elastic thickness)
- In planetology, data are generally limited…. All sources of information must be used

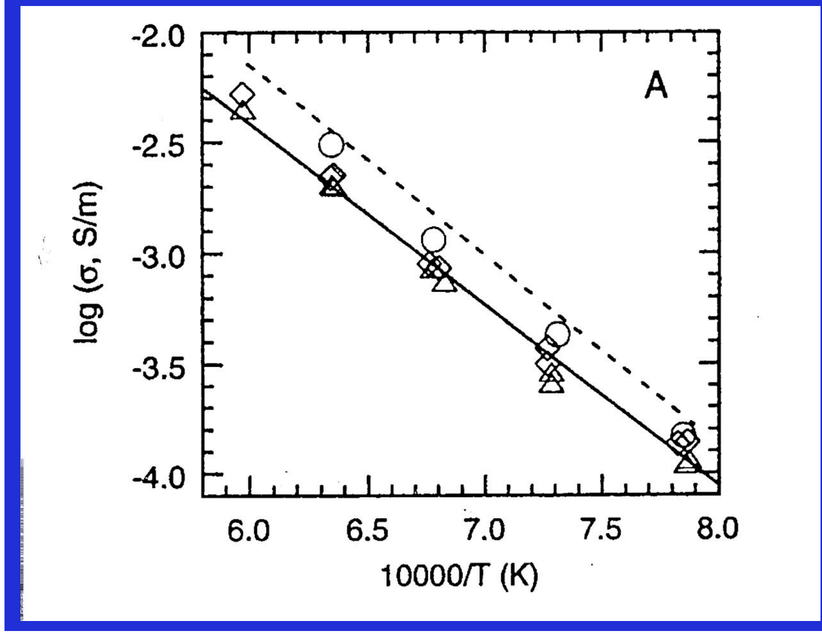
What is the mineralogy, structure and temperature?



Mantle seismic velocities versus Mantle iron content (Mocquet et al., 1996)



Core density versus core composition
(Bertka and Fei, 1998)



Mantle electrical cnductivity versus electrical conductivity
(Xu et al., 1998)

Magnetic sounding

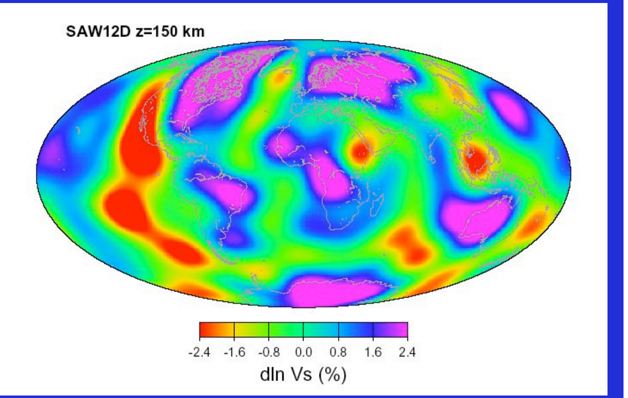
- Principle:
  - An external, time dependant, magnetic field penetrates in an planet
  - Time variation of the magnetic field inside the planet generates currents in the conductive areas canceling partially the field
  - The induced magnetic field is measured by magnetometers
- Limitation
  - Magnetic field is diffusing inside the body

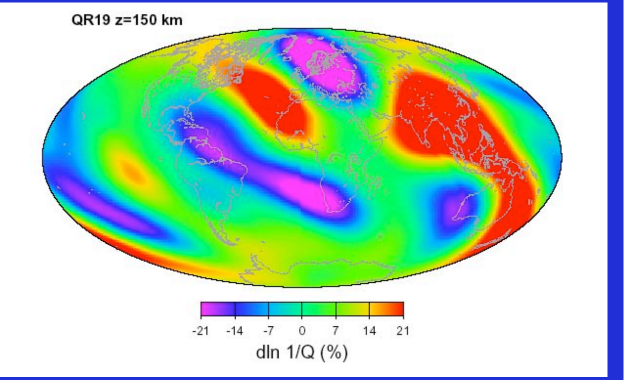
$$\nabla^2 \mathbf{B} = \mu\sigma \frac{\partial \mathbf{B}}{\partial t},$$

  - Only long periods magnetic field (hours) are sounding deep
  - Diffusion makes the reconstruction of discontinuities difficult
- Success
  - Detection of highly conducting part of a planet (iron metallic core, low velocity zone in a mantle associated to partial melting, liquid in the crust or below a crust)
- Sources
  - Moon: Magnetic field variations associated to the displacement of the Moon in the Earth magnetosphere
  - Moon: Magnetic field variations associated to the solar wind
  - Jovian satellites: Magnetic field variations associated to the tilted rotating Jovian magnetosphere

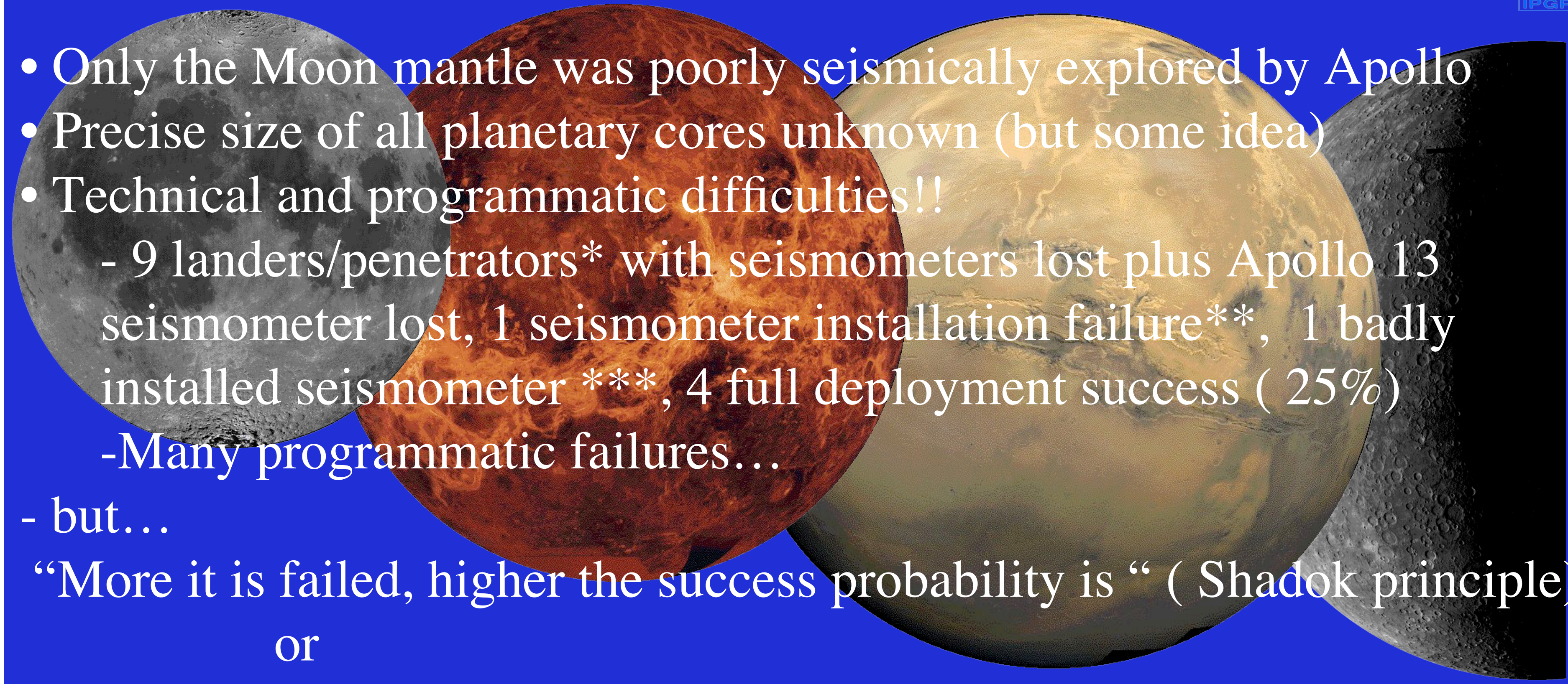
Seismic sounding

- Principle
  - Use active (impactors, explosive) or passive (quakes, meteorite impacts, crack) seismic sources
  - Record and analyse the seismic signals
- Key dates
  - Earth:
    - Von Reben Paschwitz, 1889, first signal
    - Oldham, 1906, discovery of the core
    - 1960, discovery of the normal modes
    - 1980+ Tomographic models (i.e. details of a few %)
  - Sun:
    - Leighton et al., 1962, discovery of the normal modes
  - Moon:
    - Latham et al., 1969, Apollo 11, first records and deep moonquakes
  - Jupiter:
    - Hammel et al., 1995, observation of Atmospheric Tsunamis
  - Mars:
    - Anderson et al., 1976, First installation of seismometer, single observation?





Deep interior of telluric planets: a real challenge



- Only the Moon mantle was poorly seismically explored by Apollo
- Precise size of all planetary cores unknown (but some idea)
- Technical and programmatic difficulties!!
    - 9 landers/penetrators* with seismometers lost plus Apollo 13 seismometer lost, 1 seismometer installation failure**, 1 badly installed seismometer ***, 4 full deployment success (25%)
    - Many programmatic failures…

- but…

"More it is failed, higher the success probability is " (Shadok principle)

or

" If the success probability is 10%, we have to hurry up in failing the first 90% of the trials " (Another Shadok principle)

* Rangers 3-4-5, Phobos whopper 1-2, Mars96 SSS1-2, Penetrator 1-2

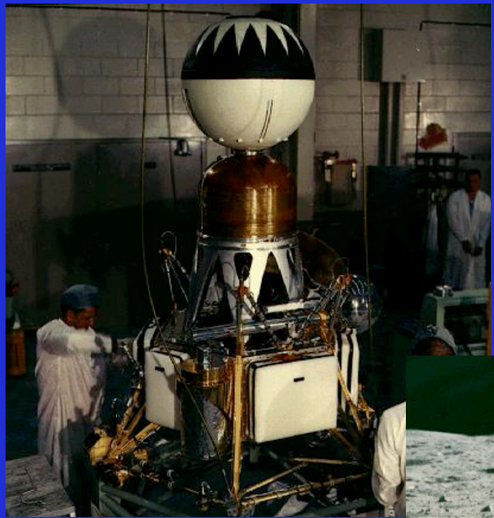
** Viking Lander 1

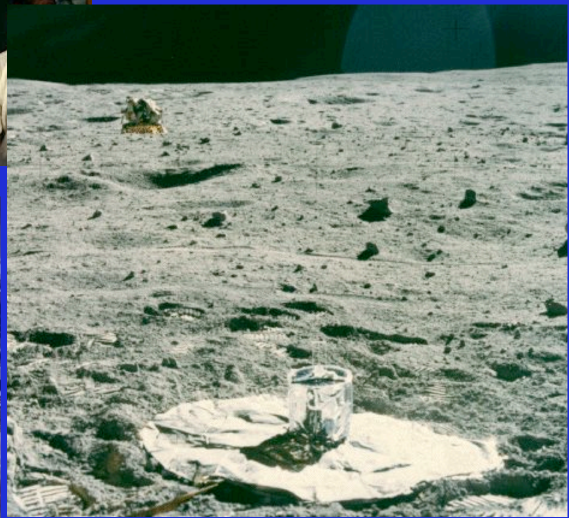
*** Viking Lander 2

Seismology outlines

- How to do Planetary seismology: the exemple of the Lunar case
- How to prepare future Mars seismic network mission
- Future seismology perspectives

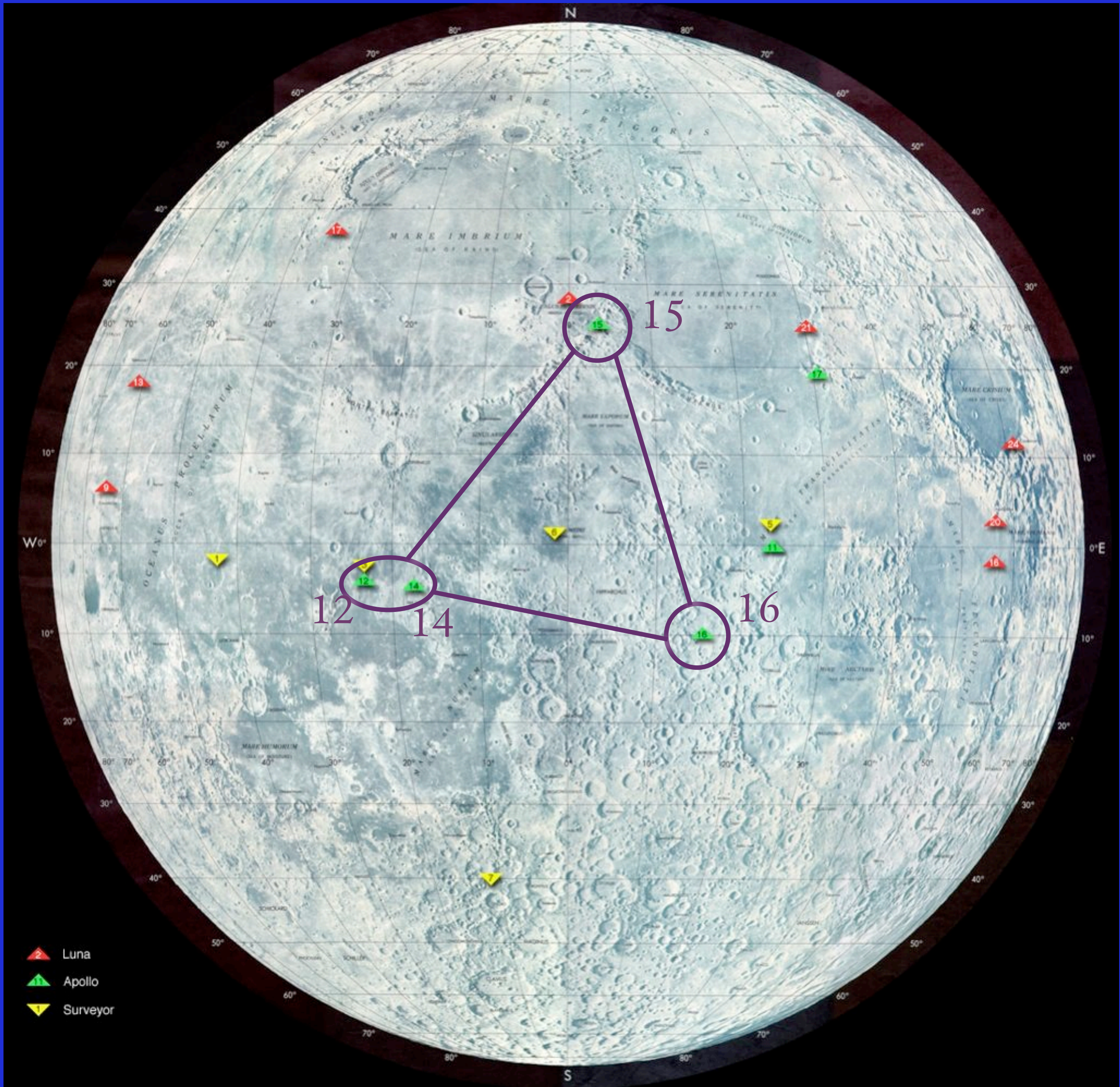
Moon Seismology: history





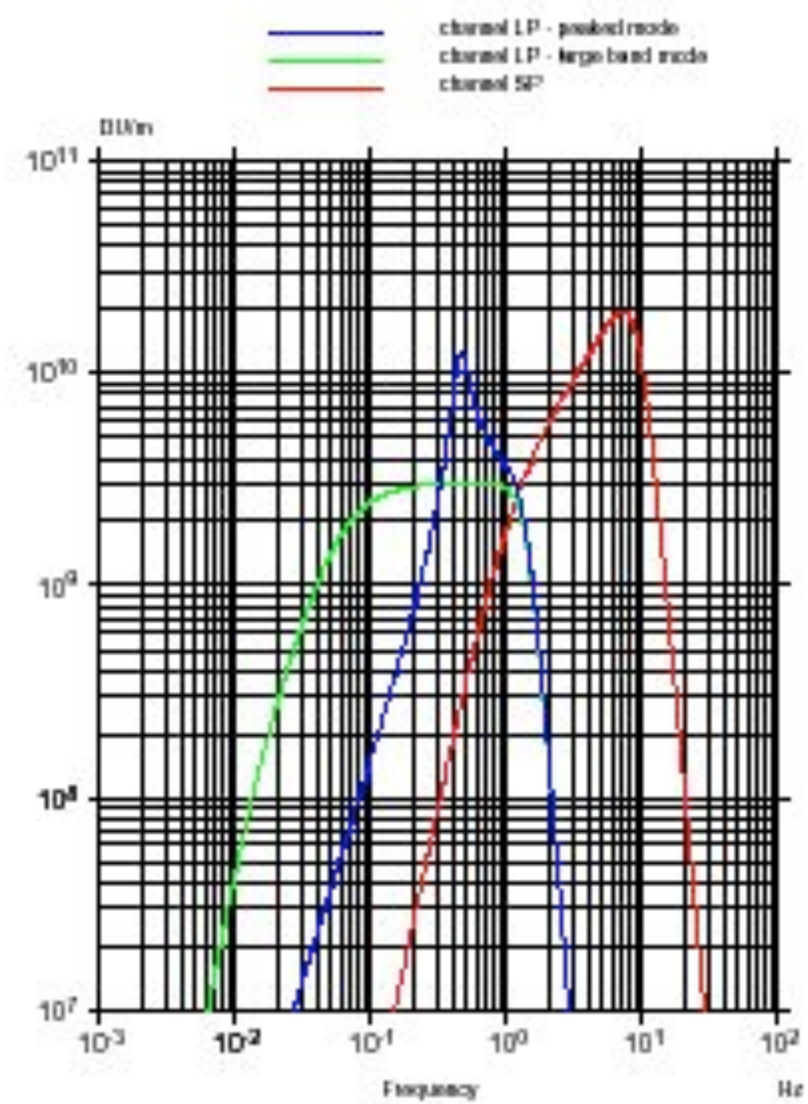
- 1961-1962: Failure of Rangers 3-5 all with seismometer launched to the Moon
- 1969-1973: Success of Apollo Seismic network with operation up to 1977
- 2006+: 2 antipodal seismic penetrators to be launched by Japan-ISAS (Lunar A)
- Other data for deep interior:
  - Density
  - Inertia factor
  - Love number (real and imaginary part)
  - Heat flux (2 measurements) + surface content in Th

Apollo Seismic Network



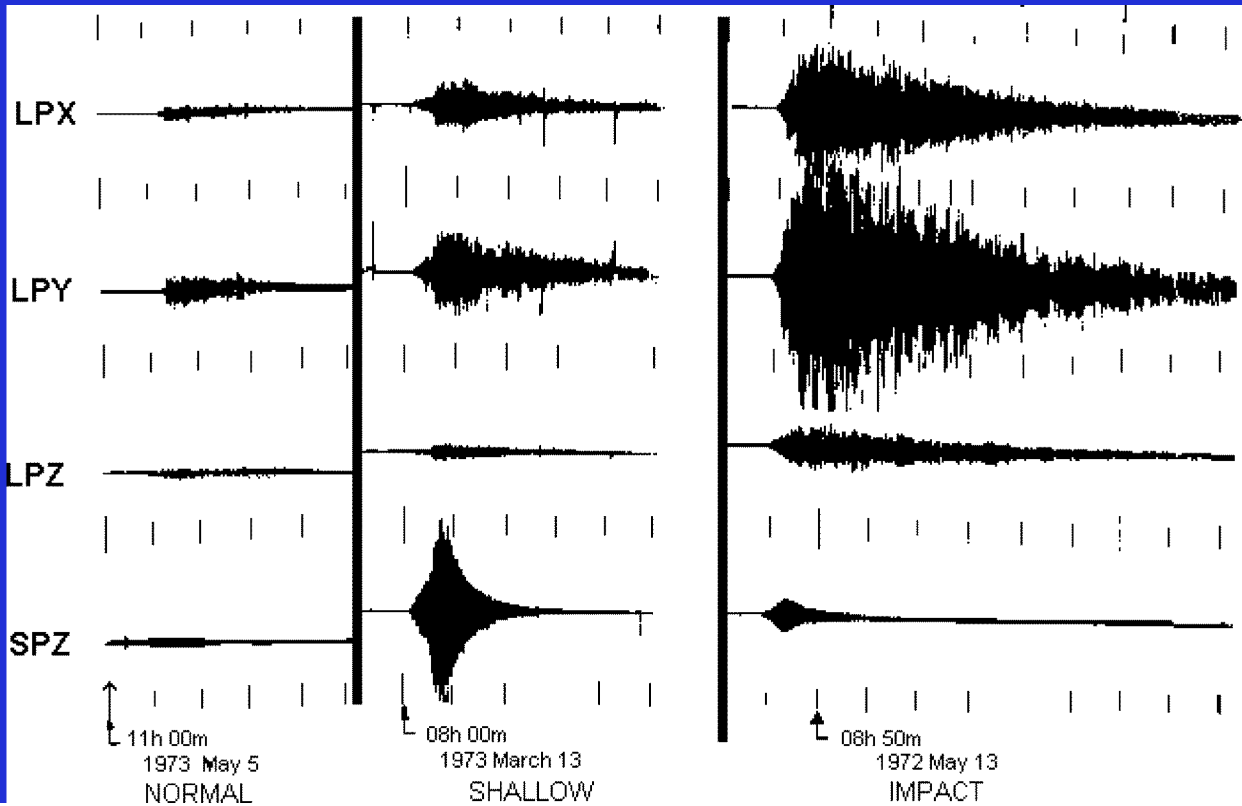
- 4 stations: Apollo sites 12, 14, 15 and 16
- installed between 1969 and 1972
- turned off in 1977

sensitivity

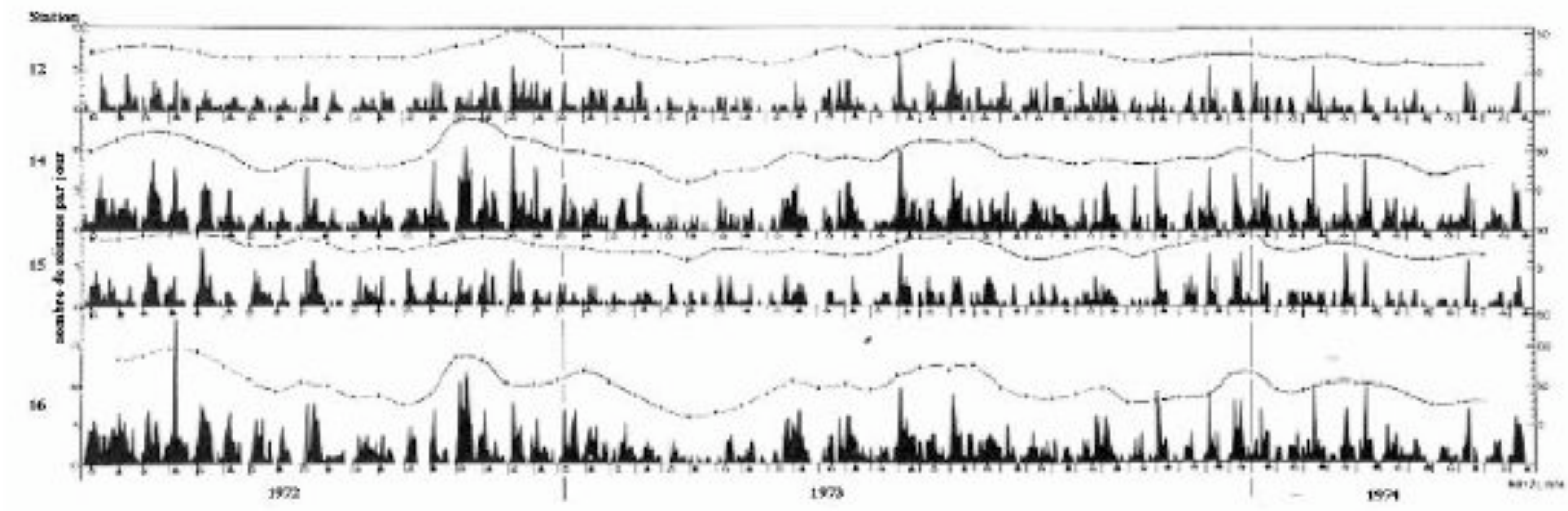


Moonquakes

- deep moonquakes
  - 700-1000 km depth, near just at the bottom of the elastic lithosphere of the Moon
  - Very small magnitudes quakes
  - Origin: accumulation of stress related to the thermoelastic cooling of the planet triggered by the Earth tide
  - Several faults identified where quakes occur repeatedly
- superficial moonquakes
- meteoroid & artificial impacts

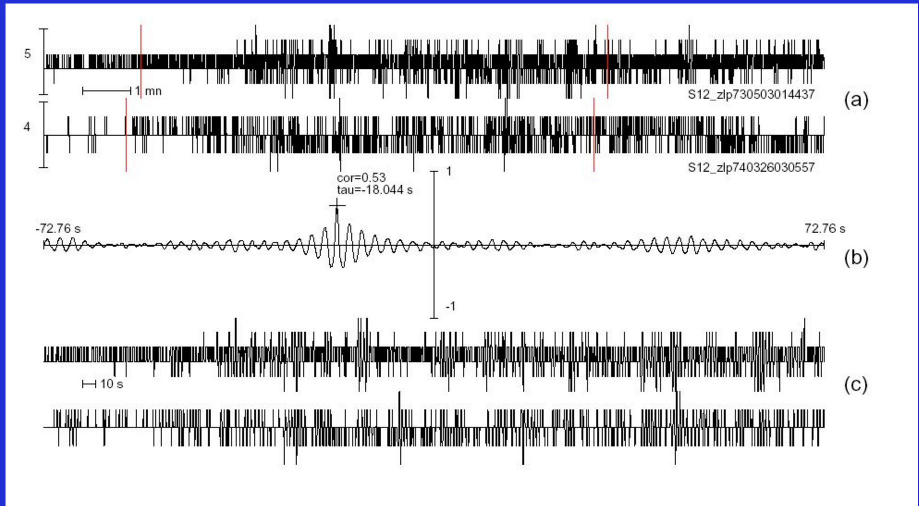


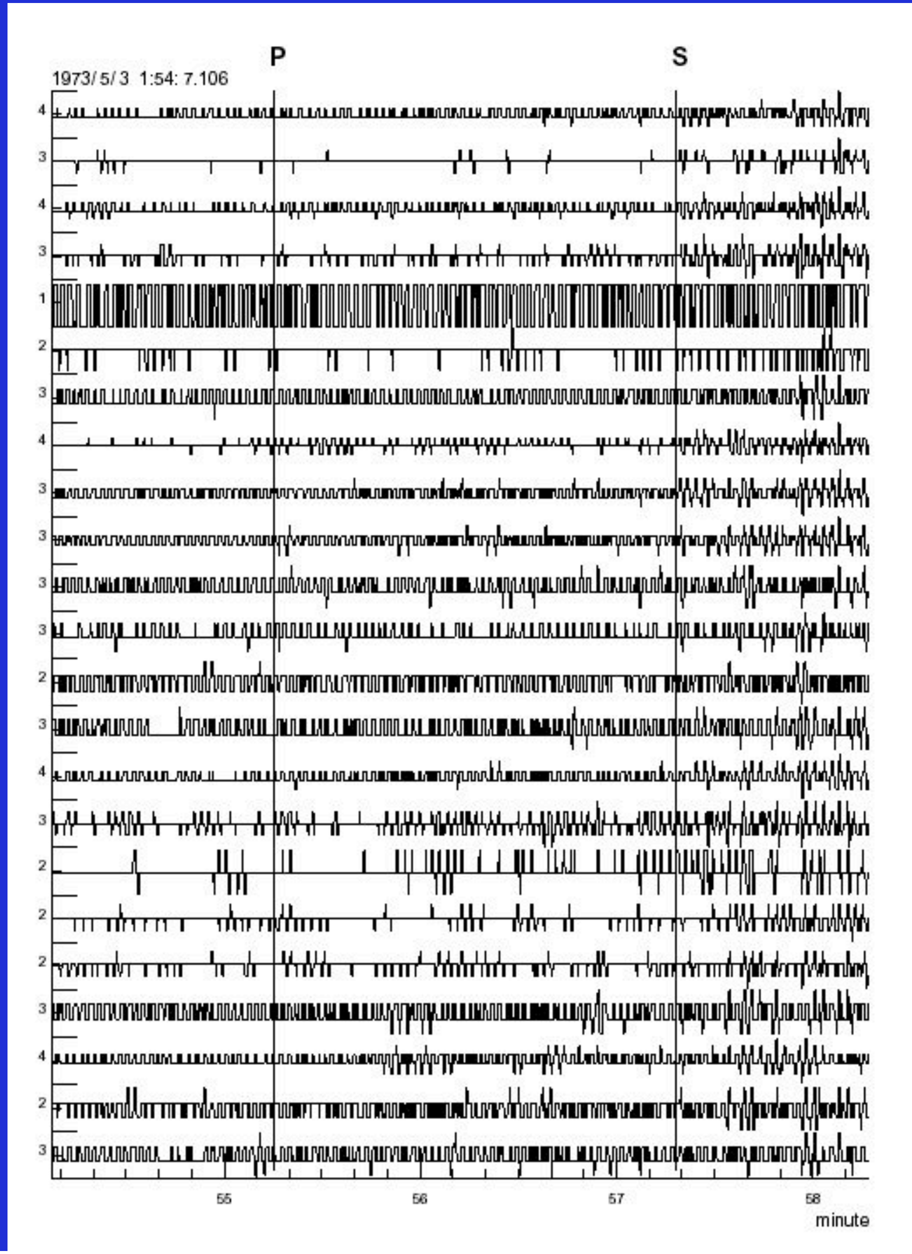
Deep moonquakes



- Number and amplitude of quakes is related to the amplitude of tide
- About 50 active faults detected
- Quakes occur at the same fault regularly but with very low amplitudes, with ground displacement of a few Angströms at 2 sec (0.5 10^{-9} ms^{-2} of ground acceleration)

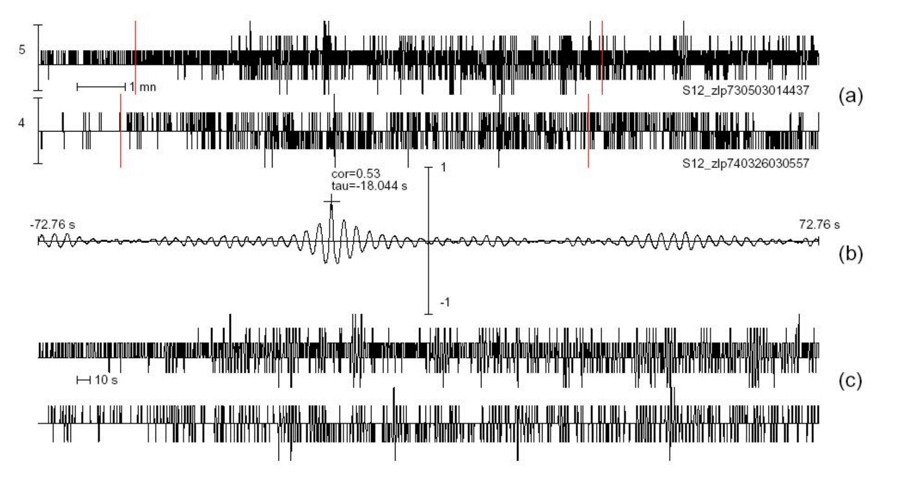
Deep Moonquake





- example of two quakes from the same deep focus and their cross-correlation
- cross-correlation provides the time shift necessary to align the arrival times
- stacking can then be done

Deep Moonquake



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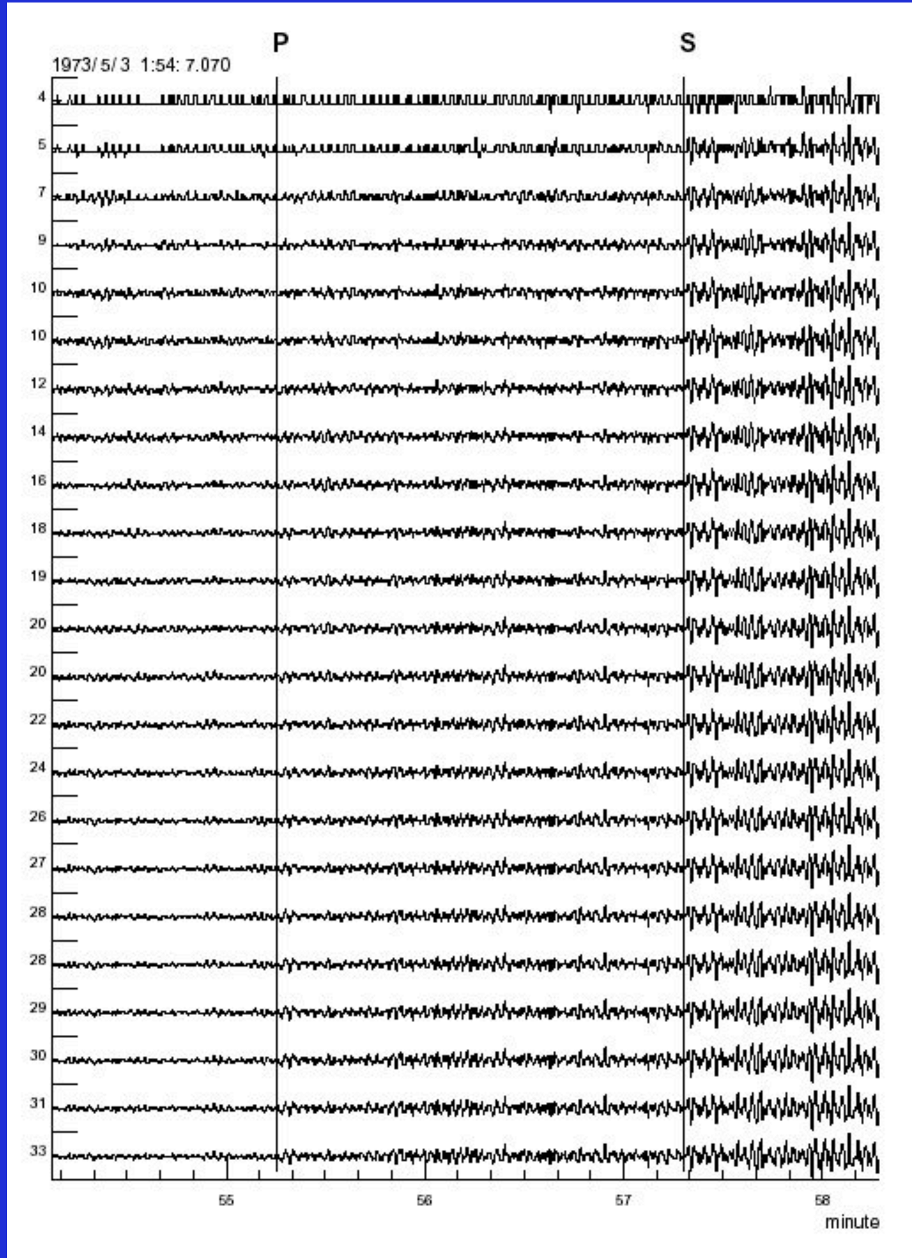
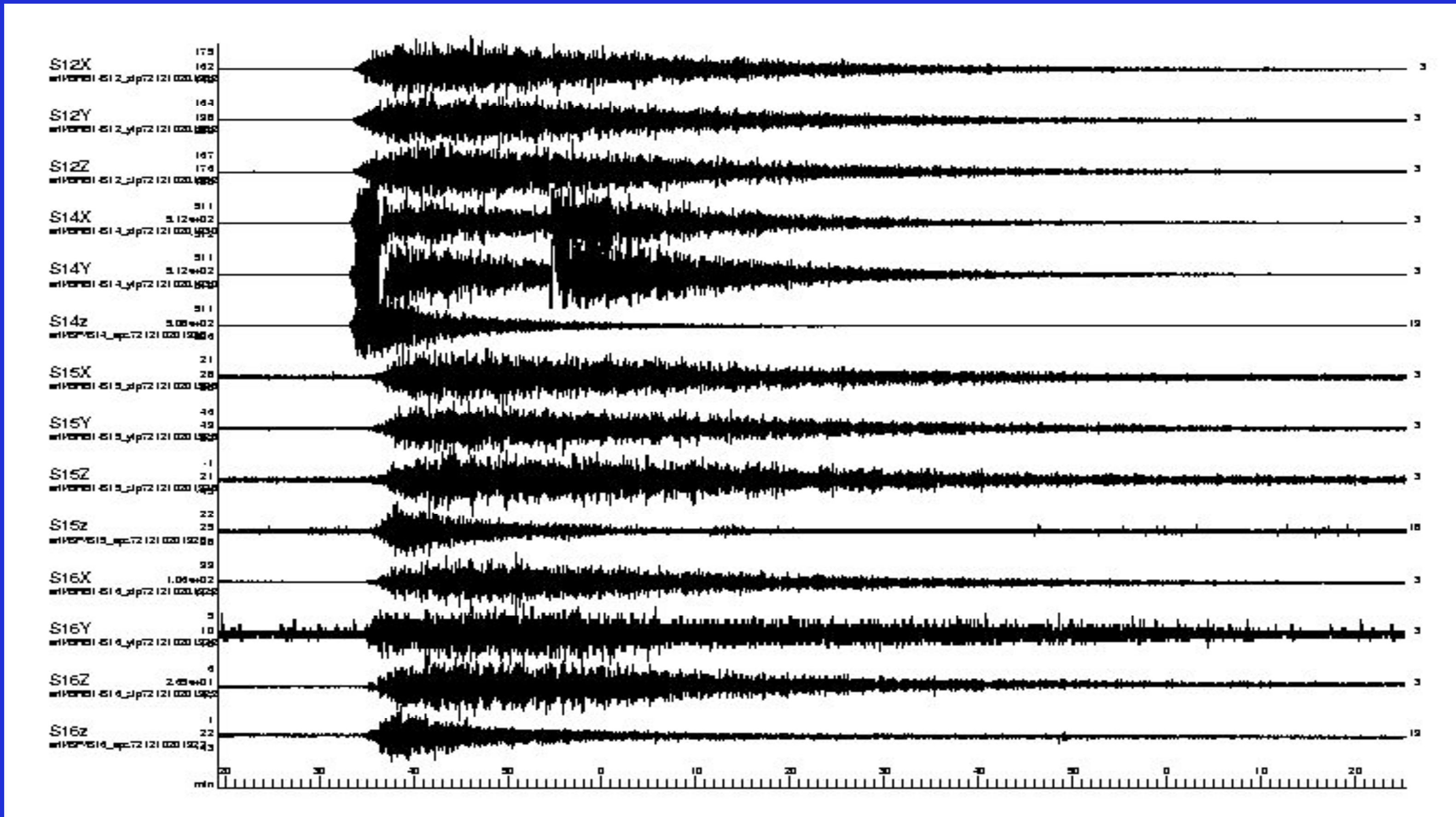


Active source: impacts

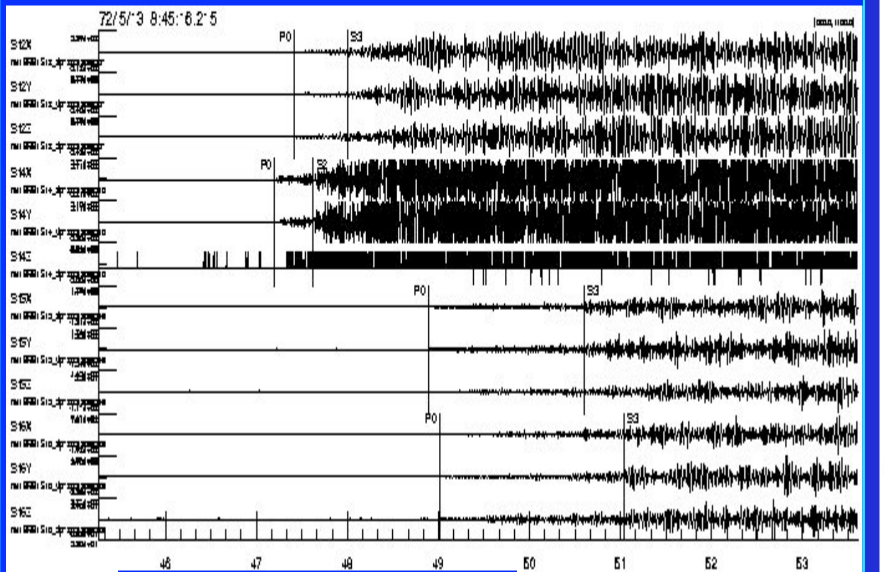


- Impact of the Apollo 17 Saturn V upper stage (Saturn IVB) on the Moon on 10 December 1972 at distances of 338, 157, 1032 and 850 km from the Apollo 12, 14, 15 and 16 stations, respectively. Amplitudes at Apollo 14 station, 157 km from impact, reach about 10^{-5} m s^{-2}
- Known time and location: all arrival times give information on the structure

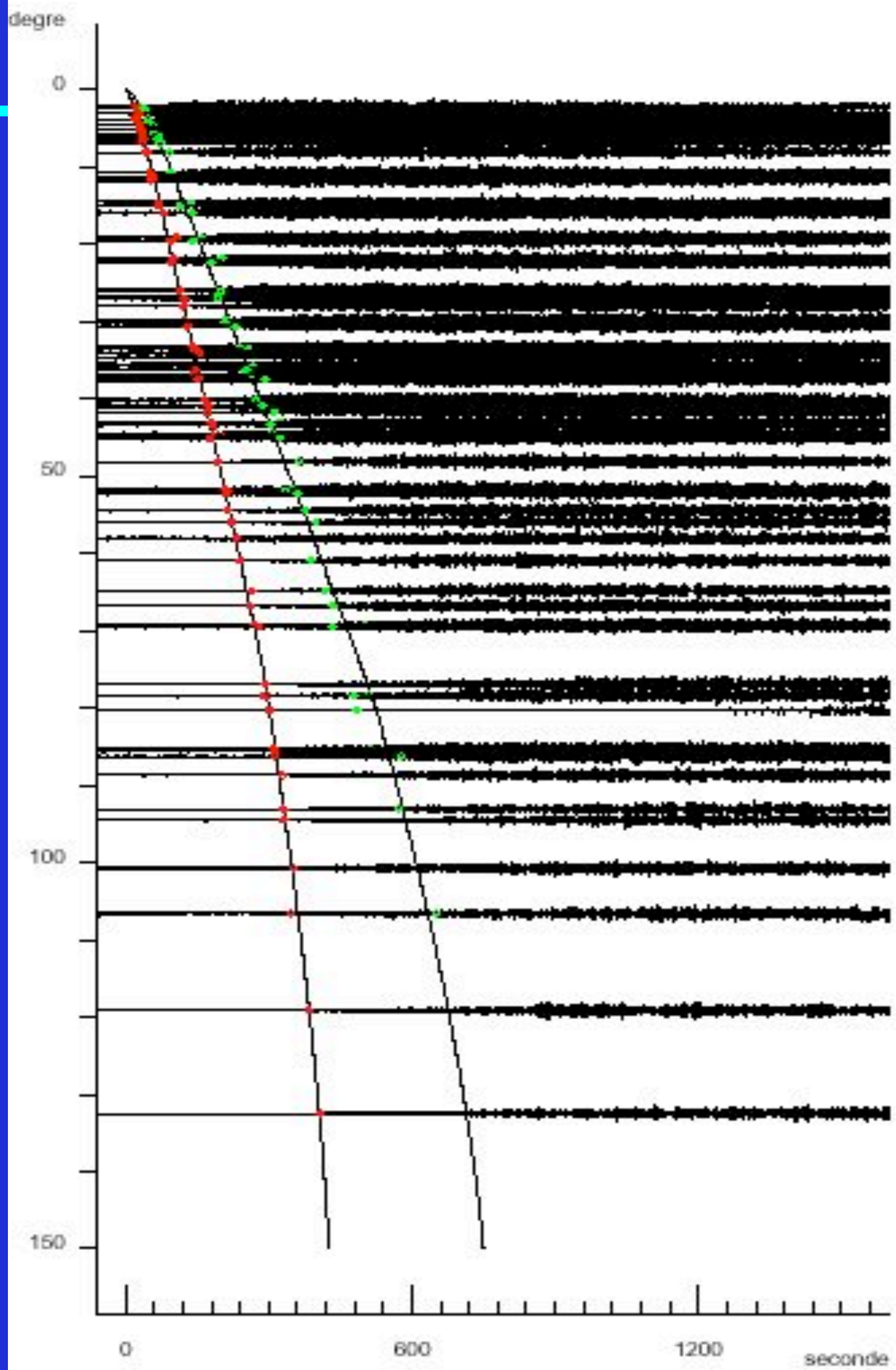
Making seismic models

- Seismic data can be used in different way
  - Easy and robust inversions are based on secondary data, i.e. data obtained by the processing of seismograms
    - Arrival times of the body waves (P and S)
    - Arrival times of secondary waves (Conversion of P and S…)
    - Arrival times (with frequency) of the surface waves
    - Azimuth of the waves and polarisations
    - For large quakes, free oscillations periods
  - Complex and less robust inversions are based on waveform inversion
- Quality of the inversion is related to
  - Error in secondary data determinations
  - Effect of non-inverted parameters, such as
    - Mantle 3D lateral variations
    - Crustal stucture below the stations and near the seismic source, for crustal sources

Example of arrival time determination



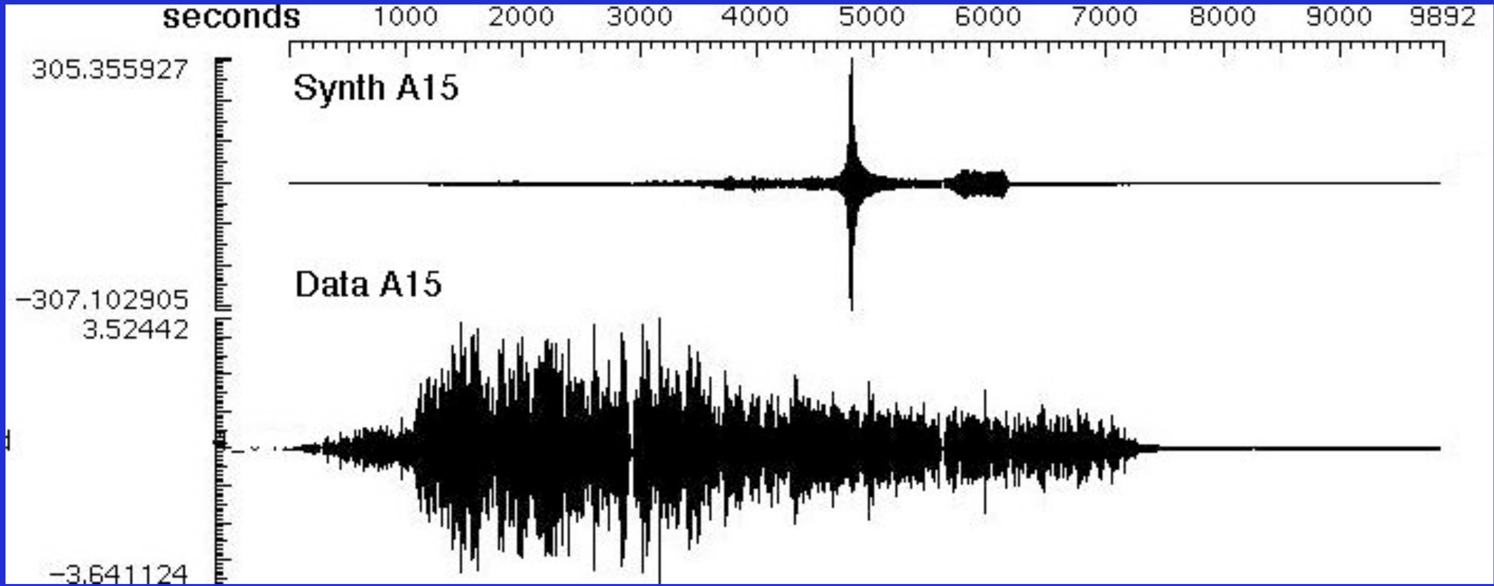
Example of seismic traces : a near impact recorded at all stations on 3 components



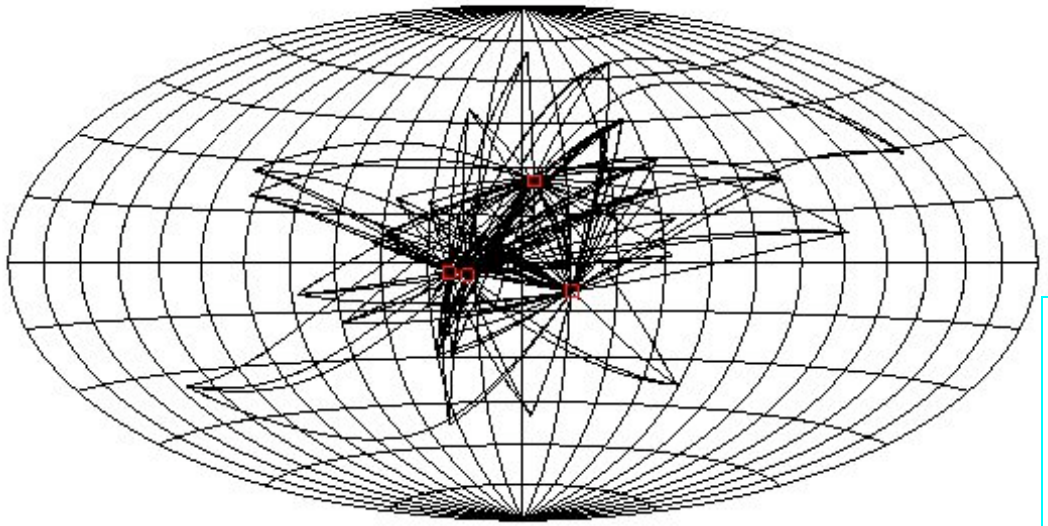
• In many case, diffraction is making the determination of arrival times difficult, with error up to 10sec (mean error is about 2s)

Diffraction

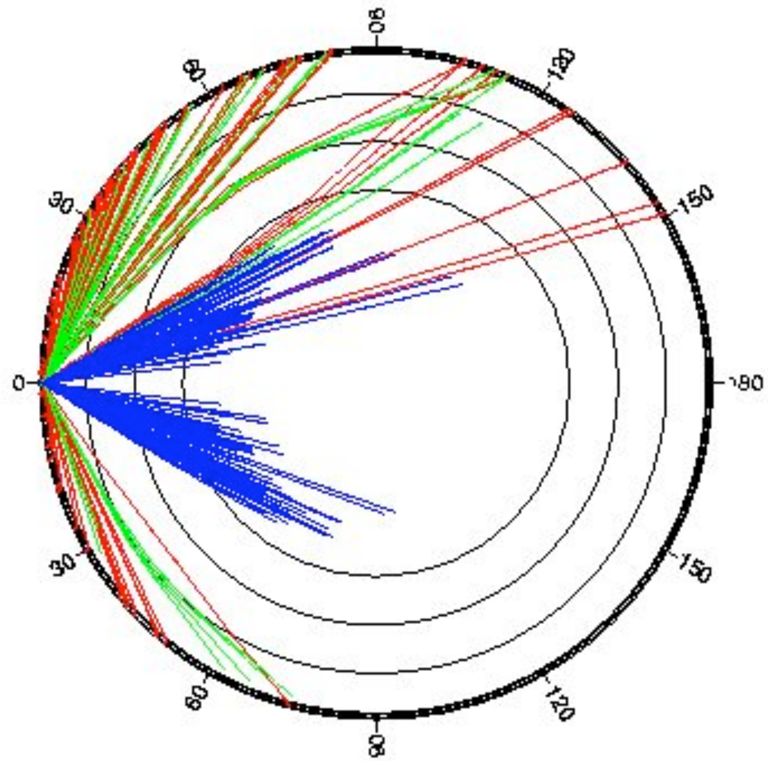
- The Moon subsurface is highly fracturated, as a results of non-resurfacing and of a long impacting history
- Propagation equation is not valid anymore for waves propagating in the crust and diffusion equation must be used
- Scattering destroy short period surface waves and is able to transfer energy from P to S waves up to an equipartition given by $E_p = v_s^3 / v_p^3 \, E_s/2$, where E_p, E_s are the energy of P and S waves, v_p , v_s are the velocities of P and S waves



Rays sampling



Ray path of 59 events used for global inversion.



All ray paths available in the Moon. Blue is for deep events, red for impacts, green for superficial moonquakes.

• recording events with different epicentral distances give access to the structure with depth

• the inverted model is however not a mean model of the planet, but a mean model of the area where the network is deployed

Principle of the inversion

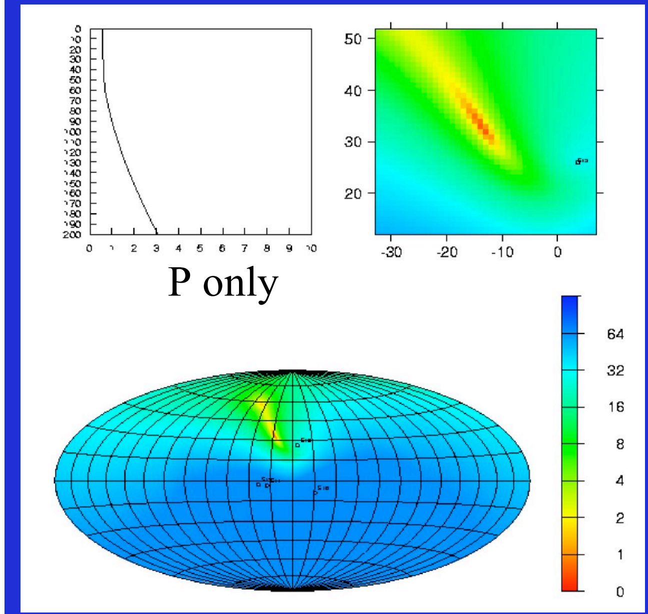
Seismic data, i.e. arrival times at the stations

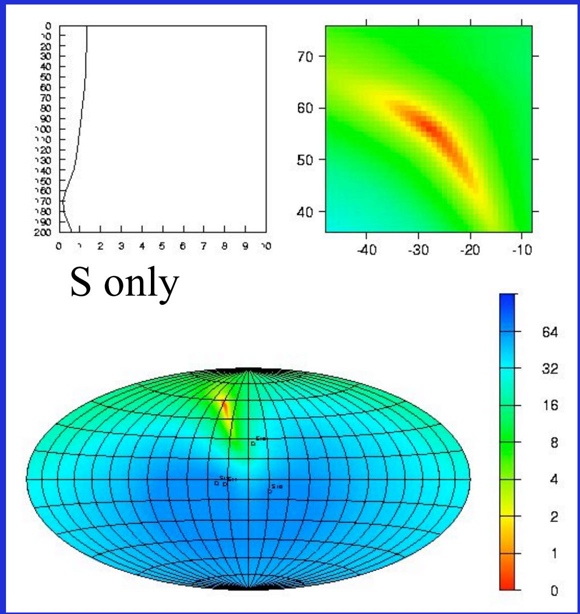
Source parameters, i.e. position and times of the quakes

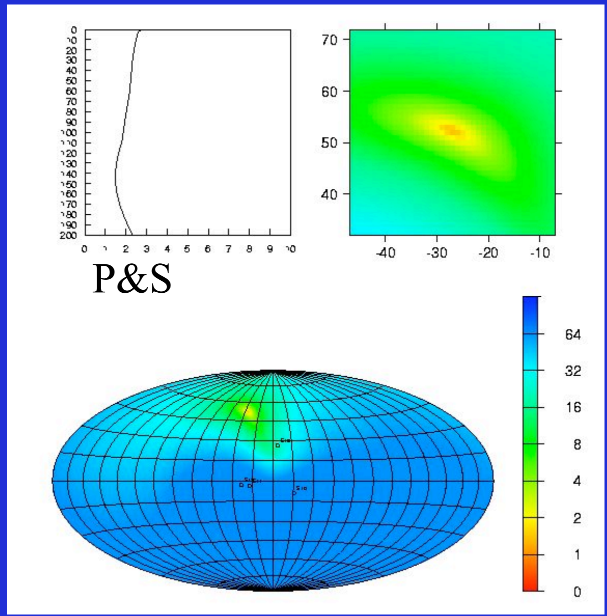
Model parameters, i.e. P and S seismic velocities with depth

Sources relocalisation

- Source localisation is done iteratively in the inversion: for each new structure new model, a new localisation of the sources is done and then used for next inversion of structure







Principle of the inversion

Seismic data, i.e. arrival times at the stations
Nx6

Source parameters, i.e. position and times of the quakes
Nx4

Model parameters, i.e. P and S seismic velocities with depth
< 2N

- N is the number of quakes
-The seismic model must
    -be limited to depth seen by the seismic rays
    -if errors are high, an oversampling is mandatory to reduce the impact of errors on the data
    -Number of layers with V_p and V_s inverted is therefore $N_l << N$

Inversion in the Appolo case

Seismic data, i.e. arrival times at the stations
Nx6

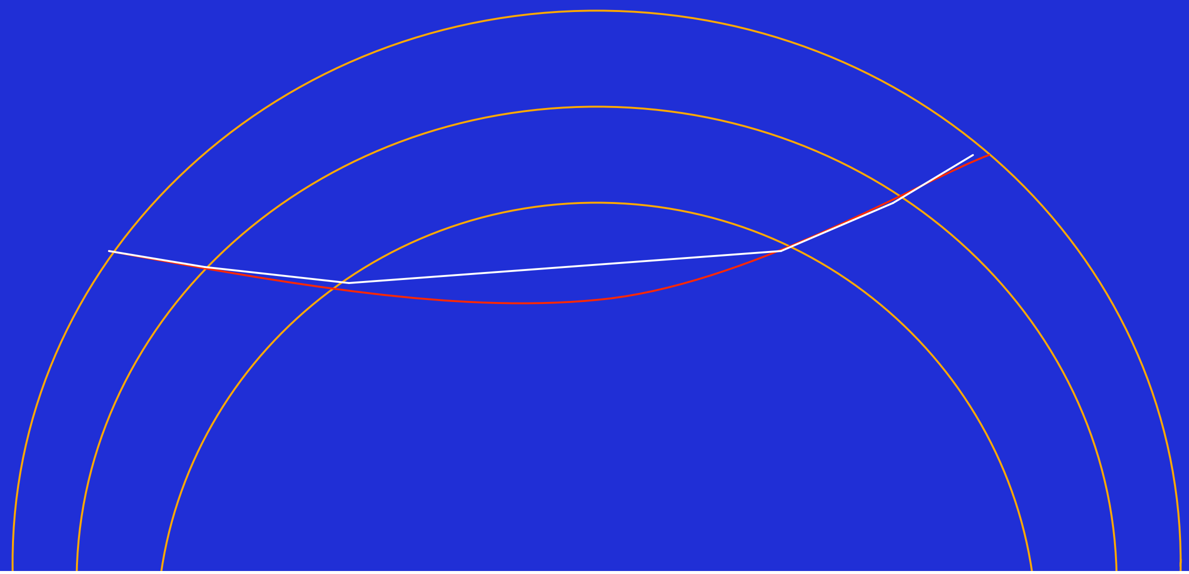
Source parameters, i.e. position and times of the quakes
Nx4

Model parameters, i.e. P and S seismic velocities with depth
< 2N

-Practically, in total 319 P & S arrival time data where used to constrains 59 seismic sources, including 185 source parameters and 134 degree of freedom available for internal structure

-Mean error is 2 sec for arrival times

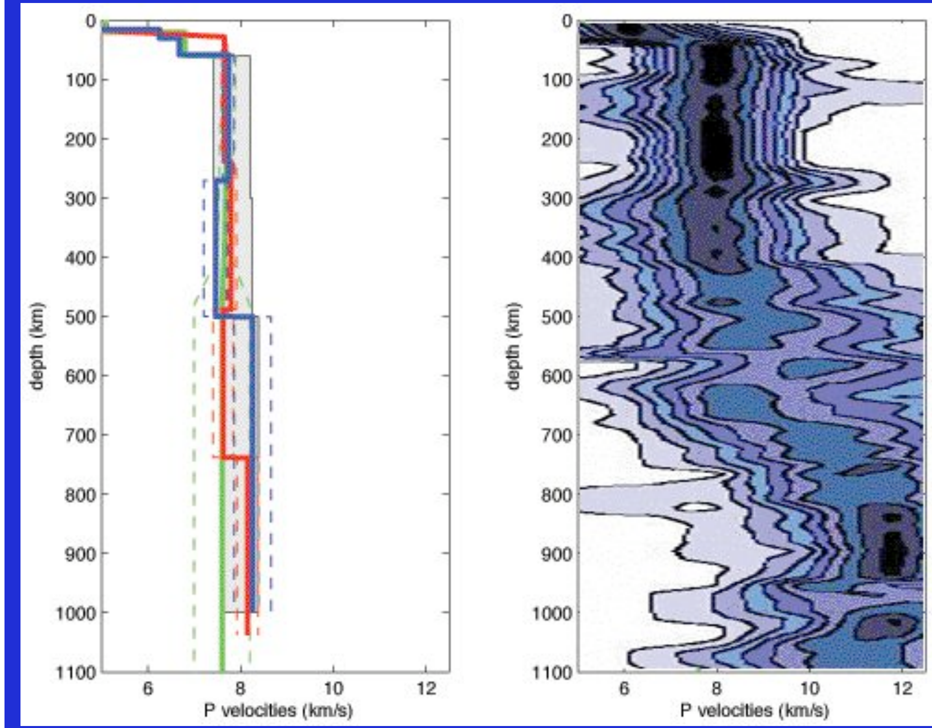
- 2 possible inversion strategy
- Inversion with a limited number of layers (typically about 5-10)
  - Inverted parameters are not the true velocities but the mean velocities in a layer
  - Some error is done in the theory
  - When sdata > stheory, the error on the inverted models is improved
- Inversion with a large number of layers (typically 50)
  - Inverted velocities have error directly related to the mean quality of data

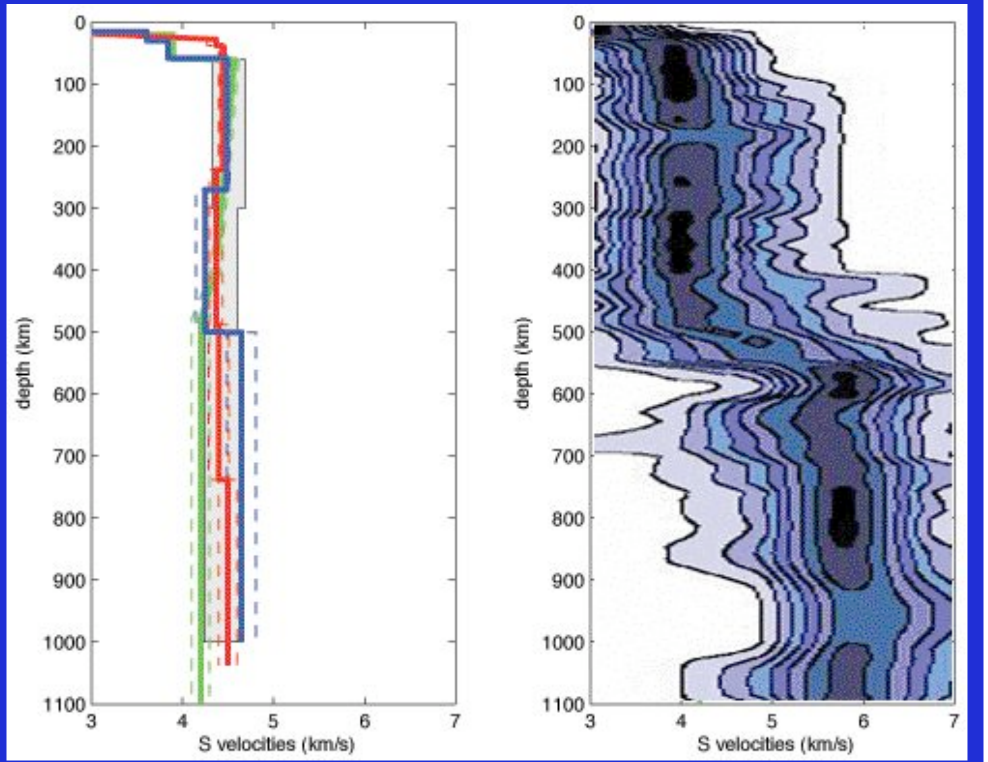


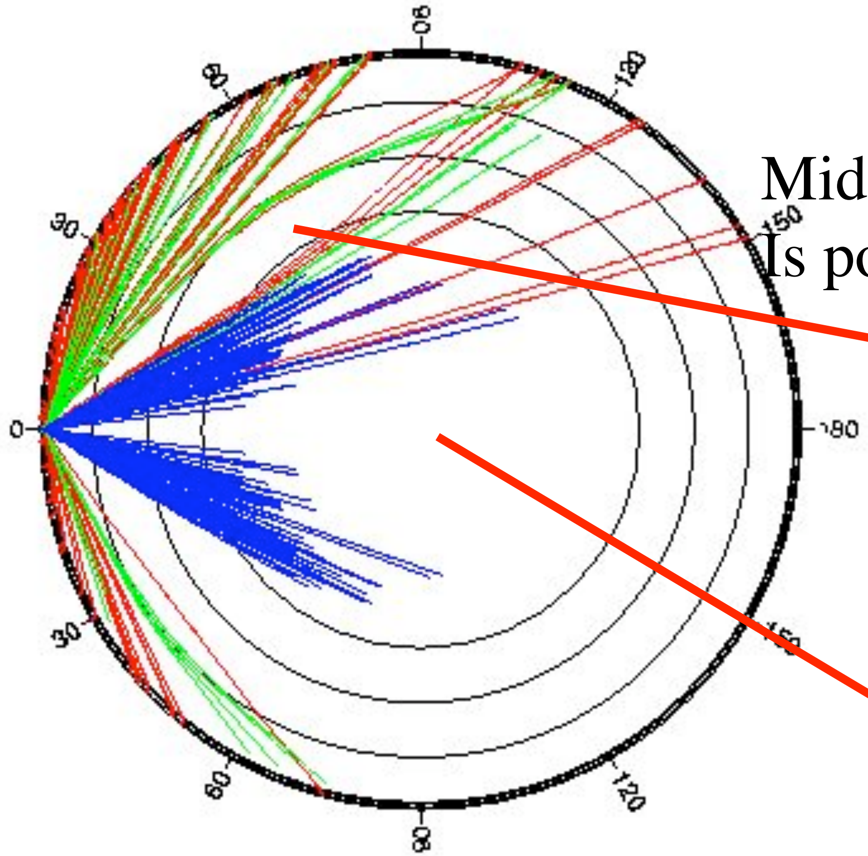
Inversion results

• right: highly layered model (Khan et al., 2000, 2002) with unselected data

• left: weakly layered model (Lognonné et al., 2003) with selected data

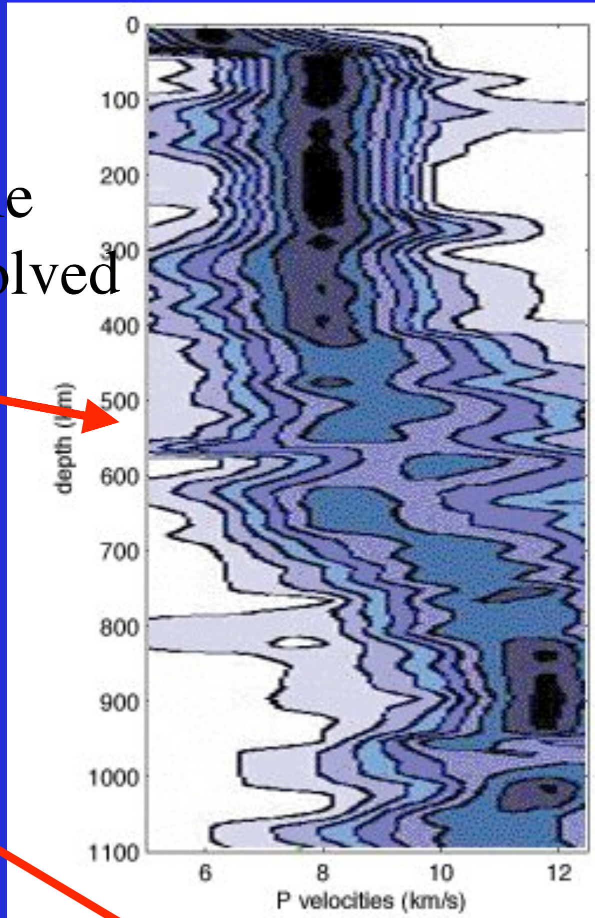






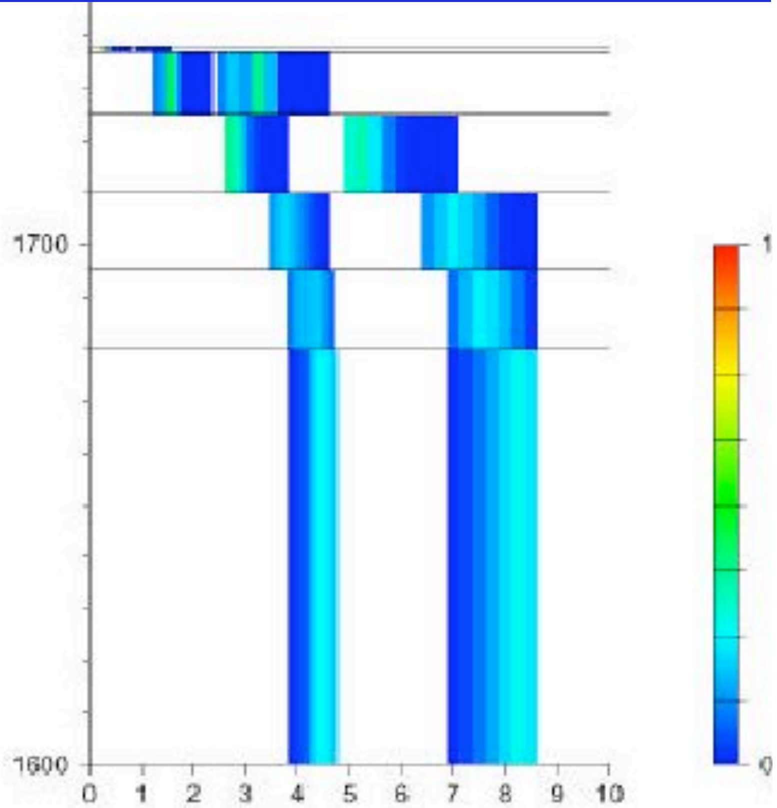
All ray paths available in the Moon. Blue is for deep events, red for impacts, green for superficial moonquakes.

Middle mantle
Is poorly resolved

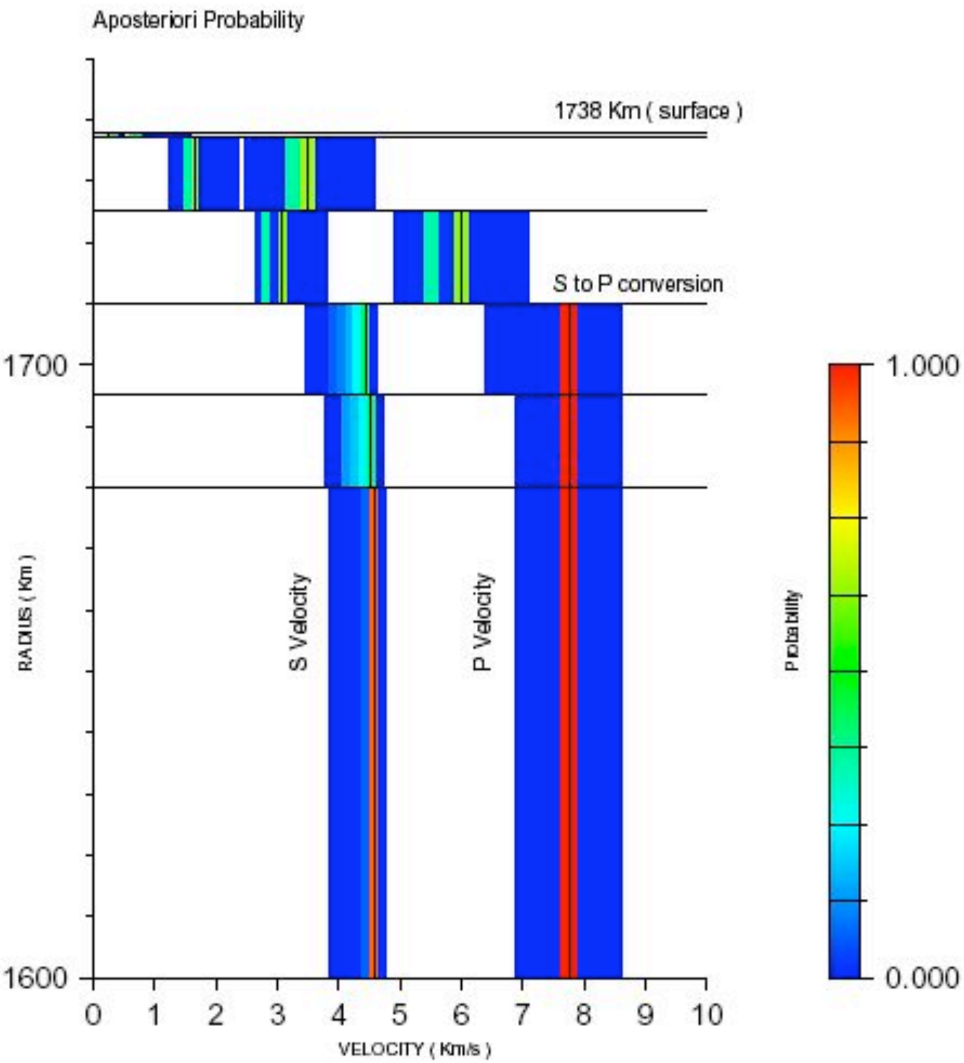


The lunar core was not seen by the Apollo Network

A priori and a posteriori models

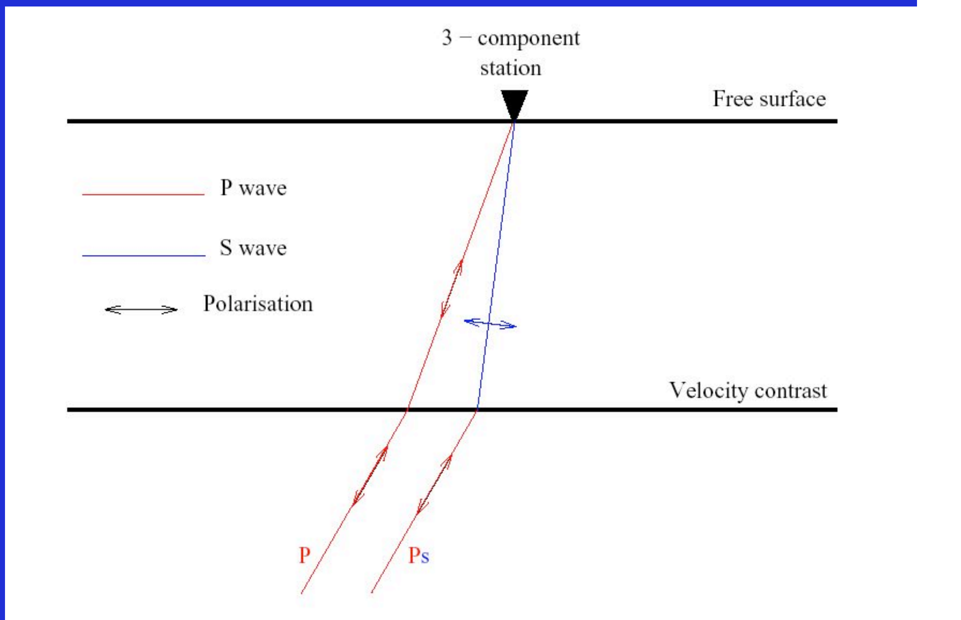


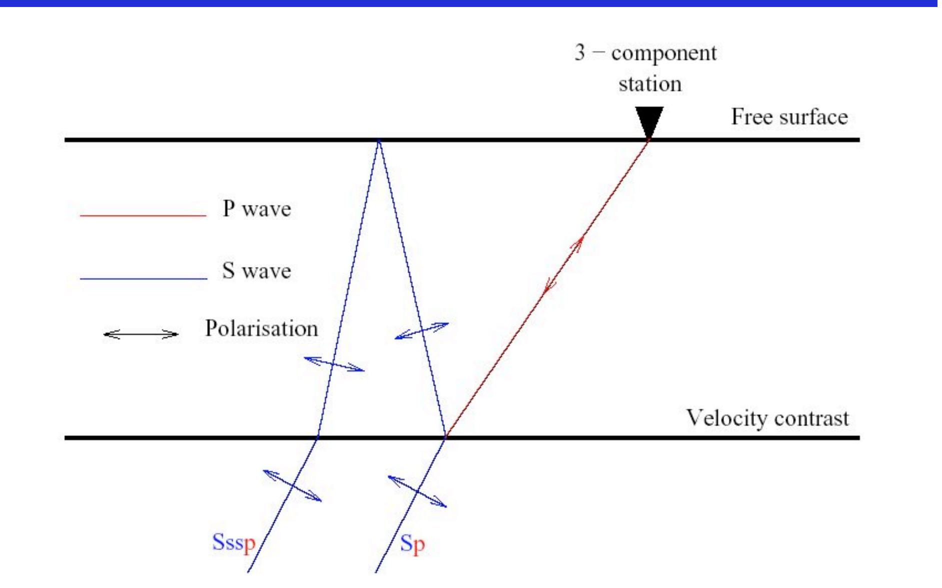
(a) a priori model



Inversion with some 3D effects: crustal structure

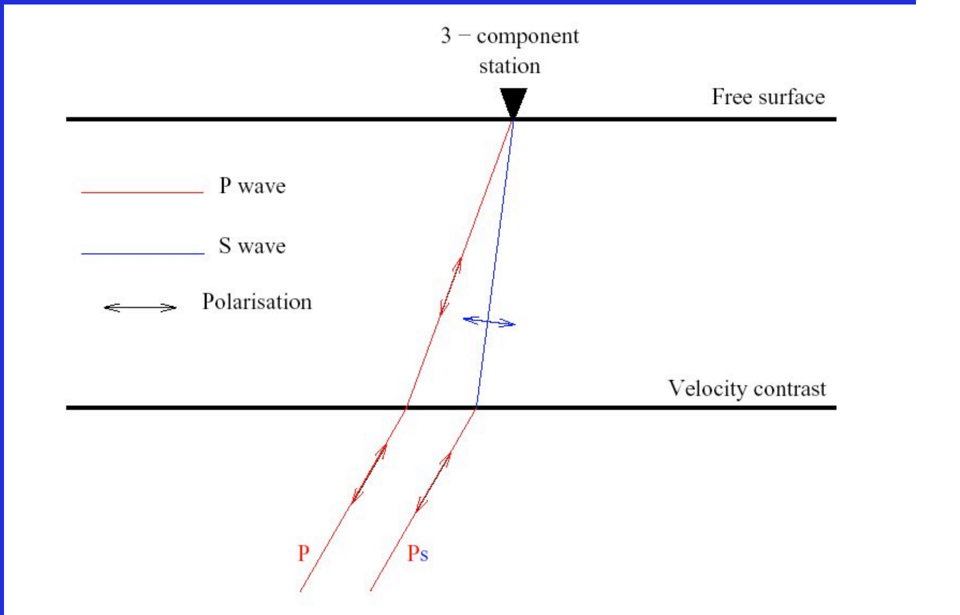
- The crustal structure leads to conversion and reverberations
  - Primary wave arrival ~$P(t-t_p)$ x T
    - P(t) is the amplitude in of the P wave below the crust, depending on the mantle propagation and of the seismic source, T the transmission coefficient to the crust and tp the transmission time through the crust
  - Converted wave ~$P(t-t_c)$ c C
    - C is the transmission coefficient of the crust from Primary wave to converted wave and tc the transmission time through the crust

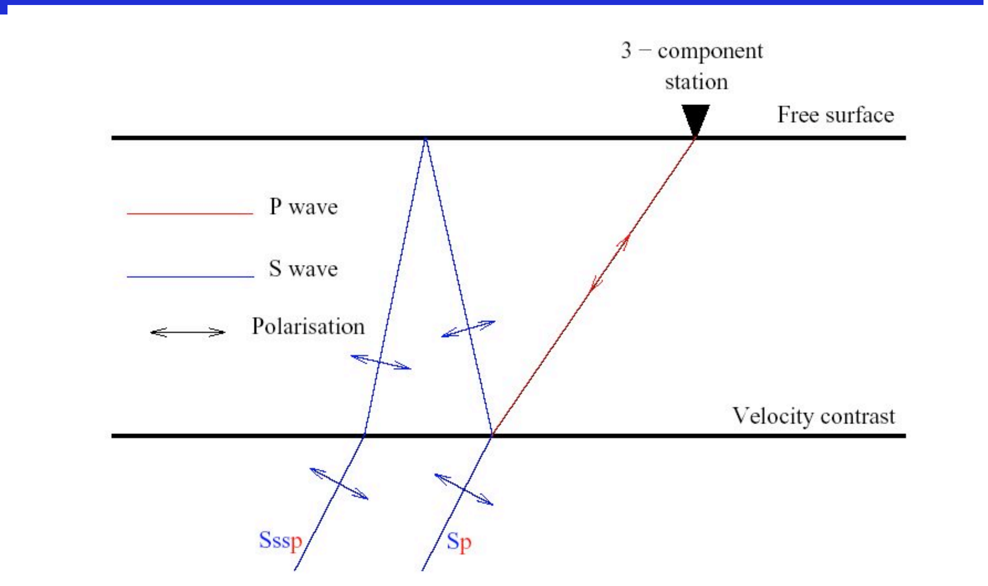




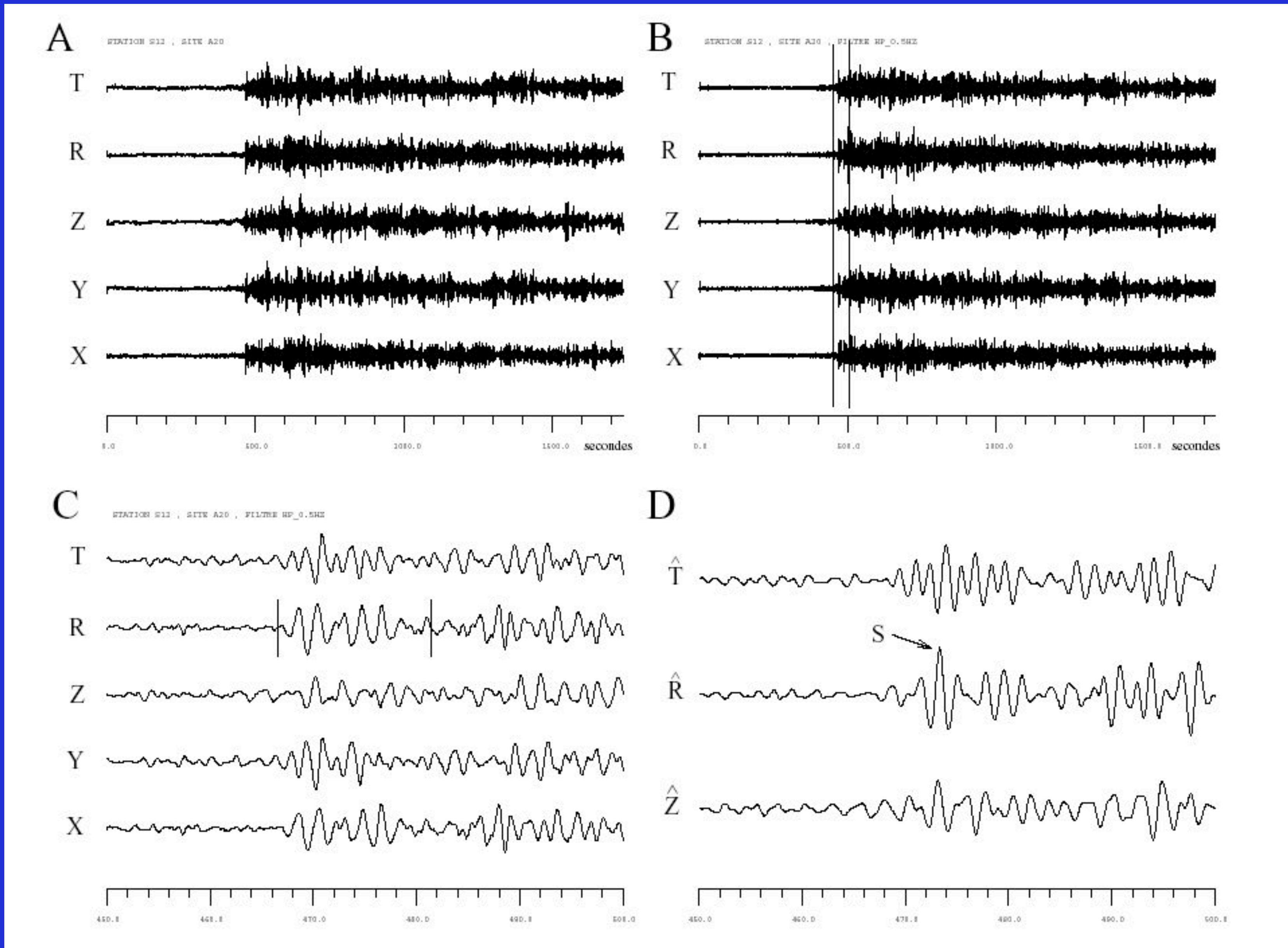
Receiver function method

- 1st step : make the Fourier transformation of the arrivals
  - Primary wave arrival Fourier Transformation ~ $T\ P(\omega)\ \exp(i\omega t_p)$
  - Converted wave ~$C\ P(\omega)\ \exp(i\omega t_x)$
- 2nd step: perform the deconvolution of the converted wave by the primary wave in frequency domain
  - $R(\omega)= [T\ P(\omega)\ \exp(i\omega t_p)] / [C\ P(\omega)\ \exp(i\omega t_x)] = T/C\ \exp(i\omega(t_p - t_x))$
- 3rd step: perform the inverse Fourier transformation
  - $R(t)=T/C\ \delta((t_p - t_x)$

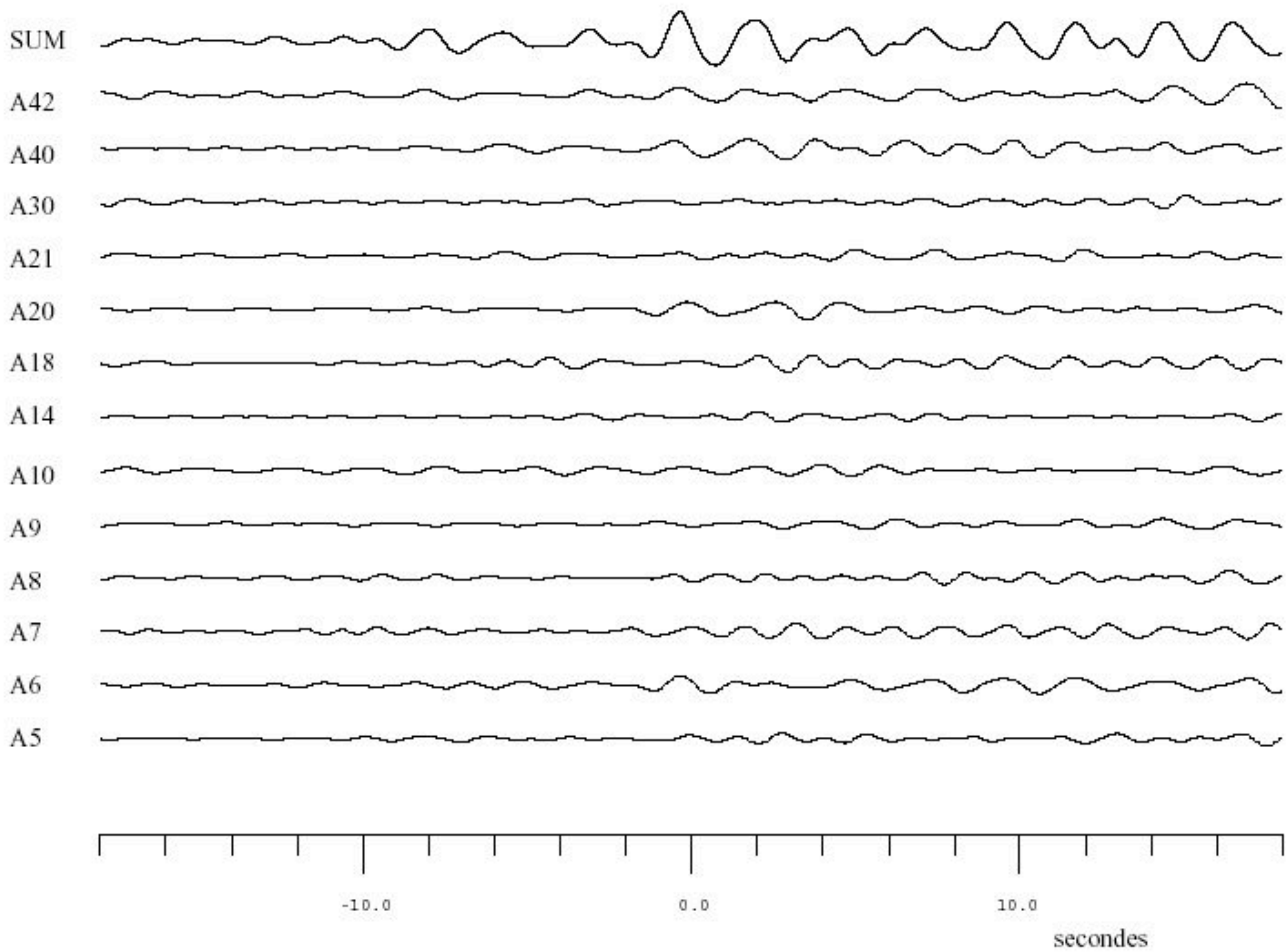




Deconvolution process



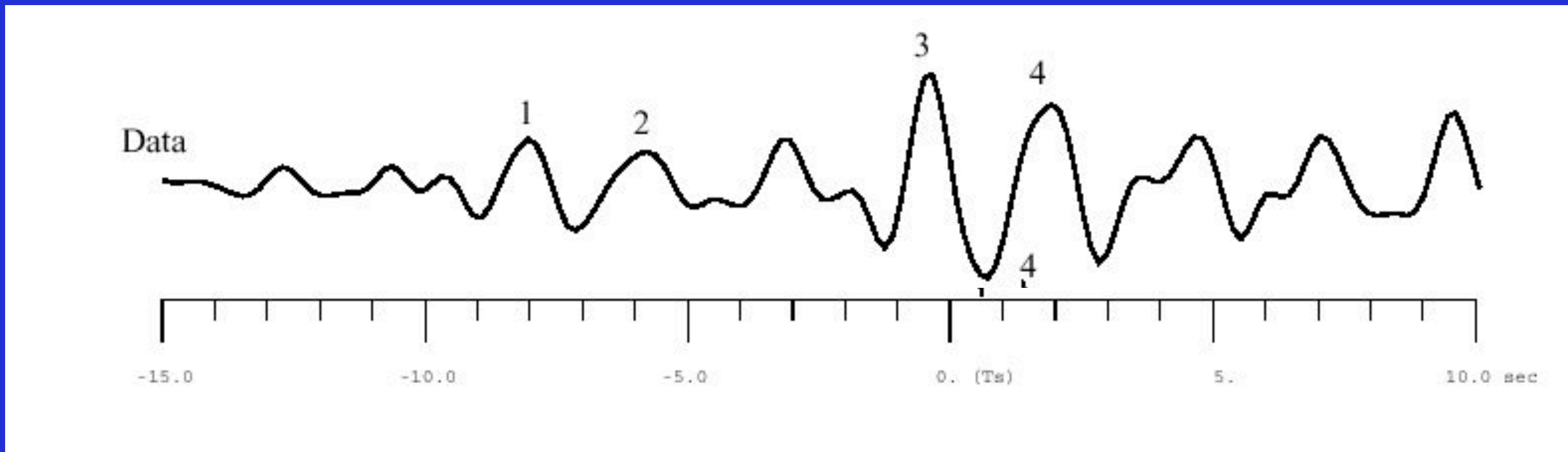
Improving the signal to noise ratio with stack



Moon receiver function (Apollo 12 site)

S->P conversion at the Base of the crust

Subsurface/regolith delay and reverberation

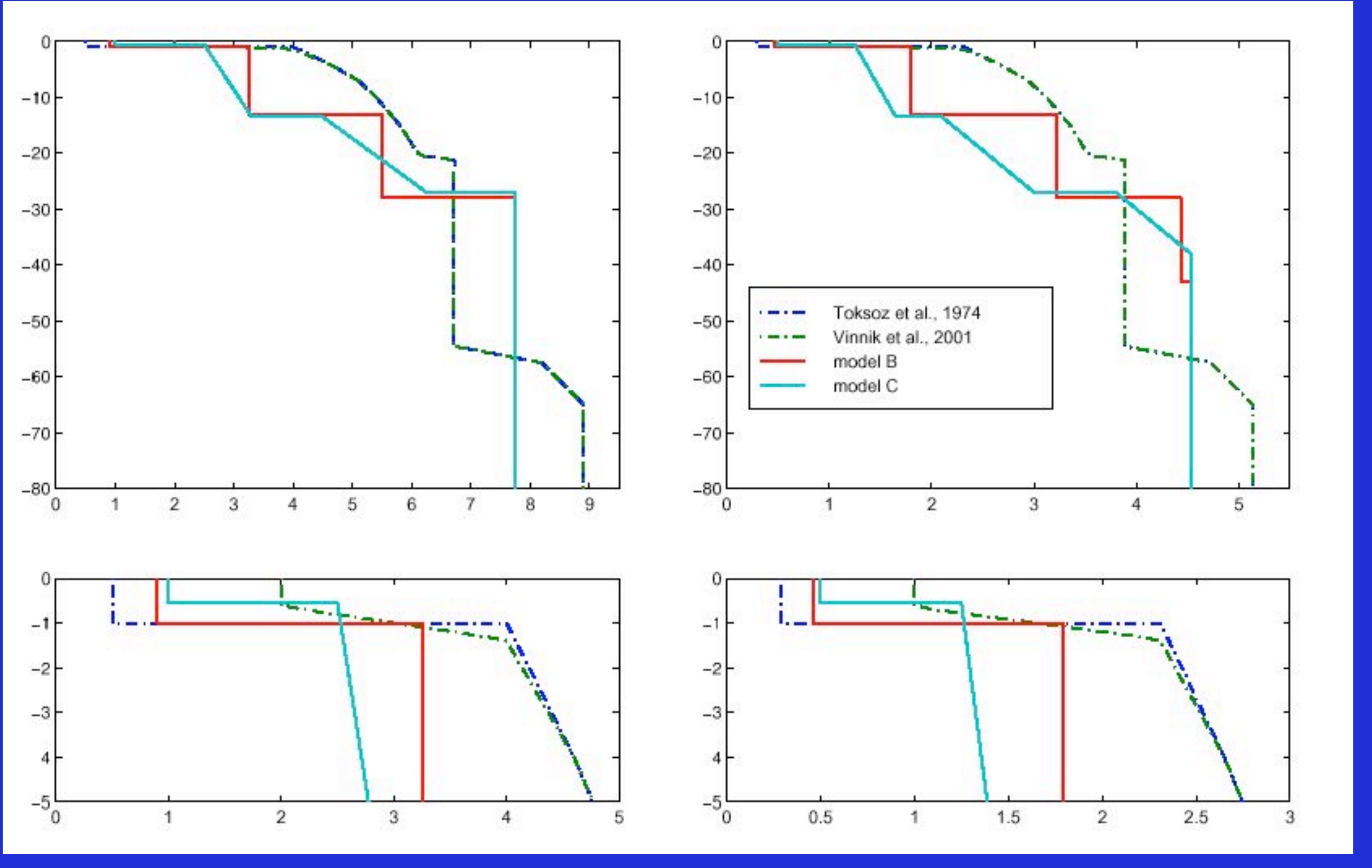


• S-P delay is equal to

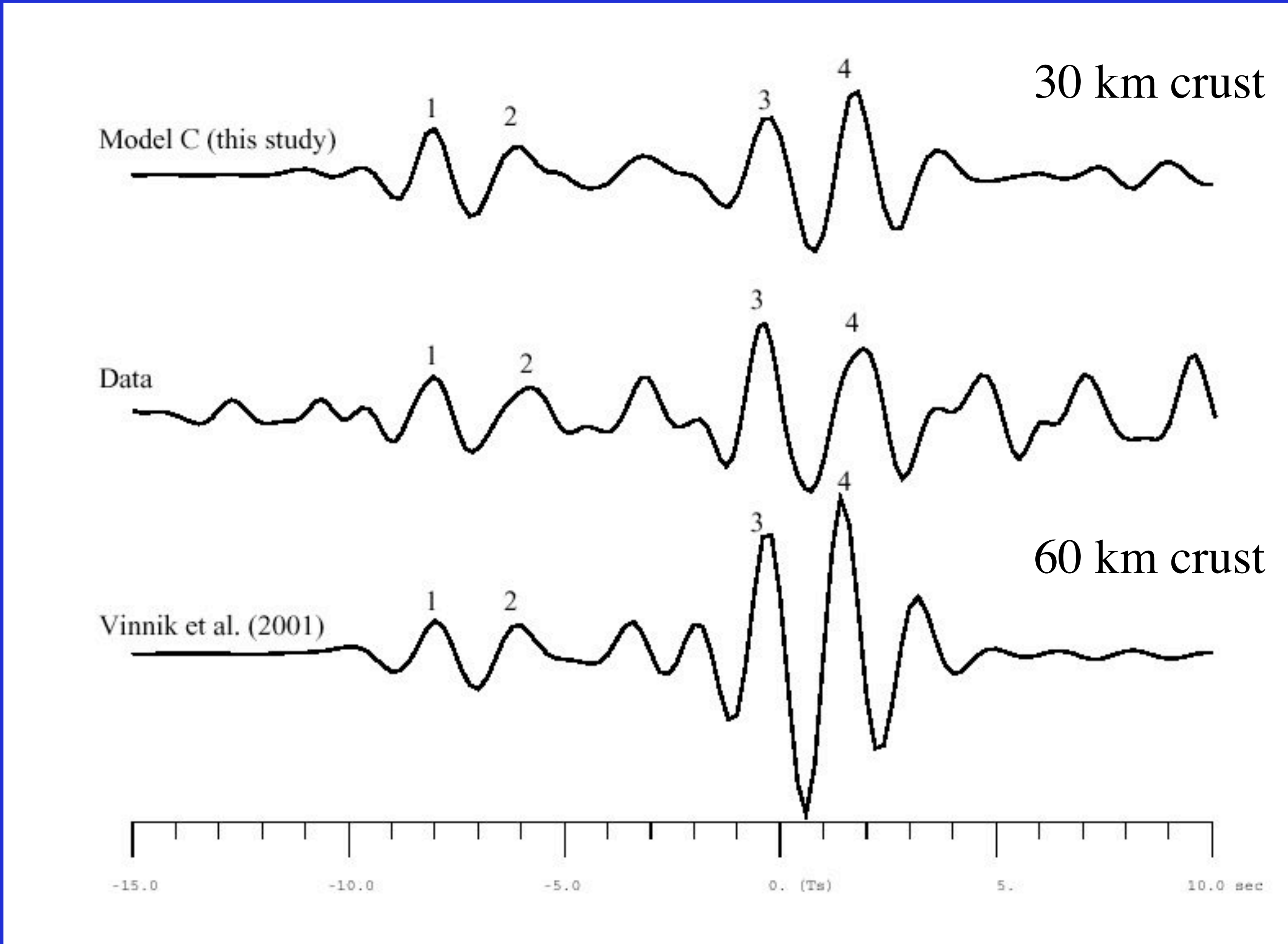
$$\Delta t = t_s - t_p = D(\frac{1}{v_s} - \frac{1}{v_p})$$

And therefore does not give a unique solution

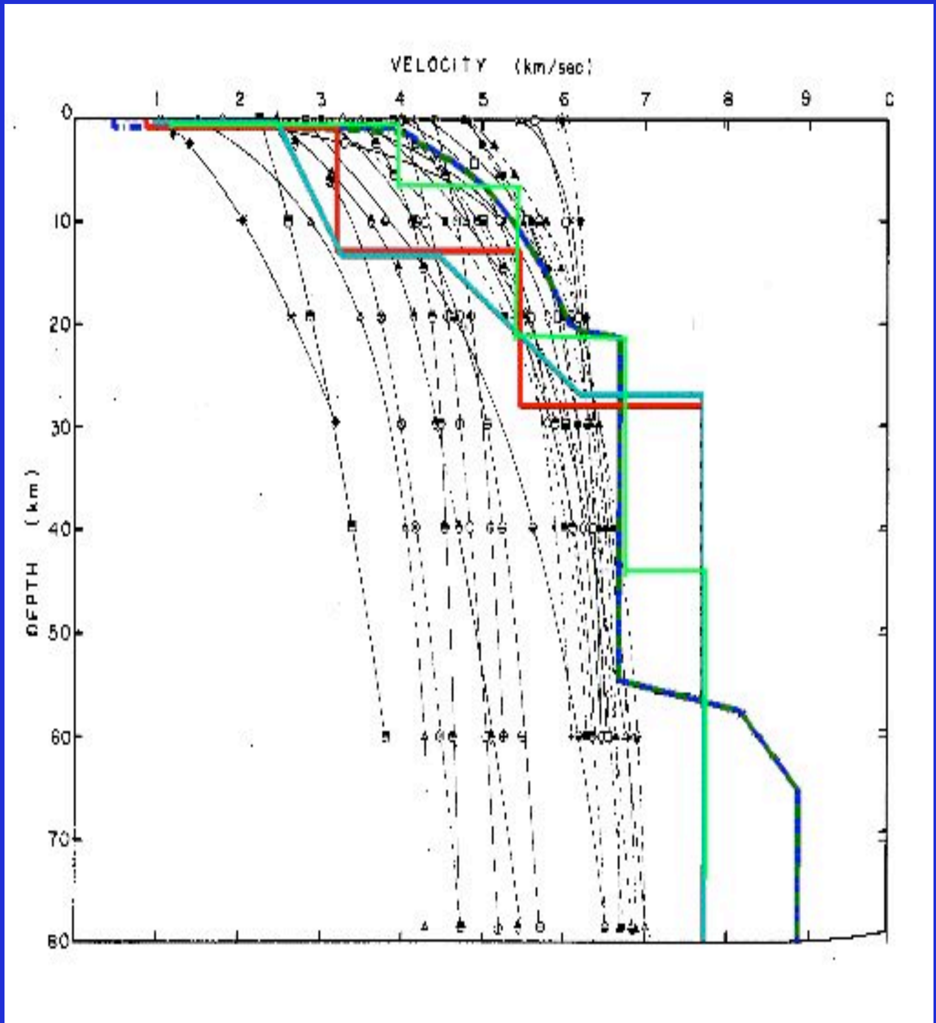
• other informations are needed (amplitude conversion coefficient)

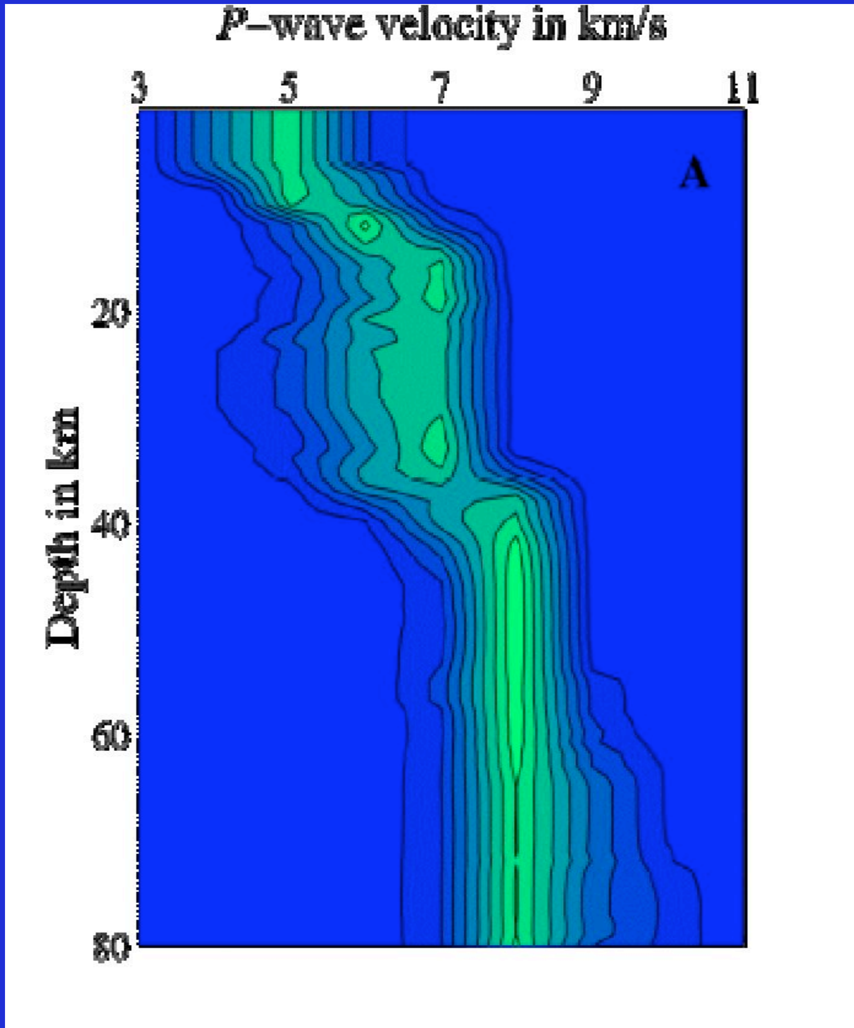

Toksoz et al., 1974
Vinnik et al., 2001
model B
model C
0
-10
-20
-30
-40
-50
-60
-70
-80
0
1
2
3
4
5
6
7
8
9
-1
-2
-3
4
-5
0.5
1.5
2.5

Moon receiver function (Apollo 12 site)



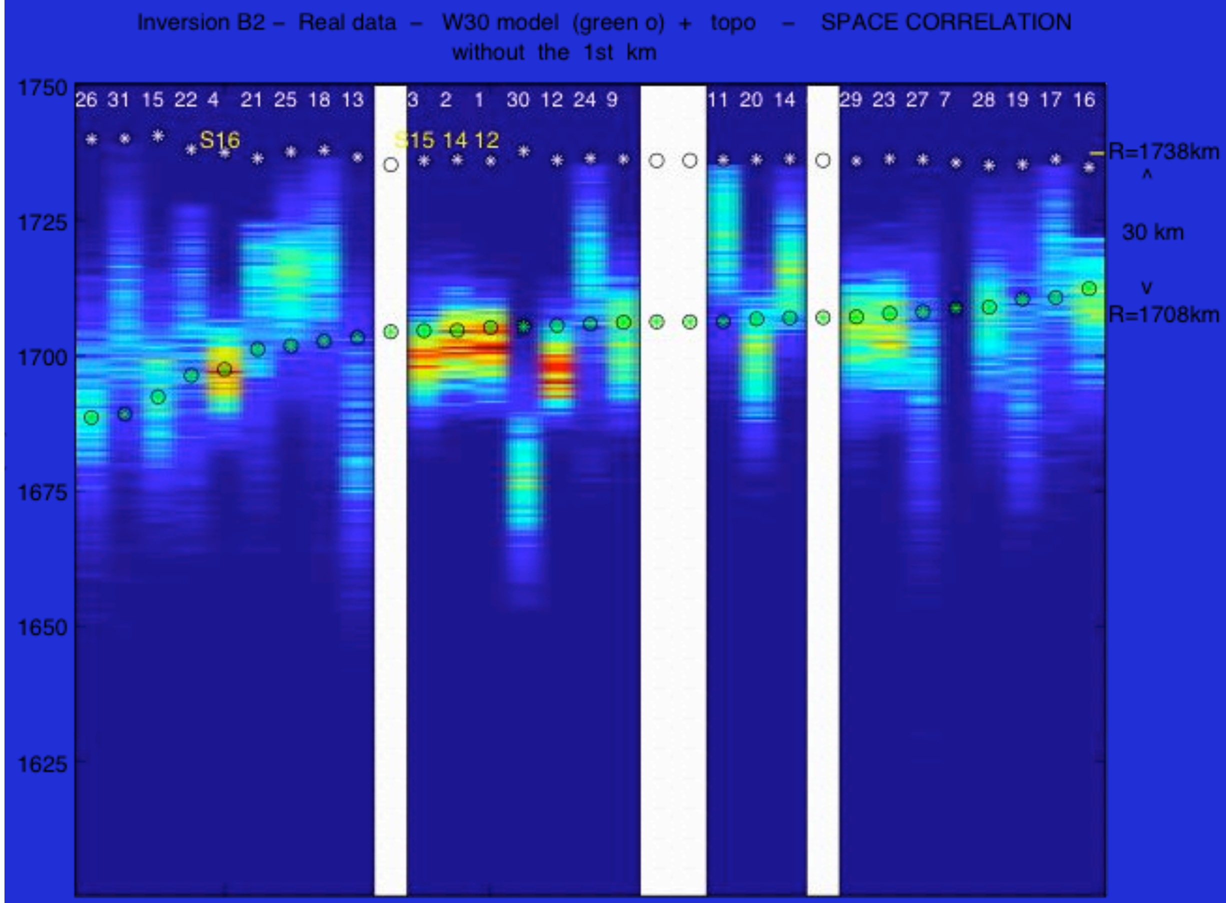
Crustal models





Lateral variation

- Idea: use of the meteorite impacts (only 3 source parameters)
- homogeneous crust with measured topography

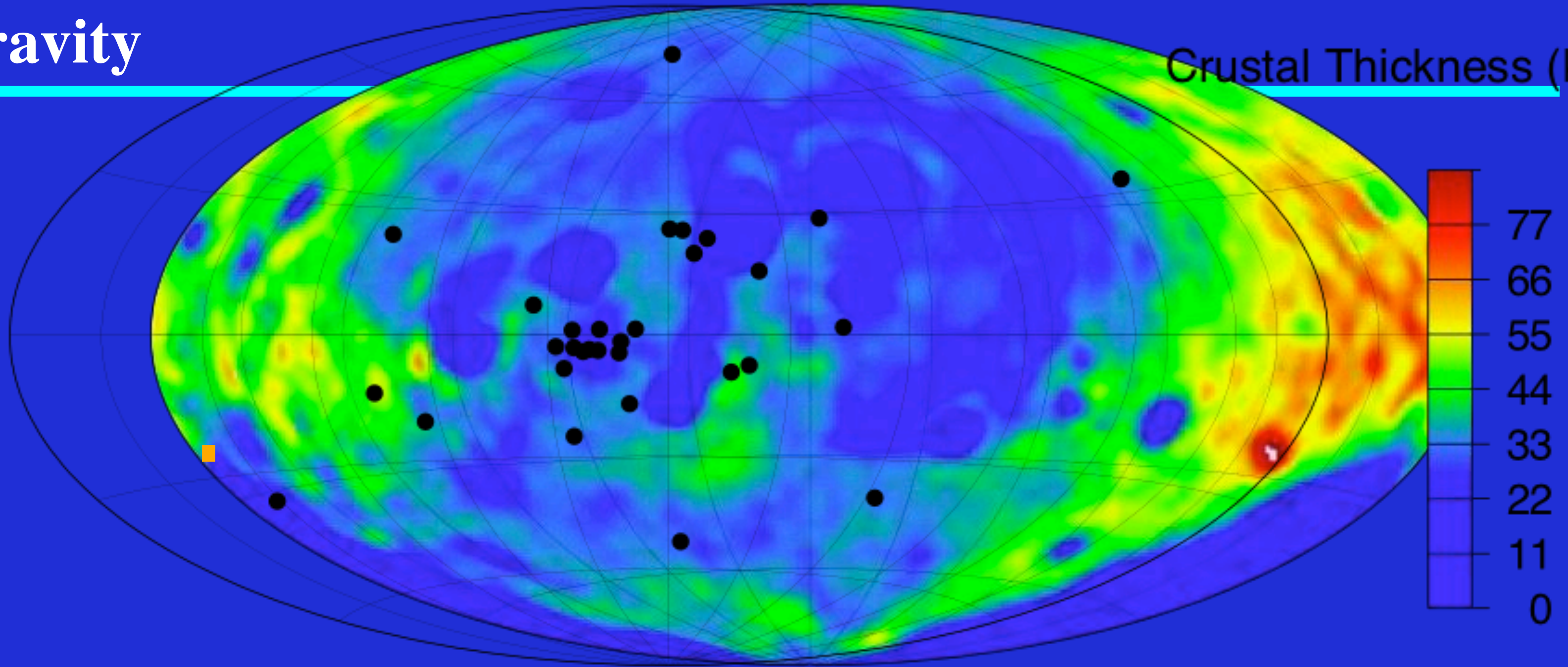


•Monte Carlo inversion of the crustal thickness from arrival time for all impacts and stations

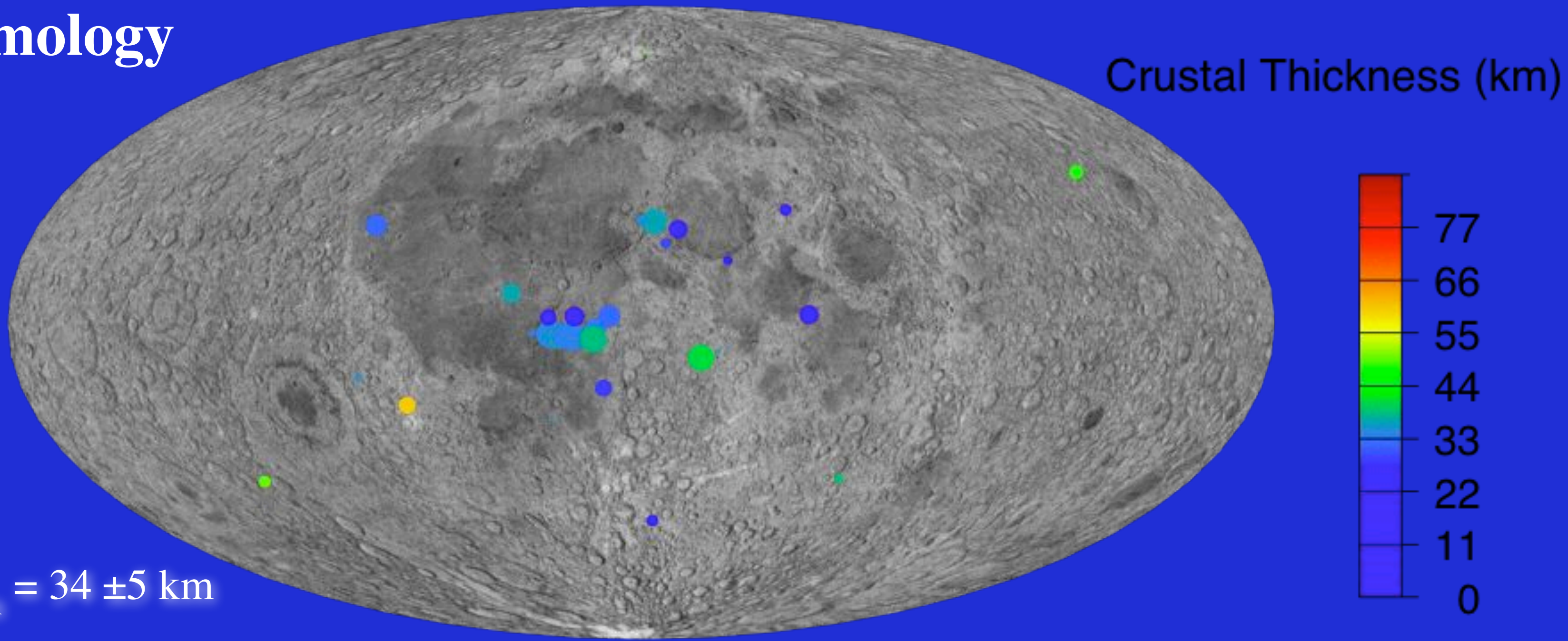
•Comparison with the estimation of crustal thickness (Airy hypothesis)

•Chenet el al. (2004)

Gravity

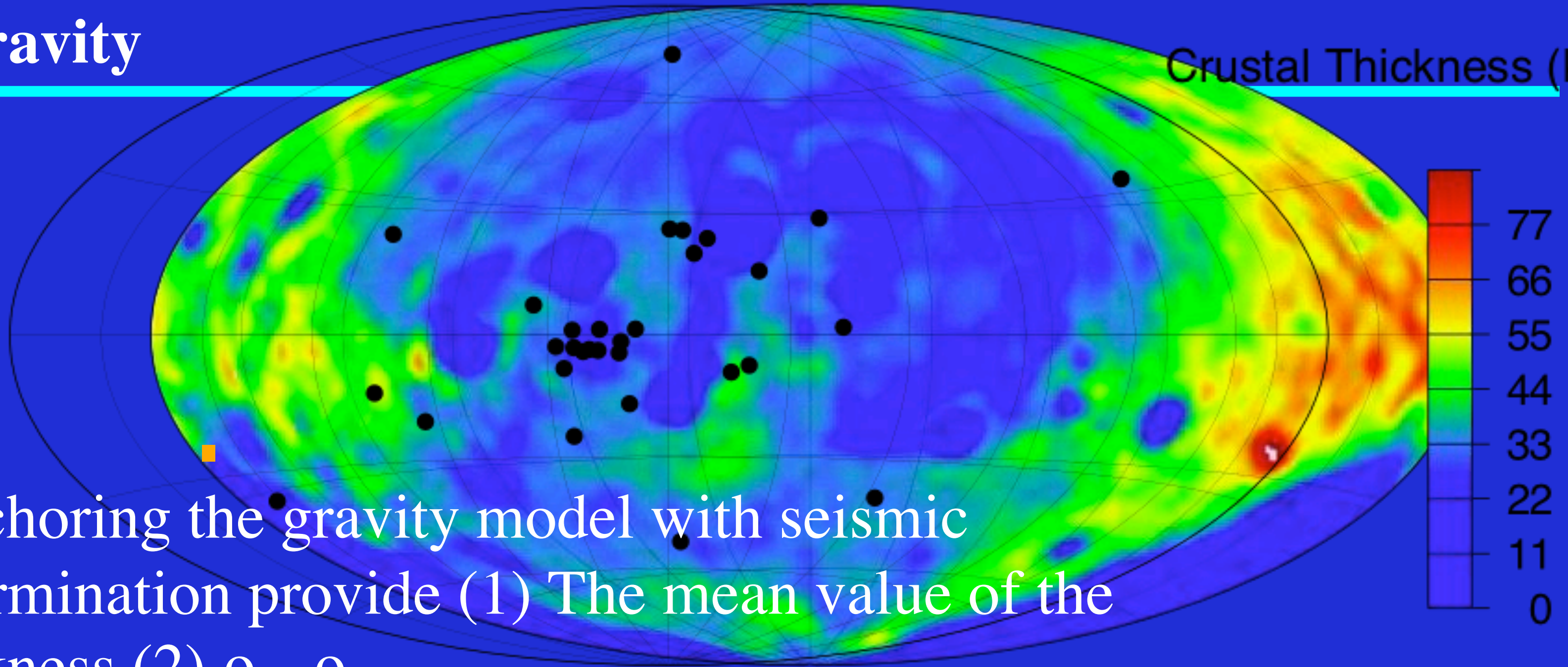


Seismology



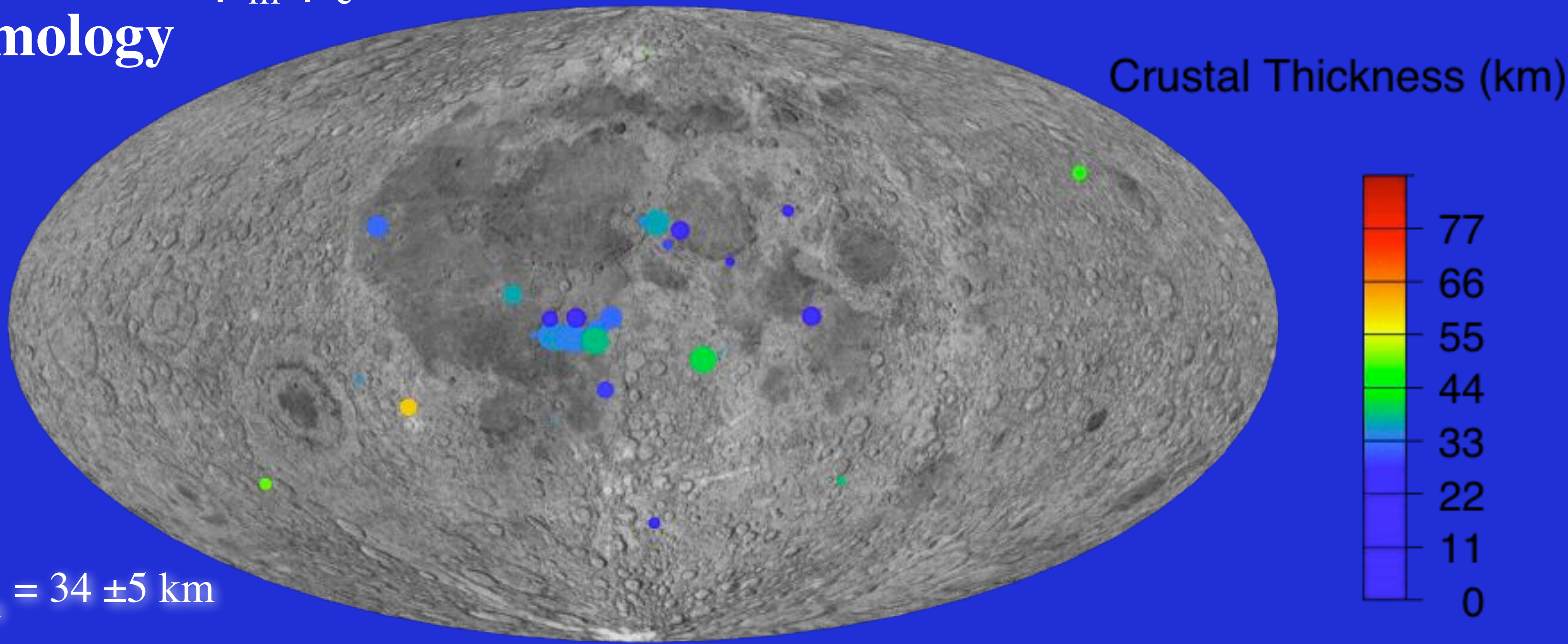
$T_{mean} = 34 \pm 5$ km

Gravity



- anchoring the gravity model with seismic determination provide (1) The mean value of the thickness (2) ρ_m-ρ_c

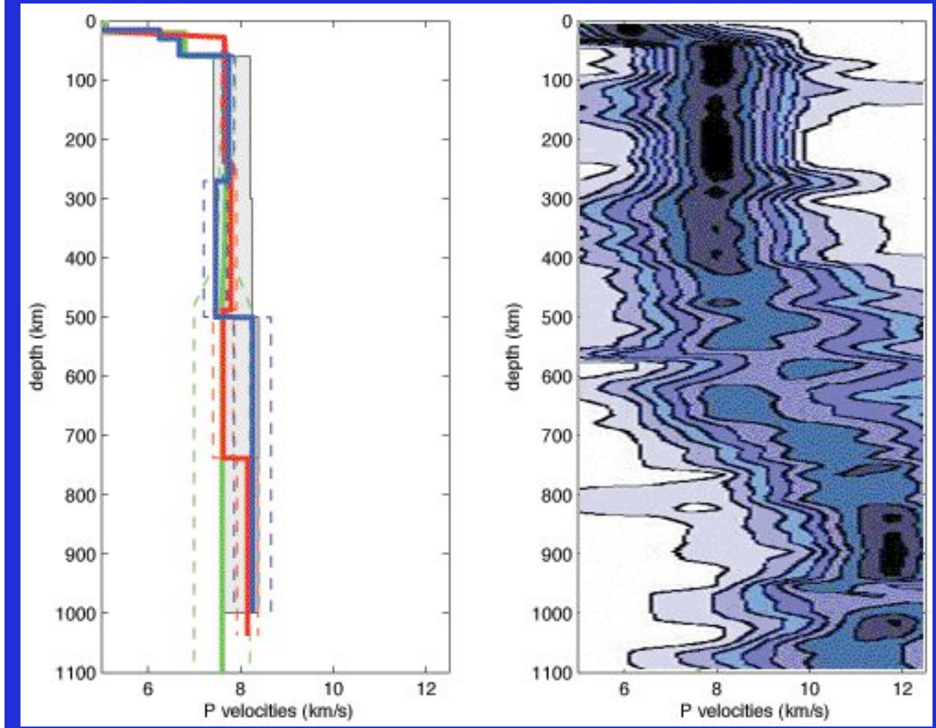
Seismology

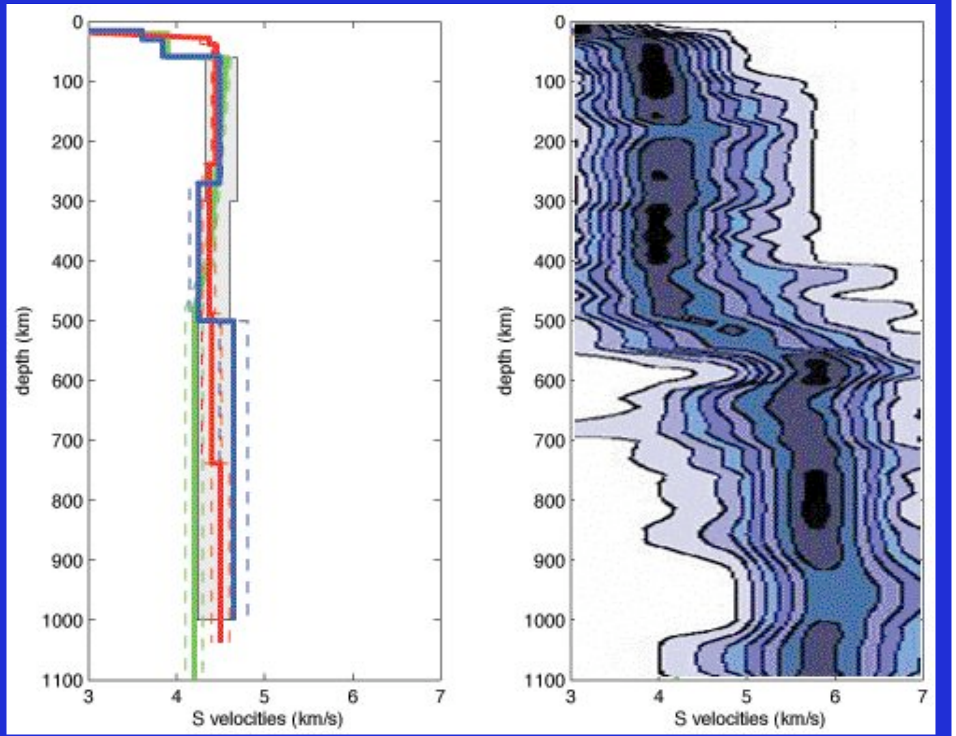


T_{mean} = 34 ±5 km

Interpretation of the seismic models

Mantle seismic model





Left figure shows the layered models of *Goins et al*. [1981] (Green), *Nakamura et al* [1983] (blue) and *Gagnepain-Beyneix et al* [2004] (red). Right figure shows the probability distributions of *Khan et al*. [2002]

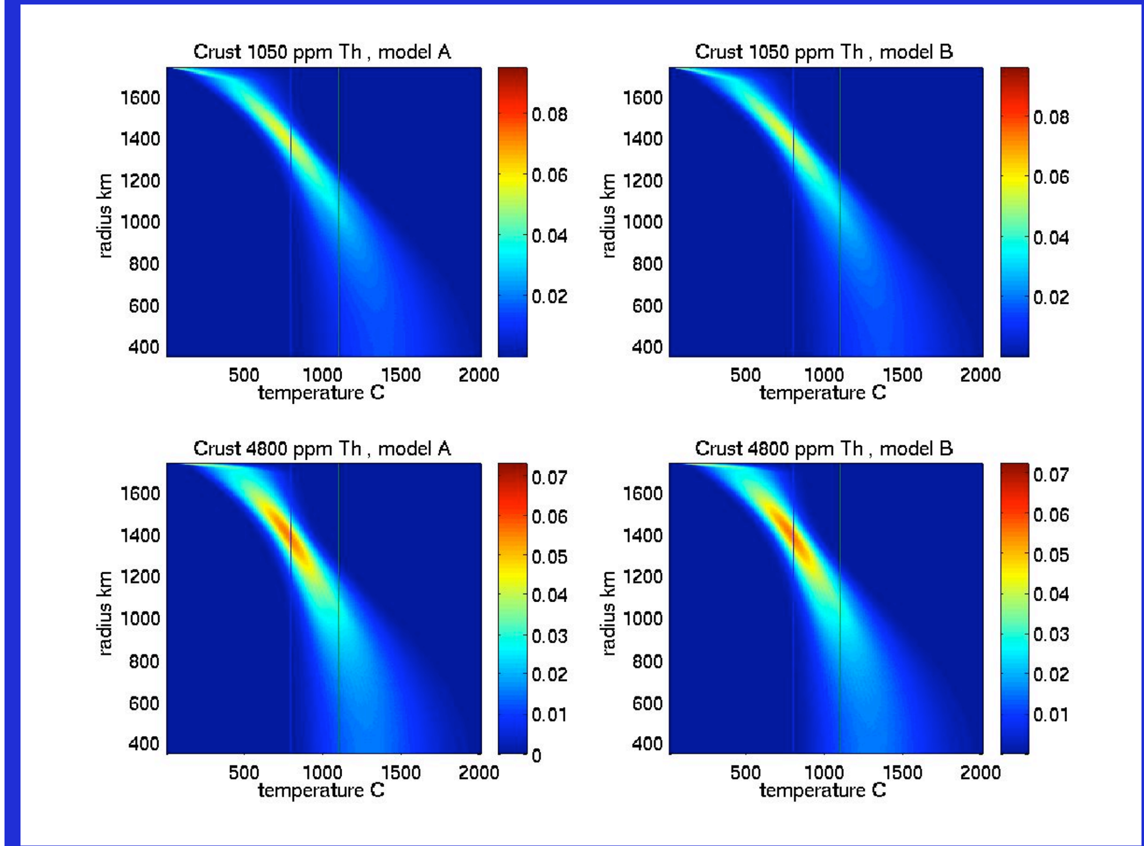
Mineralogical interpretation

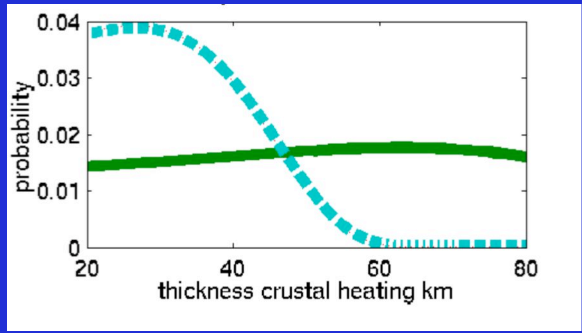
	1	2	3	4	5	6	7	8
SiO2	43,69	42,3	46,1	44,78	54,13	52,3	50,2	54
Al2O3	7,65	3,62	3,51	4,32	5,1	4	4	4
FeO	13,12	16,62	12,62	9,14	13,76	20,7	17,6	23,8
MgO	29,36	34,54	34,97	38,25	22,94	20	25,2	15,2
CaO	6,18	2,92	2,8	3,51	4,07	3	3	3

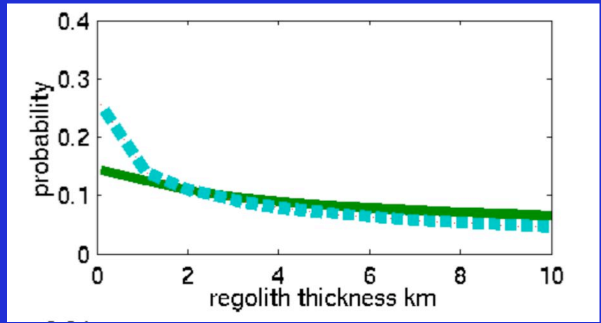
- We use 8 mineralogical models listed by Kuskov, 1995
- Model 5 and 7 are those with smaller temperature differences with respect to the a priori temperature profile
- But…
  - incompatibility with the mantle density (as constrained by the inertia factor) (cold temperature for seismic velocities, hot for density for 7)
  - Model 5 is constrained by the composition of mare basalts (Ringwood and Essene, 1970)
  - Incompatibility with the crustal thickness from temperature modelling

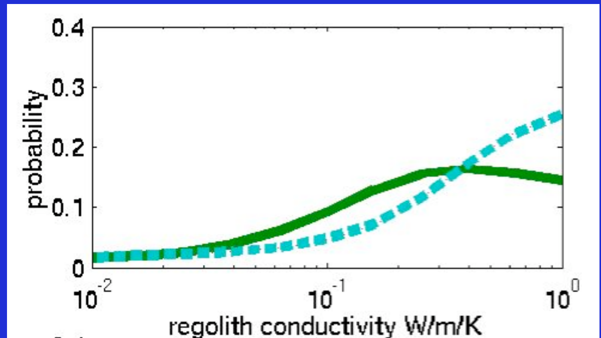
Temperature modeling

- Fit of seismic velocities for a known mineralogy
  - Seismic velocities are mainly a thermometer constraining the temperature
- Temperature model with regolith insulation, crustal heating, and upper/lower mantle heating









Sounding the Lunar core

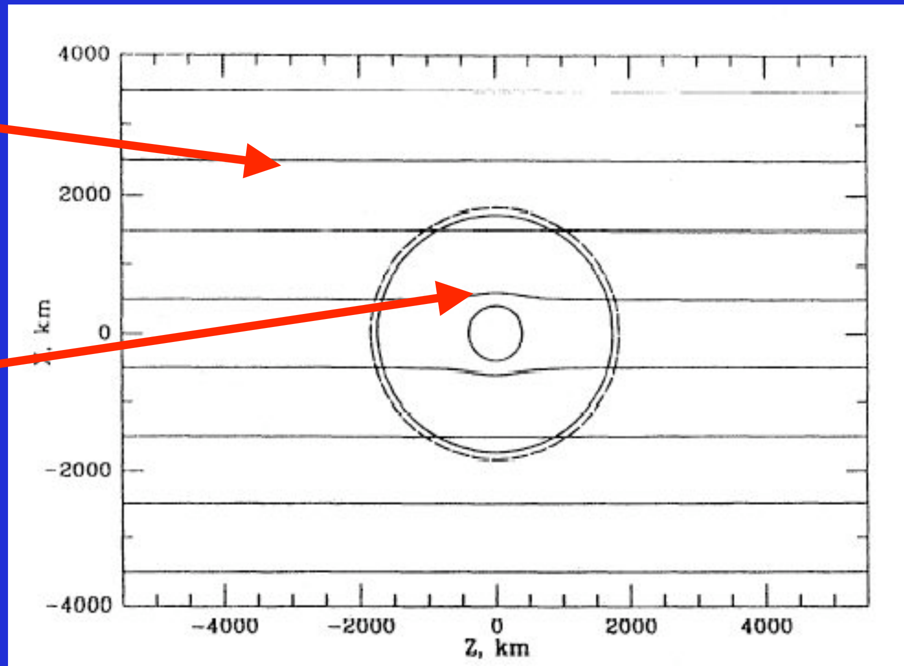
- no (direct) data from the Apollo seismic data
- data available:

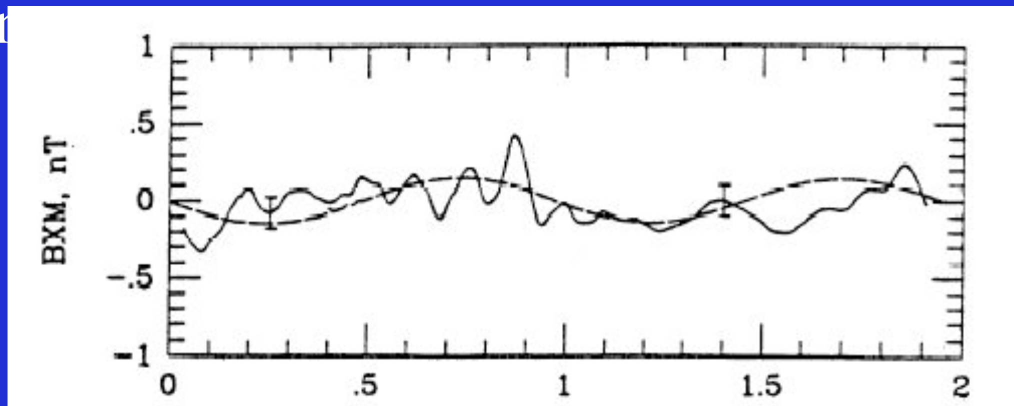
Density and moment of inertia

Magnetic sounding

Lunar prospector magnetic sounding

- Primary magnetic field is the magnetic field of the geotail (12-16nT)
- Magnetic field is slighly expulsed from the iron moon core
- A low altitude orbiting satellite with magnetometer (Lunar Prospector) measure the small (0.4 nT) perturbation
- Best fit is achieved with a metallic core of 400 km





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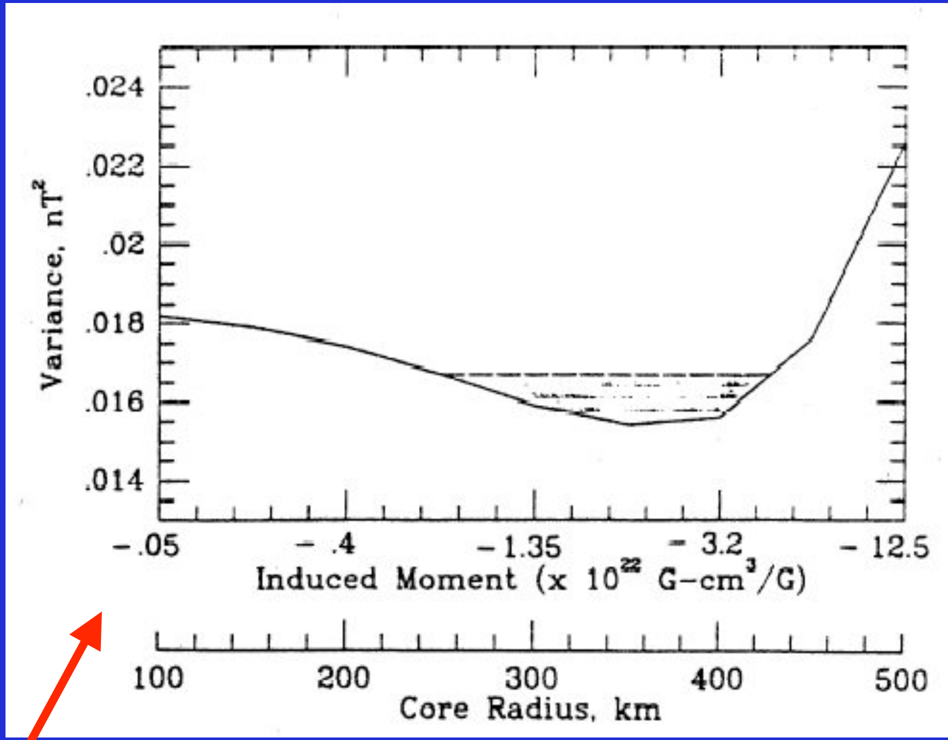
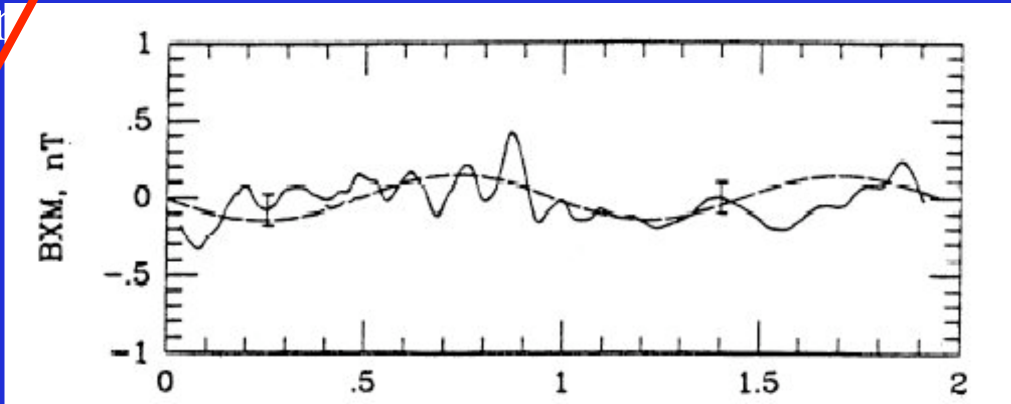




Geodetic and tidal informations

- Density: 3346.5±1.5 kg/m3
- Normalised moment of inertia: 0.3935±0.0002
- Love number
  - Amplitude of k_2 = 0.0227±0.0025 and associated tidal Q = 33 ±4

Interpretation

$$I = \int dV \rho d^2 \approx \frac{2}{3} \int dV \rho r^2$$

Homogeneous case

$$\frac{I}{Ma^2} = 0.4$$

$$-\frac{dP}{dr} + \rho g = 0$$

$$M(r) = \frac{4\pi}{3} \rho r^3$$

$$g = -G\frac{M(r)}{r^2} = -\frac{4\pi G}{3}\rho r$$

thus

$$P = P_s + \frac{2\pi G}{3}\rho^2(a^2 - r^2)$$

Incompressible planet

$$\frac{I}{Ma^2} = 0.4[1 - \frac{4}{35}\zeta + \frac{8}{1575}(9+10b)\zeta^2] = 0.3986$$

$$\rho\vec{g} - \nabla p = \vec{0}$$

$$\frac{1}{K} = \frac{1}{\rho}\frac{\partial \rho}{\partial P}$$

$$\nabla.\vec{g} = -4\pi G\rho$$

thus, if $K=K_0+bp$ ($K_0=1.5\ 10^{11}$ Pa, b=8)

$$\rho = \rho_0\left[1 + \frac{\zeta}{a^2}(a^2 - r^2) - \frac{\eta}{a^4}(a^2 - r^2)^2\right]$$

with

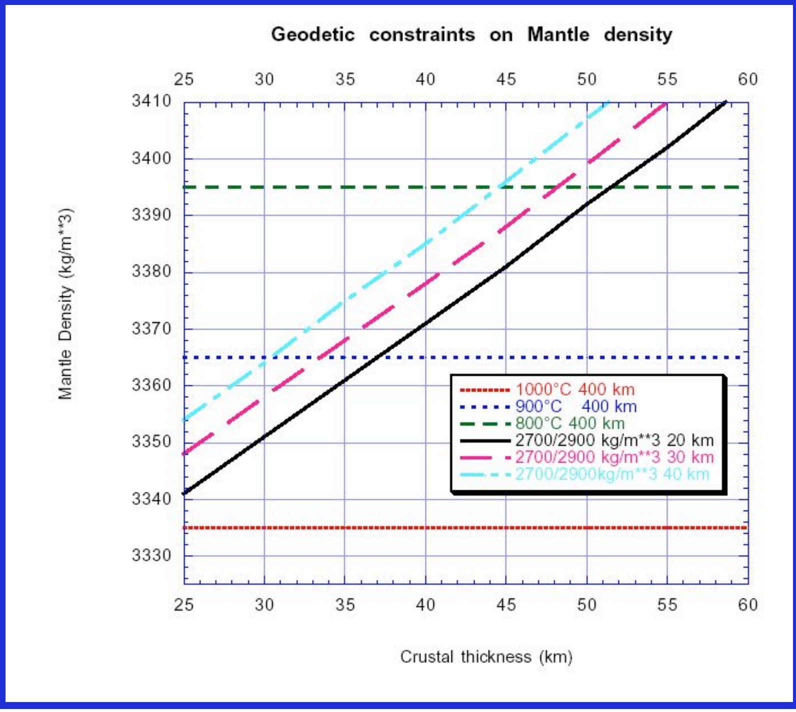
$$\zeta = \frac{2\pi}{3}\frac{G\rho_0^2 a^2}{K_0}, \eta = \frac{2\pi^2}{9}\frac{bG^2\rho_0^4 a^4}{K_0^2} = \frac{1}{2}b\zeta^2$$

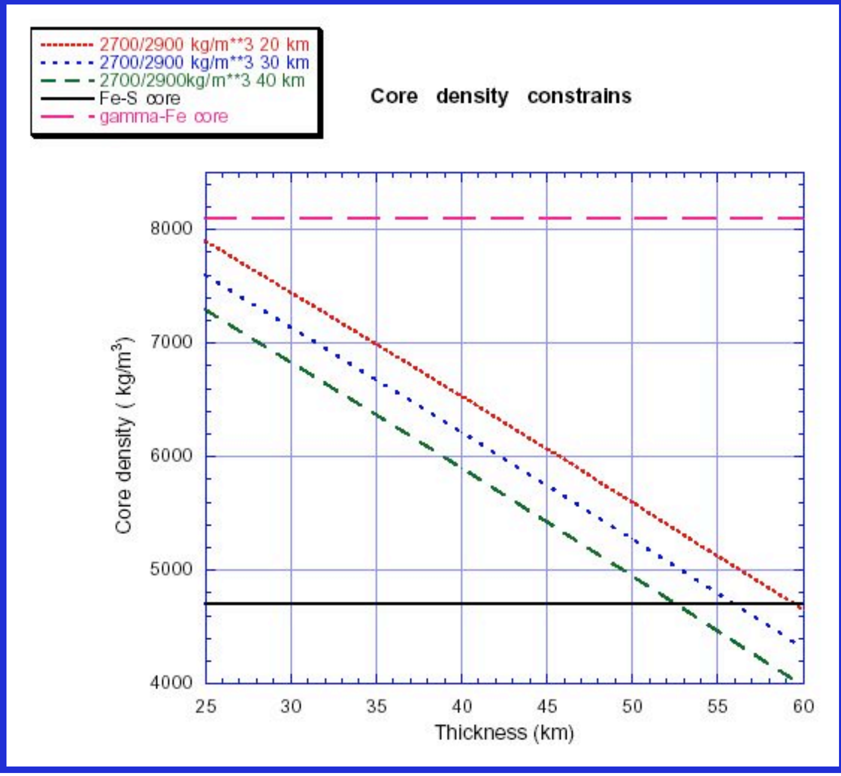
and

$$M = \frac{4\pi}{3}a^3\rho_0\left(1 + \frac{2}{5}\zeta - \frac{8}{35}\eta\right), I = \frac{8\pi}{15}a^5\rho_0\left(1 + \frac{2}{7}\zeta - \frac{8}{63}\eta\right)$$

Moon ….

- The moment of inertia must therefore be decreased with structure with increasing density with depth
  - crust (2900-3000 kg/m^3, 40-70 km)
  - core (4000-8000 kg/m^3, 300-500 km)
- Strong indetermination remains for a core,mantle,crust model (5 unknown for two data)
  - Crustal thickness and density must be constrained by seismic data and gravity
  - Love number can be added to inversion

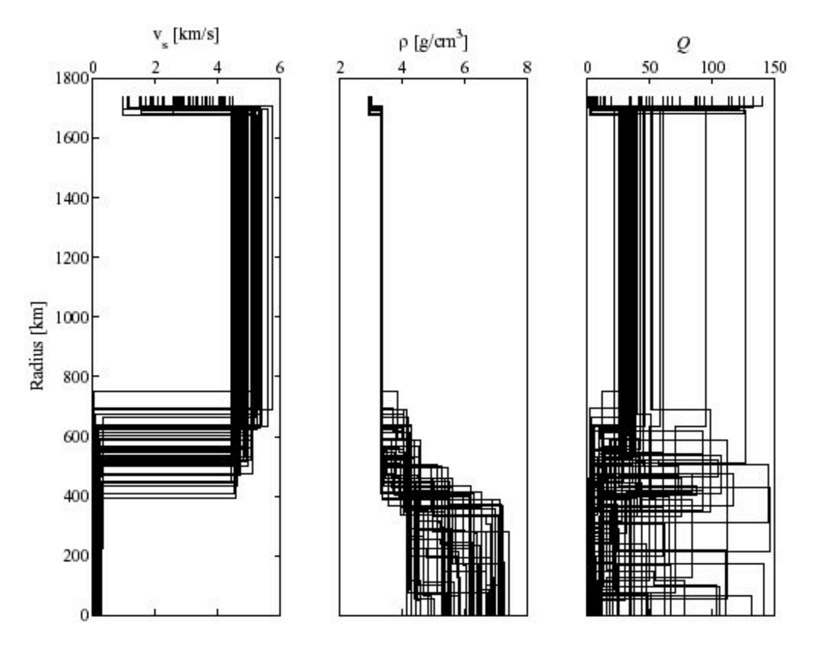




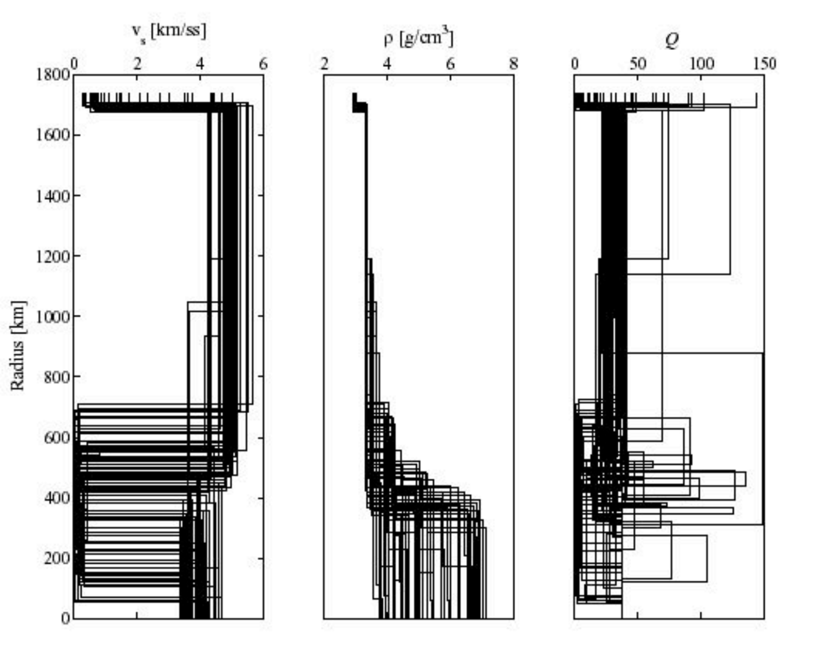
Deep interior and state of core (1/3)

- Liquid core models and solid core model can be tested
- Monte Carlo inversion of the density, Love number and Inertia factor (Khan et al., 2004)
- A posteriori probability favour a liquid core of about 350 km

Liquid core models

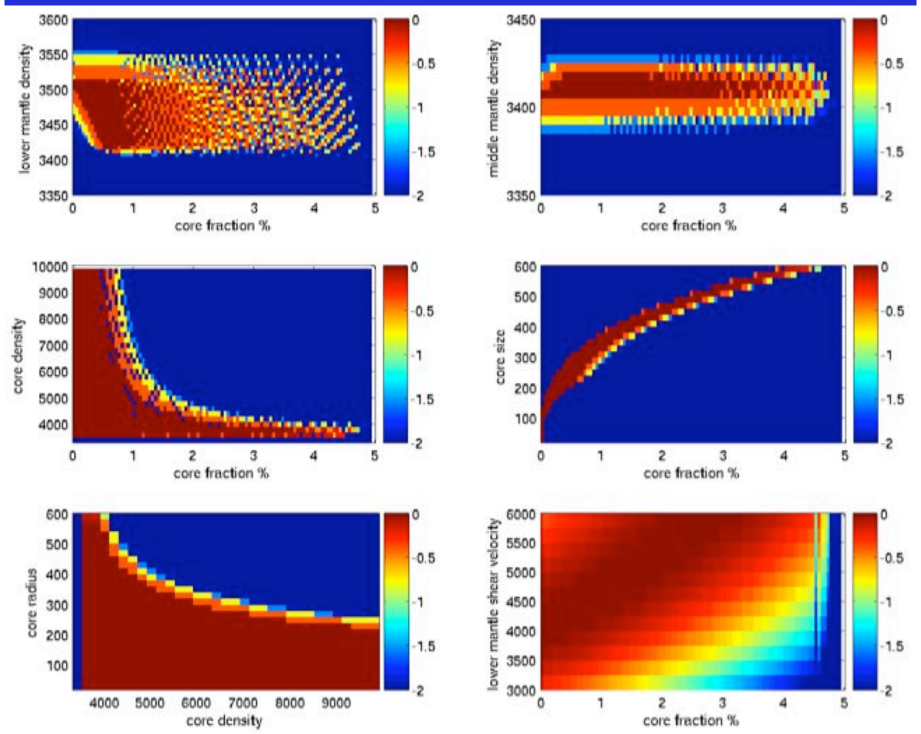


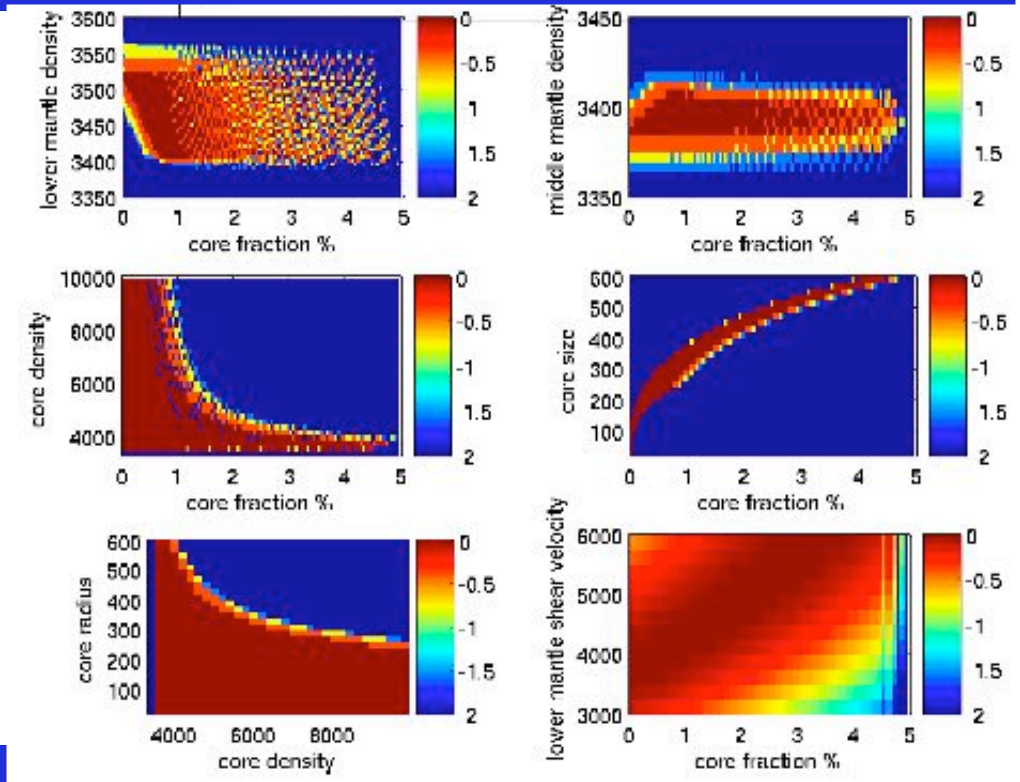
Solid core models



Deep interior and state of core (2/3)

- Upper structure is constrained by seismic data
- Invert only for the structure not resolved by seismic data



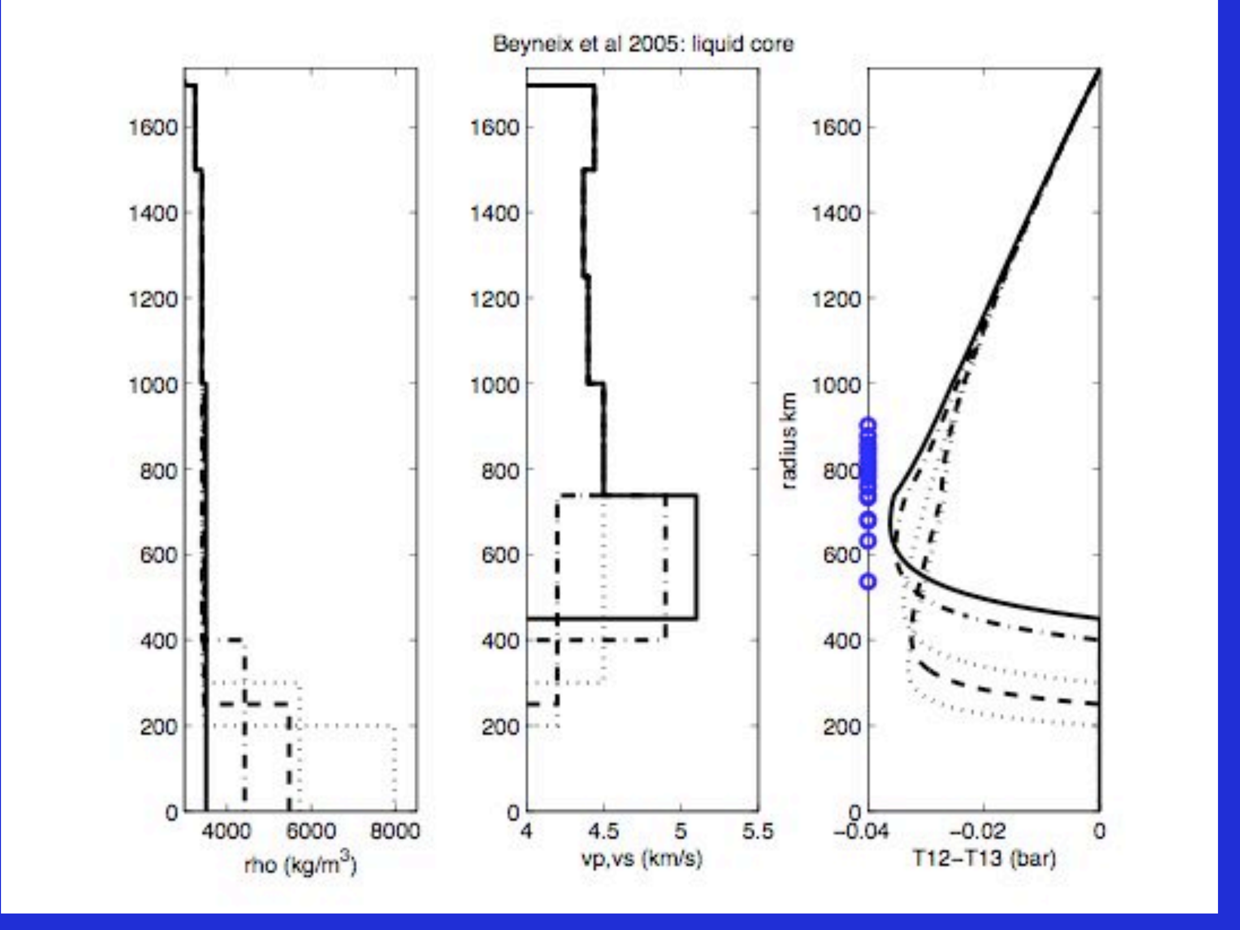


Crust 40 km (Beyneix et al., 2005)

Crust 70 km (Nakamura. 1983)

Deep interior and state of core (3/3)

- Deep moonquakes and maximum of stress

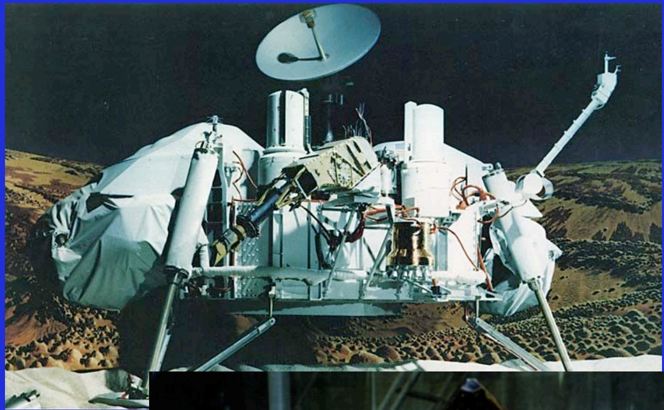


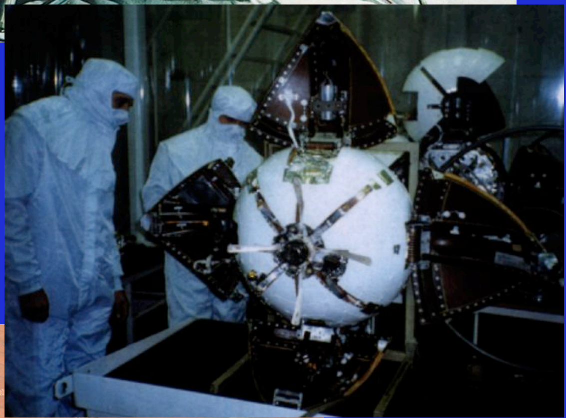
- Large core seems more likely > 350 km with therefore relatively low density

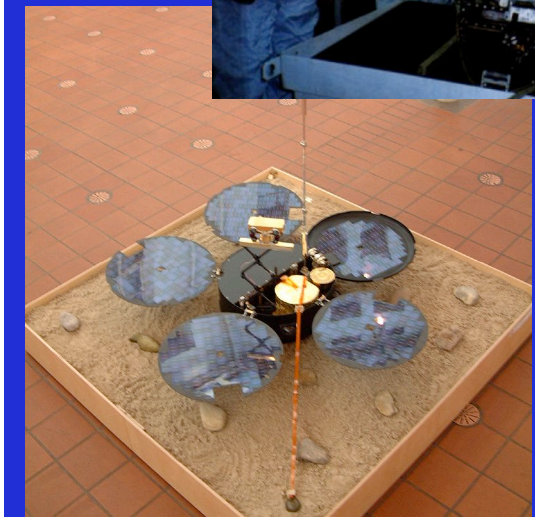
Conclusion for the Moon

- Mean Crustal thickness is about 40 km
- The crust is mainly an anorthosite crust with low density (r~2800-2900 kg/m3)
- Pyroxenite upper mantle resulting of a magma ocean in the early moon
- Possible more primitive lower mantle
- Core of 350-400 km, probably liquid and probably with light elements

Mars seismology: history



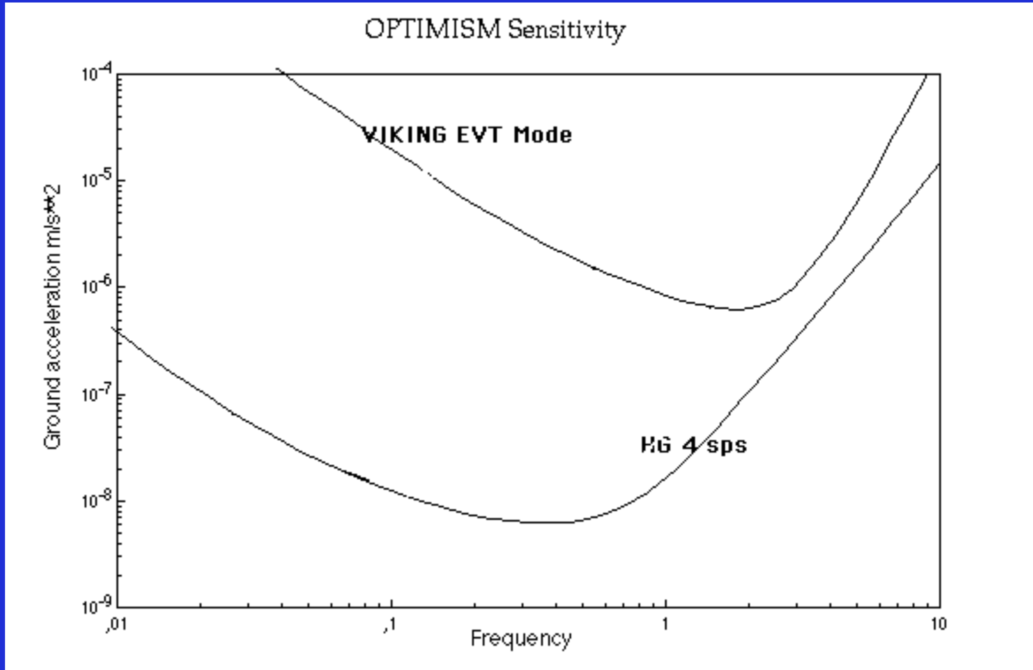


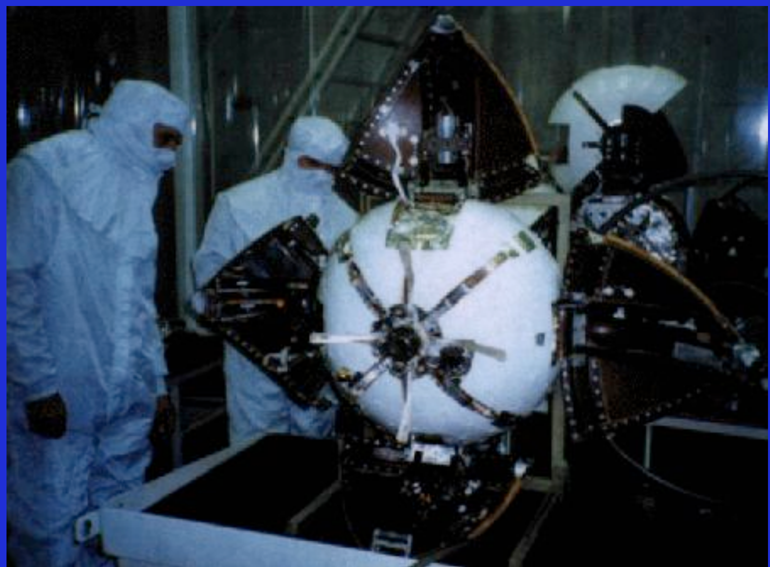


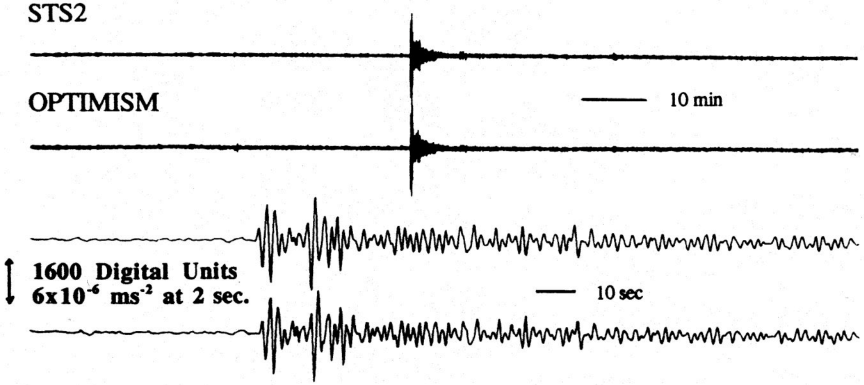
- 1975: 2 Viking landers equipped with seismometers. Possible detection of one quake on one lander
- 1996: Failure of the launch of Mars96, with 2 surface stations equipped with BRB Z axis seismometers and 2 penetrators with SP geophones
- 2003: The NetLander mission is stopped by CNES and NASA before phase B completion.
- Other data for deep interior:
  - Density
  - Inertia factor
  - Love number (real and imaginary part)
  - surface content in Th soon

PREVIOUS FAILED/LOST EXPERIMENTS (OPTIMISM/ VIKING)

• Viking Seismometer was too high frequency
Vertical seismometers in the two small Stations of Mars96
• about 0,8 kg including the electronics,
• 16 bits A/D and 12 bits D/A for thermal drift control
• lost after launch, 11/1996

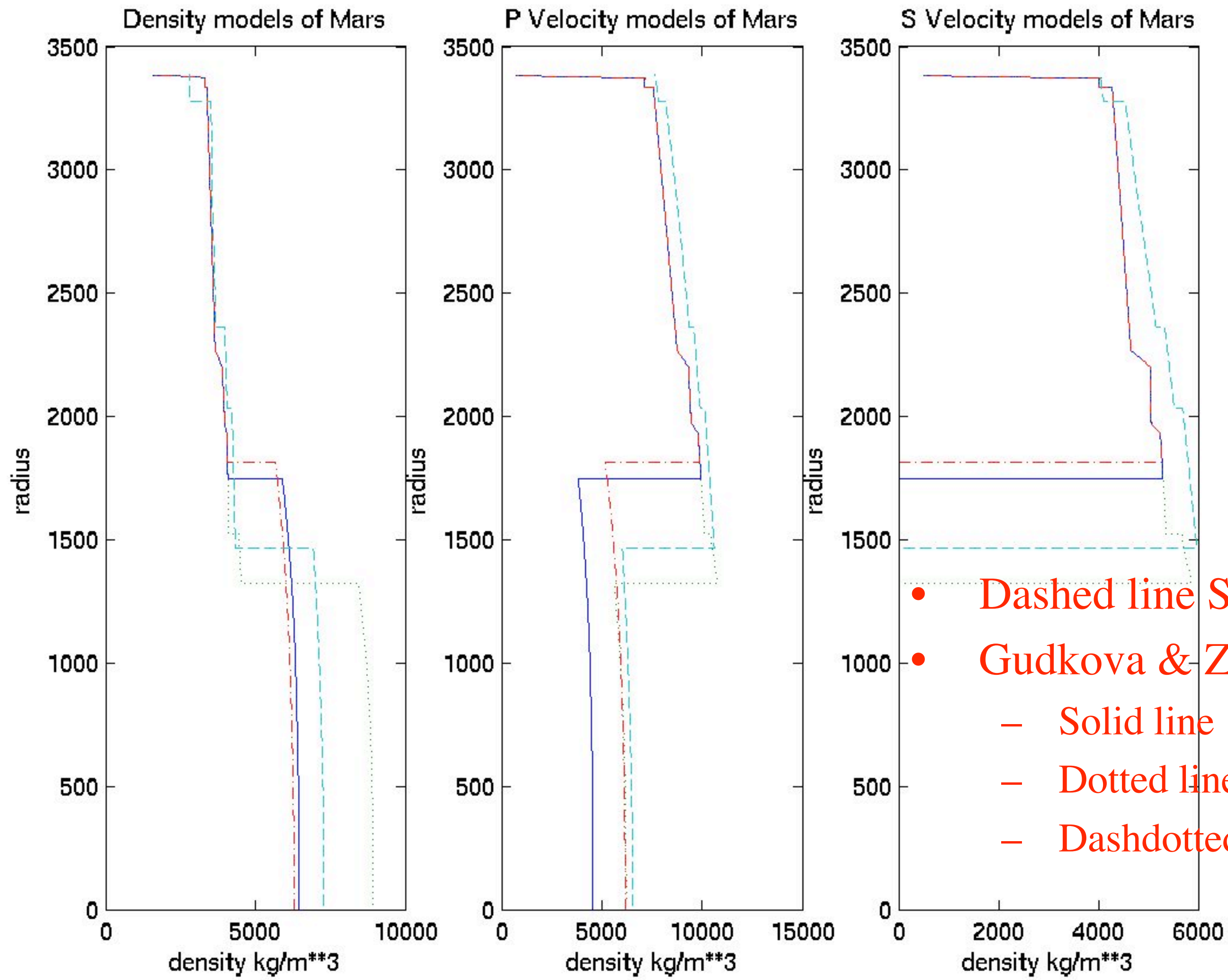






Ms=6.1 Quake recorded at 55° of epicentral distance in Pinion Flat Observatory, California.

Interior models: core size



- Dashed line Sohl and Spohn [1997]
- Gudkova & Zharkov (2004]
  - Solid line 50%H, 14% S
  - Dotted line 70% H, 14% S
  - Dashdotted line 35% S