

CRUST AND UPPER MANTLE SEISMIC SURVEYS (2)

OUTLINE

LARGE-SCALE REFRACTION SURVEYS

Interpretation of refraction profiles

- multilayered models, detection of a low-velocity layer
- dipping refracting boundaries, reversed profiles
- estimation of velocities, use of reduced travel-times

Crustal refraction surveys

- the Mohorovičić discontinuity (the “Moho”)
- continental crust: velocities, thickness, the Conrad discontinuity
- oceanic crust: velocities, layering, thicknesses

Seismic velocities in the upper mantle

- interpretation of travel-times through the upper mantle
- evidence for the low-velocity layer, regional variations

Background reading: Fowler §4.4, Lowrie §3.6 & 3.7

REFERENCES

- Blundell, D., Freeman, R. & Mueller, S., 1992. A Continent Revealed: The European Geotraverse. Cambridge University Press. *Summary of seismic refraction and reflection results across Europe.*
- Open University Course Team, 1990. Lithosphere Geophysics in Britain. Study Unit S339 1B, Open University Educational Enterprises. *Good review of seismic refraction and reflection results in Great Britain.*

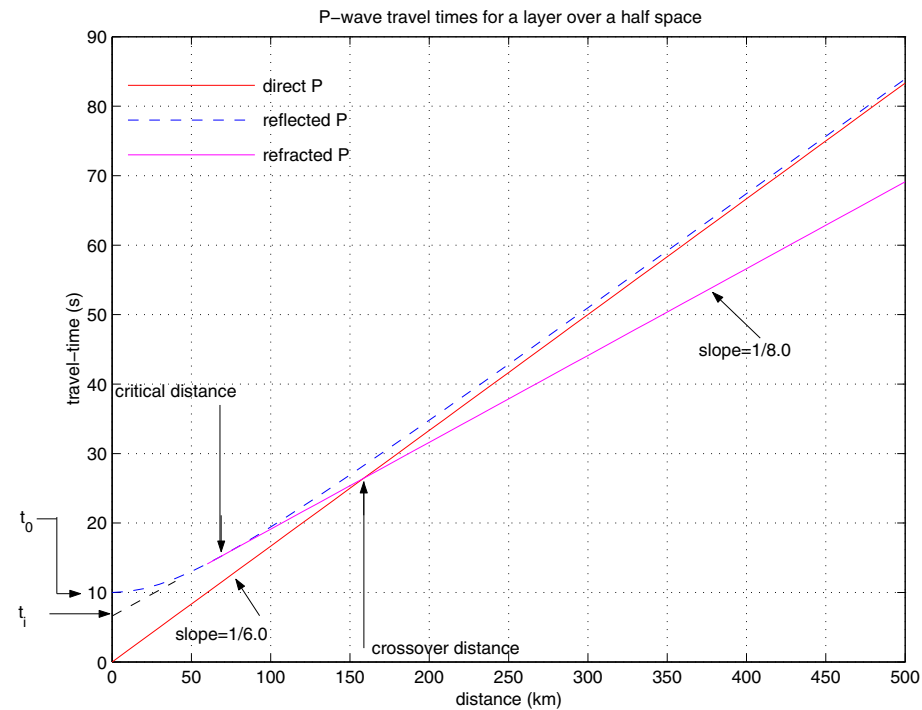
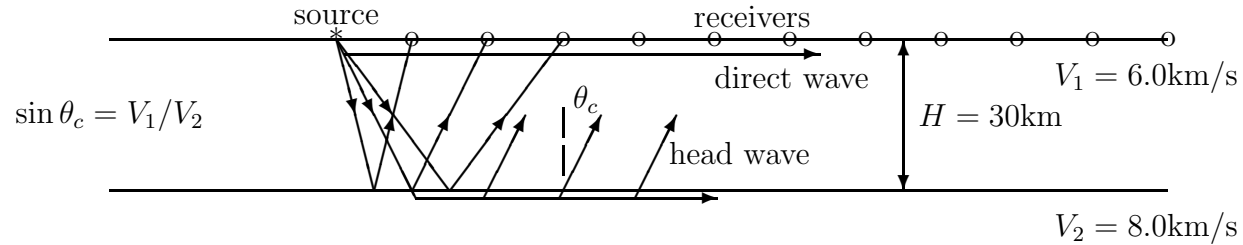
SEISMIC REFRACTION SURVEYS

Large scale seismic refraction are a primary source of information on the structure of the crust and upper mantle. At a much smaller scale they are used for engineering purposes (e.g. depth to bedrock) and in hydrology (e.g. depth to the water table). A seismic source is fired into a *spread* of seismometers out to a distance of typically 10 times the depth of investigation or more. So, assuming a depth of 30 km to the Moho, a crustal refraction experiment would ideally record data to 300 km or more. Recording seismic waves to this distance requires a sufficiently large source (e.g. depth charges). When this strong a source is not available, the experiment can make use of the post-critical reflections from the Moho as well as any refractions from beneath it that are recorded. The wide angles of these post-critical reflections are outside the range of conventional reflection surveying techniques.

The P-wave travel times are plotted against distance and interpreted in terms of depths to refracting interfaces and velocities of the refracting layers. In more advanced interpretation, *synthetic seismograms* are computed from a model of interface depth profiles and layer velocities and the model is altered until the synthetic seismograms match the refraction seismograms.

THE LAYER OVER A HALF SPACE

The model of uniform layer over a half space, described in "Seismic Travel Time Curves" of Lecture 4, encapsulates the basic principle of refraction seismology. It is the prototypal model for the interpretation of refraction profiles. The model and its P-wave travel-time curves, which should be memorised, are repeated below using idealised crustal refraction parameters.



INTERPRETATION OF REFRACTION PROFILES (1)

Multilayered structures, velocity gradients, low velocity layers, irregular and dipping layers introduce complicated formulae into interpretation, the details of which are of specialist interest only. The following outlines the general form of the travel time curves from these complexities.

Multilayered structures introduce additional segments into the travel time curves. Where **one** layer over a half space gives **two** P-wave first arrival straight line travel time segments (one for the direct, and one for the refracted arrivals), **two** layers over a half space produce **three** segments, **three** layers over a half space produce **four** segments, and so on. This assumes that the velocity of the layers increases with depth so that each extra layer introduces an extra refracted arrival. No refraction can be recorded from a layer whose velocity is less than any velocity above it. Consequently *low velocity layers* cannot be observed directly from refraction surveys and must be inferred by modelling. Where they introduce a clear break in the first arrival travel times, they can be detected and modelled fairly readily but their existence is often not obvious and, if masked, can confuse the interpretation. The problem of a *hidden layer* is not confined to low velocity layers since not every layer in a multilayered structure necessarily produces first arrivals or even identifiable later arrivals.

Velocity gradients can be approximated by a finely layered structure or they can be modelled by ray tracing. The rays are curved and the travel time curves are also curves. Strong velocity gradients can introduce triplications into the travel time curves.

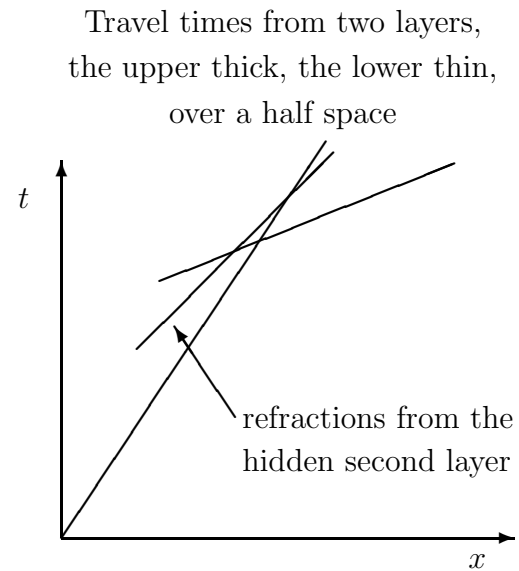
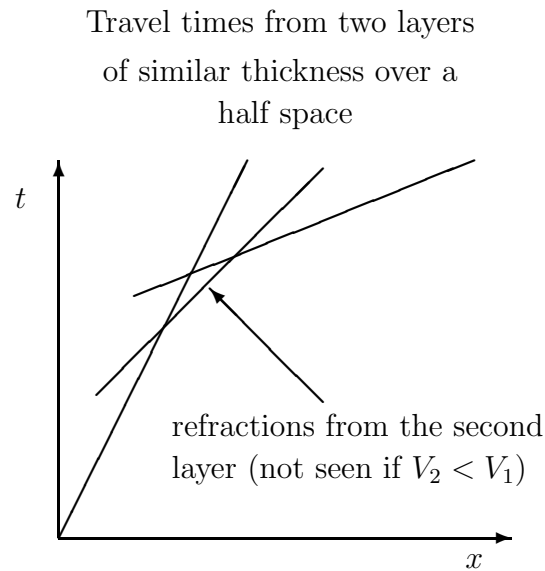
Irregular layers, such as the undulating weathered layer at the Earth's surface, introduce varying time delays into refraction travel times. Exploration geophysics texts (e.g. Parasnis 1997) describe a number of techniques for interpreting these types of data. Travel times from a series of layers of varying thickness are modelled by ray-tracing software.

REFERENCE

Parasnis, D.S., 1997. Principles of Applied Geophysics. Chapman & Hall.

REFRACTION TRAVEL TIMES FROM MULTILAYED STRUCTURES

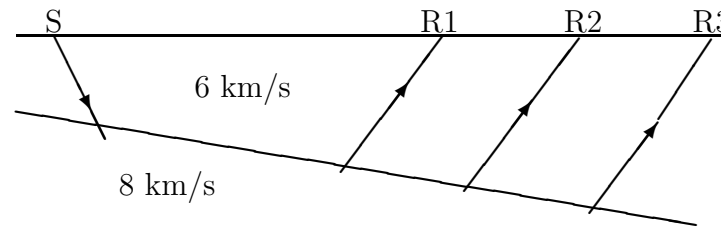
The figures below show schematic travel times for a two layers over a half-space. In the figure on the right the second layer is too thin for refractions through the second layer ever to be first arrivals. There would be no refraction at all from a low velocity layer; it would be directly detectable only if it generated measurable reflections.



INTERPRETATION OF REFRACTION PROFILES (2)

DIPPING LAYERS

These alter the apparent refractor velocity. The increasing delay with distance to the receiver when shooting down dip lowers the apparent down-dip refractor velocity. Similarly the decreasing delays shooting up dip increases the apparent up-dip velocity. For example, in the simple single layer case illustrated below, a dip of 1 in 8 (7.1°) makes the velocity estimated from the slope of the refraction times recorded down dip 7.26 km/s whereas that from the up-dip times is 9.06 km/s. In order to avoid velocity errors, refraction profiles should be shot in both directions, either as a split-spread or preferably as a reversed profile.

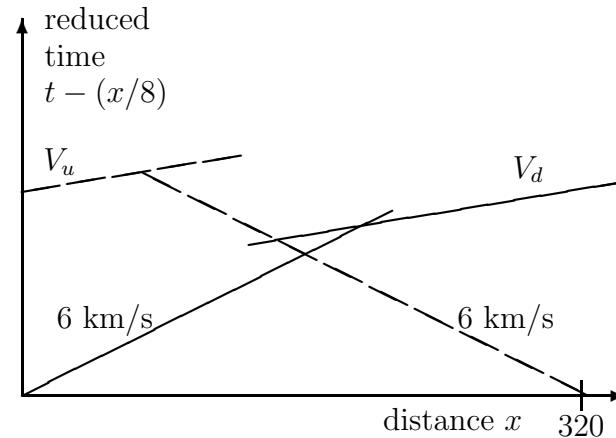


Shooting down dip,
the receiver delay
increases with distance,
i.e. $V_d < 8\text{ km/s}$.
Similarly $V_u > 8\text{ km/s}$.

INTERPRETATION OF REFRACTION PROFILES (3)

USE OF REDUCED TRAVEL TIMES

It is often convenient to plot refraction travel times or display recordings using reduced travel times relative to a convenient reference velocity V_r . That is, results or records are displayed using the reduced time $t - (x/V_r)$, where x is source-to-receiver distance, instead of the actual time t . This reduces the size of the plot and makes it easier to assess, for example, how well the travel times fit those computed from a model. The figure below illustrates the reduced travel time plot from a reversed profile shot over the model in the previous figure. The reference velocity V_r is 8 km/s.



Solid lines: shooting down dip

from a source at $x = 0$

Dashed lines: shooting up dip

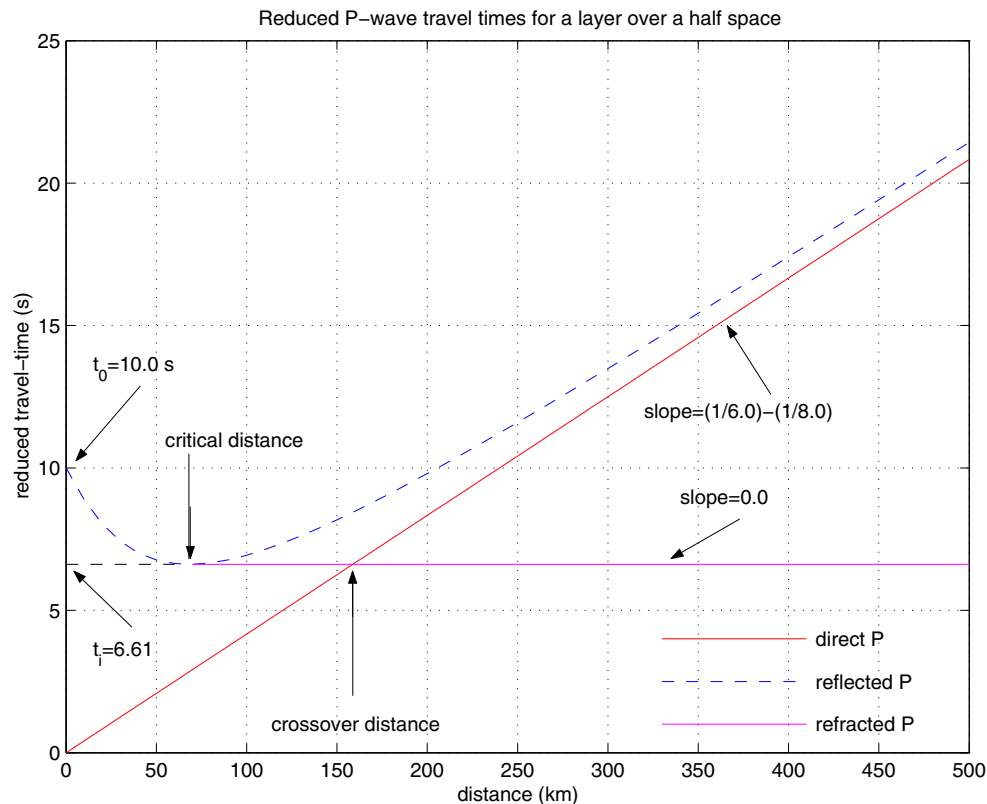
from a source at $x = 320 \text{ km}$

Note that the distance origin for the up-dip times is $x = 320 \text{ km}$ and that distance increases to the left.

REDUCED TRAVEL TIMES FOR A LAYER OVER A HALF SPACE

The figure below shows the reduced travel times for the model of uniform layer over a half space. The model is the same as the one used earlier to illustrate the P-wave travel-time curves from a layer over a half space. The reduction velocity applied to the travel times is 8 km/s.

It is important to realise that the slopes are no longer the reciprocals of the velocities but the velocities can still be found from the slopes and the reduction velocity. Can you see that it is easier to read t_i accurately from this plot than the travel time plot. As an exercise, try calculating the depth of the Moho from t_0 and t_i using the formulae in "Seismic Travel Time Curves" from Lecture 4.



CRUSTAL REFRACTION STUDIES (1)

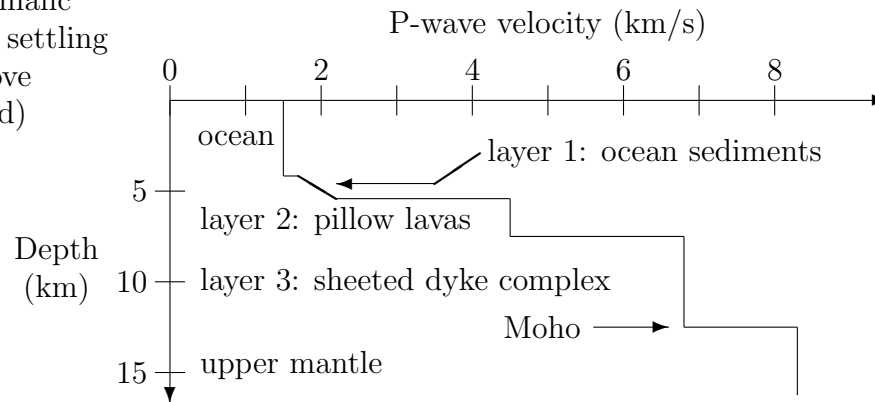
THE MOHOROVIČIĆ DISCONTINUITY

Crustal refraction surveys show a distinct jump in velocity between crust and mantle. The boundary between them is named after its discoverer, A.Mohorovičić, who found it from investigating the Croatian earthquake of 8 October 1909. The depth to the “Moho” varies from area to area, and particularly between continents and oceans, but it is found almost everywhere.

OCEANIC CRUST

This is only $\sim 5 - 11$ km thick and much more uniform in structure than continental crust. It has three distinct layers. Beneath the deep sea sediments of layer 1 come the basaltic pillow lavas of layer 2. The sheeted dyke complex of layer 3 is gabbroic. The upper mantle

comprises an ultramafic cumulate (crystals settling under gravity) above tectonite (deformed) ultramafics.



CRUSTAL REFRACTION STUDIES (2)

CONTINENTAL CRUST

The continental crust varies in thickness and velocity distribution. Each geological province tends to have its own characteristic thickness and velocity profile.

Thickness: commonly 30 to 45 km but it can be anywhere in the range 20 to 90 km.

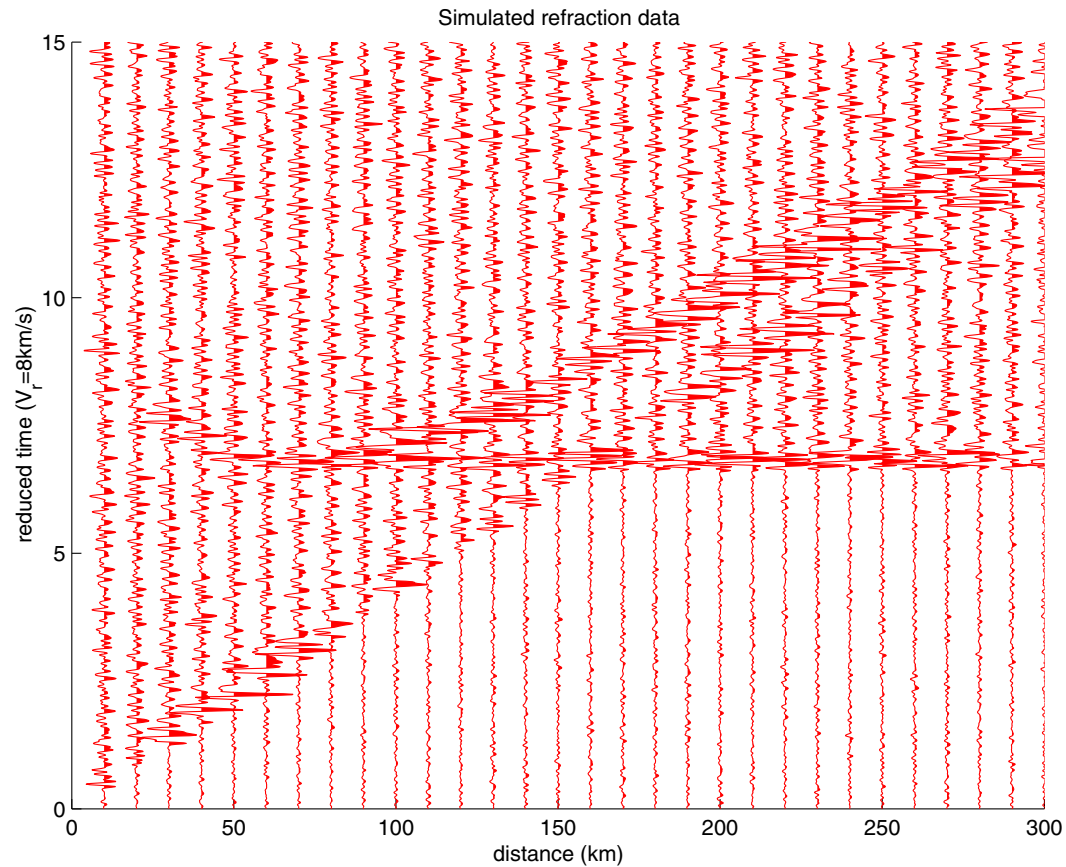
Velocities: Apart from velocities attributable to sedimentary rocks close to the surface, P-wave velocities in the upper crust are typically in the range 5.8 – 6.3 km/s. P-waves that penetrate deeper into the crust often exhibit higher velocities, in the 6.6 – 7.2 km/s range.

Layering: the evidence for distinct layering of the continental crust is often a matter of interpretation: the modelling of seismic refraction data is rarely unambiguous and generally requires some interpretational choices. For example, the scatter in the travel times and noise in the data makes it difficult to distinguish a gradual increase in velocity with depth from a series of step increases. Crustal layering may say more about the number of straight line segments chosen to approximate scattered crustal P-wave travel times than about genuine layering in the crust. Nonetheless crustal refraction studies often do show genuinely different characteristics in the upper and lower crust. There is also often evidence for a low-velocity layer in the upper crust.

- Conrad (1923) postulated a two-layer crust. When there appears to be a distinct boundary between the upper and lower crust, it is called the Conrad discontinuity. Refractions from beneath the Conrad discontinuity are frequently second arrivals and difficult to disentangle from other later arrivals such as scattered waves and reflections from the Moho.
- Lowrie (p157) attributes the low-velocity layer in the crust to a zone containing granite laccoliths. Beneath this he puts a mid-crustal layer of migmatites (mixed high-grade metamorphic-igneous, gneissose-granitic rocks) and then a high-velocity ‘tooth’ of amphibolites (metamorphosed basic rocks).

SIMULATED EXAMPLE OF CRUSTAL REFRACTION DATA

The figure below shows a simulated P-wave refraction section plotted in reduced time $t - (x/8)$ from the idealised crustal model presented earlier. Real refraction data is messier than this. For real examples refer to the papers in Blundell et al. 1992 (e.g. in class we shall look at Luosto & Korhonen, Crustal structure of the Baltic shield based on off-Fennolora refraction data).



WORKED EXAMPLE: CRUSTAL THICKNESS FROM A REFRACTION EXPERIMENT

The following table gives the shot-to-station distances and the travel-times of P-wave phases picked from recordings along a refraction profile running north from Lake Superior. The profile was part of a major collaborative project, called Project Early Rise. The project was designed to investigate variations in crustal and upper mantle structure of North America.

Station	Distance Δ (km)	P-wave travel-times (s)
1	163.2	23.8
2	206.0	30.2
3	231.4	35.3
4	260.9	41.0
5	295.0	43.8
6	335.9	49.1
7	374.0	53.7
8	446.2	62.6
9	483.4	67.1
10	524.6	72.1
11	564.1	77.2
12	605.1	82.2
13	639.5	86.4

The distances have been corrected for the Earth's ellipticity and converted to km.

REFERENCE:

R.F.Mereu and J.A.Hunter, 1969, Crustal and upper mantle structure under the Canadian Shield from Project Early Rise data, Bull.Seism.Soc.Am. 59, 147-165.

WORKED EXAMPLE: PROCEDURE

We shall work through this example using the following procedure

1. Plot travel-times against distance and, by drawing best-fit straight lines, identify the P_1 and P_n travel-times on it.
2. Compute reduced travel times using a reduction velocity of 8.0 km/s. (*A reduction velocity of 8 km/s is convenient since this is approximately the velocity of P-waves beneath the Moho*).
For example, the reduced P-wave time at station 1 is $23.8 - (163.2/8) = 3.4$ s.
3. Plot reduced travel-times against distance and find the slopes of the P_1 and P_n reduced travel times and the intercept time t_i of the P_n phase.
4. Calculate the velocities, v_1 and v_n , of the P_1 and P_n waves from the slopes.
5. Estimate the thickness of the crust H from the intercept time t_i of the P_n waves.

TRAVEL-TIME EQUATIONS:

We shall use the following equations, assuming a constant velocity crust over a flat Moho.

$t(P_1)$, the travel-time of the direct crustal P-wave P_1 is :

$$t(P_1) = \frac{\Delta}{v_1}$$

Assuming a constant thickness H for the crust, the travel-time of the head P-wave P_n refracted along the base of the crust, $t(P_n)$, is:

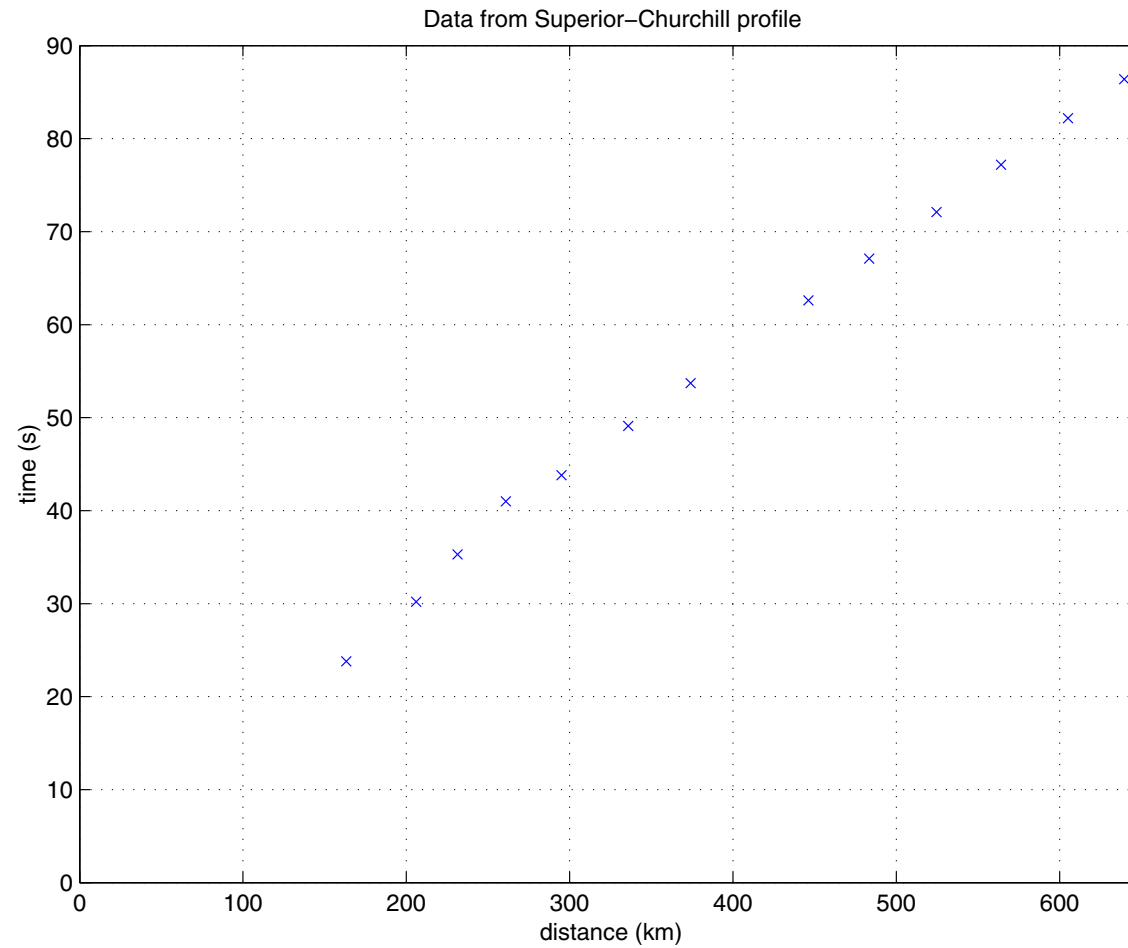
$$t(P_n) = t_i + \frac{\Delta}{v_n} \quad \text{where} \quad t_i = \frac{2H\sqrt{v_n^2 - v_1^2}}{v_1 v_n}$$

The cross-over distance, the distance at which the P_1 and P_n travel-time lines intersect is:

$$\Delta_d = 2H\sqrt{\frac{v_n + v_1}{v_n - v_1}}$$

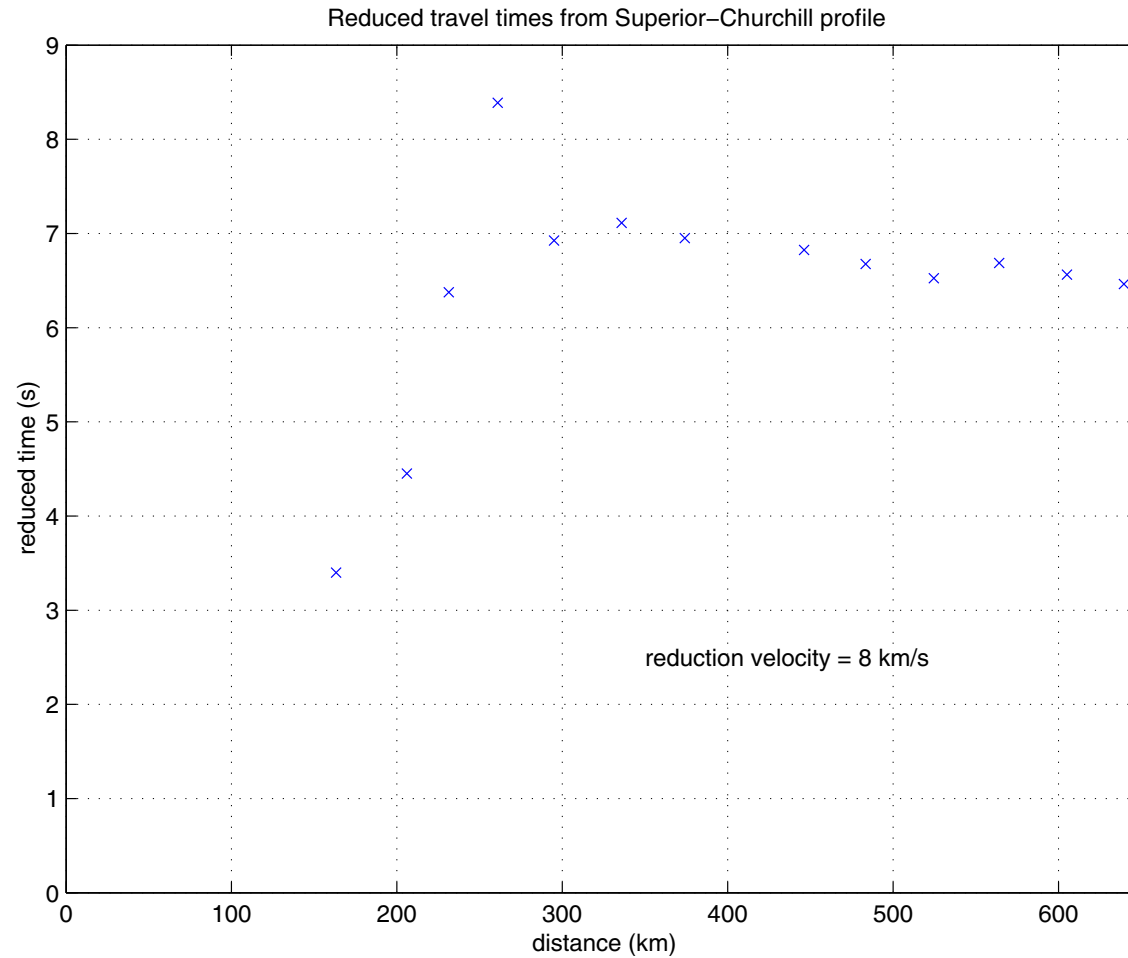
TRAVEL-TIME VERSUS DISTANCE FROM SUPERIOR-CHURCHILL PROFILE

Is the 41.0 s P-wave arrival at 260.9 km a P_1 phase or a P_n phase? It is not clear from this plot.

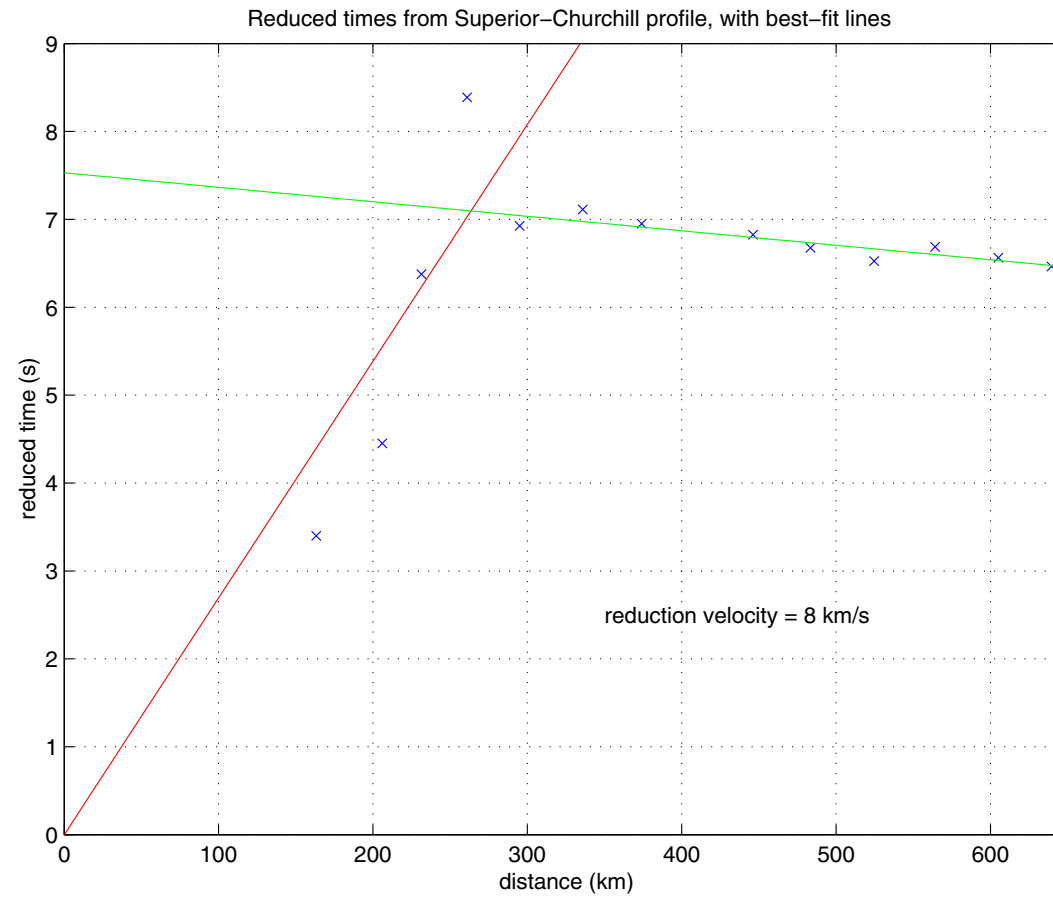


REDUCED TRAVEL-TIME PLOT FROM SUPERIOR-CHURCHILL PROFILE

On this plot it appears that the 41.0 s P-wave arrival at 260.9 km is more likely to be a P_1 phase than a P_n phase. The 41.0s was therefore taken to be a P_1 time.



REDUCED TRAVEL-TIME PLOT FROM SUPERIOR-CHURCHILL PROFILE
(with least-squares best-fit lines)



WORKED EXAMPLE: CALCULATIONS

The fit to the reduced times and distances from the four P_1 observations was made to pass through the origin ($t = 0$, $\Delta = 0$). A least-squares fit gave a slope of 0.02693. Because the reduced times are $t(P_1) - 0.125\Delta$, the P_1 travel times therefore fit

$$t(P_1) - 0.125\Delta = 0.02693\Delta, \quad \text{i.e.} \quad t(P_1) = 0.15193\Delta = \Delta/6.582$$

The least squares fit to the reduced times of P_n is

$$t(P_n) - 0.125\Delta = 7.5293 - 0.00165\Delta, \quad \text{i.e.} \quad t(P_n) = 7.5293 + 0.12335\Delta = 7.5293 + \Delta/8.107$$

The velocities are therefore $v_1 = 6.58$ km/s and $v_n = 8.11$ km/s.

Of course least-squares fitting the original times and distances gives the same velocities and intercept time as fitting the reduced times:

$$t(P_1) = 0.15193\Delta = \Delta/6.582 \quad \text{and} \quad t(P_n) = 7.5293 + 0.12335\Delta = 7.5293 + \Delta/8.107$$

The thickness of the crust is found from the intercept time t_i :

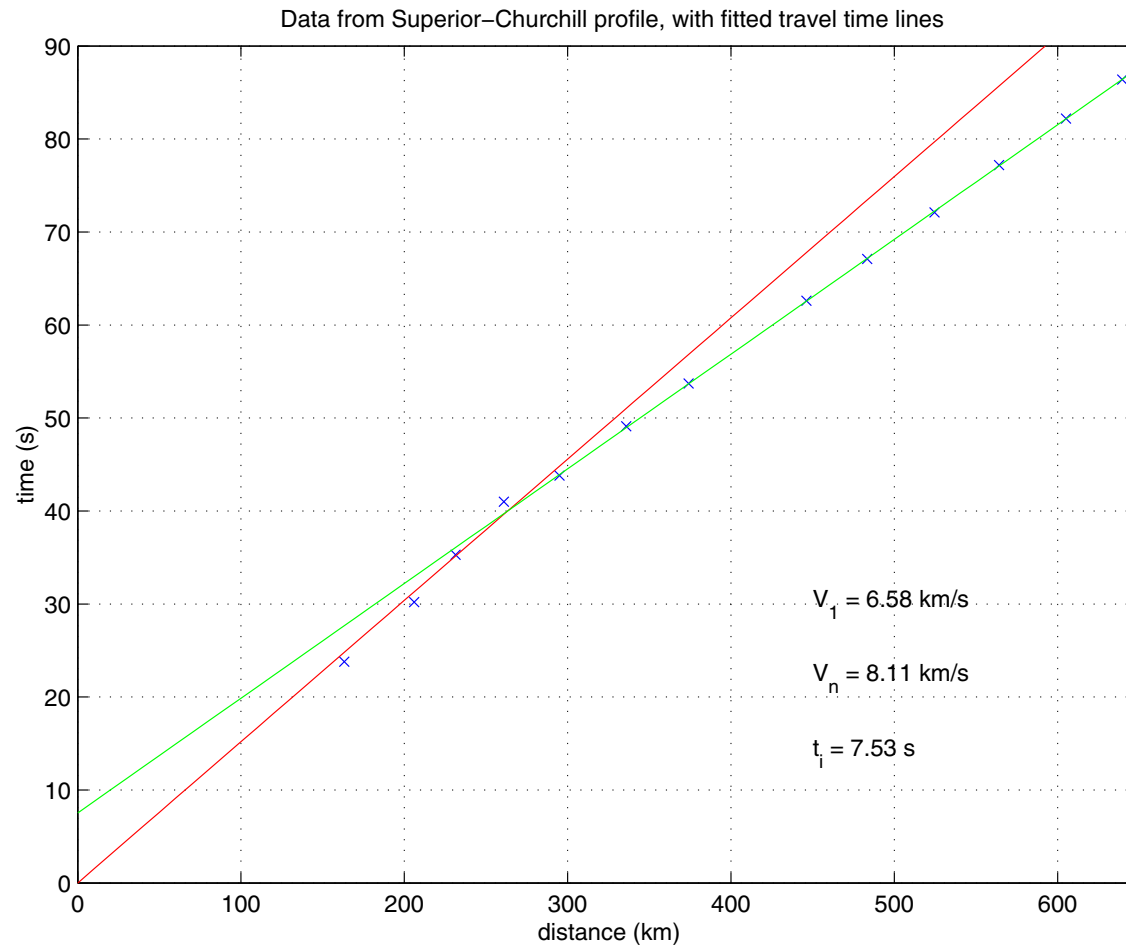
$$t_i = \frac{2H\sqrt{v_n^2 - v_1^2}}{v_1 v_n}$$

$$H = \frac{t_i v_1 v_n}{2\sqrt{v_n^2 - v_1^2}} = \frac{7.5293 \times 6.582 \times 8.107}{2\sqrt{8.107^2 - 6.582^2}} = 42.4 \text{ km}$$

Use of the cross-over distance (263.4 km) gives the same result.

$$H = \frac{\Delta_d}{2} \sqrt{\frac{v_n - v_1}{v_n + v_1}} = \frac{263.4}{2} \sqrt{\frac{8.107 - 6.582}{8.107 + 6.582}} = 42.4 \text{ km}$$

TRAVEL-TIME VERSUS DISTANCE FROM SUPERIOR-CHURCHILL PROFILE
(with least-squares best-fit lines)



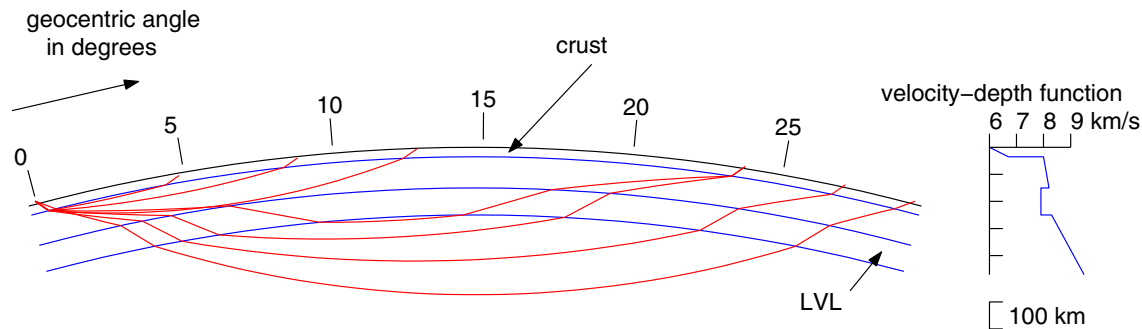
TRAVEL-TIMES THROUGH THE MANTLE

The curvature and ellipticity of the Earth and gradual variations of seismic velocity with depth have to be taken into account in modelling refraction travel times through the mantle to distances of the order of 500 km or more. These factors cause the raypaths to be curved; travel times are no longer linear with distance. At these larger distances also, it is no longer accurate to use distances calculated over the surface of the Earth. Instead distances are calculated in terms of the *geocentric angle* between the seismic source and recording station. The geocentric angle is often denoted by the Greek capital letter Δ (delta). It can be converted to an approximate distance in kilometers using 1 geocentric degree $\simeq 111.2$ km. Really the surface length of a geocentric degree varies with latitude and with the azimuth (bearing) from source to recording station. This variation is compensated using ellipticity corrections.

Although the theory is more complicated than that for refraction in a "flat Earth" model, the raypaths are still governed by Snell's law and many features of the travel time curves are similar to those met in a "flat Earth".

RAY PATHS THROUGH A MANTLE MODEL HAVING A LOW-VELOCITY LAYER

The sub-Moho refraction emerging at $\sim 12.5^\circ$ geocentric distance grazes the top of the low-velocity layer (LVL). The slightly deeper ray that dives into the LVL emerges at $\sim 24^\circ$ indicating how a LVL causes a break in the travel time curve.



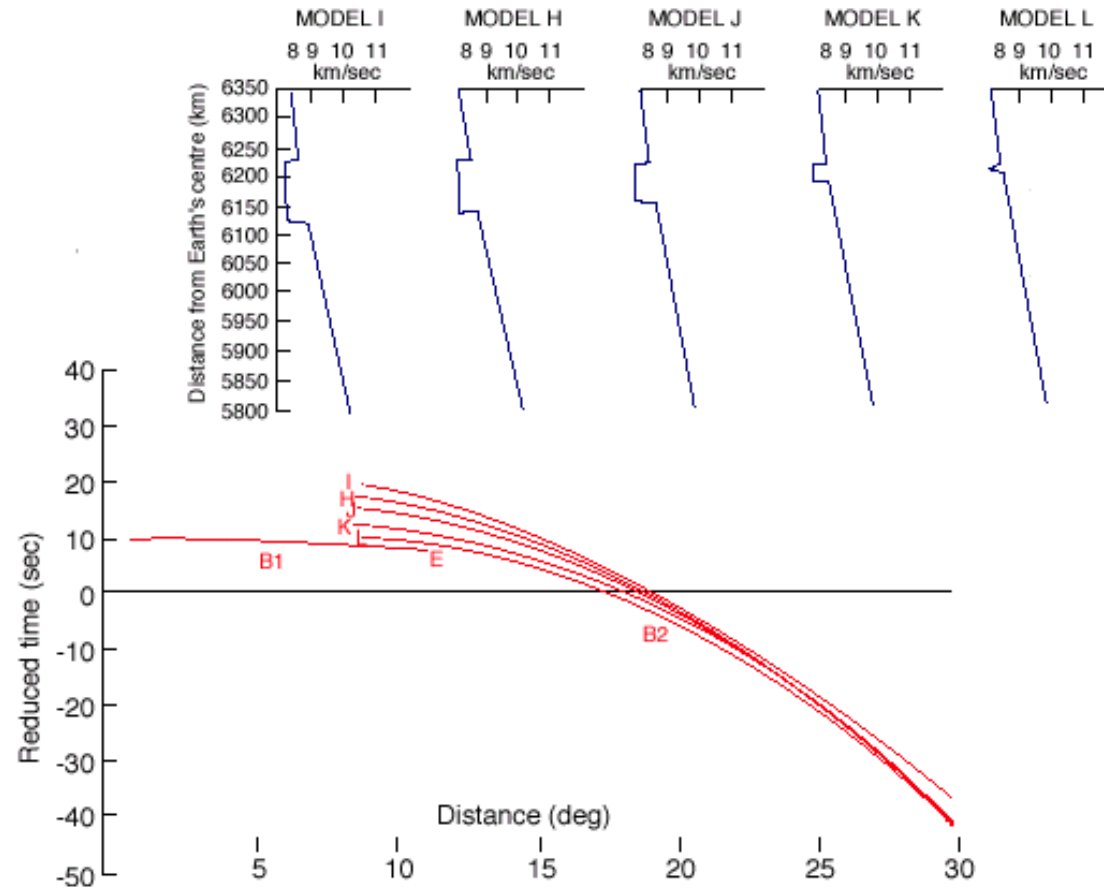
INTERPRETATION OF TRAVEL-TIMES THROUGH THE MANTLE

The upper mantle contains a low-velocity layer under the oceans and under younger parts of the continents. A low-velocity layer makes it difficult to determine the velocity-depth distribution at and below this layer directly. A typical procedure would be to:

1. construct a set of Earth models covering the likely range of velocity-depth curves;
2. compute the travel-time curve for each model by tracing rays through the model and calculating times along the ray paths;
3. compare the computed travel-time curves with the observed times and select the best-fit model;
4. adjust the model in such a way as to improve the fit if possible; if necessary this can be done by creating a new set of models around the best-fit model;
5. repeat steps 1 to 4 and iterate until the error of fit is less than the timing error in the observed travel-times.

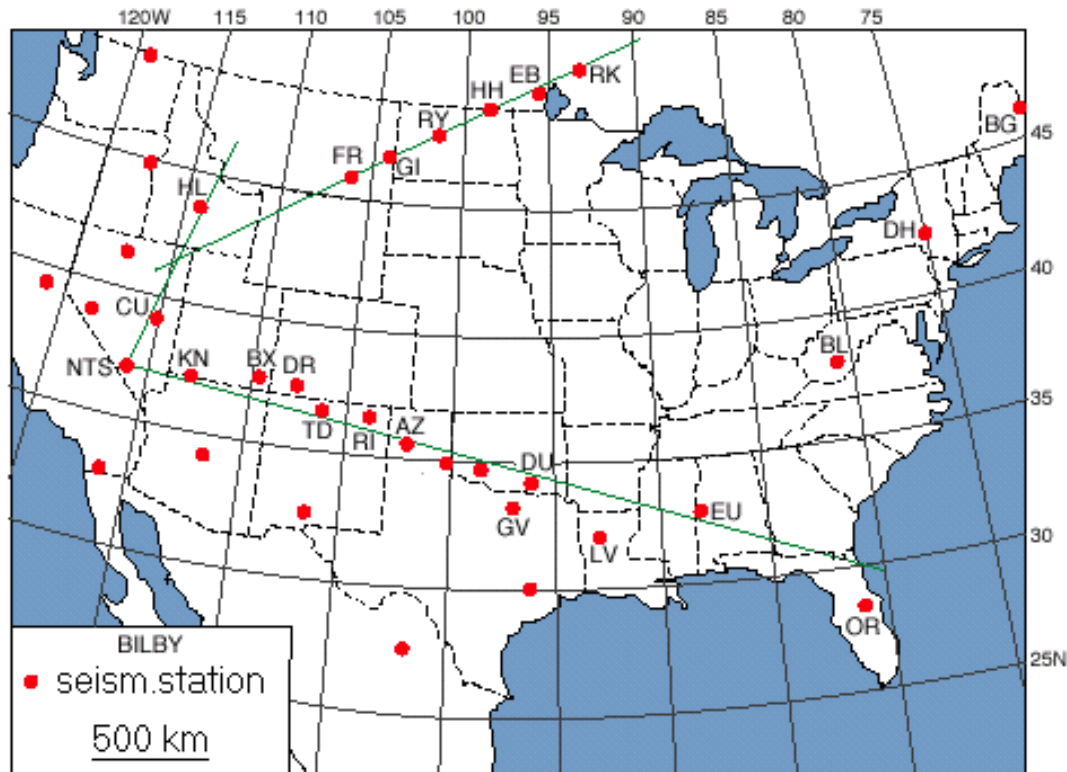
THE EFFECT OF A LOW-VELOCITY LAYER ON TRAVEL TIMES

The figure below illustrates the effect of a low-velocity layer on travel times through the upper mantle. The low-velocity layer causes a break in the travel time curve. The thicker the low-velocity layer, the larger the jump in travel times.



MAP OF A LARGE SCALE SEISMIC EXPERIMENT

The testing of nuclear explosions gave seismologists the opportunity to conduct large scale refraction experiments that determined the travel times of seismic waves that penetrated below the low-velocity layer in the upper mantle. One such experiment used the September 1963 BILBY explosions in Nevada. The map below shows the location of the seismic recording stations.



TRAVEL TIMES SHOWING A LOW-VELOCITY LAYER IN THE MANTLE

This example shows travel-times along the NE profile from the 1963 BILBY nuclear explosion in Nevada. The break between the two branches of the travel times at $\sim 14^\circ$ is indicative of a thin (~ 20 km) low-velocity layer at ~ 150 km depth.

