What Can Seismology Say About Hot Spots?

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Seismology offers the highest-resolution view of mantle structure. In the decades since *Morgan* [1971] first proposed deep-mantle plumes, seismologists have used increasingly sophisticated methods to look for evidence of such structures, but so far they have had little success. This abstract outlines the relevant seismological methods for non-specialists and summarizes the current state of knowledge about structure beneath hot spots, to set the stage for the seismological component of this conference.

Factors Affecting Seismic-Wave Speeds

Direct thermal effect – If thermal plumes exist in the mantle, they would have lower seismic wave speeds than their surroundings. In the upper mantle, a 100 K temperature rise lowers the compressional-wave speed, V_P , by about 1%, and the shear-wave speed, V_S , by about 1.7%. In the deep mantle, this effect is several times weaker. The temperature anomalies proposed for plumes are about 200 to 600 K.

Indirect thermal effect – Temperature variations also cause variation in the depths of polymorphic phase boundaries in the transition zone between the upper and lower mantle. These are places where pressure causes certain minerals to change their crystal structure, and these changes are accompanied by jumps in density and seismic wave speed [*Anderson*, 1967]. Two such zones in particular, at depths of about 410 and 650 km, are global features and fairly easily detectable. A 100 K temperature rise would depress the “410-km” discontinuity by about 8 km, and raise the “650-km” discontinuity by about 5 km. (Both of these numbers are based on the assumption that olivine is the main mantle mineral, and are subject to significant uncertainty.) Thus a high-temperature anomaly would produce negative wave-speed anomalies at 410 km and positive ones at 650 km. The depths to these phase changes can also be measured directly using waves reflected from them (see **Receiver Functions**, below).

Chemical effect – If a plume has a different composition from the surrounding mantle, this alone will cause a seismic wave-speed anomaly. The sign and magnitude of the anomaly will depend on what minerals are involved. As a rule of thumb more buoyant materials have lower wave speeds, but an exception to this rule is mantle residuum – peridotite from which partial melt has been removed. Residuum is buoyant, but has a higher wave speed than its parent fertile peridotite.

Melting – The presence of even a small amount of melt in a rock can have a large effect on the seismic-wave speeds. Partial melting may reflect either thermal (high temperature) or chemical (low melting point) effects. The magnitude of the effect on seismic wave speeds depends strongly on the geometric form of the melt bodies. Thin films on grain boundaries have the largest effect, and approximately spherical melt bodies have the smallest effect [*Goes et al.*, 2000].

Anisotropy – Seismic wave speeds and other properties of rocks vary with direction, and this can be as strong an effect as spatial heterogeneity. Olivine, in particular, becomes strongly anisotropic when flow causes crystal to align. Most studies of Earth structure ignore this effect, and their results probably are biased by this oversimplification. Studies dealing explicitly with anisotropy are becoming more common [*Montagner*, 2002].

Anelasticity – Many physical processes remove energy from seismic waves and convert it to heat, causing the waves eventually to die away. Most of these processes are thermally activated, so hotter regions are expected to exhibit stronger attenuation (high Q^{-1}). A side effect of anelasticity is to introduce a weak frequency dependence of the wave speeds, which must be accounted for in studies of Earth structure.

Seismic Tomography

The travel time of a seismic wave through the Earth gives an average of the wave speed along the ray path (but see **Bananas & Doughnuts**, below). If travel times are available for enough ray paths, passing through all parts of a region in many different directions, it is possible to un-scramble the times to determine the three-dimensional wave-speed distribution. The term tomography, borrowed from medicine, is given to such seismic techniques. Seismic tomography is much more difficult than X-ray tomography, because the ray paths are curved and initially unknown, and in some cases the locations of the sources are poorly known. Three seismic tomography techniques are particularly useful in studying mantle structure:

Teleseismic Tomography – In order to study the structure immediately under an area, one can deploy an array of seismometers and record waves from distant earthquakes (>~ 2,500 km away). Such waves arrive at angles within about 30° of the vertical, so crossing rays sample the structure down to depths comparable to the array aperture. The ray directions are not isotropically distributed, however; no ray paths are ever close to horizontal. Consequently, compact structures tend to be smeared vertically in images obtained by this technique [*Keller et al.*, 2000]. This smearing is unfortunate, because it generates artifacts that can be mistaken for real structures. It is possible to estimate quantitatively the severity of the smearing, however, and if due attention is paid to this error source, teleseismic tomography is the best technique available for studying the upper few hundred kilometers of particular regions.

Figure 1 shows results of one of the most detailed teleseismic tomography studies, of the structure beneath Iceland [*Foulger et al.*, 2001; *Foulger et al.*, 2000]. A strong low-wave-speed anomaly in the upper mantle, 200 to 250 km in diameter, extends to the deepest well-resolved depths, about 400 km. Significantly, though, the shape of the anomaly changes strongly and systematically below about 250 km, becoming tabular and parallel to the mid-Atlantic Ridge. This shape strongly suggests that the anomaly is related to plate-tectonic processes, a conclusion supported by whole-mantle tomography, which shows no continuation of the wave-speed anomaly beneath the transition zone [*Ritsema et al.*, 1999].

Whole-Mantle Tomography – There have now been thousands of seismometers deployed globally for decades, and millions of travel-time observations have accumulated and been used to derive three-dimensional models of the whole mantle. Some studies use enormous data sets obtained from seismological bulletins such as that of the International Seismological Centre, but these data are subject to large and systematic observational errors. Others use data measured in more objective and consistent ways, usually using digitally recorded seismograms. Most whole-mantle models agree about the largest-scale anomalies (thousands of kilometers in size), but for a long time this was not so. The model that currently has the best resolution at depths of a few hundred kilometers is described by *Ritsema et al.* [1999]. This model shows, among other things, a strong low-wave-speed anomaly in the upper mantle beneath Iceland, which extends down to the transition zone but not to greater depths, confirming inferences drawn from teleseismic tomography of the region.

The resolution of whole-mantle tomography models is limited both by the ray distributions and by the state of computer technology. The smallest anomalies currently resolvable are 500 km or more in size. Furthermore, ray paths fall far short of sampling the Earth uniformly. Both earthquakes and seismometers are distributed irregularly over the Earth, and some places within the Earth are sampled poorly or not at all *e.g.*, the southern hemisphere, and particularly the south Pacific and Indian Oceans. The uneven ray distribution also systematically distorts anomalies in the Earth. As with teleseismic tomography, this distortion can be assessed quantitatively, but not by the general reader unless considerable information on this subject is given in the paper in question.

Surface-wave Tomography – Tomographic methods can also be applied to surface waves, low-frequency seismic waves that propagate in the crust and upper mantle and owe their existence to the presence of the free surface. The depth to which surface waves are sensitive depends on frequency, with low-frequency waves "feeling" to greater depths and therefore propagating at higher speeds. It is a rule of thumb that surface waves "feel" down to about a quarter of their wavelength. They also propagate at about 4 km/s, so this depth, in kilometers, is about 1/frequency (Hz).

Because of the distribution of earthquakes and seismometers, surface waves can often sample regions of the crust and upper mantle that body waves do not. They are also expected to be highly sensitive to plume heads, which are predicted to flatten out in the upper mantle, producing low wave speed regions that extend for thousands of km [*Anderson et al.*, 1992]. Body-wave and surface-wave data are often combined in whole-mantle tomography studies, such as that of *Ritsema et al.* [1999].

Bananas & Doughnuts

The statement above, that travel times are averages along ray paths, is a simplification. In reality, seismic waves "feel" the structure in a finite volume, and in fact *Dahlen et al.* [2000] have recently shown that travel times are most sensitive near a hollow surface around the ray, whose shape reminds them of certain snack foods. Figure 2 shows examples of the spatial distribution of sensitivity according to the new theory. Incorporation of frequency-dependent kernels into tomographic practice will significantly improve the quality of three-dimensional Earth models. The first results of such studies [*Montelli et al.*, 2003] show vertical low-wave-speed anomalies in the south Pacific, notably beneath Hawaii, Samoa-Tahiti, and Easter Island, but these are also regions of sparse data or clumps of data surrounded by regions lacking data (Figure 3), a circumstance that would tend produce spurious structural features in tomographic images. The reality of these plume-like anomalies is a critical question at present.

Multiple *ScS*

Because they have limited resolution and can distort anomalies in complicated ways, tomographic results often are difficult to interpret. It would be much better if seismic waves sampled precisely a region of interest, and nothing else. Happily, nature occasionally arranges an experiment for us in just this way. For example, the seismic phase *ScS*, a shear wave reflected from the core-mantle boundary (CMB), when observed close to the epicenter of an earthquake, has a nearly vertical ray path through the entire mantle [*Anderson and Kovach*, 1964]. Such waves are ideally suited to looking for narrow vertical structures such as plumes.

On April 26,1973, a magnitude 6.2 earthquake occurred in Hawaii, and the records from seismometers on Oahu show an usually clear train of multiple-*ScS* phases, reflected repeatedly between the Earth's surface and the CMB [*Best et al.*, 1975]. These waves are sensitive to structure in a vertical cylinder with a diameter of about 500 to 1000 km extending down to the CMB. They show no indication of a plume. The wave speed V_S in the upper and middle mantle inferred from arrival times is higher than the average for the southwestern Pacific [*Katzman et al.*, 1998], and the propagation efficiency (Q) is also high [*Sipkin and Jordan*, 1979; *Sipkin and Jordan*, 1980]. Figure 2 shows the sensitivity kernel for these *ScS* waves at different depths in the mantle. The location of a possible plume in the lower mantle might be far enough from Hawaii that these *ScS* waves would not sample it, but these observations argue strongly against a large region of unusually high temperature or extensive melting in the upper mantle beneath Hawaii. In particular, an upper-mantle anomaly similar to the one beneath Iceland appears to be ruled out.

Receiver Functions

When a compressional or shear seismic wave strikes a discontinuity in the Earth, it generates reflected and transmitted waves of both types. Because of this, waves from distant earthquakes passing through a layered medium such as the crust or upper mantle generate complicated seismograms containing many echoes. To interpret these records, seismologists process them to generate simplified artificial waveforms, somewhat inscrutably called *receiver functions*. These can be inverted to yield the variation of V_S with depth, and they are particularly sensitive to strong wave-speed discontinuities. Receiver functions are particularly powerful for studying the depths to the Moho and the "410-km" and "650-km" discontinuities, which may provide evidence about crustal thickness and temperature at these depths [*Du et al.*, 2002].

One of the most detailed receiver-function studies done to date took the form of a profile across the eastern Snake River Plain, the suggested track of a mantle plume now beneath Yellowstone, which lies at the northeastern end of the Plain [*Dueker and Sheehan*, 1997]. The results illustrate the complexity of structures revealed by receiver functions, and some of the difficulties of interpreting them. Several "discontinuities" are present, in addition to the major ones near 410 and 650 km. Even these two major features are not continuous, and the 410-km discontinuity appears to split in two near the left end of the profile. The depths to the 410- and 650-km discontinuities are expected to be negatively correlated if their topography results from temperature variations, but actually they are weakly positively correlated. The receiver functions thus provide no evidence of elevated temperatures

“Plume waves” (“Fiber waves”)

Zones of low wave-speed trap energy and act as waveguides, along which waves propagate efficiently for great distances. This is the principle behind fiber-optic communication. The narrow low-wave-speed anomalies expected for plumes would be ideal for supporting such waves (J. R. Evans, personal communication, 2002), but no such plume waves have ever been noticed. This failure might indicate merely that nothing excites plume waves efficiently (there are no earthquakes in the lower mantle), or it might mean that plume waveguides do not exist. Quantitative theoretical investigation of the excitation and propagation of plume waves would be highly worthwhile.

Summary

The main methods for studying Earth structure in a way that is useful in the search for plumes include seismic tomography, studying the transit times and attenuation of individual waves that penetrate the volume of interest, and the use of receiver functions to study topography on the boundaries of the transition zone. A potentially new and interesting approach is the search for “plume waves”. Whereas downgoing slabs in subduction zones and their effects on the transition zone have been easy to detect, the same cannot be said about plumes, heads or tails, and promising images often have not proved reproducible by later, more detailed studies. It will be interesting to follow what the next decade brings.

Figure captions

Figure 1: Compressional-wave speed (V_P) anomaly in the upper 500 km beneath Iceland, imaged by *Foulger et al.* [2001] using teleseismic tomography techniques. Inside the green surface, V_P exceeds the average value at a given depth by 0.5%. At shallow depths the anomaly is approximately cylindrical, but beneath 250 km it changes shape, and becomes tabular, with its long dimension parallel to the mid-Atlantic Ridge.

Figure 2: Frechet kernels giving the sensitivity of the travel time of the seismic phase *ScS*, a shear wave that reflects from the core-mantle boundary, to changes in the shear-wave speed at different depths, computed using the theory of *Dahlen et al.* [2000] for waves with dominant frequency 0.035 Hz. The geometry is appropriate for waves generated by the Hawaii earthquake of April 26, 1973 (star) and recorded at station KIP (triangle), as reported by *Best et al.* [1975] and *Sipkin and Jordan* [1980]. The maps have been corrected for convergence of the verticals, which otherwise exaggerates the size of deeper features; thus the distance scale, not than the geographic coordinates, correspond to reality in the deep mantle. The two shallowest maps have attenuated color scales.

Figure 3: Sampling of the mantle by the travel time data set of *Bolton and Masters* [2001], which was used, in addition to high-frequency data from the ISC Bulletin, in the recent tomographic study of Montelli [2003]. Squares show surface projections of turning points of rays in the indicated depth ranges. *P* phases are shown on the left, and *S* phases on the right. In the central and south Pacific, coverage is sparse and observations occur in clumps that correspond closely to the plume-like anomalies reported by Montelli [2003].

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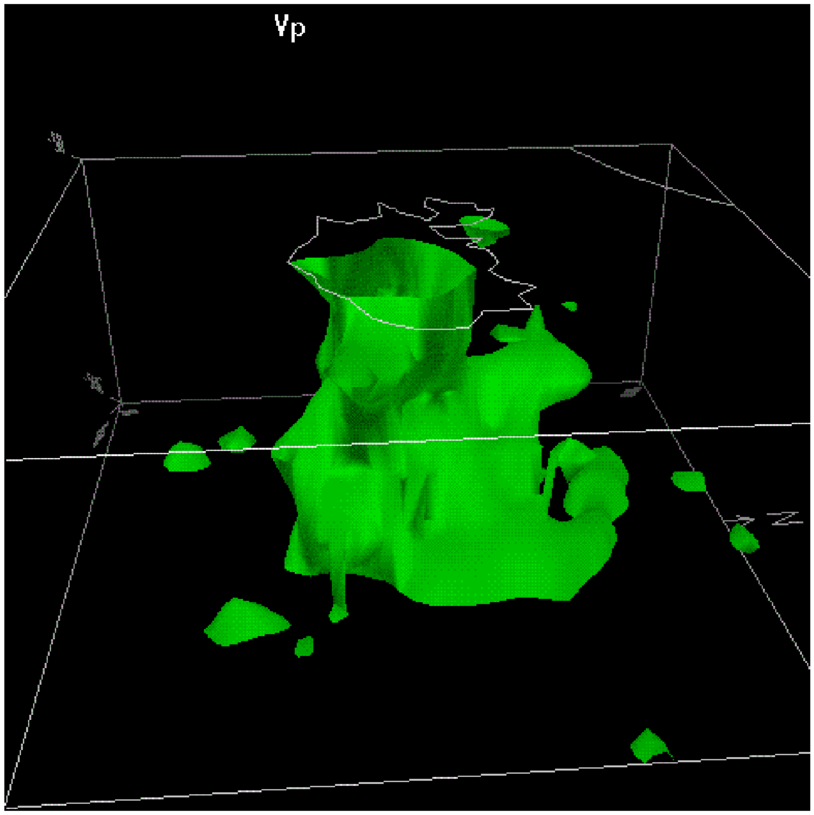
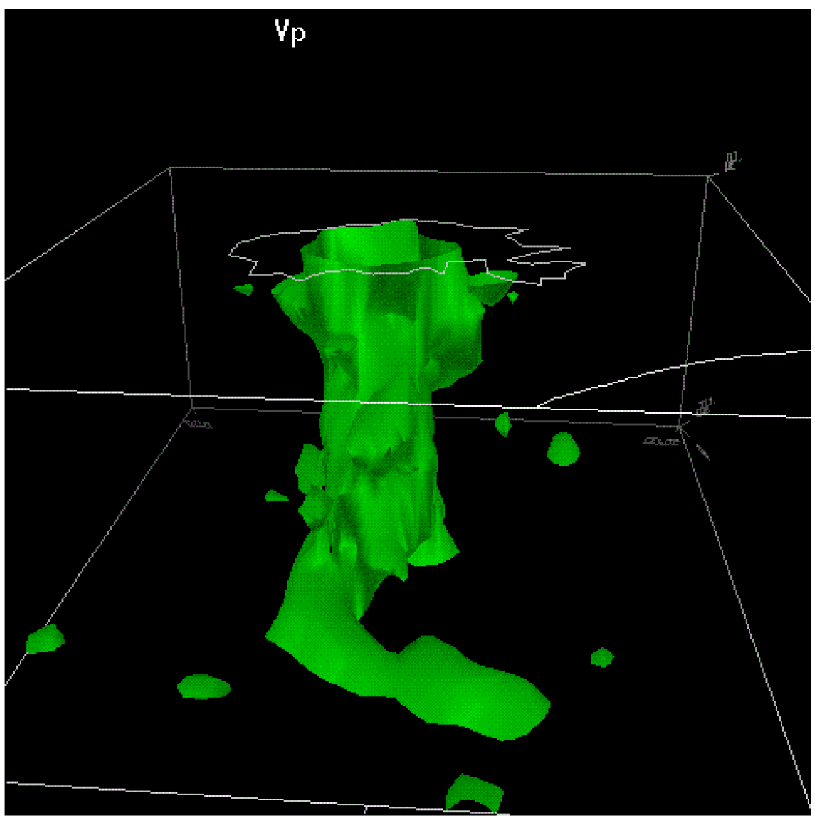
a)



Looking west

b)



Looking south

Figure 1: Views of Iceland upper-mantle anomaly

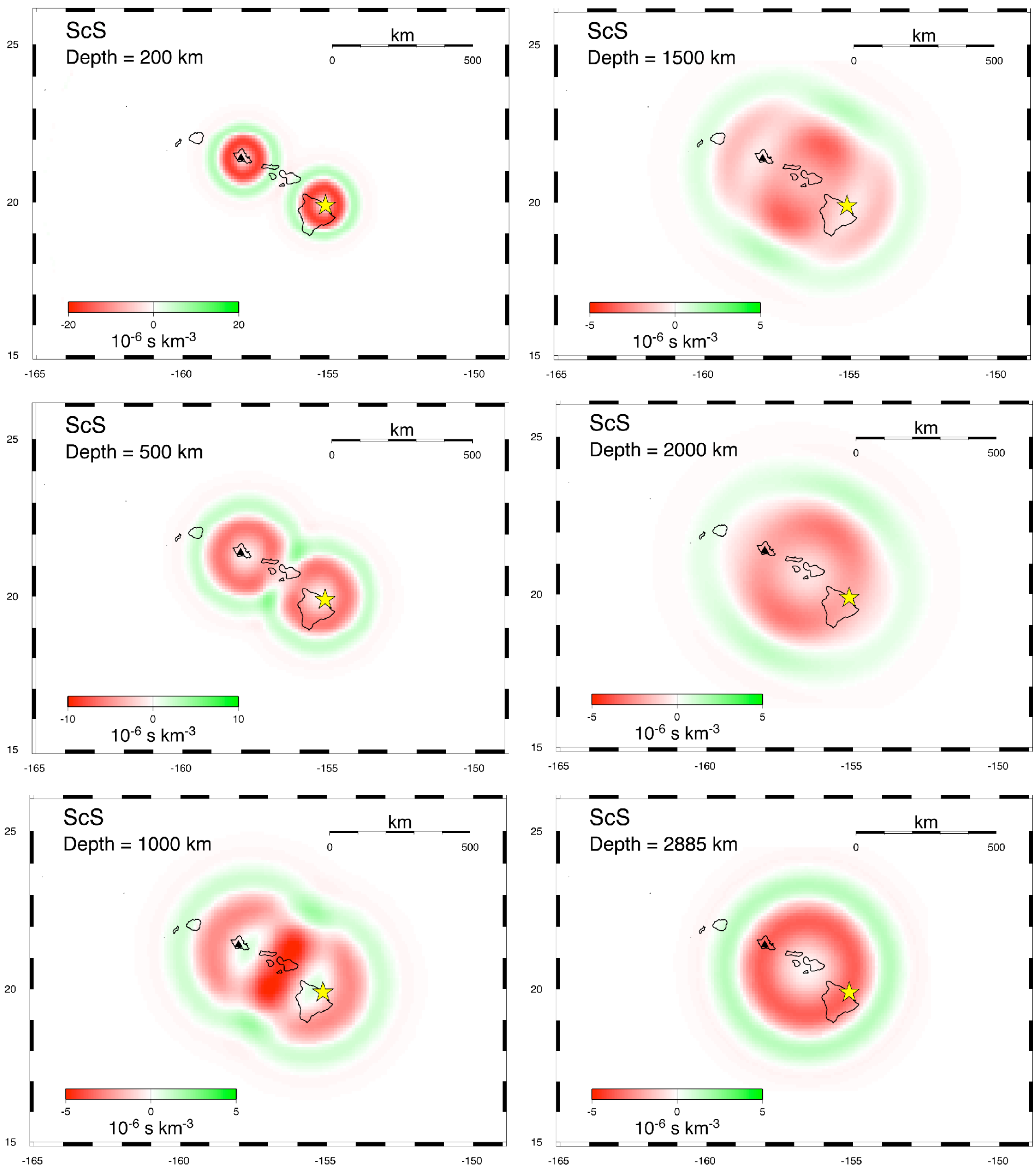


Figure 2: ScS travel-time data kernels

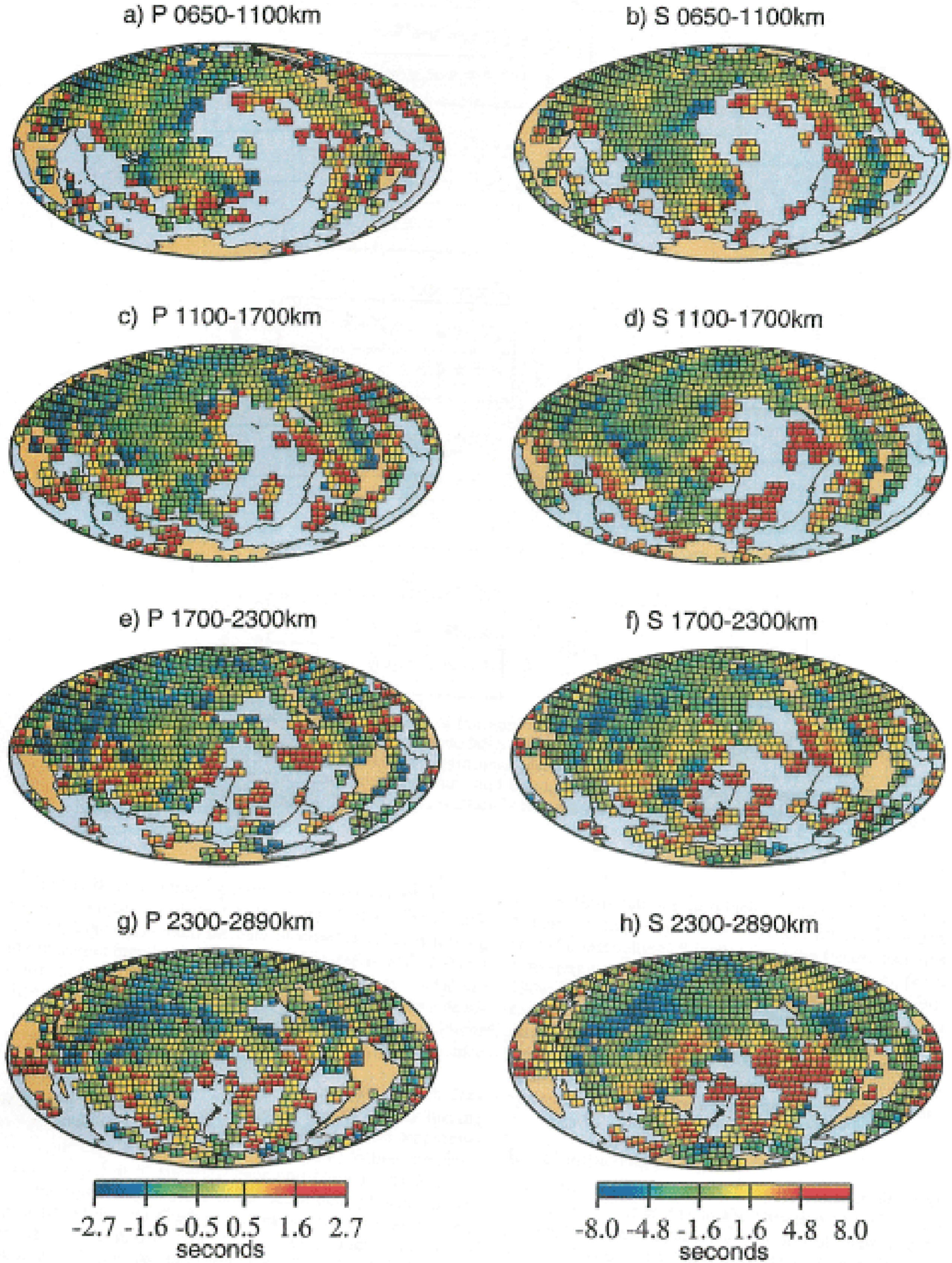


Figure 3: Geographic sampling of Bolton-Masters dataset