The deep structure of continents

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The phase velocity of Rayleigh waves has been used to study the deep structure of continents for more than fifty years. In the last ten years the widespread installation of broad band digital seismometers has generated more than 10^6 seismograms that can be used for surface wave tomography, and so obtain maps of the thickness of the continental lithosphere. Such maps show that the tectonic architecture of continents is controlled by variations in lithospheric thickness, and that many regions of flat-lying phanerozoic sediment are underlain by thick lithosphere. Perhaps the most surprising discovery is that the thickest lithosphere that has yet been mapped lies beneath Tibet.

The large-scale evolution of oceanic plate boundaries has been understood for more than forty years (Fowler 2005). Ridges are the result of plate separation, and add sea floor to the two plates on either side at approximately the same rate. Therefore ridges axes move with respect to both plates at half the spreading rate. Many trenches move rather slowly, but some, especially in the western Pacific, are moving rapidly in the direction of the subducting plate, and by so doing produce back arc spreading and marginal seas. These simple rules control the evolution of existing oceanic plate boundaries. New ridges form when a plate breaks and the two pieces separate. New trenches are much more difficult to produce, and little is still known about the processes involved.

Our understanding of continental tectonics is quite poor compared with that of the oceans. Continental deformation is commonly distributed over wide zones whose internal deformation is not simply related to that of the rigid regions on either side. Small continental fragments often move parallel to the boundaries of the deforming zones, and also undergo rapid rotation about vertical axes. Excellent maps of the kinematics of continental deformation are now available, generated from earthquake mechanisms and velocity vectors from the Global Positioning System. The problem is that the observed deformation is presumably controlled by the mechanical properties of the entire continental lithosphere but, in contrast to oceanic lithosphere, little is known about its deep structure. Small nodules of continental mantle are sometimes brought to the surface by alkali basalts and other alkaline rocks from depths as great as 200 km. They can be dated by measuring their Os isotopic composition, and often give ages similar to those of the last major metamorphic event that affected the crust on top. Like the crust, the mantle part of continental lithosphere is likely to have been formed by the coalescence of lithospheric fragments of widely differing age and history. Though this background has been widely accepted for perhaps 30 years, it is little help in understanding why continents behave as they do. What we need for this purpose are maps of the thickness of the continental lithosphere, which we can then compare with the geometry of present and past deformation. We cannot make such maps from mantle nodules alone, because they are too rare. So there seemed no way of attacking this important problem.

As so often happens in the Earth Sciences, the whole situation has been changed by a technological advance, in this case in seismology. In the 1960s, seismometers recorded ground motion on

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photographic film. The great advance that played a critical part in the development of plate tectonics was the installation of a standard world wide network of seismometers, the WWSSN, which for the first time allowed reliable fault plane solutions for large \((M > 5.5)\) earthquakes to be obtained, and from these the slip vectors between plates. But reading the film was slow and tedious. The obvious next step was to record the ground motions digitally. But doing so requires huge amounts of storage. Recording the three components of ground motion at one site every second for a year requires about 1/2 a gigabyte. The storage and transfer of digital seismograms has only become possible because of the huge advances that have occurred in networks and storage.

Seismologists would like to be able to use every wiggle on every seismogram to study the internal structure of the Earth. Doing so is obviously much easier if the seismogram is in digital form than if it must be digitized from a photographic record. There are now more than 1000 digital seismometers operating continuously whose records are available on the internet. This enormous digital data base now allows seismologists to carry out projects that they could only dream of doing 20 years ago.

Figure 1 shows a seismogram from South Africa. Early seismologists constructed velocity models of the Earth by measuring the arrival times of P- and S-waves, but could not make use of other features of the seismogram. A striking feature of seismograms that record vertical ground motion is the train of large amplitude Rayleigh waves that arrive after the S-waves. The velocity of such surface waves is not constant, but depends in a complicated way on the velocity at depth along the path that these waves follow from the earthquake to the seismometer. Over the last 50 years a great deal of effort has gone into studying this problem. The forward problem involves calculating the surface wave velocity from a three dimensional model of the Earth. However, what we really would like is to be able to use the surface wave velocities, which we can measure from seismograms, to calculate the three dimensional seismic velocity structure within the Earth. This is the inverse problem of surface wave tomography. Several groups (Ritsema and van Heijst 2000; Debayle and Kennett 2000; Priestley and Debayle 2003) have worked on this problem, and can now produce three dimensional velocity models of most parts of the Earth’s surface to depths of about 350 km, with a horizontal resolution of between 300 and 800 km and a vertical resolution of between 30 and 50 km. Such studies require powerful computers and large amounts of storage. The velocity structure of north America discussed below used about 20,000 seismograms, and required a few terabytes of disc storage.

At present, such studies only use waves with periods between 50 and 160s, but the shorter period waves contain a wealth of information about the thickness and velocity structure of the crust, and the groups involved are working to extend the techniques to include periods of 20–30s. It is particularly important to use the velocities of the higher modes of Rayleigh waves (figure 1) which arrive at the seismometer after the S-waves but before the fundamental Rayleigh waves. The fundamental Rayleigh wave mode has its greatest amplitude (figure 1a) near the Earth’s surface and its amplitude decays exponentially with depth (figures 1b, 1c). The vertical resolution of the S-wave velocity model obtained from inversion using only fundamental modes therefore becomes less and less good as the depth increases. But the higher modes behave quite differently. For instance, the first higher mode in figure 1(b) is most sensitive to the velocity at a depth of about 140 km, where its amplitude is greatest. The vertical resolution of the velocity models can therefore be increased by requiring them to agree with the velocities of the modes shown in figures 1(b) and 1(c). The maps shown were obtained from the phase velocities of the fundamental and first four higher mode Rayleigh waves.

Velocity models obtained from surface wave tomography look quite different from those from P- and S-wave tomography because they have good vertical resolution but poor lateral resolution. Structures like high velocity subduction zones are often almost invisible in models from surface waves but dominate models from P- and S-wave tomography. This difference has resulted in considerable controversy (Priestley and McKenzie 2002), because the velocity models from body waves look so different from those from surface waves. However, the structure of the lithosphere is better resolved by surface waves than by P- and S-wave tomography, especially if higher modes are included.

The velocity of surface waves is principally controlled by the S-wave velocity of the mantle, and depends little on either the density or the P-wave velocity. Furthermore, the S-wave velocity of the mantle at surface wave periods is now known to be principally controlled by temperature, and to be largely independent of composition. This behaviour only became clear when the velocity at long periods was measured in the laboratory, by Jackson and his colleagues (Jackson 2000; Jackson et al 2002) and by Gribb and Cooper (1998, 2000). At the ultrasonic frequencies generally used to measure the S-wave velocities the temperature dependence is small, and only becomes large at seismic frequencies. This important discovery shows that the regions of low S-wave velocity in the
Figure 1. (a) A seismogram from a vertical instrument in South Africa written by an earthquake in Turkey, with focal depth given as 33 km. The great circle distance between the source and receiver is 7325 km ($\Delta = 66^\circ$) and the path is almost entirely continental. The fundamental mode Rayleigh waves show dispersion, with the longer period waves travelling faster, because the S-wave velocity increases with depth. The higher modes arrive between the S-wave and the fundamental mode Rayleigh waves, and exist only because the S-wave velocity varies with depth. The amplitude of the vertical displacement of various Rayleigh wave modes at periods of 20 s, (b), and 30 s, (c), scaled to make their maximum amplitude unity, as functions of depth. The fundamental mode is labelled 0, and the $n$th higher mode $n$. The amplitude of the fundamental mode is greatest near the surface, and its velocity is therefore controlled by the shallow velocity structure. The amplitudes of the higher modes have more complicated depth dependence, and therefore modelling their velocities can resolve the details of the S-wave depth dependence.

The obvious next step is to use the S-wave velocity to calculate the temperature. Unfortunately, such a calculation is not straightforward, because the relationship between the two depends on the grain size of the material. Though we don’t yet fully understand why the S-wave velocity at seismic frequencies depends so strongly on temperature, we believe it is because the stresses between the grains have time to relax, by diffusion of material along the grain boundaries. At ultrasonic frequencies there is no time for such diffusion to occur, so the S-wave velocity is little changed from its low temperature value. But at periods of 50 s or more we believe that diffusion becomes important at high temperatures. However, we don’t yet have...
Figure 2. (a) The thermal structure of the oceanic lithosphere that agrees with the observed variation of depth and heat flow with age (McKenzie et al. 2005). (b) The average S-wave velocity structure of the Pacific Ocean, plotted as a function of age. (c) The observed variation of S-wave velocity as a function of temperature (solid lines), obtained by combining (a) and (b). The dashed lines are calculated from an empirical expression that fits the observations (Priestley and McKenzie 2006).

A theory that is good enough to predict the change in velocity as a function of grain size. Furthermore, we don’t know what the grain size is in the mantle, though we believe it is much larger than those of the materials used in the laboratory. Different groups have taken two approaches to overcoming this problem. One approach (Goes et al. 2000; Faul and Jackson 2005) is to parameterize the laboratory experiments and use the resulting expressions to extrapolate the behaviour to an estimated mantle grain size. An alternative (Priestley and McKenzie 2006) is to use the geophysical information in figure 2 to plot the shear wave velocity as a function of temperature (figure 2c), and then find an expression that fits the resulting curves. Neither approach is really satisfactory: what we need is a proper theory, of which as yet there is no sign. Fortunately the two approaches give similar results (McKenzie and Priestley 2008; Shapiro et al. 2004). Furthermore, we can test their accuracy...
by using the compositions of minerals in nodules brought to the surface by alkali basalts and kimberlites. These compositions provide estimates of the temperature and depth at which the nodules equilibrated (figure 3a, b). These in turn can be used to determine the geotherm. Figures 3(c) and 3(d) show the temperatures at intervals of 25 km determined from the S-wave velocities at the two kimberlites locations. At temperatures above about 1000°C the agreement between the temperature estimates from the nodule mineralogy and the S-wave velocity is excellent. At lower temperatures the agreement is less good, and that from the S-wave velocity is everywhere less than that from the nodules. This behaviour suggests that temperature estimates of 1000°C and greater from the S-wave velocities will provide more reliable constraints on the geotherm than the lower temperature values, and therefore that the lithospheric thickness should be estimated from these high temperature values alone.

Figure 4 shows maps of the lithospheric thickness from North America, together with some estimates from nodule mineralogy and the locations of diamond-bearing kimberlite pipes. Most diamonds in such pipes come from the lower part of the lithosphere, and are only present if this region is in the diamond stability field. Figure 3
shows that the lithosphere needs to be thicker than about 160 km before this condition is satisfied. Because of the great effort that has been devoted to finding diamonds, the locations of many diamond-bearing kimberlites are known from which no nodules have been analysed. Figure 4(b) shows that there is excellent agreement between the locations of such diamond-bearing kimberlites and regions of thick lithosphere, and that the thickness estimates from seismology agree well with those from the nodule mineralogy. The Canadian Shield is underlain by thick lithosphere, as is most of the central part of the US. Seismology cannot determine the age of rocks or their composition. Priestley and McKenzie (2006) therefore preferred to call contiguous regions of thick lithosphere ‘cores’ rather than ‘cratons’ when they are mapped using surface wave tomography. One striking feature of figures 4(a) and 4(b) is the location of the fold mountain belts. Both the Rockies and the Appalachians wrap round the margins of the North American Core. In SW Canada and Georgia thrusts carry the folded rocks hundreds of kilometres from the margins towards the interior of the Core. Clearly, the architecture of the North American fold and thrust belts has been controlled by the geometry of the Core, presumably because its cold, thick lithosphere strongly resists deformation.

A map of Africa, figure 5(a), shows similar features. The estimates of lithospheric thickness from

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**Figure 4.** (a) The major structural features of North America (Holmes 1965). (b) Map of the lithospheric thickness beneath North America from surface wave tomography, contoured at intervals of 20 km. The numbers in white boxes show thicknesses estimated from nodule mineralogy. Red circles show the locations of diamond-bearing kimberlites (Janse and Sheahan 1995), yellow circles show those with no diamonds. The yellow continuous line shows the boundary of the Canadian Craton.

**Figure 5.** (a) As for figure 4(b). The yellow lines mark the craton boundaries: a West African Craton, b Angolan Craton, c Tanzanian Craton, d Kalahari Craton. (b) The craton boundaries are: a East European Craton, b Siberian Craton, c North China Craton, d South China Craton, e Bastar, Singhbhum and Aravalli Cratons and f Dharwar Craton.
seismology agree well with those from nodules, and the diamond-bearing kimberlites are largely confined to regions where the lithospheric thickness is 160 km or more. Such kimberlites are less common in regions of very thick lithosphere, presumably because there is less melt generation beneath such regions. Like North America, the thick lithosphere extends beyond the craton boundaries.

A particularly interesting region is northern Tanzania, where there is a rapid variation of lithospheric thickness. It changes from less than 100 km in southern Kenya to more than 300 km in Tanzania over a distance of about 200 km. The slope of the base of the lithosphere must therefore be at least 45°.

A more surprising map is that of Eurasia in figure 5(b). As expected, the East European and Siberian cratons are underlain by thick lithosphere. But so is the West Siberian Basin, where there has been extensive subsidence since the Jurassic to form its oil reservoirs. The Urals extend N–S along 60°E and mark a step in the lithospheric thickness. Since the lateral resolution of the surface waves is about 400 km, or less than the width of the Urals, it is not yet clear whether the thick lithosphere is continuous beneath them. But the two unexpected features of figure 5(b) are the thin lithosphere beneath the Chinese Cratons and the thick lithosphere beneath Tibet. The thinning of the Chinese lithosphere occurred in late Jurassic and Early Cretaceous, and the diamond-bearing kimberlites were emplaced during the Ordovician. Though the mechanism by which the thinning occurred is still controversial, there is no conflict between the seismological and petrological estimates of lithospheric thickness. The lithosphere beneath the Indian Peninsula is also surprisingly thin. The diamond-bearing kimberlites there were emplaced in the Precambrian, when the lithosphere may have been thicker.

The most unexpected feature of figure 5(b) is the thickness of the lithosphere beneath Tibet and Iran. Before it became possible to map its thickness with surface waves, the general view was that the lithosphere was thinner beneath Tibet than elsewhere, because it was thought that thick lithosphere was convectively unstable, and that the lower part would therefore detach and sink into the mantle. This has not happened. The map suggests that thick lithosphere can be produced by the collision between continents, and that this process may have been responsible for generating the thick lithosphere found beneath many older parts of continents. Clearly, we need a better understanding of the evolution of the temperature structure and the transport of radioactivity when large scale continental shortening occurs. Figure 5(c) shows that the western part of Australia is underlain by thick lithosphere, which extends beyond the boundaries of the cratons.

Figure 5. (c) The craton boundaries are: a Darwin Craton, b Kimberley Craton, c Pilbara Block, d Yilgarn Block, e Gawler Craton.
Iran (McKenzie and Priestley 2008). We are also modelling the composition of alkaline volcanics to discover that of their sources. Because the S-wave velocity is essentially unaffected by the removal of large melt fractions, the only obvious method of finding the composition of the mantle part of the lithosphere is by using that of the melts produced from it, and from the composition of the nodules transported to the surface. The melts we have examined so far from regions of thick lithosphere have all come from mantle that has been depleted by the removal of large amounts of melt. The density of such material is reduced by melt removal, which is presumably why the thick lithosphere is not removed by convective instabilities.

As has so often happened in the past, a technological advance, in this case in seismology, has thrown a new light on problems that have concerned earth scientists for many years. Like most such advances, it is of limited use by itself. Much more can be discovered about structure and evolution of the lithosphere by combining the new seismological information with existing ideas from materials science, mineral physics, petrology and mainstream geophysics.

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