

RESEARCH NOTE

Regional phases in continental and oceanic environments

B. L. N. Kennett¹ and T. Furumura²

¹Research School of Earth Sciences, Australian National University, Canberra ACT 0200, Australia. E-mail: brian@rses.anu.edu.au

²Earthquake Research Institute, University of Tokyo, 1-1-1 Yayoi, Bunkyo-ku, Tokyo 113-0032, Japan

Accepted 2001 March 14. Received 2001 March 14; in original form 2000 November 27

SUMMARY

The very different crustal thicknesses between continental and oceanic crust lead to the establishment of distinct styles of regional seismic wavefield. Observations and numerical simulations of events near subduction zones in Japan allow a direct comparison of seismic wave propagation in oceanic and continental structures. The thick zone of lower seismic velocities in the continental crust acts to trap *S*-wave energy in particular and generates crustally guided phases such as *Pg*, *Lg*. In the oceanic environment there is no efficient *S*-wave trapping and crustal energy progressively leaks into the mantle. Reverberations in the crust and the water above help to sustain the amplitude of the mantle arrivals *Pn*, *Sn* to large distances.

Key words: continental crust, oceanic crust, regional seismic phases, seismic wave propagation.

1 INTRODUCTION

There are major differences in the nature of the regional seismic wavefield in continental and oceanic structures. For continental regions both crustally guided phases (*Pg*, *Lg*) and mantle phases (*Pn*, *Sn*) appear in the records, and in shield structures with low attenuation relatively high-frequency waves can travel substantial distances. Commonly the *Lg* phase with a group velocity close to 3.5 km s^{-1} is the dominant feature of the wave train (see e.g. Campillo 1990). In contrast, for pure oceanic paths, *Lg* is not found (Knopoff *et al.* 1979) and the mantle phases (*Pn*, *Sn*) can be very prominent and travel long distances (Walker 1977, 1981; Sereno & Orcutt 1985, 1987). *Lg* is extinguished after propagation through more than 100 km of oceanic structure when emerging from a continent (Press & Ewing 1952).

Zhang & Lay (1995) have provided a comparison of regional *S* propagation in continental and oceanic structures using both modal representations and finite difference synthetic seismograms. The relatively thick lower-velocity waveguide formed by the continental crust supports a large number of higher-mode surface waves at frequencies around 1 Hz. The *Lg* phase then corresponds to the superposition of group velocity extrema for many modes within a narrow band of group velocities. The presence of sediments modifies the detail of the dispersion behaviour and mode counts, but does not alter the constructive interference of the modes. An alternative representation for the *Lg* phase is as the sum of multiple post-critical reflections

within the crust (see e.g. Kennett 1985; Campillo 1990). As the distance from the source increases, successive multiple *S* reflections become most significant close to the critical angle of reflection. The envelope of the traveltimes of the multiple *S* reflections gives an onset for *Lg* with a group velocity of 3.5 km s^{-1} .

For the basaltic oceanic crust with a thickness of about 7 km, the influence of sediments is stronger. At a given frequency, far fewer modes are supported by the crustal structure than for continents, and there is not the consistent pattern of group velocity extrema needed to create the *Lg* phase (Knopoff *et al.* 1979; Zhang & Lay 1995). From a body wave viewpoint, the multiple reflections within the crust are very closely spaced. For the phase velocities that form continental *Lg*, the reflections in the oceanic structure are pre-critical at the crust–mantle boundary. Thus there is transmission of *S* waves into the mantle at each bounce, and this transmitted energy serves to reinforce the *Sn* wave train.

There has been considerable discussion of the role of the continent/ocean transition in the suppression of *Lg* in oceanic structures (e.g. Ewing *et al.* 1957; Kennett 1986; Regan & Harkrider 1989; Zhang & Lay 1995). The mismatch of the crustal thickness is similar to joining electromagnetic waveguides with different impedances, and in consequence reflection of *Lg* may occur. The gradients associated with the transition from continent to ocean can lead to energy previously trapped in the crust now being transmitted into the mantle (see the ray simulations of Kennett 1986). However, the main reason for the

reduction of Lg amplitude with distance propagated through oceanic crust is the failure to set up a set of crustally trapped S waves (even for SH).

In regions adjacent to subduction zones, such as Japan, shallow sources in the subduction zone, which lies off-shore from the island chain, can give rise to clear regional arrivals including Lg at land stations. The wavefield on the continental side is established despite the structural complexity of the region.

2 OBSERVATIONS FROM A COMMON SOURCE

Even in regions next to subduction zones it is not easy to find situations where it is possible to follow seismic wave propagation in both continental and oceanic structures from a single source. However, the high station density in Japan provides some unusual opportunities.

Fig. 1 shows the locations of the sources and stations selected for the record sections of regional phases displayed in Figs 2 and 3. The vertical-component velocity data come from

the JARRAY operated by a consortium of Japanese Universities, which includes ocean bottom hydrophones off the southern coast of Japan as well as island stations in the Izu-Bonin arc.

The first event lies in the subduction zone off Kyushu, which provides clear oceanic paths, although there are no close stations on the oceanic side. The record sections of regional phases are displayed in Fig. 2. The propagation to stations on land in Kyushu and southern Honshu (profile A) is efficient and shows clear crustally guided waves Pg , Lg , with less prominent mantle arrivals Pn , Sn , because the distance span is not large. In contrast, the long oceanic paths to the stations on profile B show very clear Pn , rather patchy Sn and no sign of Lg at the expected group velocity of 3.5 km s^{-1} .

The second event lies in the subduction zone off-shore from Hokkaido. The record sections in Fig. 3 show clear crustal phases propagating across Hokkaido even though they pass through a volcanic belt (profile C). The amplitude of Lg is quite large for stations near 200 km from the source. The paths to coastal stations in eastern Honshu have a large oceanic component and show rather poor crustal phases, and Lg is particularly weak (profile D).

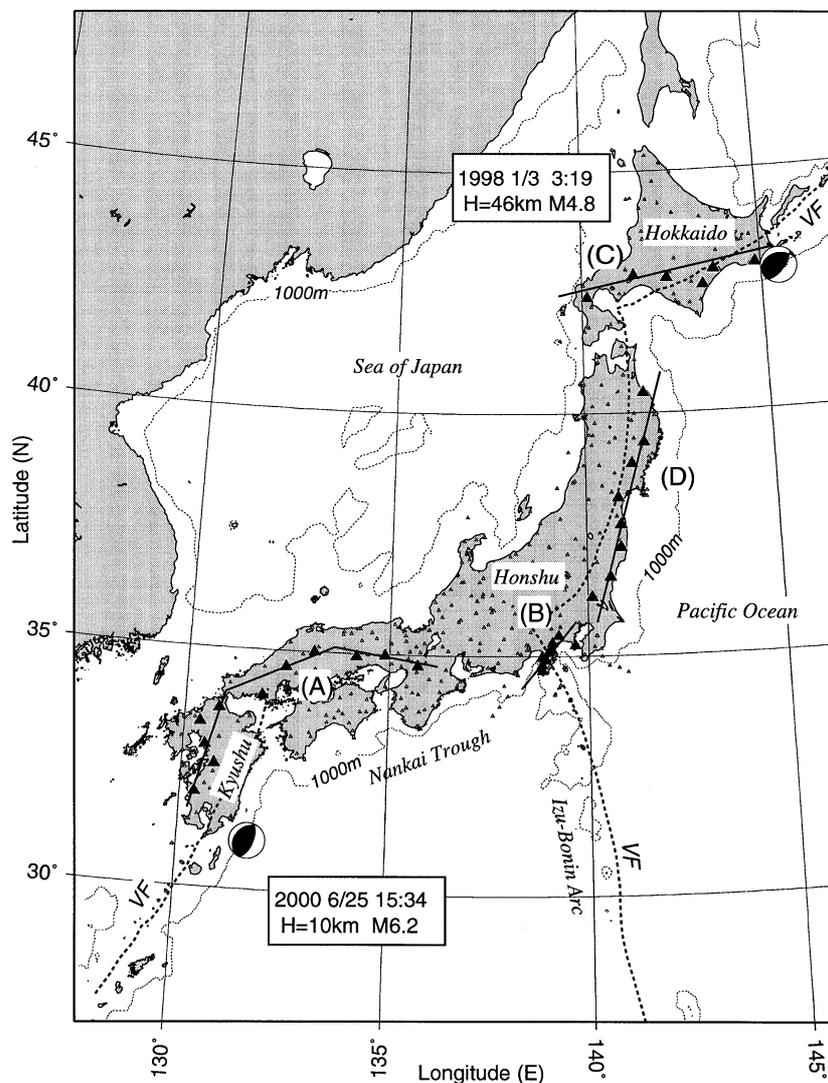


Figure 1. Locations of the two events and the stations in the JARRAY shown as small triangles. The larger triangles denote the stations displayed in Figs 2 and 3. VF indicates the volcanic front associated with the subduction zones.

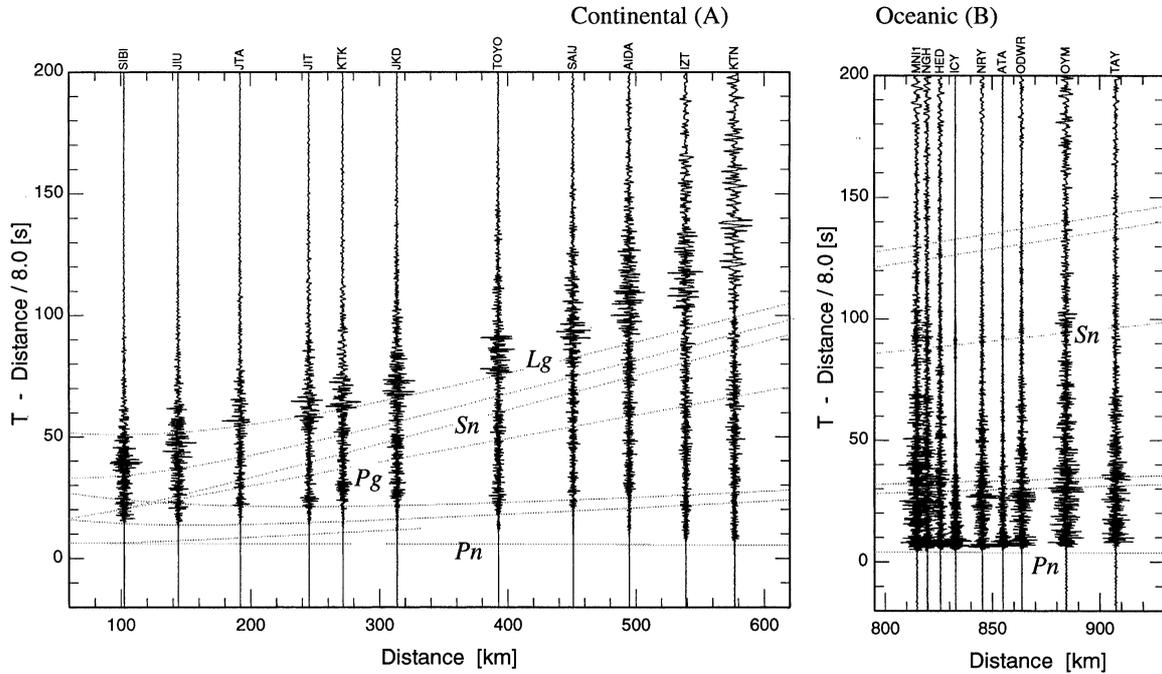


Figure 2. Regional wave propagation from an event in the subduction zone off Kyushu to both land stations (A) and oceanic stations (B). Traces are normalized to a common amplitude.

The mantle phases for the oceanic paths to profiles B and D show noticeable differences. Across the younger Philippine plate the paths from Kyushu to profile B lie nearly parallel to the trench in the Nankai Trough and *Pn* is most prominent with only a hint of the presence of *Sn*. The paths for the deeper event in Hokkaido to profile D lie along the margin of the much older Pacific plate and are oblique to structural trends, with *Sn* much stronger than *Pn*. It is difficult to judge whether these differences lie in the structure through variations in the sub-Moho gradients for the two groups of oceanic paths or in the interaction of the sources with subduction zone structure.

3 NUMERICAL SIMULATION OF SOURCES IN THE CRUSTAL TRANSITION ZONE

A good illustration of the differences between continental and oceanic paths is provided by numerical simulation of the wavefield for sources close to the continent–ocean transition. We employ the pseudo-spectral method for *P–SV* waves with a staggered grid representation for displacements and tractions (Furumura *et al.* 1998). The model (Fig. 4) has both oceanic and continental components, including a water layer with depths

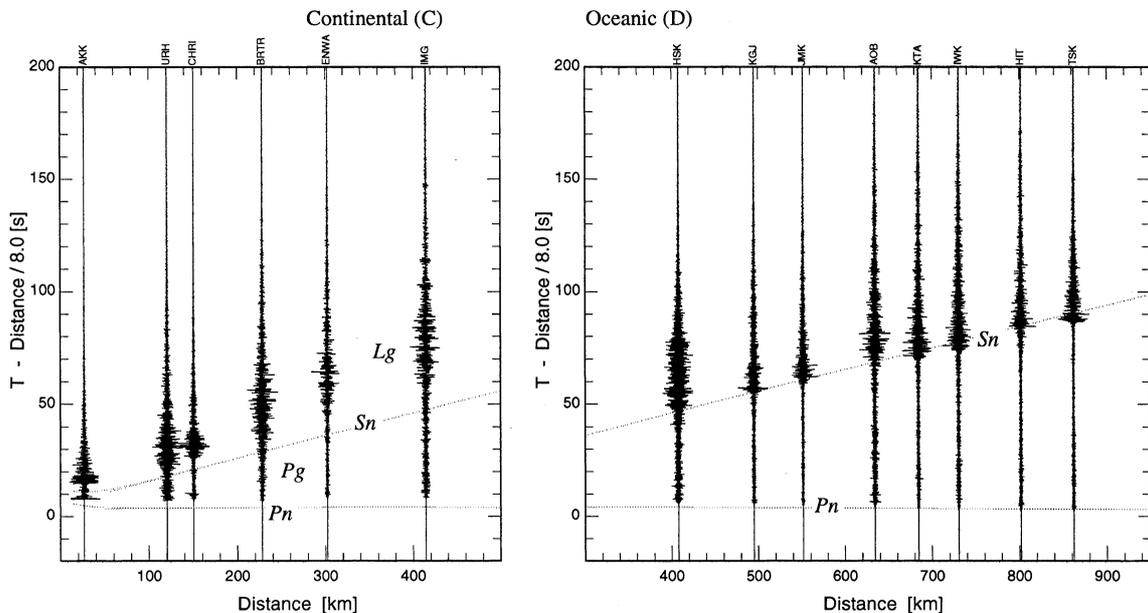


Figure 3. Regional wave propagation from an event in the subduction zone off Hokkaido to stations in Hokkaido on continental paths (C) and to stations on the east coast of Honshu on dominantly oceanic paths (D). Traces are normalized to a common amplitude.

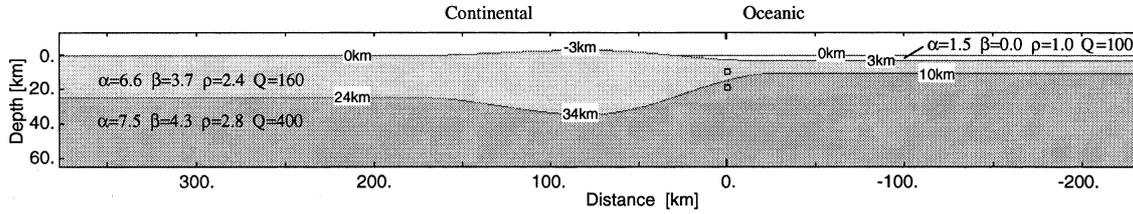


Figure 4. Model configuration for numerical simulation; the P velocity (α) and S velocity (β) are specified in km s^{-1} , and the density ρ in Mg m^{-3} .

up to 3 km. The continental side has a region of thickened crust close to the continent–ocean margin and then grades into a homogeneous section. The thickness of the crust on the continental side has been chosen to be quite thin (24 km with a thickening to 37 km) but still sufficient to display the strong differences from the oceanic regime. The model can be regarded as a simplified representation of crustal thickness variation adjacent to a subduction zone. We have deliberately taken a simple model without the complications of the subduction zone itself so that the structure of the regional wavefield can be clearly seen. We have taken a simple homogeneous mantle across the full span of the model to emphasize the features associated with the variations in crustal structure. As a result we have no reinforcement of Pn , Sn phases from mantle structure and the mantle phases in the numerical simulations are less prominent than in the observations shown in Figs 2 and 3. For a 45° dip-slip source at 10 km depth in the crustal transition zone the wavefield rapidly establishes a very different character on the continental and oceanic sides, as can be clearly seen in the snapshots displayed in Fig. 5.

For the oceanic structure the patterns of multiple S reflections within the crust are closely spaced in distance (see for example

the frames at 30 and 44 s) and transmission into the mantle occurs in each of the pre-critical reflections, helping to enhance the Sn field. The S waves in the oceanic crust are linked to reverberations of P waves in the ocean. These P multiples build a strong wave train following Sn that extends nearly to group velocities of 3.5 km s^{-1} (Fig. 6). This sequence of multiples in the coda of Sn attenuates with distance from the source, as can be seen in the 30–58 s snapshots in Fig. 5, and the peak amplitude moves further back in the multiple sequence. The presence of the strong reverberations is associated with the location of the source in the crust beneath shallow water. There is also a clear Pn contribution on the oceanic side, as can be seen in the 16 s frame, followed by the pPn arrival and P reverberations. Sereno & Orcutt (1985, 1987) pointed out the importance of the oceanic and sediment reverberations in building a significant oceanic Pn and Sn and Fig. 5 shows how these reverberations link to lithospheric structure.

On the continental side there is a smooth passage of the wavefield through the zone of thickened crust, and by 44 s in Fig. 5 we see the development of clear crustally guided Pg and Lg . There is significant conversion from Pg to S waves at the free surface, so that energy is being steadily extracted from Pg ,

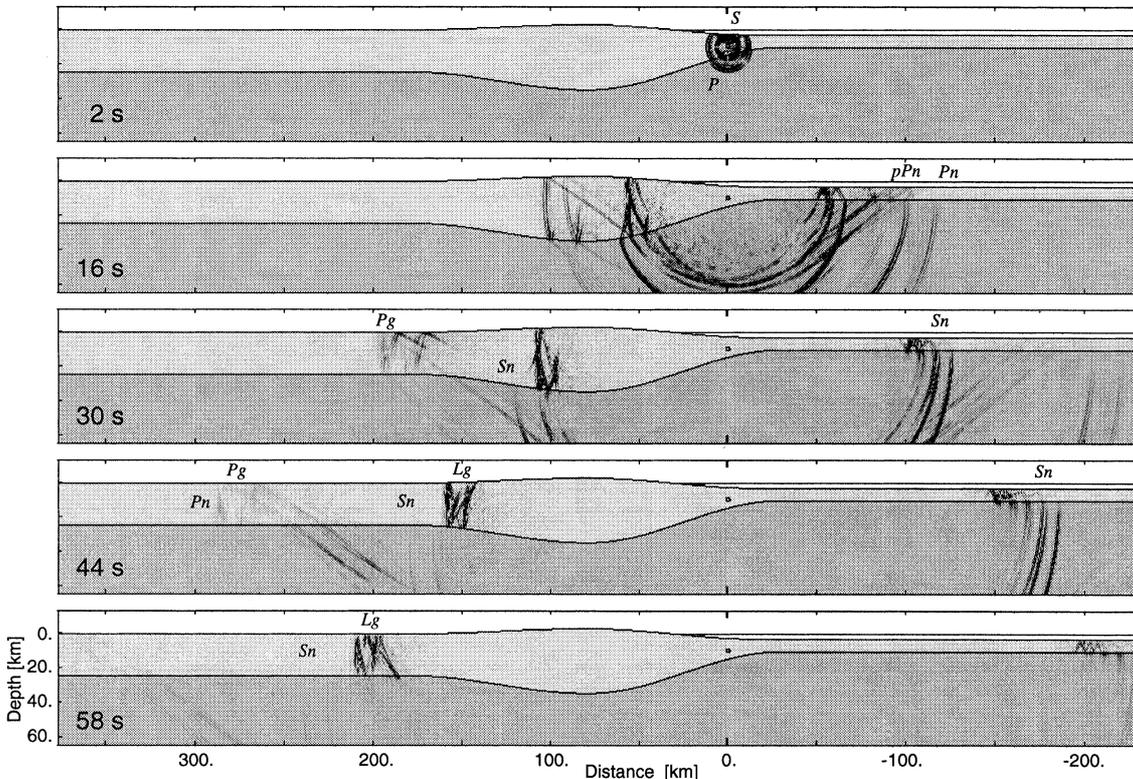


Figure 5. Snapshots of the seismic wavefield for a crustal source at 10 km depth, with a 2 Hz dominant frequency for the source pulse.

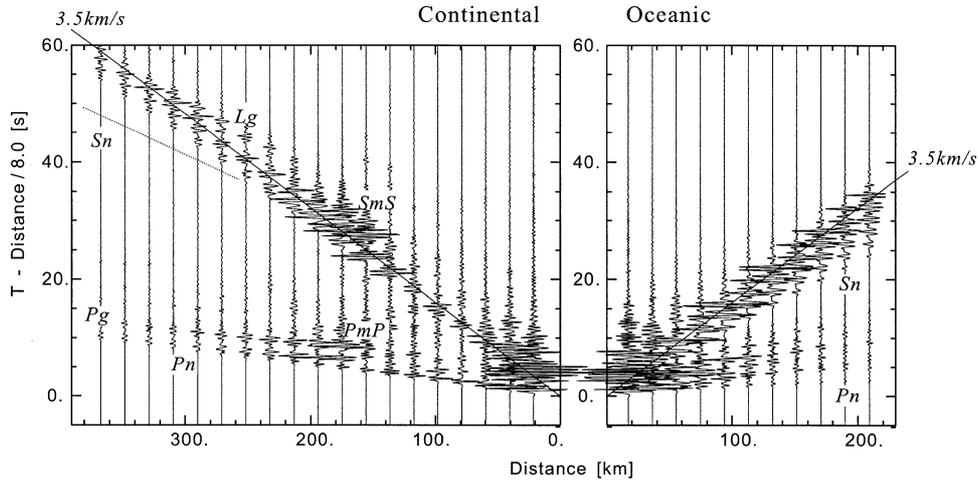


Figure 6. Vertical-component seismograms for a crustal source at 10 km depth, with a 2 Hz dominant frequency for the source pulse.

leading to strong apparent attenuation of this phase. The converted *S* waves have phase velocities such that they will be turned back from the upper mantle transition zone or below and so will emerge at the surface beyond 2000 km.

In the zone of thickened crust the *P* waves in the crust link to clear *S* head waves in both crust and mantle, which are recognizable by their V-shaped profile for both the direct *P* contribution and *pP* in the 16 s frame in Fig. 5. The complex pattern of *S* crustal reflections though the zone of crustal thickening tends to amplify as the waves enter the more homogeneous crustal region, giving a distinct group of energy associated with *Lg*. The shape of the crust–mantle boundary leads to constructive interference by focusing of the *PmP* and

SmS reflections near 200 km range (Fig. 6). This effect may help to explain the clear observation of waves with the group velocity of *Lg* in western Hokkaido (profile C) at about 200 km from the subduction zone, despite the crossing of the attenuative structure associated with the volcanic front. As the *S* waves emerge from the thickened crust there is some transfer from crustal to mantle *S* waves and this is visible in the 58 s frame of Fig. 5. The mantle phases *Pn* and *Sn* are not very pronounced on the continental side and the dominant contribution in this simple model comes from the crustally guided phases.

We consider a source at greater depth below the crust–mantle boundary in Figs 7 and 8; this case would correspond to a source at the top of a subduction zone. The behaviour of

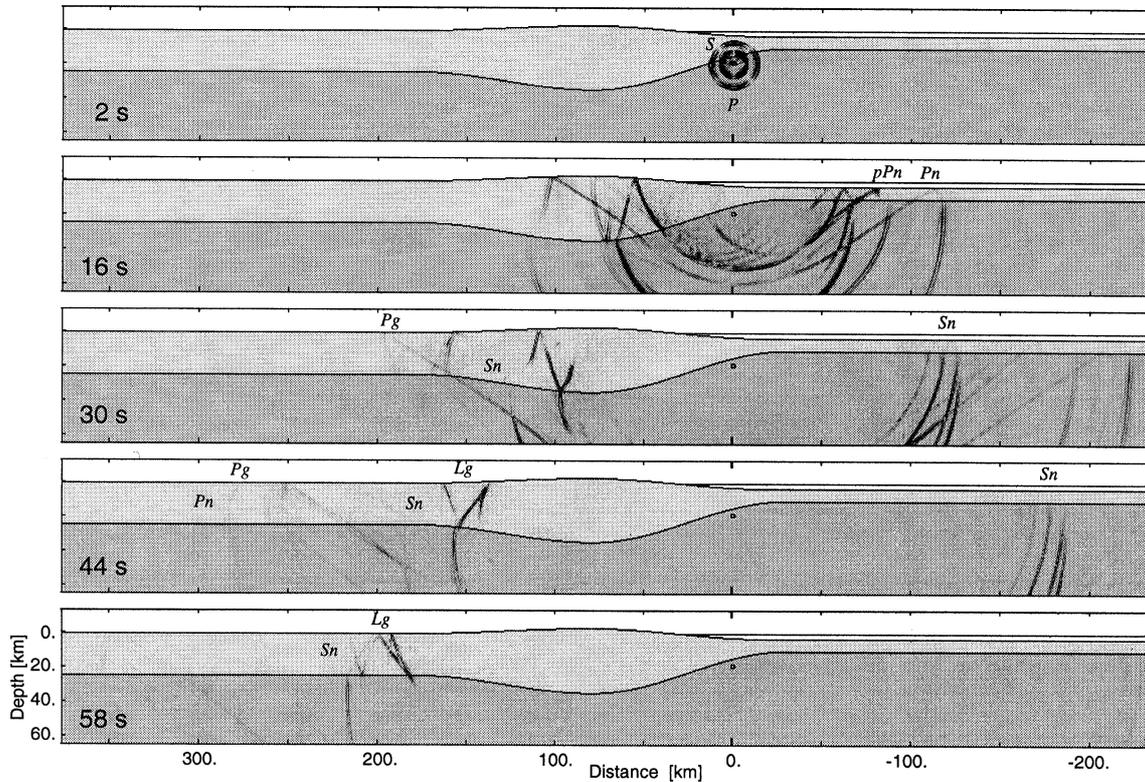


Figure 7. Snapshots of the seismic wavefield for a deeper source at 20 km depth, with a 2 Hz dominant frequency for the source pulse.

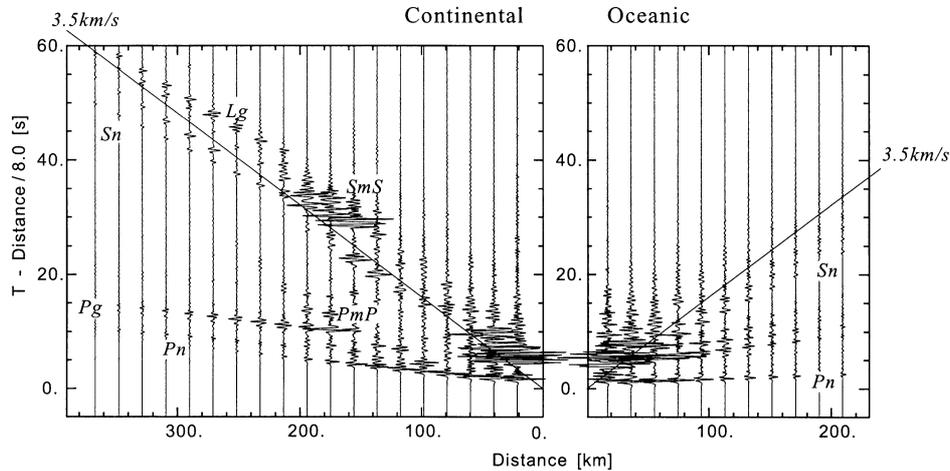


Figure 8. Vertical-component seismograms for a deeper source at 20 km depth, with a 2 Hz dominant frequency for the source pulse.

the wavefield is generally similar to the shallow source but the greater time separation between waves that have propagated directly from the source and the surface reflections also produces spatial separation in the snapshots and hence sharper phase groups.

The mantle source injects more energy into the mantle arrivals P_n , S_n than the shallower source. Also, the mantle phases arising from the surface reflections are more distinct, as for example for S_n on the continental side at a range of 300 km and reduced time of about 42 s. The group of mantle reflections that add to produce the crustally guided L_g phase are very clearly seen for the continental side in the 44 and 58 s frames of Fig. 7. Once again there is a strong amplitude increase in the crustal phases near 200 km range associated with the influence of crust–mantle topography. On the oceanic side the mantle phases are quite strong but there is a very weak crustal contribution and the water reverberations are much less significant than for the shallower source.

4 DISCUSSION

The set of observations and numerical simulations we have presented gives a very clear representation of the differences in the character of the regional wavefield between oceanic and continental regimes, and the way in which the different wave types interact with crustal structure. Both oceanic and continental regimes can show distinct P_n , S_n arrivals to considerable distances. These mantle phases are reinforced by the nature of the velocity distribution in the uppermost mantle. In the case of oceanic P_n , S_n , reverberations in the water layer help to sustain the amplitude (*cf.* Sereno & Orcutt 1985, 1987). Surface multiples of the mantle phases can also be important for longer-range propagation in the continental regime (Kennett 1985).

Continental structures can develop an efficient set of guided S waves trapped by total reflection at the free surface and the crust–mantle boundary for a wide range of frequencies. The train of multiples encompasses many different orders of reflection, and as the distance from the source increases, successive multiples become important, with enhanced amplitude near the critical angle for the reflection process at the crust–mantle boundary. This gives rise to the characteristic trajectory of the

L_g phase. In a modal viewpoint there is a skein of overlapping group velocity extrema in the range $3.5\text{--}3.0\text{ km s}^{-1}$, over a large range of frequencies, which collectively represent the L_g phase.

If the crust is too thin, then the spatial and temporal separation of individual multiples is lost. The pattern of modal group velocities then breaks up into isolated contributions reflecting the variation in propagation characteristics with frequency (see e.g. Fig. 7 of Zhang & Lay 1995). Low-frequency waves are not influenced much by the presence of the thin oceanic crust. Only at very high frequencies is there any degree of crustal trapping, even for SH waves that are not coupled into the water layer. Leakage of S energy from the crust into the mantle is the normal situation in oceanic structures.

ACKNOWLEDGMENTS

BLNK would like to thank the Japan Society for the Promotion of Science for an Invitational Fellowship to the Earthquake Research Institute, University of Tokyo, during which much of this work was carried out. The various groups contributing to the JARRAY coordinated by the Earthquake Research Institute, University of Tokyo, are acknowledged for the cooperation that has made such a dense data set readily available for a wide variety of studies.

REFERENCES

- Campillo, M., 1990. Propagation and attenuation characteristics of the crustal phase L_g , *Pure appl. Geophys.*, **132**, 1–9.
- Ewing, M., Jardetsky, W. & Press, F., 1957. *Elastic Waves in Layered Media*, McGraw-Hill, New York.
- Furumura, T., Kennett, B.L.N. & Takenaka, H., 1998. Parallel 3-D pseudospectral simulation of seismic wave propagation, *Geophysics*, **63**, 279–288.
- Kennett, B.L.N., 1985. On regional S , *Bull. seism. Soc. Am.*, **75**, 1077–1086.
- Kennett, B.L.N., 1986. L_g waves and structural boundaries, *Bull. seism. Soc. Am.*, **76**, 1133–1141.
- Knopoff, L., Mitchel, R.G., Kausel, E.G. & Schwab, F., 1979. A search for the oceanic L_g phase, *Geophys. J. R. astr. Soc.*, **56**, 211–218.
- Press, F. & Ewing, M., 1952. Two slow surface waves across North America, *Bull. seism. Soc. Am.*, **42**, 219–228.

- Regan, J. & Harkrider, D.G., 1989. Numerical modelling of *SH Lg* waves in and near continental margins, *Geophys. J. Int.*, **98**, 107–130.
- Sereno, T. & Orcutt, J.A., 1985. The synthesis of realistic oceanic *Pn* wavetrains, *J. geophys. Res.*, **90**, 12 755–12 776.
- Sereno, T. & Orcutt, J.A., 1987. Synthetic *Pn* and *Sn* wavetrains and the frequency dependence of *Q* of oceanic lithosphere, *J. geophys. Res.*, **92**, 3541–3566.
- Walker, D., 1977. High frequency *Pn* and *Sn* phases recorded in the western Pacific, *J. geophys. Res.*, **82**, 3850–3360.
- Walker, D., 1981. High frequency *Pn* and *Sn* velocities: some comparisons for the western, central and south Pacific, *Geophys. Res. Lett.*, **9**, 207–209.
- Zhang, T. & Lay, T., 1995. Why the *Lg* phase does not traverse oceanic crust, *Bull. seism. Soc. Am.*, **85**, 1665–1678.