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***Subducted Lithosphere or 530 km
discontinuity?***

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Subducted lithosphere or 530 km discontinuity?

Serge Oreshin¹, Lev Vinnik¹, Alexandr Treussov¹ and Rainer Kind²**Abstract.**

Tomographic images of the mantle west of Japan and south Kuril islands suggest that in this region the subducted lithospheric slabs instead of penetrating the lower mantle, lie flat on top of it [van der Hilst *et al.*, 1991; Fukao *et al.*, 1992]. To test these models with independent data and techniques, we analyse the slowness of the P arrivals of Kuril-Japan seismic events at a seismic network in Baikal region. Relative to IASP91 standard earth model, the flat lying slab should generate an additional refraction line at epicentral distances less than 30°. The presence of this line is very clear in our data. The slope of the line and its position relative to the other lines are qualitatively consistent with the estimates of the slab velocity and thickness in the tomographic models. However, a similar wavefield can be generated by 530 km discontinuity, the presence of which is established by long-range refraction profiling in the neighbouring regions of northern Eurasia.

Introduction

Tomographic images trace the subducted plates in several regions as continuous high-velocity anomalies from the earth's surface to depths exceeding 800 km. However, in the north-west Pacific, some plates, instead of penetrating the 660 km discontinuity, lie flat on top of it [van der Hilst *et al.*, 1991; Fukao *et al.*, 1992; Tajima and Grand, 1995]. Distribution of seismic sources and receivers is critical for resolution of traveltime tomography. Hypocenters and the absolute travel times can be biased due to uneven station distribution. Waves sensitive to structure near 600 km depth are recorded as first arrivals only in a narrow distance range, and the number of stations available in this interval can be small. In this study we test the models of the upper mantle beneath easternmost Eurasia with other data and techniques. The idea is to use dense arrays of seismic sources and receivers (Figure 1), and to infer the upper mantle velocity model not from the absolute travel times which can be easily biased, but from the slopes of the time-epicentral distance plots.

Data and techniques

We use a network of analogue short-period stations in the Baikal rift region (Figure 1) and a set of 23 seismic events in and around Japan (Figure 1 and Table 1). For the parameters of the stations see Fedorova [1985]. Arrival times of the P waves can be evaluated accurately by correlating similar features of the recorded waveforms. Most of the epicenters are located in a vicinity of 42°N, 142°E, and the related P waves propagate to NWW. Figure 2 shows the 2-D model of

the subduction zone for these wavepaths and the results of raytracing for a shallow event in the slab. The dip of the subducted plate between the surface and 600 km depth is inferred from the hypocentral data given by the ISC. The length of the flat lying slab is adopted from the published tomographic images. Relative to IASP91 [Kennett and Engdahl, 1991], the flat lying slab generates an additional loop with the refraction line in first arrivals at distances 21° – 24°. The slope of this line is directly related to the slab velocity, while the distance of the point where this line emerges in first arrivals is sensitive to slab thickness. This point is marked by a sharp change in slowness of the first arrivals. Two other loops are related to global 410-km and 660-km discontinuities.

Corrections of two kinds are applied to the raw travel time measurements for each event: (1) correction for lateral heterogeneity beneath the stations, and (2) correction for source depth. Correction (1) is necessary, because apparent slowness can be biased by the crust and mantle heterogeneity beneath the stations. To quantify this effect, the time residuals relative to IASP91 were determined for the P waves of a number of seismic events in two distant seismic zones (Figure 1). Lateral heterogeneities in the lower mantle are weak, and the residuals reflect lateral heterogeneity beneath the network rather than in the lower mantle. The residuals at the stations of the Baikal network for each event were normalized by subtracting their average and then averaged for every station over many events. The obtained average values (Figure 1) were subtracted from the raw P travel times of the events in and around Japan.

Correction (2) is required, because the data of an individual event are usually not sufficient for a dense coverage of the epicentral distance range of interest. Therefore we combine the data for several events into single time - epicentral distance plot. This is illustrated in Figure 3 with the traveltime data for a shallow event in

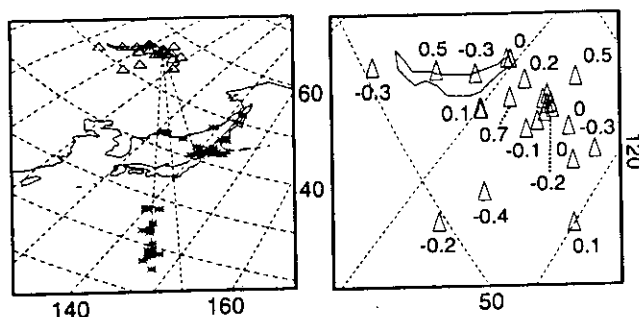


Figure 1. Map of the region with seismograph stations (triangles) and epicenters of seismic events (stars). Blow-up of the receiver network with the teleseismic P traveltime residuals in seconds is shown on the right. Epicenters north of 30°N are used to measure mantle P velocity beneath easternmost Eurasia. Epicenters south of 30°N and in the New Hebrides region (indicated only by projection of the wavepath) are used to calibrate the network.

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Table 1. List of representative events

	Date	Lat.	Lon.	Dep.	Mb
1	16-05-78	40.99	141.26	15	5.7
2	28-07-79	49.99	142.58	15	5.5
3	25-12-79	43.23	131.19	543	5.0
4	27-11-81	42.88	131.11	546	5.7
5	17-03-83	51.60	142.45	15	5.6
6	11-02-87	43.15	132.14	504	5.4
7	21-04-90	47.46	138.96	520	5.2
8	12-06-91	42.76	143.34	114	5.8

the slab, and one at a depth of 114 km. All variations of slowness related to the bottom of the mantle transition zone in the raw data for a deeper event are shifted towards smaller epicentral distances relative to those for a shallow event. The correction which eliminates this effect is evaluated by raytracing for the model in Figure 2. In Figure 3, the data for the second event are corrected by assuming that velocity in the dipping slab is 3% higher than in IASP91. If, instead of 3%, 1% was assumed, the epicentral distance shift for the second event would be 1.25° , instead of 1.1° . The difference is clearly insignificant, which means that the correction is

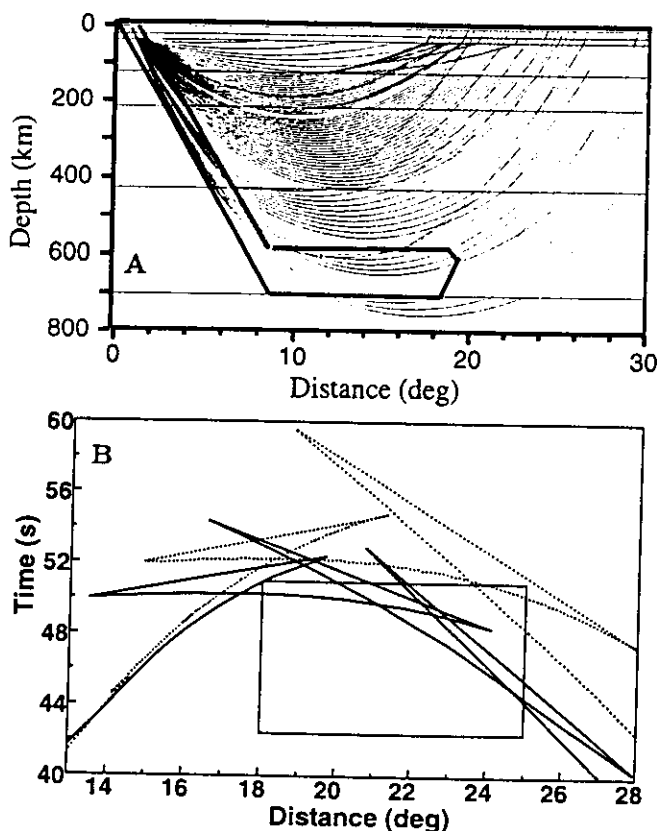


Figure 2. Subducted slab model and seismic rays, with flattening transformation applied (a), and the related P traveltimes (b, solid line). The traveltimes are calculated for the P velocity in the slab 3% higher than in IASP91 [Kennett and Engdahl, 1991] and IASP91 outside of the slab. Traveltimes for IASP91 are shown for comparison by dash line. Reduction slowness for travel-times here and in Figures 3, 4 is $11\text{s}/^\circ$. The box in (b) is the time and distance window for the data in Figures 3, 4a and 4c.

only weakly dependent on the assumed velocity in the dipping slab. To combine the data of the two events in Figure 3 into one time-epicentral distance plot, the travel times of the second event are incremented by the value, which provides the best fit (minimum rms difference) between the data of the two events in their common distance intervals. The corrections make the absolute travel times meaningless, but preserve the slope of the time-epicentral distance slope.

Similar procedures were applied to the other events in the slab at 15 km - 372 km depths, starting from shallow events, and complementing them with the deeper ones. The depth corrections were calculated assuming 3% velocity anomaly in the slab. The resulting time - epicentral distance plot is shown in Figure 4a. Two other panels in Figure 4 correspond to the events at 500 - 550 km depths and shallow events in the Sakhalin island. The plot for the deep events is obtained like in Figure 4a, but for 500 km reference depth. Even if the assumed geometry and the P velocity of the dipping slab deviate from reality, the related errors in the data are small, because the wavepaths for the deep events in the dipping slab are short. The data for the two Sakhalin events are important, because both are at the same depth (15 km), and their P waves propagate outside of the dipping slab.

Interpretations

The corrections for the the data in Figure 4a for events at different depths (shown by different symbols) differ by up to an order of magnitude. The corrected data don't show any systematic discrepancy, which implies that the corrections are accurate enough. At small distances, the times are practically constant, which means that the apparent slowness of the first arrivals is close to $11\text{s}/^\circ$. At larger distances, the data suggest

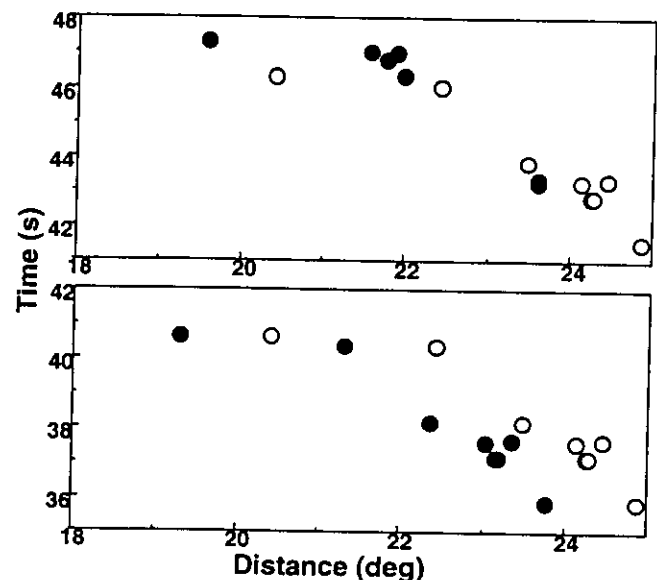


Figure 3. Illustration of correction for event depth. Traveltimes for events 1 of Table 1 (15 km depth, upper panel) and 8 (114 km, lower panel), with the teleseismic residuals subtracted are shown by filled circles. To obtain the composite traveltime plot for 15 km reference depth, traveltimes of event 8 are shifted, first, in distance (open circles, lower panel) and, second, in time (open circles, upper panel).

appearance of a branch with a higher apparent velocity. The sharp change of the slope occurs at around 21° . We compare these data with the theoretical data for tree models. The theoretical data are shifted along the time axis to obtain the best fit (minimum rms difference) between the observed and theoretical data.

For the line labeled 'IASP91', the model of the dipping slab shown in Figure 2 (without flat lying part), and IASP91 velocity model outside of the dipping slab are adopted. The P velocity in the dipping slab is taken 3% higher than in IASP91. The agreement between the two kinds of data is rather poor: the sharp change of the slope in the theoretical curve occurs at 23° , contrary to 21° in the observations. As a result of this, at distances less than 22° , most of the observed data lie well above the theoretical line, whereas at distances between 22° and 24° , they lie mostly below the line. The theoretical line labeled 'Slab' is obtained for the slab geometry shown in Figure 2 (including flat lying part) and IASP91 outside of the slab. P velocity in the slab is taken 3% higher than in IASP91, and thickness is fixed at 100 km. The adopted parameters of the slab minimize the variance of the data relative to the line 'Slab' (Figure 5a). This model is clearly superior to the first one, but it is not unique. Seismic refraction profiling sometimes reveals a discontinuity at 520–540 km depth, which is missing in IASP91 and other global models. This discontinuity is especially clear in the records of long-range profiling in northern Eurasia [Egorkin *et al.*, 1987; Mechie *et al.*, 1993; Ryberg *et al.*, 1997]. Fig. 5b shows a model of 530-km discontinuity [Ryberg *et al.*, 1997], with a velocity step of 0.25 ± 0.05 km/s at a depth of 530 km and a velocity gradient of 0.001 1/s below the discontinuity. To improve fit between our observations and the theoretical line, this model was slightly modified by adopting a velocity step of 0.3 km/s at a depth of 540 km (Figure 5b, solid line). The travel times were calculated as for 'IASP91', but with the model shown in Figure 5b outside of the dipping slab. For this model, variance of the data relative to the theoretical line ('Discontinuity' in Figure 4) is reduced practically to the level provided by the line 'Slab'.

The data for the deep events (Figure 4b) demonstrate striking difference between the observed slope and that corresponding to IASP91. This difference is equivalent to $3 \pm 0.5\%$ difference in P velocities in the bottom layer of the upper mantle. Lines 'Slab' and 'Discontinuity', calculated for the same models as in Figure 4a, are in agreement with the observations. The P waves from the shallow Sakhalin events don't propagate in the dipping slab. Therefore, the models for them (Figure 4c) were modified accordingly. Here, again, IASP91 is in a strong disagreement with the observations, whereas the other models fit them very well.

Discussion and conclusions

Three independent data sets in Figure 4 are well consistent with each other, and require similar interpretations. One of these interpretations is a flat lying slab. The P velocity anomaly in the slab (3%) is 2–3 times stronger than in the tomographic models, but otherwise the optimum parameters of the slab are broadly consistent with those in the tomographic models. On the other hand, the same features can be interpreted equally well as an indication of the presence of 530-km discontinuity. The optimum velocity step at 530-km discontinuity (0.3 km/s) required by our data is somewhat larger

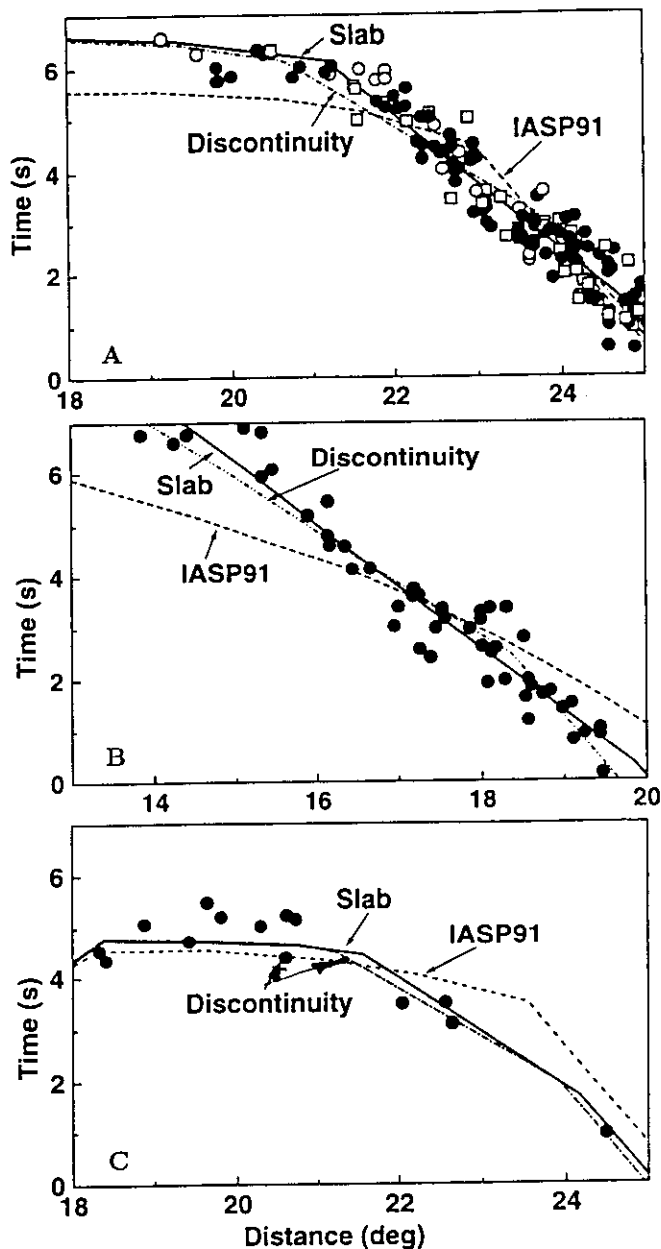


Figure 4. Composite traveltimes and theoretical curves for the models discussed in the text: a - events in the slab at 15 - 85 km depths (open circles), 88 - 178 km (filled circles), and 248 - 372 km (squares); b - deep events 3, 4, 6, and 7 of Table 1; c - events 2 and 5. Reference depths are 15 km for (a) and (c), and 500 km for (b).

than reported by Ryberg *et al.* [1997], but within their error bounds. The P velocity step at 660 km discontinuity in this model is smaller than in IASP91. Qualitatively, this reduction is consistent with the indications [Estabrook and Kind, 1996], that the actual P velocity step at 660 km discontinuity is smaller than in IASP91.

Some constraints on the properties of the bottom layer of the upper mantle can be provided the Ps converted phases observed in the same region. If the anomalously high P velocity in this layer is caused by low temperature, the related S velocity anomaly (in % of normal velocity) should be of about the same magni-

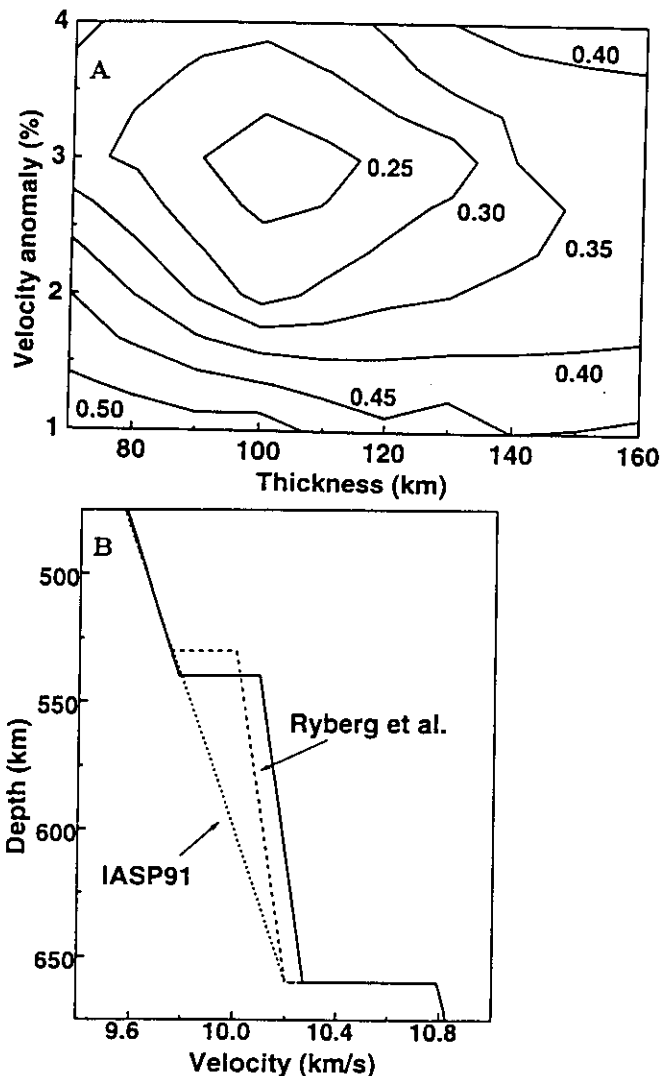


Figure 5. Variance of the traveltimes in Figure 4a in s^2 relative to theoretical traveltime curve for the slab in Figure 2 as a function of slab velocity anomaly and thickness (a), and the P velocity model for 530 km discontinuity (b, solid line). IASP91 and model by Ryberg *et al.*, [1997] in (b) are shown for comparison.

tude. Then in the region of the flat lying slab, instead of one strong Ps phase converted from 660 km discontinuity, two weaker phases should be present: one converted from the top of the slab, and the other related to its bottom at about 660 km depth. This, however, is not observed: the Ps converted phases in the region of interest are very similar to those recorded elsewhere [Vinnik *et al.*, 1996]. Moreover, a cooling effect of the slab, which should be manifested by a depression on 660 km discontinuity, is not observed, as well. On the other hand, whereas indications of 530 km discontinuity in the P wave field are reported frequently, there is no evidence of it in the published data on the mantle Ps phases in various regions. Thus, the observations of

the mantle Ps phases, if combined with the data of this study, seem to favour 530 km discontinuity rather than the flat lying slab beneath easternmost Eurasia.

According to high-pressure data, 530 km discontinuity can be related to the phase transition from β to γ spinel in olivine. P velocity contrast related to this transition is 0.8% – 1.5% [Ridgen *et al.*, 1991], much lower than in the seismic observations in northern Eurasia. For this reason, it is conceivable that 530 km discontinuity is produced not only by the phase transition in olivine, but also by a change of composition and/or texture. These changes could be induced by ancient episodes of subduction, with remnants of the paleoslabs being trapped in the mantle transition zone.

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