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Crustal structure of the Bay of Bengal and adjacent seas

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ABSTRACT. The time-distance curves, plotted from the first onsets of *P*-waves, recorded by the coastal seismograph stations for the earthquakes originating from the Bay of Bengal and Andaman and Nicobar Islands, indicate a single layered crust in this region. The depth to the Moho has been determined as 25·4 km. The velocities at the upper layer and at the top of the mantle have been determined as 6·28 ± 0·04 and 7·97 ±0·01 km/sec respectively. The granitic layer was found totally absent.

1. Introduction

The region investigated is roughly bounded by the 4°N and 22°N parallels and by the 80°E and 96°E meridians. The southern part of the Burmese arc is largely submerged in the Bay of Bengal. The Andamans and Nicobars are the unsubmerged peaks of a group of ridges. The bottom configuration of this region is very imperfectly known. 43 earthquakes originating from this region during the period 1950 to 1967 have been selected for the present study. A great earthquake occurred in the Bay of Bengal on 31 December 1881 which caused a lot of devastation in the coastal regions.

2. Data and analysis

The coastal seismographic stations whose record was used in this study are Port Blair, Chittagong, Calcutta, Visakhapatnam, Madras and Colombo as shown in Fig. 1.

P-times of all the earthquakes have been taken from the monthly *Seismological Bulletins* published by India Meteorological Department. The epicentres, origin times, magnitudes and depths have been taken from different sources. The distance of the epicentre from the recording station has been determined by the direction cosine method as given below.

$$\cos\triangle = aA + bB + cC \qquad (1)$$

where a, b, c are the direction-cosines of the stations and A, B, C are the direction-cosines of the epicentres. These are related to the Lat. ϕ and Long. λ by the relations :

$$\left.\begin{array}{l} a = \cos\phi_s .\cos\lambda_s \\ b = \cos\phi_s .\sin\lambda_s \\ c = \sin\phi_s \end{array}\right\} \qquad (2)$$

$$\left.\begin{array}{l} A = \cos\phi_E \cos\lambda_E \\ B = \cos\phi_E \sin\lambda_E \\ C = \sin\phi_E \end{array}\right\} \qquad (3)$$

where ϕ_s and λ_s are the latitude and longitude of the station respectively and ϕ_E and λ_E are those of epicentre. Only geocentric latitude and longitude of the station as well as of the epicentre were used. Knowing the arrival times and the distance from various seismographic stations the time-distance curves have been drawn by the method of least squares as shown in Fig. 2. The reverse of the slopes or the velocities has been determined.

3. Travel-times and depth

The travel-time of a wave can be represented as

$$T = V^{-1}\triangle + I \qquad (4)$$

where V is the wave velocity and I is the intercept, the wave makes in the Y-axis which depends upon the thickness of the layer and the velocity V. Thus in Fig. 2, the travel-times of the two waves may be represented by the following two linear equations.

$$T_1 = (0\cdot1593 \pm 0\cdot0010)\triangle + 1\cdot6 \pm 1\cdot0 \qquad (5)$$

$$T_2 = (0\cdot1254 \pm 0\cdot0010)\triangle + 5\cdot0 \pm 1\cdot3 \qquad (6)$$

The reduced travel times ($T - V^{-1}\triangle$) *versus* $\triangle$ have been plotted as shown in Figs. 3 and 4 to show clearly the cross over points from one straight line segment to another.

The wave velocities, intercept times and slopes are presented in Table 1.

The depth H to the discontinuity can be determined by the relations

$$I = \frac{(2H - h).\ \cos_{12}}{V_1} \qquad (7)$$

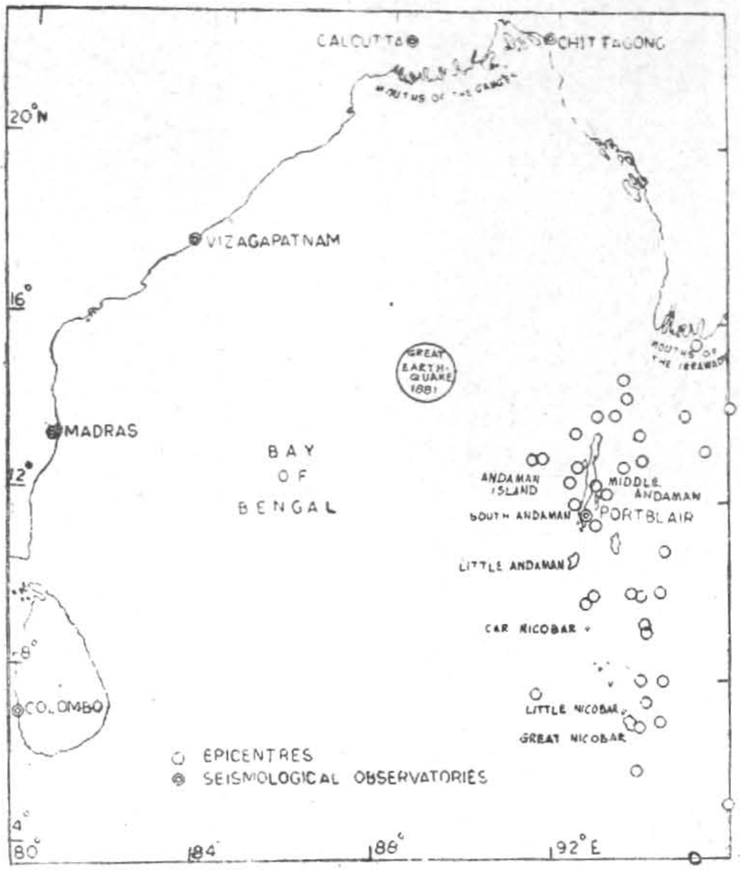

Fig. 1
Epicentre of earthquakes and the Seismological Observatories around the Bay of Bengal

$$\cos_{12} = \sqrt{1 - \left\{ \frac{V_1}{V_2} \right\}^2} \qquad (8)$$

where V_1 and V_2 are the velocities of P-waves above and below the discontinuity and h is the mean focal depth.

Even at this advanced stage of Seismology, the the absolute depth still remains a puzzling factor to the seismologists. Many agencies differ much in depth determinations. Take the case of the present paper, some of the earthquakes as examined were depthed as 18, 14 and 33 km respectively by U.S.C.G.S. from the data of distant stations, while I.S.S. and U.R.S.S. have computed them as of zero focus. Similarly, the depths of a few earthquakes used in this paper as computed by various agencies using the data of distant stations differ much from one another. The mean focus, therefore, lies somewhere in the layer of thickness H. Thus due to uncertainty of the depth of focus, three cases arise as follows :

(1) When the mean focus lies at the top of the layer ($h=0$)

(2) When the mean focus lies at the mid-point of the layer ($h=H/2$).

(3) When the mean focus lies at the bottom of the layer ($h=H$)

The depth of discontinuity has been determined for each case using the relation (7) and (8) as presented in Table 2.

The author was, however, handicapped in the absence of sufficient seismological observatories in the vicinity of earthquake belt in the Bay of Bengal. The line, therefore, representing a velocity of 6·28 km/sec does not contain sufficient points. The main contribution in it was from Port Blair, situated in the heart of this region.

4. Interpretation of the result

Studies have been carried out for investigating the crust in the ocean at different parts of the world by many workers using body waves. The result they obtained was quite similar to what has been obtained in the present study.

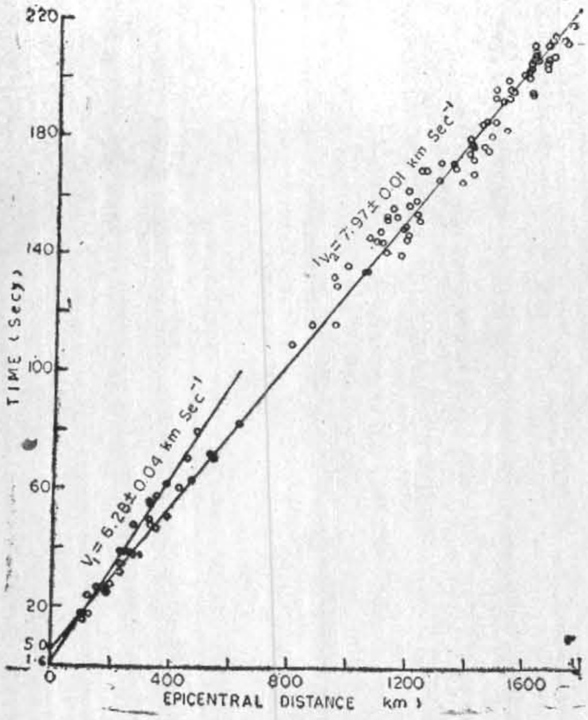

Fig. 2

Travel time curves obtained from the first onset of the *P*-waves in the Bay of Bengal and adjacent seas

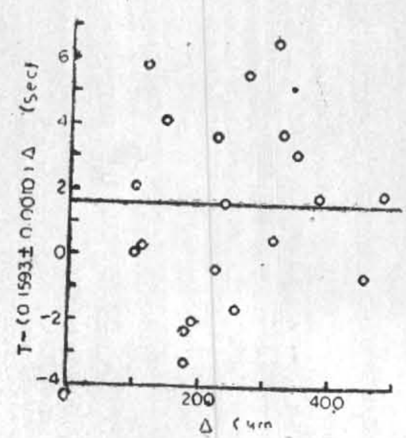

Fig. 3

Reduced travel time plot. Reducing velocity 6·28 km/sec

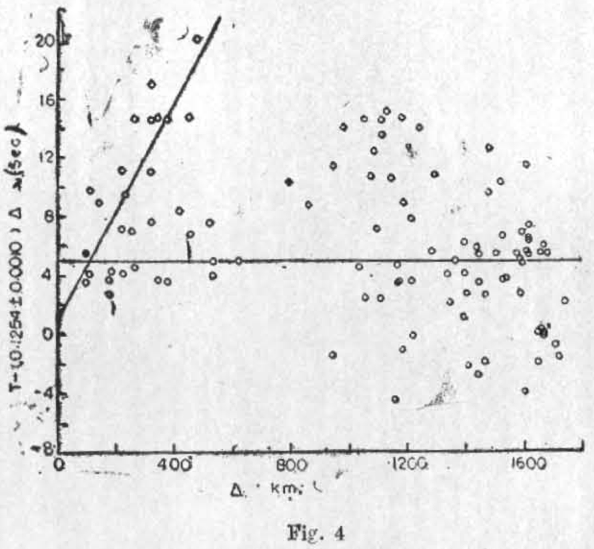

Fig. 4

Reduced travel time plot. Reducing velocity 7·97 km/sec

TABLE 1

Wave velocities, intercept-times and slopes

Wave	Velocity (km/sec)	Intercept (sec)	Slope (sec/km)
P^*	$6{\cdot}28 \pm 0{\cdot}04$	$1{\cdot}6 \pm 1{\cdot}0$	$0{\cdot}1593 \pm 0{\cdot}0010$
Pn	$7{\cdot}97 \pm 0{\cdot}01$	$5{\cdot}0 \pm 1{\cdot}3$	$0{\cdot}1254 \pm 0{\cdot}0010$

TABLE 2

S. No.	h (km)	H (km)
1	0	25·4
2	$H/2$	33·9
3	H	50·9

NOTE—In determining the thickness of layer it was assumed that the velocities are constant in each layer and the layers are lying flat.

For instance, Green and Hales (1968) found from Early Rise (Project) data of lake Superior, a *P*-wave velocity of 6·3 km/sec at zero depth having layer thickness of 20 km.

Lehmann and Bullen found an oceanic velocity of 6·30 km/sec at a depth of 5 km and a velocity of 8·01 km/sec at a depth of 10 km to 60 km.

Similarly, Lomnitz and Bolt (1967) determined a velocity of *P*-wave as 6·3 km/sec having a layer thickness 15 km and velocity of P_n as $7{\cdot}98 \pm 0{\cdot}05$ km/sec to the coast ranges southwards from San Francisco bay to Priest.

Richter and Nordquist (1951) in the study of Manix earthquake of California found the *P*-wave velocity of 6·34 km/sec.

The author in the present study plotted the first onsets of P-waves irrespective of their nature (P_g, P^*, P_n), the apparent velocity of 6·28 km/sec may therefore be interpreted as that of P^* in the basaltic layer as found by many other workers for other regions as mentioned above and the velocity 7·97 km/sec is an average upper mantle velocity immediately below Moho. The velocities beneath Moho vary at sea. At a depth of less than 400 km, possibly at 140 km depth, the difference between the oceanic and continental model disappears. The crustal velocity 6·28 km/sec and the upper mantle velocity 7·97 km/sec appear much low than those of surrounding continental areas. But it is not strange. The continental velocities are much higher than the oceanic velocities. In Kurile Islands area, P-velocity under Moho boundary is extra low, about 7·6 km/ sec, much lower than that in surrounding areas of Pacific Ocean where it amounts to about 8·0 km/sec. For oceanic areas between Greenland and Norway and north of Iceland, the upper mantle P-wave velocity is 7·4 km/sec to a depth of 140 km when the velocity increases to 8·2 km/sec.

The results in Table 2 (S. Nos. 2 and 3) do not appear to be reasonable, as even in the nearby continental areas the thickness of a single layer was not found so much high of the order of 34 to 51 km by any worker. Such a huge thickness of the crust in an oceanic region is not possible. Therefore the result as derived, assuming the earthquakes to be of surface focus and giving the thickness of the layer as 25·4 km is the most probable. Jeffreys (1946) had himself admitted while studying the seismic waves in Western and Central Europe that most of the small earthquakes have their foci near the surface. This fact was further corroborated by the records of Port Blair giving clear near earthquake phases. The transmission of near earthquake phases particularly in a region of thinner crust is out of question if the focal depth is of the order of 34 to 51 km. Transmission of these phases is possible only when focus is very much shallow and lies in the topmost layer.

As the crust was short of granitic block, the total crustal thickness was less and hence the delay times caused by the P_n wave in starting from the focus is also less.

The small intercept of P^* may mainly be due to the fact that most of the recording stations are situated on continental areas where the granitic layer is present while the epicentral region is without the granitic layer. This might have caused a weak refraction delaying the wave. It is, therefore, quite clear from the present study that the crust in this region is single layered and is of the order of 25 km.

5. Conclusion

The result so obtained may be summarized as follows :

(*i*) The Bay of Bengal rests on a 25·4 km thick layer of basalt having the velocity 6·28 km/sec.

(*ii*) The depth to the Moho is 25·4 km and the upper mantle velocity of P-wave immediately below Moho is 7·97 km/sec.

(*iii*) The granite block is almost completely absent, however some unconsolidated sediment (1-2 km) is always possible in the form of mud as generally found in oceanic areas.

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