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Anantnag earthquakes (February to April 1967)

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ABSTRACT. A study has been undertaken of the Anantnag earthquake of 20 February 1967 and 80 aftershocks which followed the main shock upto April 1967. The results indicate that the main shock occurred at a depth of about 25 kilometres on a reverse fault striking N 330°E and dipping at an angle of 57° towards N 240°E, having motion along the direction of dip. The after shocks were also located on the same fault plane at depths varying between 15 and 35 km. The decay of the sequence of aftershocks can be represented fairly accurately by the relation $N(t)=15/(t-0\cdot5)$ where, t is in days and $N(t)$ is the number of aftershocks after t days. The aftershock sequence reveals a 'b' value of 0·6 as compared to 1·0 obtained in the case of highly seismic areas, which can be attributed to the greater depth of the shock and also to a lower seismogenic potential of the area.

The total energy released in the aftershocks has been found to be 0·47 times the energy released in the main shock inspite of the fact that the stored creep strain was 2·3 times that of the elastic strain showing that only 20 per cent of the creep energy was radiated as elastic energy, the remaining being dissipated as heat.

The travel time data of the various crustal phases recorded at near stations indicates that both the Granitic and the Basaltic layers in the region are about 29 km thick *i.e.*, the average depth of the Moho is about 58 km, a result consistent with the authors earlier conclusion of nearly 60 km from the data of surface wave dispersion.

1. Introduction

During the period 20 February to the end of April 1967, the area near Anantnag in the Kashmir valley was subjected to a large number of earth tremors, some of which were strong enough to cause minor damage to buildings in Anantnag town and the villages located near the epicentre. The main shock occurred in the evening of 20 February 1967 at 1519 GMT and was followed by two strong shocks at 1238 GMT on 21 February and at 1735 GMT on 24 February. Available seismograph records also show that the main shock on the 20th was preceded by a foreshock at 1424 GMT on the same day. Aftershocks continued to be recorded with diminished frequencies at least upto the first week of April 1967. Due to the absence of an observatory in the Kashmir valley all the aftershocks could not be recorded but 80 shocks were strong enough to be picked up at seismograph stations located more than a hundred kilometres away from the epicentre. The stronger shocks could be recorded by sensitive stations throughout the world and this made possible an accurate determination of their parameters. The object of this paper is to present the results of analysis of the seismograph records at some Indian stations with a view to study the origin and nature of the main shock and the aftershock sequence. For this purpose the data collected from the network of seismological observatories started in connection with the investigation of earthquakes in the Beas and Sutlej catchments, has been most useful. These stations are located at Dalhausie, Nurpur, Mukerian, Pong, Jwalamukhi and Ghaggar. Fig. 1 shows the location of these stations on a tectonic map. In addition to the data of these observatories the records of the observatories at Bhakra, Dehra Dun, Rohtak, Sonepat and Delhi have been used whenever available. The observatories at Dalhausie, Nurpur, Pong and Jwalamukhi have sufficiently high magnification to pick up the first arrivals due to all strong aftershocks.

It is seen that practically all the stations are located in a southeasterly direction from the epicentre, a condition which is not conducive to accurate location of the epicentres. Great care had therefore to be undertaken in order to ensure the correctness of the epicentral parameters. Another disadvantage had been the lack of knowledge about the actual seismic wave velocities and crustal structure in the region, which had to be determined first, using data of strong shocks recorded throughout the world.

During the third week of March 1967 a party led by Shri S. N. Wakhaloo (1967) of the Department of Geology Jammu & Kashmir University, visited the affected parts and submitted a report. According to this report the shocks were felt at several places in the Anantnag district and were most severe at Shangas and Sap villages. In Anantnag town there were cases of objects over thrown from table and fall of plaster. Two walls of the first floor of a building collapsed. Even in the most affected villages the damage was limited to cracks in some buildings, being more severe in *kutcha* houses built of sunbaked bricks. No substantial building showed any significant damage. All these observations show that even in the most affected area the maximum intensity of the earthquake was

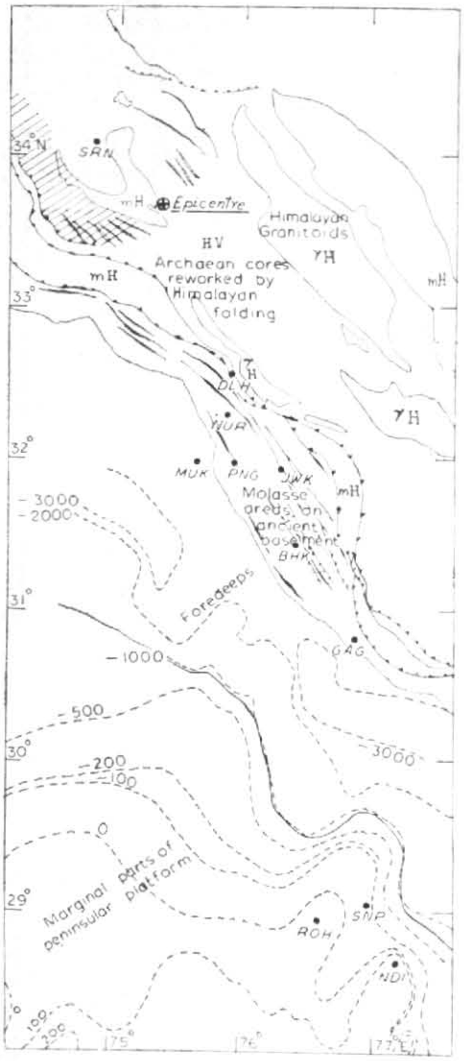

Fig. 1

only VI on the Modified Mercalli Scale which would put the Richter magnitude of the shock around $5 \cdot 5$.

2. Seismic history of the Kashmir Valley

The Kashmir valley, due to its proximity to the chain of the great Himalayan Mountains and the Hindukush, lies in a seismically active region and has been subjected to moderate to heavy damage due to earthquakes originating there. The most disastrous earthquake in the valley occurred on 30 May 1885 with its epicentre about 20 km west of Srinagar. According to available reports 6000 persons perished in this earthquake. A number of earthquakes also had their epicentres in the valley itself. Bhan (1955) has given a list of 176 shocks felt during the 30-year period (1923—1952) at Srinagar. None of these caused any significant damage to structures. Only 41 caused minor craks in *kutcha* buildings. Among the recent shocks originating in the Kashmir valley mention may be made of the Badgam earthquake of 2 September 1963 which has been studied in detail by Srivastava *et al.* (1964).

3. Epicentre, origin times and other parameters of the principal shocks

It has been mentioned earlier that the principal shocks on the 20 and 21 February were recorded widely. The epicentres and other parameters for these shocks have been determined with the aid of electronic computors by the United States Coast and Geodetic Survey(USCGS). The USSR Academy of Sciences has also determined these parameters using their network of stations. Table 1 gives the epicentral parameters as determined by these organisations.

It will be seen that the origin times for the two shocks as determined by the two organisations are the same while the epicentres differ by about 10 km. An examination of the data of near stations in India (Dalhausie, Nurpur, Mukerian, Pong, Jwalamukhi, Bhakra, Ghaggar, Dehra Dun and Delhi) revealed that the S–P intervals recorded at these were the same for both the shocks, and the time interval between the arrival times of the P wave at different stations remained the same for both the shocks. It is therefore apparent that both the shocks originated from nearly the same source and had nearly the same focal depth. This aspect became more apparent on plotting the travel times *vs* distance of the arrivals of the P and S phases as recorded at the observatories mentioned above. The best result was obtained when the epicentre of both the shocks was taken to be at Lat. 33.61°N and Long. 75·33°E, *i.e.*, the epicentre of the shock of 21 February 1967 as determined by USCGS. The origin times of both the shocks had also to be reduced by $0 \cdot 5$ seconds, to make the intercept of $\overline{P}$ and $\overline{S}$ waves with the time axis as zero.

Since there were no stations very close to the epicentre, the depth of focus for the shocks could not be determined directly from seismograph records. The depths of focus for the shocks as determined by USCGS are given in Table 1. The depth of focus can also be determined from macroseismic data using the maximum felt intensity I_0 and the magnitude of the shock. Various formulae are available, for this purpose. Karnik gave the formula (Zatopek 1968)—

$$M = 0 \cdot 6\, I_0 + \log h + 0 \cdot 4$$

and Shebalin (1961) has found the following relationship.

$$M = 0 \cdot 7\, I_0 + 2 \cdot 3 \log h — 2 \cdot 0$$

Gutenberg and Richter (1956) had earlier given several formulae to evaluate focal depth from

TABLE 1

Date Feb 1967	Epicentre		Origin time		Depth (km)		Magnitude	
	USCGS	USSR	USCGS	USSR	USCGS	USSR	USCGS	USSR
20	33·69°N $\pm$2·6 km 75·28 E	33·6°N 72·5 E	151839·9	151840	24$\pm$8·8	—	5·7	5½
21	33·61°N $\pm$6·3 *L, s* 75·33 E $\pm$4·3 km	33·5 N 75·4 E	123744·5	123744·0	31$\pm$15·2	—	5·1	5

macroseismic observations. They used the radius of perceptibility r of the shock.

The two formulae given by them are—

$$6 \log (r/h) = I_0 - 1\cdot5$$

$$3\cdot6 \log (r/h) = M - 2\cdot2$$

The magnitude of the main shock as determined by the records of standard Wood Anderson seismographs of Pong and Delhi comes out to be 5·5 as against the USCGS determination of 5·7 and USSR value of 5½. No observations are available for the radius of perceptibility and so Gutenberg's formulae could not be used. The maximum intensity I_0, as described earlier, did not exceed VI M.M. The two formulae given by Karnik and Sheblin give the values 31·6 and 27·5 km respectively for the depth of focus, an average of nearly 30 km. These considerations as well as the evidence provided by the data of near stations particularly show that the depth of focus was about 25 km.

4. Direction of faulting in the main shock of 20 February

Out of a total of 140 observations of the P-arrival times available for the main shock of 20 February from published bulletins, 51 observations gave the direction of the onset of the P-wave from the vertical seismograms. These observations have been utilised for determining the direction of strike dip and motion along the fault by using a method due to Byerly (1938). For the Indian stations, the original seismograms were seen. These observations are given in Table 2 along with the azimuth of the observing stations, their distance from the epicentre and the calculated extended distance. In Fig. 2 the observations of compressions and dilatations are plotted on a stereographic projection (with the anti centre as the pole of projection) at the appropriate azimuth and extended distance. It will be seen that out of a total of 51 observations only 6 stations reported dilatation and the remaining compression. Two out of the six observations reporting dilatations were inconsistent and could

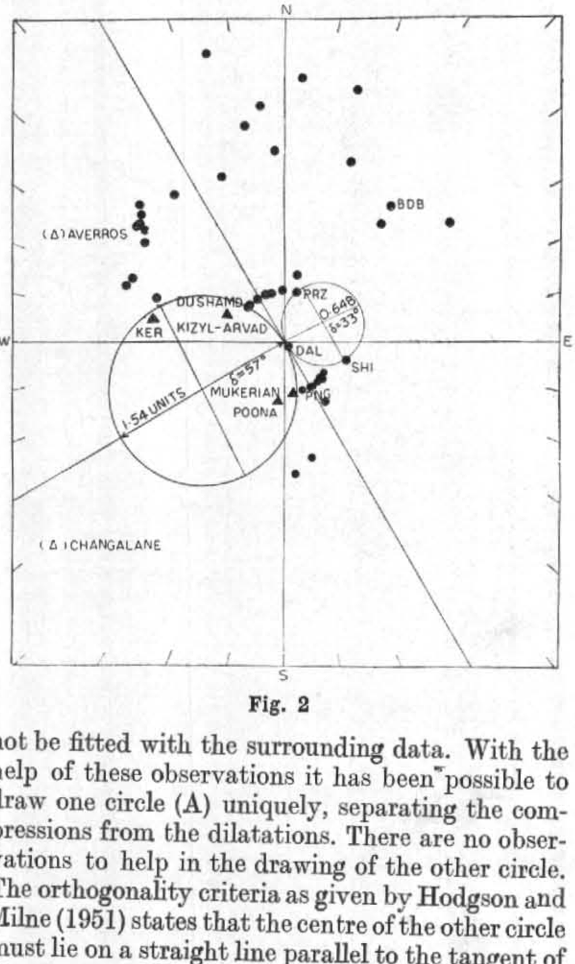

Fig. 2

not be fitted with the surrounding data. With the help of these observations it has been possible to draw one circle (A) uniquely, separating the compressions from the dilatations. There are no observations to help in the drawing of the other circle. The orthogonality criteria as given by Hodgson and Milne (1951) states that the centre of the other circle must lie on a straight line parallel to the tangent of the first circle at the origin and separated from it by a distance equal to $1/4r$ units, where r is the radius of the first circle. This limitation makes the circle B as the only possibility, inspite of the fact that there are no observations available within the circle. If the radius of this circle is reduced, the centre will not lie on the required line, and if it is increased the available observations of compression will become inconsistent. According to theory either of the circles A or B can represent the fault plane. If the circle A represents the

TABLE 2

S. No.	Station	△ (Deg)	Z	Comp. or Dir	Extended distance
1	Dalhausie	1·30	154	C	·021
2	Pong	1·85	161	C	·048
3	Mukerian	1·72	171	D	·043
4	Jwalamukhi	2·02	153	C	·054
5	Rohtak	4·98	163	C	·151
6	Dehradun	4·10	146	C	·127
7	Sonepat	5·0	162	C	·151
8	Delhi	5·27	161	C	·158
9	Khorongan	6·2	316	C	·188
10	Garm	6·7	326	C	·198
11	Dushambe	7·2	315	C	·217
12	Fergana	7·3	338	C	·217
13	Andijan	7·5	346	C	·225
14	Tashkent	9·0	329	C	·274
15	Frunze	9·2	358	C	·280
16	Przhavlak	9·2	16	C	·280
17	Quetta	7·9	246	C	·23[illegible]
18	Chatra	12·4	126	C	·383
19	Bokaro	13·2	136	C	·409
20	Poona	15·0	187	D	·467
21	Kizyl-Arvat	16·2	297	D	·510
22	Shillong	16·5	115	C	·520
23	Semipalatinsk	17·2	10·5	C	·550
24	Vizag	17·6	145	C	·572
25	Kodaikanal	23·5	175	C	1·065
26	Madras	21·1	166	C	·973
27	Kermendes	23·3	279	D	1·075
28	Tabriz	23·8	289	C	1·092
29	Irkutsk	28·0	39	C	1·210
30	Ankara	34·3	293	C	1·336
31	Bodayabo	35·6	37	C	1·357
32	Kishinev	37·3	306	C	1·382
33	Cine	38·2	290	C	1·394
34	Puekovo	39·3	324	C	1·409
35	Apatity	41·6	338	C	1·439
36	Uzhgorad	41·8	307	C	1·441
37	Bld	44·7		C	1·484
38	Prague	46·8	309	C	1·518
39	Pruhonice	46·8	309	C	1·518
40	Tiksi	47·2	20	C	1·527
41	Kheis	47·6	357	C	1·534
42	Moxa	48·6	310	C	1·554
43	Bensberg	51·4	311	C	1·612
44	Yuzno Sakhalin	51·6	54	C	1·618
45	Nord	57·2	350	C	1·764
46	Alert	62·2	354	C	1·894
47	Averros	67·0	296	D	2·035
48	Barrow	69·1	16	C	2·097
49	Mould Bay	70·0	14	C	2·123
50	Ghangalane	72·3	220	D	2·197
51	Forbishbay	78·7	344	C	2·426

fault plane, then faulting took place along a reverse (thrust) fault striking N 330°E and dipping at an angle of 57° towards N 240°E. The direction of motion was almost along the direction of dip. If the circle B represents the fault plane, then it will represent a thrust fault striking N 336°E and dipping at an angle of 33° towards N 66° E.

The strike direction is in general agreement with the strike direction af major thrust faults in the region. For example the Panjal thrust and the Murray thrusts striking in a NW-SE direction was located about 45 km to the SSE of the epicentre location. These thrusts are normally dipping in a northeasterly direction and it would thus appear that the circle B would represent the correct solution. It would be shown later that this finding is not consistent with the location of the hypocentres of the aftershocks which unmistakably point out to the fact that the circle A represents the correct solution. It may be mentioned here that the depth of focus of the main shock has been estimated to be about 25 km, and on the basis of this, the depth of focus of most of the aftershocks has been found to be between 15 and 35 km; the shocks with larger depth of focus have all been located to the west.

5. Aftershocks (Epicentral parameters)

Between 20 February and 5 April a total of 80 shocks, including the main shock were recorded by the network of observatories in the Beas and Sutlej Link project areas. At least one foreshock was clearly recorded about an hour before the occurrence of the main shock. Out of the 80 shocks, 43 were recorded by four or more stations, 23 by three stations and 14 by only one or two stations. Epicentral parameters could be determined in the case of 64 shocks only, which were well recorded by 3 or more stations. As mentioned earlier all the observatories are located to the southeast of the epicentral region with practically no control from any other direction. Lack of knowledge about the regional wave velocities and crustal structure added further difficulties in assessing the parameters accurately. The following method was adopted in locating the hypocentres. Two shocks, namely, the main shock at 1519 hours GMT on the 20 Feb and the principal aftershock at 1238 GMT on the 21 Feb were recorded at a large number of observatories throughout the world and their epicentre, origin time, and depth of focus was determined by USCGS and the seismological service of USSR. These shocks were also well recorded by Indian observatories located within $\triangle = 10°$ of the epicentre. On the assumption of a three layered model, approximate, travel time curves were drawn for the various crustal phases from which the wave velocities, and the crustal structure could be found. Since these shocks occurred at a depth of approximately 25 km, travel time curves were also drawn for depths ranging from 0 to 35 km at 5 kilometres intervals. These travel time curves were then used to locate the hypocentres of the aftershocks. After locating the hypocentres, travel times of the different crustal phases were again plotted and a fresh assessment was made of the wave velocities and crustal structure. Using the revised travel times the location of hypocentres and the assessment of the origin time was repeated.

Since the observatories are all located to the southeast of the epicentre the various arcs drawn for epicentral location tended to graze each other causing considerable uncertainty in the assessment of the longitude of the epicentre. Several checks had therefore to be provided. For example the first arrival at Mukerian was always earlier by about a second than at Pong, located about 30 km to the east. Only in a very few cases the first arrival was the same, and in no case was it later. This ensured that all the epicentres were to the west of the perpendicular bisector of the straight line joining Mukerian and Pong. Similar checks were provided with respect to other pairs of stations where the first arrival was the same type of wave, such as at Pong and Jwalamukhi, where the first arrival was invariably *Pn*. The shocks could usually be classified into two categories namely those in which the difference in time between the first arrival at Nurpur and Dalhousie was 4 to 5 seconds and the others where this difference ranged between 2 and 3 seconds. In the first case, the first arrival was identified as $\overline{P}$ at Dalhousie and in the second category as *Pn*. The hypocentre of the first category shocks lay within the granitic layer while those of the second category below the granitic layer. The distance of Jwalamukhi from the epicentre was such that the first arrival was always *Pn*. The difference in arrival times of the first arrival at Jwalamukhi and Dalhousie could thus be used for the assessment of focal depth. The depth was usually 25 km or less if this difference was from 11 to 13 seconds and more if the difference was less than 11 seconds. The arcs drawn from Dalhousie and Jwalamukhi would never intersect unless the depth was properly chosen. With these precautions it is considered that the epicentral parameters have been located with an accuracy of $\pm$ 5 km, the basic assumption being the correctness of the epicentral parameters of the master shocks.

Table 3 gives the epicentre, origin time, depth of focus and the magnitude of all the shocks. The magnitude of the shocks was mainly computed

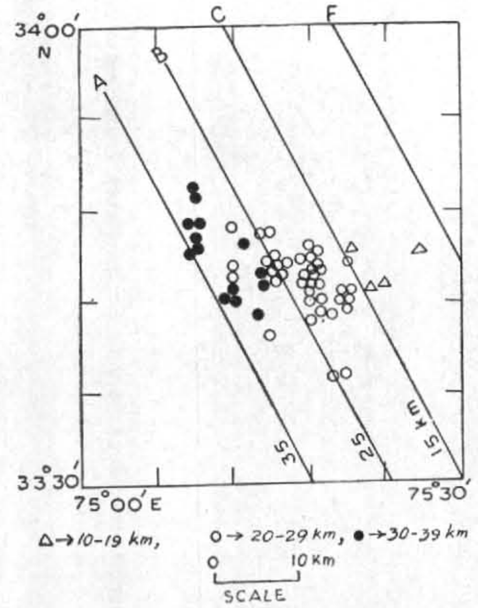

Fig. 3

from the records of the standard Wood Anderson seismograph operating at Pong. The magnitudes were further checked with the records of Hagiwara seismographs operating at Jwalamukhi. The assessment usually agreed within $\pm 0{\cdot}2$ units.

In Fig. 3 the epicentres of shocks, thus determined have been plotted. Solid circles denote those shocks whose depth is between 30 and 35 km, while hollow circles represent shocks whose depth of focus ranged from 20 to 29 km. The triangles represent shocks with depth ranging from 10 to 19 km. It may be noted that the epicentre of the main shock lies about 7 km to the south of the cluster of aftershocks, indicating that the activity shifted northwards after the main event. Shocks with deeper hypocentres are located to the west and the shallower ones to the east, showing that the fault plane along which movement took place was dipping in a westerly direction. Line B in Fig. 3 is an imaginary line passing through the hypocentre of the main shock and parallel to the strike of the fault plane as determined earlier (N 330°E) at a depth of 25 km. Considering that the dip of the fault plane was 57°, similar lines at depths of 35 and 15 km are shown by lines A and C. The line F is the imaginary intersection of the fault plane with the horizontal surface.

Fig. 4 is a vertical section perpendicular to the strike of the fault. All the epicentres plotted in Fig. 3 have been projected on this vertical plane at the appropriate depth. The slanting line represents the line of intersection of the fault plane with the vertical. It may be seen that all the hypocentres

TABLE 3

Sl. No.	Origin Time (h m s)	Epicentre Lat. (°N)	Epicentre Long. (°E)	Depth (km)	t (hr)	Mag.	$E \times 10^{13}$ (erg)	$E^{\frac{1}{2}} \times 10^{6}$ (erg)$^{\frac{1}{2}}$	$S \times 10^{6}$ (erg)$^{\frac{1}{2}}$	Remarks
					20 February 1967					
1	14 23 45·9	33·65	75·33	20	—0·91	4·8	43650	661	—	Fore shock
2	15 18 39·5	33·61	75·33	25	0	5·5	794300	2818	—	Main shock
3	15 25 —			—	0·1	4·3	5495	234	234	Record mixed up with other shocks
4	15 30 1·5	33·42	75·54	5	0·2	4·0	1585	126	360	
5	15 30 13·0	33·77	75·25	25	0·2	4·2	3631	191	551	
6	15 39 45·3	33·74	75·25	25	0·35	4·7	28840	537	1088	
7	15 46 6·0	33·75	75·25	25	0·45	3·4	131·8	36	1124	
8	15 53 57·9	33·66	75·25	25	0·58	4·0	1585	126	1250	
9	16 23 49·5	33·72	75·23	30	1·1	2·3	1·38	3·7	1253·7	
10	16 27 20·8	33·72	75·12	>35	1·13	3·3	87·10	29·5	1283·2	
11	16 37 12·0	33·68	75·23	30	1·32	3·1	38·02	19·5	1302·7	
12	16 41 31·3	33·73	75·23	30	1·4	3·0	31·62	15·8	1318·5	
13	17 40 50·0	33·76	75·30	20	2·35	3·9	1047	100	1418·5	
14	18 04 45·9	33·73	75·25	25	2·77	3·5	199·5	44·7	1463·2	
15	18 15 44·3	33·73	75·25	25	2·95	3·2	57·54	24·0	1487·2	
16	18 19 47·0	33·70	75·20	30	3·0	2·6	4·79	6·9	1494·1	
17	18 27 36·6	33·68	75·30	20	3·15	2·3	1·38	3·7	1497·8	
18	18 32 37·5	33·77	75·15	35	3·23	3·1	38·02	19·5	1517·3	
19	18 33 37·7	33·77	75·15	35	3·25	3·3	87·10	29·5	1546·5	
20	18 38 28·3	33·83	75·15	35	3·3	3·3	87·10	29·5	1576·3	
21	19 28 47·5	33·72	75·38	15	4·17	3·9	1047	100	1676·3	
22	19 52 13·7	—		—	4·73	2·6	4·79	6·9	1683·2	Data discrepant epicentre not determinable
23	21 32 3·2	33·77	75·15	35	6·22	2·8	10·96	10·0	1693·2	
24	22 56 18·6	33·78	75·15	35	7·63	3·4	131·8	36·3	1729·5	
25	23 22 9·2	33·77	75·21	30	8·07	2·4	2·09	4·6	1734·1	
26	23 59 42·8	33·75	75·15	35	8·7	2·8	10·96	10·0	1744·1	
					21 February 1967					
27	01 20 4·0	33·77	75·27	25	9·0	4·5	12590	355	2099·1	
28	01 29 38·0	—		—	9·2	2·6	4·79	6·9	2106·0	
29	01 49 44·0	33·68	75·36	25	9·5	3·3	87·1	29·5	2135·5	
30	03 34 19·4	—		—	12·3	2·9	16·60	12·9	2148·4	
31	04 28 48·1	33·74	75·35	20	13·2	3·1	38·02	19·5	2167·9	
32	05 18 32·0	33·73	75·40	15	14·0	2·5	3·16	5·6	2173·5	
33	06 02 1·2	33·77	75·20	27	14·72	3·3	87·10	29·5	2203·0	
34	07 20 30·0	33·78	75·15	35	16·03	3·1	38·02	19·5	2222·5	
35	09 50 2·0	33·75	75·45	10	18·52	4·2	3631	191	2413·5	
36	11 14 51·0	33·71	75·36	20	19·93	4·5	12590	355	2768·5	
37	11 29 21·2	33·68	75·31	25	20·18	4·7	28840	537	3305·5	
38	12 37 44·0	33·61	75·33	25	21·32	5·2	229100	1514	4819·5	
39	14 44 8·2	—		—	23·42	2·8	10·96	10·5	4825·0	
40	15 42 0·0	33·70	75·35	20	24·38	3·4	131·8	36·3	4861·0	
41	17 24 31·0	33·72	75·20	25	26·1	2·8	10·96	10·5	4871·5	
42	17 30 43·0	33·72	75·30	20	26·2	3·6	302·0	55·0	4926·5	
43	18 18 44·6	33·82	75·15	35	27·0	3·4	131·8	36·3	4962·8	
44	18 30 51·0	33·73	75·30	20	27·2	3·2	57·54	24	4986·8	
45	21 32 42·1	33·72	75·30	20	30·23	3·7	457·1	67·6	5054·4	
46	22 20 30·0	—		—	31·03	2·7	7·24	8·5	5062·9	
47	23 28 5·7	33·75	75·30	20	32·17	3·3	87·10	29·5	5092·4	

TABLE 3 *(contd)*

Sl. No.	Origin Time h m s	Epicentre Lat. (°N)	Long. (°E)	Depth (km)	t (hr)	Mag.	$E \times 10^{13}$ (erg)	$E^{\frac{1}{2}} \times 10^{6}$ (erg)$^{\frac{1}{2}}$	$S \times 10^{6}$ (erg)$^{\frac{1}{2}}$	Remarks
					22 February 1967					
48	04 36 24·8	33·70	75·35	20	37·3	3·2	57·54	24	5116·4	
49	04 39 —	—		—	37·35	2·7	7·24	8·5	5124·9	
50	09 28 3·9	33·73	75·30	20	42·15	3·0	31·62	15·8	5140·7	
51	18 34 57·0	—		—	51·27	3·7	457·1	67·6	5208·3	
52	20 48 10·8	—		—	53·48	3·2	57·54	24	5232·3	
					23 February 1967					
53	01 16 43·0	—		—	57·97	3·2	57·54	24	5256·3	
54	08 30 7·7	33·70	75·20	35	65·18	3·2	57·54	24	5280·3	
55	11 01 34·0	33·65	75·35	20	67·72	3·1	38·02	19·5	5299·8	
56	11 51 4·5	33·75	75·25	25	68·53	2·9	16·60	12·9	5312·7	
57	19 14 11·1	33·70	75·35	20	75·93	2·8	10·96	10·5	5323·1	
					24 February 1967					
58	00 17 34·8	33·70	75·32	25	80·98	4·5	12590	355	5678·2	
59	04 50 10·7	33·70	75·35	20	85·53	2·6	4·79	6·9	5685·1	
60	04 43 39·6	—		—	86·42	2·4	2·09	4·6	5689·7	
61	10 20 56·8	33·70	75·13	25	91·03	4·1	2399	155	5844·7	
					25 February 1967					
62	23 37 50·1	33·75	75·30	25	104·32	3·2	57·54	24	5868·7	
63	08 59 20·0	33·72	75·30	20	113·68	3·7	457·1	67·6	5936·3	
64	12 41 30·2	33·72	75·30	20	117·38	3·6	302·0	55	5991·3	
65	12 48 40·1	—		—	117·50	2·7	7·24	8·5	5999·8	
66	19 37 37·7	33·75	75·30	20	124·3	3·5	199·5	44·7	6044·5	
					27 February 1967					
67	17 18 13·5	33·75	75·25	25	170·0	2·4	2·09	4·6	6049·1	
					1 March 1967					
68	05 10 —	—		—	205·87	4·3	54·95	234	6283·1	
69	08 08 35·0	—		—	208·18	2·8	10·96	10·5	6293·6	
					3 March 1967					
70	04 26 35·4	—		—	253·13	3·0	31·62	15·8	6309·4	
71	04 32 3·3	—		—	253·23	3·5	199·5	44·7	6354·1	
72	19 40 58·1	33·75	75·25	25	268·37	2·9	16·60	12·9	6367·0	
					4 March 1967					
73	08 19 6·5	33·73	75·30	20	281·02	2·7	7·24	8·5	6375·5	
					5 March 1967					
74	17 28 12·5	33·72	75·30	20	314·17	3·6	302·0	55	6430·5	
75	23 08 23·4	33·72	75·20	25	319·83	3·3	87·10	29·5	6460·0	
					7 March 1967					
76	15 33 11·9	—		—	360·25	2·6	4·79	6·9	6466·9	

TABLE 3 (*contd*)

Sl. No.	Origin Time h m s	Epicentre Lat. (°N)	Epicentre Long. (°E)	Depth (km)	t (hr)	Mag.	$E \times 10^{13}$ (erg)	$E^{\frac{1}{2}} \times 10^{9}$ (erg)$^{\frac{1}{2}}$	$S \times 10^{6}$ (erg)$^{\frac{1}{2}}$	Remarks
					10 March 1967					
77	01 33 23·5	33·75	75·35	15	418·25	3·2	57·54	24	6490·9	
					20 March 1967					
78	09 17 47·0	—		—	656·98	2·9	16·60	12·9	6503·8	
					26 March 1967					
79	22 38 36·0	33·77	75·25	25	823·33	4·0	1585	12·6	6629·8	
					5 April 1967					
80	19 27 35·0	33·77	75·17	30	1060·15	3·5	199·5	44·7	6674·5	

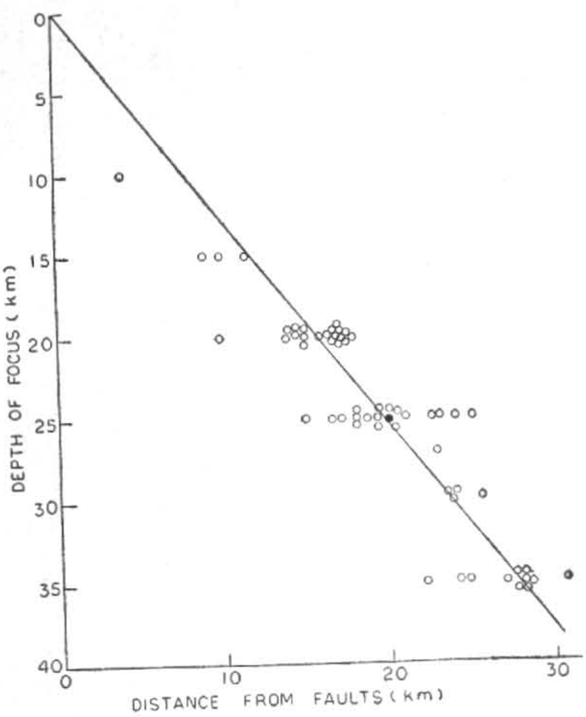

Fig. 4

are located close to the fault line. The scatter represents errors in the location of the hypocentres.

6. Magnitude frequency relationship in the aftershock sequence

Omori first observed that the decay of aftershock activity can be represented by the hyperbolic relation—

$$N(t) = \frac{K_1}{t + C_1}$$

where $N(t)$ is the number of aftershocks in a given interval of time t after the mainshock, and K_1 and C, are constants. This relationship holds good in most of the cases. In order to make the relationship more general Utsu (1955) modified the formula to the following—

$$N(t) = \frac{K_2}{(t + C_2)^p}$$

which becomes the same as Omori's relationship if $p = 1$. In the present case, out of a total of 80 shocks including one foreshock and the mainshock, 28 occurred within the first 24 hours of the occurrence of the main shock, 12 during the next 24 hours and the frequency decreased to 6, 4, 4 and 1 during the subsequent twenty four hourly intervals. After this the frequency varied between 0 and 2. With the exception of a few shocks the majority of aftershocks registered by the Beas network of observatories had a magnitude above 2·5. Many more would have been registered if there was a sensitive station close to the epicentral region. This however is true for the entire period of aftershock occurrence and it can be presumed that the number of unregistered smaller events was proportional to the number of events of higher magnitude, and the decay of aftershock activity can be represented correctly even if the number of shocks considered is above a minimum threshold magnitude. This is actually found to be the case for many aftershock sequences.

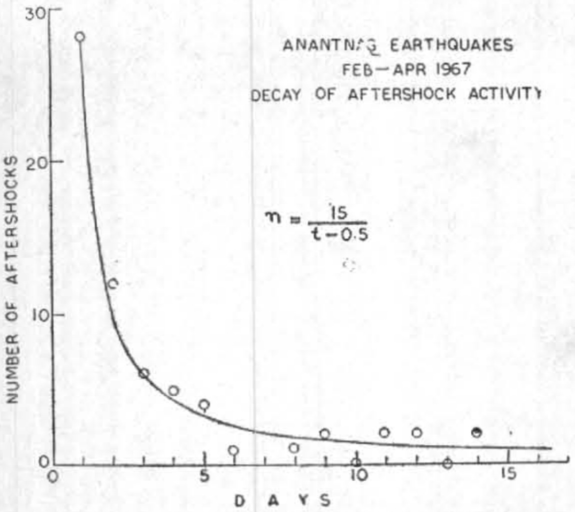

Fig. 5

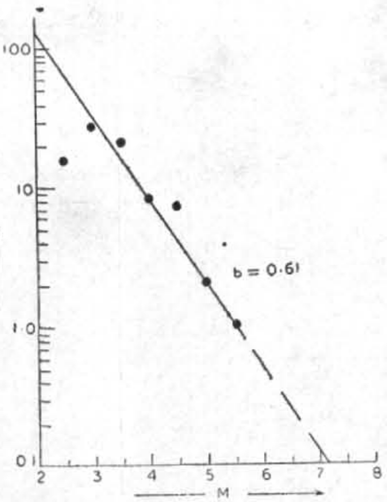

Fig. 6

The decay curve for the present sequence is shown in Fig. 5. It has been found that the following relationship based on Omori's simple formula is quite adequate to represent the decay curve.

$$N(t) = \frac{15}{t - 0 \cdot 5}$$

where, t is in days. This would mean that even after more than 100 days the average activity would still be about one aftershock per week having a magnitude of over 2·5.

Gutenberg and Richter (1954) had shown that the frequency of aftershocks of different magnitudes can be represented by the following relation ship :

$$\log N = a - bM$$

where, N is the number of aftershocks of magnitude M occurring within a specified time interval and a and b are constants. The same relationship has been found to hold good for earthquakes (other than aftershocks) occurring in a given area. Various interpretations have been given to the physical significance of the constants in order to use the above relationship for seismicity studies. Some seismologists believe that b is a universal constant while the observational evidence collected in many parts of the world shows that the value of b differs from region to region and varies when determined from aftershock sequences in the same region.

It has also been shown that b varies with the depth of focii of shocks. In a recent study Eaton *et al.* (1970) observed in the case of the aftershock sequence of the 1966 Park field Cholame earthquake in California that the value of b decreased regularly from 1·03 for 2 to 4 km interval depth to a minimum of 0·61 for the 8-10 km depth interval and then increased to 0·95 for the 12-14 km interval. They found it difficult to judge whether this variation in b was due to some accidental irregularity in the slip surface geometry or due to some local peculiarity in the rock types in contact across the slip surface or due to the variations in the properties of rock with depth.

The frequency distribution of aftershocks in the various magnitude intervals which occurred during the period 20 February to 15 April is given in Table 4.

In Fig. 6 the number of shocks belonging to each class have been plotted against the average magnitude on a semi log paper. The points corresponding to the magnitude intervals 2·3–2·7 fall very much below the line, as a large number of shocks belonging to this class have escaped detection. The value of b can also be calculated according to the relation —

$$b = \frac{\cdot 4343}{M_{av} - M_{min}} \text{ due to Aki (1965)}$$

The average value of the magnitude for 64 shocks above the magnitude 2·8 comes out to be 3·5 and gives a value of 0·6 which agrees well with the value of 0·61 obtained graphically. This value of b is lower than the average value of 0·92 obtained for the Himalayan region (Tandon and Chatterjee 1970) from seismicity considerations.

Drakopoulus (1968) observed that for the whole of Greece areas as well as for smaller sectors, the average value of b was 1·0, and attributed this effect to the non-dependence of b on the tectonic and geological setting of the focal region. He attributed the variation of b to the variations in focal depth. The mean value of b for very

TABLE 4

	Magnitude range						
	2·3–2·7	2·8–3·2	3·3–3·7	3·8–4·2	4·3–4·7	4·8–5·2	5·3–5·7
Number of shocks	15	26	21	8	7	2	1

shallow shocks was found to be 1·13 while for shocks occurring at depths between 35 and 70 km the mean value of b was 0·72.

Mogi (1962) attributed the variations in the value of b to the variations in the tectonic structure of the region. He showed experimentally that the constant b depends upon the homogeneity of the material in the seismic region and on the distribution of the applied stress. The value of b increased with increase in heterogeneity and decrease in the symmetry of the stress field.

Karnik (1964) made a very detailed study of the Magnitude—Frequency relationship and seismic activity in different regions of the European area. He divided the whole area into 39 seismic zones, keeping in view the homogeneity of data and the tectonic similarity within each region. The results indicated that the European area could be divided into two broad divisions according to the b values obtained, namely, $b=0 \cdot 95 \pm 0 \cdot 05$ and $b=0 \cdot 75 \pm 0 \cdot 1$. The higher b-value is for the Alpine folding system and the lower for the belt extending from the Balkans over north Turkey to east Turkey. For the Caucasius he found an exceptionally large value, *viz.*, $b=1 \cdot 1$. He found very low b values for earthquakes with intermediate depths of focii. The most important factor influencing the value of b is the degree of homogeneity of the focal region. Earthquakes having smaller depths ($h=20$ km) originate in the upper crust which is highly fractured and heterogeneous. The value is around 0·9, while for regions where shocks occur at higher depths (near the base of the crust) the b values usually are 0·5-0·8.

Miyamura (1962) published a world wide survey of b values. He found that b varies from 0·4–1·8. High b values of 1·0–1·8 have been found for the Circum-Pacific and Alpide belts. For continetal rift zones and platforms the b value is 0·6–0·7 while very low b values are found in shield areas.

The low b value, *viz.*, 0·6 obtained in the present investigations can be attributed to the comparatively larger depth of the focii of the aftershocks and also due to the homogeneity of fault zone at this depth. The lower value also indicates a lower seismogenic potential compared to other Himalayan areas for which the b values have been found to be larger.

7. Strain release characteristics of the aftershock sequence

Benioff (1951) had shown that the aftershocks in an earthquake are produced by the creep of the fault rocks. Strain accumulates in the fault rocks over a larger number of years. The recovery of the elastic strain takes place during the principal shock, while the creep strain is relieved during the aftershocks. The creep may be either compressional or shearing or both. When both occur the compressional phase occurs first followed after a time interval by the shearing phase. Benioff also showed that the strain released during an earthquake is proportional to the square root of the radiated energy. In an earthquake sequence the value of $E^{1/2}$ for each event represents the creep strain recovery of the fault rock. If the recovery during all the preceding aftershocks is summed up, it represents the total creep recovery upto that time. A graph of this summation against time gives the creep strain recovery curves for the rock.

The energy of an earthquake can be calculated from its magnitude by means of various empirical formulae, given by Gutenberg and Richter (1942) and other authors. For the purpose of this study the following simple formula has been used although it has been modified later (Gutenberg and Richter 1956).

$$\log E = 9 + 1 \cdot 8\, M$$

$$\log E^{1/2} = 4 \cdot 5 + 0 \cdot 9\, M$$

The values of E, $E^{1/3}$ for all the shocks studied in this paper are also given in Table 3. Fig. 7 is a graph in semi-log coordinates in which the time has been plotted as abscissa and the value of $E^{1/2}$ as ordinate. The time is measured in hours from the origin time of the main shock. It may be seen that the points fall on two distinct curves. The first one beginning with the first aftershock

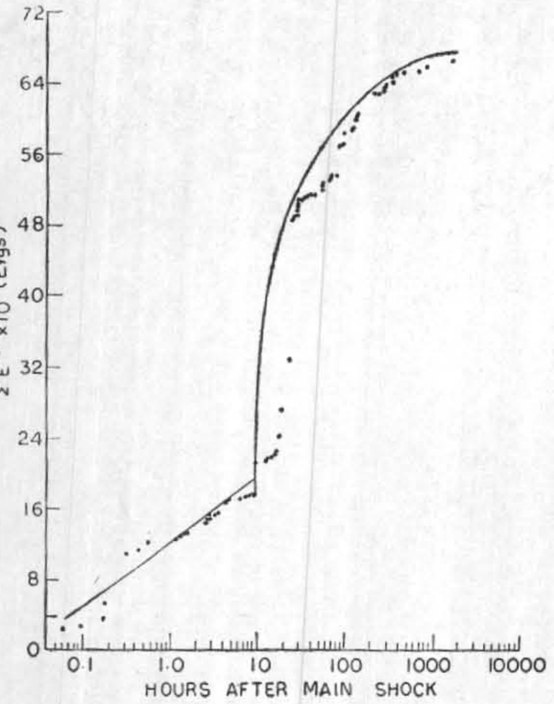

Fig. 7

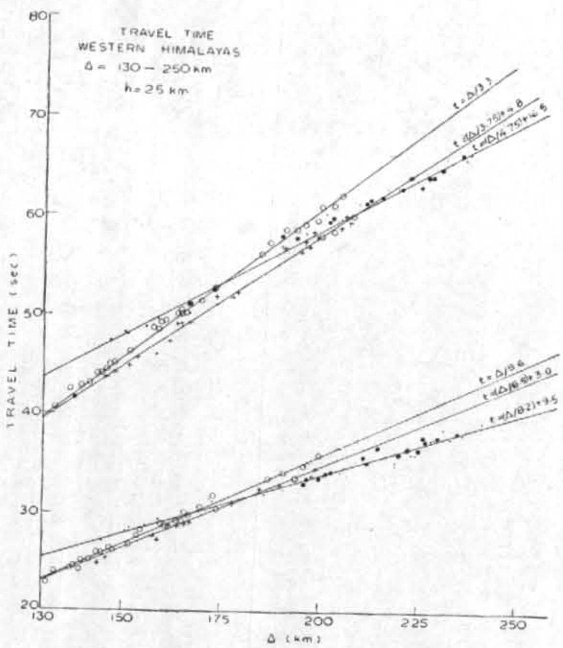

Fig. 8

is a straight line upto about 9 hours after the main shock. The curve then rises steeply on accelerated rate of creep recovery, and then diminishing exponentially. The first part of the curve represents the recovery of compressional creep while the second represents recovery of shearing strain.

Benioff (1951) had also shown that the energy W_2 stored as creep strain is related to the energy W stored in elastic strain, by the following relation —

$$W_2 = W\,(S_{\max})/E^{1/2}$$

where S_{max} is the total of the strain rebound increments for aftershocks and E, the energy released in the main shock.

In the present case —

$$S_{\max} = 6674 \times 10^6\ (\text{ergs})^{1/2}$$
$$\text{and}\quad E^{1/2} = 2818 \times 10^6\ (\text{ergs})^{1/2}$$

Therefore the energy stored in the creep strain was $6674/2818 = 2\cdot3$ times the energy stored as elastic strain.

Now the energy of the main shock is $7\cdot94 \times 10^{18}$ while the total energy released in all the aftershocks is $3\cdot58 \times 10^{18}$, which shows that the energy released in the aftershocks was $0\cdot47$ times the energy released in the main shock inspite of the fact that the stored creep strain was $2\cdot3$ times that of the elastic strain. This shows that only $0\cdot47/2\cdot3$ part of the creep energy or about 20 per cent was radiated as seismic energy, the rest being dissipated as heat.

Fig. 3 shows that the total area covered by the main shocks and the aftershocks is about 300 sq. km. Since the bulk of hypocentres were confined within a thickness of about 20 km of the fault rock, the volume of the strained rock can be taken as $300 \times 20 = 6000$ cu. km or 6×10^{18} cc. The energy released in the mainshock was $3\cdot58 \times 10^{18}$ ergs, and hence the mean energy density of the elastic energy just preceding the earthquake was $3\cdot58 \times 10^{18}/6 \times 10^{18}$ or $0\cdot6$ erg/cc.

According to Benioff (1958) the elastic strain ϵ is connected with the energy release and the elastic constant μ by the relation —

$$\epsilon^2 = \frac{2\,E_m}{\bar{\mu}\,V}$$

Taking $\bar{\mu} = 6 \times 10^{11}$ dynes/cm², $E_m/V = 0\cdot6$ as determined above,

$$\epsilon^2 = \frac{0\cdot6 \times 2}{6 \times 10^{11}} = 2 \times 10^{-12}$$

$$\text{or}\ \epsilon = 1\cdot4 \times 10^{-6}$$

Since the stress $\sigma = E\bar{\mu}$ the value of the elastic stress just before the main shock was $1\cdot4 \times 10^{-6} \times 6 \times 10^{11} = 8\cdot4 \times 10^{5}$ dynes/cm².

As shown earlier the Creep strain was $2\cdot3$ times the elastic strain. Hence the total average value of strain before the occurrence of the main shock must have been $4 \cdot 6 \times 10^{-6}$.

The total relative slip along the fault plane in the main shock can also be calculated from the value of the purely elastic strain calculated above, *viz.*, $1 \cdot 4 \times 10^{-6}$. The total area of aftershocks being 300 sq. km the average length of the fault involved can be taken to be $(300)^{1/2}$ km $= 17 \cdot 3$ km. Hence the total slip along the fault was of the order of $(17 \cdot 3 \times 10^5 \times 1 \cdot 4 \times 10^{-6})$ cm or $2 \cdot 4$ cm. In the absence of any displacement observed along the fault trace, it is difficult to verify this result, but considering the low magnitude of the shock, the value does not appear to be unreasonable.

8. Wave velocities and crustal structure of the Western Himalayas

Fig. 8 shows the final travel times of the various crustal phases for a depth of focus of 25 km on the assumption of a three-layered model. The dots denote the observed values for the waves Pn and Sn, the *plus* sign for the P^* and S^* phases and the circles the direct wave $\overline{P}$ and $\overline{S}$ through the granitic layer. The effect of the sediments has been neglected. Thick dots and circled + sign denote the average value of three or more observations. No attempt has been made to obtain a least square fit with the help of a computer but the scatter of points is seldom more than fraction of a second. The straight lines can be represented by the following equations for a depth of focus of 25 km.

$$t_{\overline{P}} = \frac{\triangle}{5.6}$$

$$t_{P*} = 3 \cdot 0 + \frac{\triangle}{6 \cdot 5}$$

$$t_{P_n} = 9 \cdot 5 + \frac{\triangle}{8 \cdot 2}$$

$$t_{\overline{S}} = \frac{\triangle}{3 \cdot 3}$$

$$t_{S^*} = \frac{\triangle}{3 \cdot 75} + 4 \cdot 8$$

$$t_{S_n} = \frac{\triangle}{4 \cdot 75} + 16 \cdot 5$$

where, t is the travel time of the appropriate wave and $\triangle$ is the distance from the epicentre in kilometres.

These values give the following results for the thicknesses of the granitic and the Basaltic layers:

$$\left.\begin{aligned} H_1 &= 29 \text{ km} \\ H_2 &= 29 \text{ km} \end{aligned}\right\} \text{From } P\text{-wave data}$$

$$\left.\begin{aligned} H_1 &= 30 \text{ km} \\ H_2 &= 28 \text{ km} \end{aligned}\right\} \text{From } S\text{-wave data}$$

The average depth of the Moho in the Western Himalayas thus comes out to be about 58 km. No data from deep seismic soundings is available to lend support to this value but earlier inferences from surface wave dispersion studies (Tandon and Chaudhury 1963) gave an average value of the thickness of the crust in the Himalayas as 60 km.

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