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550·34 : 550·346 (548·2)

Recent Mandya earthquakes and the aftershock microseismic activity

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ABSTRACT. Two recent earthquakes near Doddegoudana Koppal in the Mandya district of Mysore have been analysed in detail. To study the aftershock activity in this region, a temporary Geophone station was installed using all available facilities at the seismic array in Gauribidanur. Details of the instrumentation as well as the data reduction are discussed and possible conclusions regarding the microseismic activity drawn.

1. Introduction

Mandya region in Mysore, southern India, has experienced a few small tectonic earthquakes in the past as confirmed from Gauribidanur Array (GBA) records. For instance, in the year 1971, the following three earthquakes originating in the Mandya district were recorded :

Date	*GBA arrival (GMT)*	*Magnitude*
17 Jan 1971	14-00-14.4	4.2
6 Mar 1971	16-24-51.3	4.2
27 Mar 1971	14-48-45.4	4.3

More recently, in May 1972, recurrence of stronger seismic activity in this area resulted in some destruction of property in and around Doddegoudana Koppal village. At less than 18 hours interval, two successive shallow focus earthquakes of almost equal intensity occurred in that village on 16 and 17 May 1972. Our estimate of conventional body-wave magnitude m_b (calculated using the amplitude on the strong-motion channel with magnification only 20,000 at 1 cps) of the first shock was 4.6 and that of the second was 4.5. The events, although of moderate size, seem to have been recorded only at very few stations in the country. For example, GBA, STA (Satara), HYB (NGRI, Hyderabad), KOD (Kodaikanal) and POO (Poona) reported first P while BOM (Bombay), GOA (Goa) and MDR (Madras) could report only the first S arrival of the 16 May event. The event on 17 May was recorded at only a couple of stations (POO reported only S and P was discernible to some extent in the KOD record) other than GBA. Thus, the farthest station recording P signal from atleast the 16 May source happened to be POO at an epicentral distance $\triangle \sim 735$ km.

Perhaps for the first time, in the Mandya region, an attempt was made using the support facilities at GBA to monitor the aftershock activity by installing Geophone stations on temporary basis. The recording was continued for about a month effectively during which period 33 small shocks were registered. A scheme was worked out to calculate local magnitudes for each of the microtremors. Although definite conclusion cannot be drawn on the frequency of occurrence of these smaller events unless recordings are made over an extended period of time, atleast from the available data the seismic activity in the Mandya region appeared to have steadily decreased both in magnitude and time.

2. Basic parameters of the larger events

2.1. *Filtering and phase identification* — Determination of basic parameters of a seismic event such as hypocentre and origin time, using a single station record, requires highly reliable estimates of epicentral distance $\triangle$, focal depth d, and azimuth Z of epicentre at the recording station. While it is possible to compute Z accurately from the knowledge of time-lags across the array and the nature of the first arriving phase can be ascertained by making phase velocity measurements using the same time-lags (Arora 1967, 1970), it is the estimate of $\triangle$ which is generally in error.

TABLE 1

P arrivaltimes of the Mandya events of 16 and 17 May 1972

	Station code name	First phase and its arrival time (GMT)
16 May 1972		
	GBA	iPx 16 : 37 : 06·8
	KOD	iPn 16 : 37 : 20·8
	POO	ePn 16 : 38 : 21·2
17 May 1972		
	GBA	iPx 10 : 00 : 17·8

In the case of local events, $\triangle$ is normally obtained from the observed S-P interval, the accuracy of which calculation depends on identifying the pertinent onset times precisely. For instance, in the present case we found it difficult to pick-out the beginning of S in the background of earlier signals of high frequency (about 8 to 12 cps). The unfiltered or the wide-band records of GBA in fact contained predominant frequencies around 8 cps which represented a general feature common to both the useful signal and the uncorrelated haze due to crustal reverberation and scattering. But by pre-filtering relevant portion of the magnetic tape record through a suitable narrow passband the unwanted high frequency content was suppressed and the desired signals manifested to a reasonably good level of identification though a small loss in signal amplitude occurs in the process. The optimum filter chosen in the present cases was 0-2 Hz band-pass which yielded clear traces of P, S and even the dispersed L_R (Rayleigh wave) on the playouts with pre-set appropriate gain and chart speed. The effect of filtering is illustrated for one case in Fig. 1, and the subsequent use of these filtered records is described in the succeeding section.

The velocity measurements corresponding to first P, over all the available array channels, in both the cases, gave 6.4 km/sec which is characteristics of α_x (velocity of P_x phase) in the crustal structure model for Gauribidanur (Arora 1969, 1971). On the same model, the observed S-P interval indicated that the first arrival in the S group was S_x phase .

2.2. *Hypocentre and origin time* — The wide-band GBA seismograms gave Z=N200° E & Z=N199°E for 16 and 17 May events respectively, and the narrow-band filtered seismograms for both the events gave the following figures :

$$S_x—P_x=15.2 \text{ sec} \qquad L_R—S_x=11.5 \text{ sec}$$

The above values, in the Gauribidanur crustal model, fix $\triangle$=142 km and P_x travel-time of 24.6 sec against $d \approx 6$ km. The Rayleigh wave velocity turns out to be 2.82 km/sec as expected. Accordingly these two shocks appeared to have originated 142 km SW of GBA just close (about 5 km NE) to Doddegoudana Koppal (DGK) village in the Mandya district 24.6 sec earlier than the GBA first arrival time (*see* Table 1 for onset data).

The Precise Epicentre Determination technique described by Arora and Manekar (1969) was attempted on the first event alone, for reasons of nature as well as amount of internally consistent onset time data of other stations, using GBA, KOD and POO times (Table 1). Since the number of data points is insufficient, the results are questionable, in particular the estimate of d corresponding to normalized minimum variance obtained in the source location procedure. The results of this exercise mentioned below, however, seem to agree well with those deduced from the GBA data alone.

Date	:	16 May 1972
Epicentre	:	12.35°N, 77.01°E (12.4°N, 76.8°E)
Origin time	:	16-36-42.16 GMT (16-36-41)
Focal depth	:	7 km

Figures within brackets above refer to the corresponding deductions made by the India Meteorological Department (1972) using available P and S onset readings at 5 stations (KOD, POO, BOM, GOA, MDR).

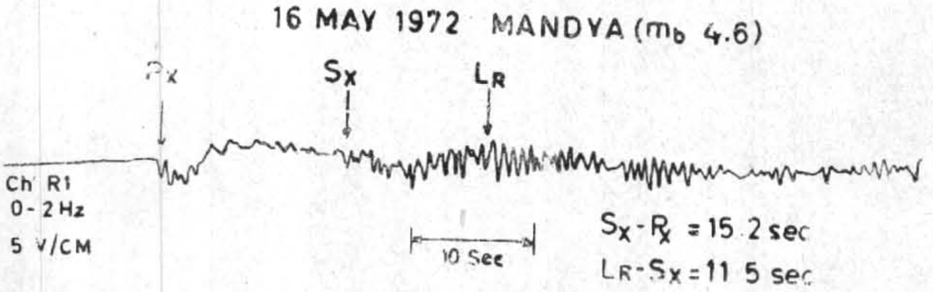

Fig. 1

Magnetic tape playout of Mandya record (16 May 1972) of one of the GBA channels with filters letting 0-2 Hz,
Chart speed 2·5 mm/sec and seismic magnification 73200 at 1 Hz.
Note the clear onsets of P_x, S_x & L_R

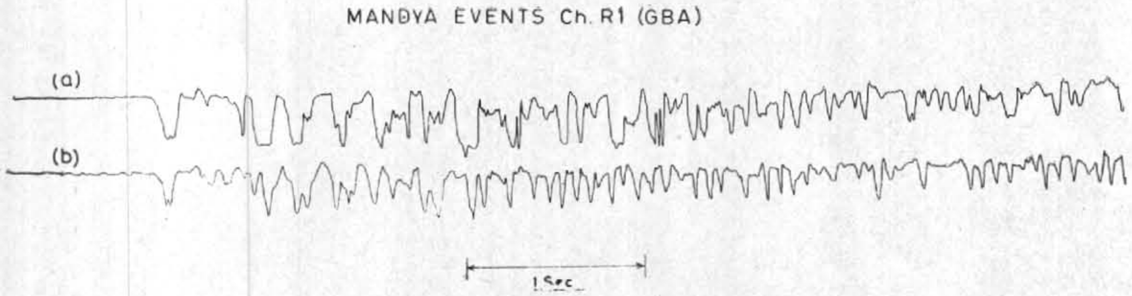

Fig. 2

Wide-band typical playouts of Mandya records; (a) 16 May 1972 and (b) 17 May 1972, taken at 50 mm/sec chart speed from the same GBA channel in both the cases to show the overall spectral similarity

In a comparison of records of the two events from channel to channel over the entire array, although exact pulse-to-pulse similarity could not be established, a great deal of overall similarity in the signal structure was, however, manifested. For illustration, a representative pair of wide-band records of the same channel played out at much faster speed (50 mm/sec) is presented in Fig. 2. It is a fit situation which calls for invoking concept of wave-form matching or spectral similarity since the two sources situated in the same area have the same strength as well. It helps one in concluding that besides the focal regions being same in both the cases, the physical processes taking place at source or the source mechanisms must be almost identical.

The shallow focus model postulated above is further supported by the presence of a good amount of energy in the 1 to 1.5 second surface waves as seen from the wide-band seismograms. The normal dispersion of Rayleigh wave is seen clearly in Fig. 1.

3. Recording and analysis of aftershock microactivity

3.1. *Instrumentation* — About three weeks later than the occurrence of the larger events a temporary Geophone station was set up near an old temple in the DGK village to monitor the aftershock activity in the vicinity of that area. The station initially commenced operation with a single sensor which, after some time, was augmented to a 3-sensor system with the aim of (*i*) providing means for rough location by triangulation, and (*ii*) discrimination against spurious signals of non-seismic origin such as cultural and electronic system noise. The elements positioned at points B, C and D (Fig.3) were spaced such that BC=CD=2 furlongs (402 m).

The field installations were powered from three 12-volt lead-acid batteries locally placed while

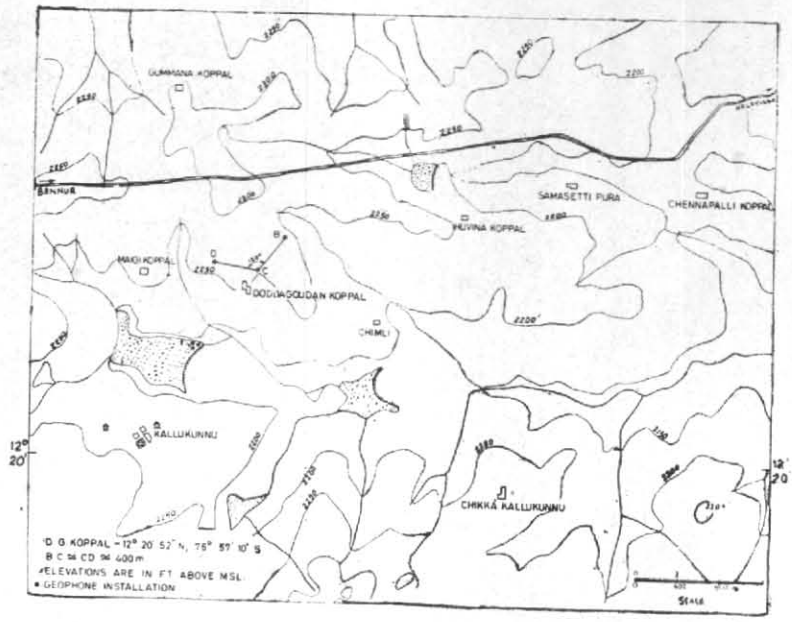

Fig. 3

Location map of the temporary Geophone station at Doddegoudana Koppal. B, C and D are the Geophone installations.

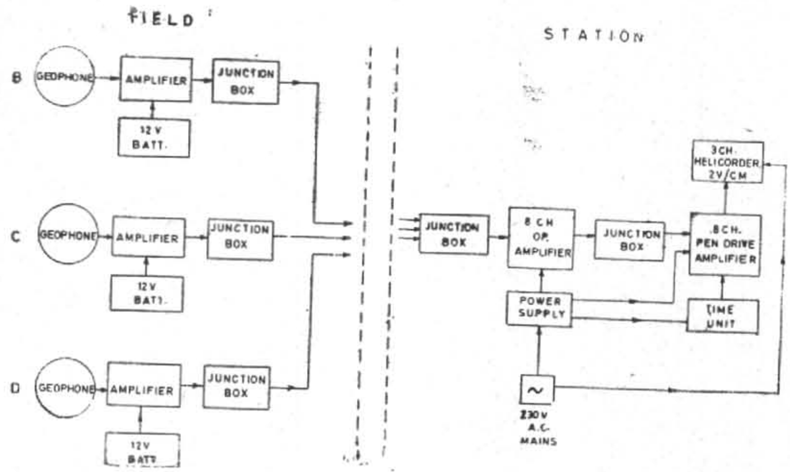

Fig. 4

Schematic block diagram of the field and recording station instrumentation for monitoring the aftershock activity in the Mandya region the instruments in the recording shed were A.C. mains powered. The outputs from the three field sensors, *viz.*, geophones, were initially fed to three head-amplifiers with gain 1.5 K each. The amplified signals were transmitted through special spiral-4 four-core (Singh *et al.* 1969) overhead cable upto the recording site where an 8-channel operational amplifier with electronic gain 20 K further amplified the signals before routing them to the recorder. Time pulses every $\frac{1}{2}$ minute from a local electronic clock were impinged on the seismic trace. The complete set-up is shown in Fig. 4.

Following are the important specifications of the main instruments of the installation :

(*i*) Type of sensor : Hall-sears HS 10 geophone

(*ii*) Output sensitivity of the sensor : 0·323 volts/cm/sec

(*iii*) Geophone system response : Flat between 5 and 300 cps

(*iv*) Type of recorder : 3-channel heated stylus helical recorder

(*v*) Recorder sensitivity : 2 volts/cm

(*vi*) Recording speed : 25 mm/min.

(*vii*) Electronic system gain : 20,000

3.2. *The data* — Mainly owing to frequent power failures at the recording site, the aftershock recordings could not be made continuously uninterrupted. The station, though remaining in existence for a much longer time (9 June to 21 July 1972), actually operated for 32 days during which period 33 small tremors were recorded. Observed duration of these events was on the average about 5 sec and the predominant signal frequency appeared to lie between 20 and 30 cps which is to be expected from localized shocks. Table 2 shows the complete set of microtremor data.

Approximate $\triangle$ was obtained for each microtremor using the following empirical relation :

$$\triangle\ (\text{km}) = 10 \times (S\text{-}P \text{ interval in sec}) \qquad (1)$$

wherein the compressional and the shear wave speeds are assumed to be 5.8 and 3.5 km/sec respectively consistent with the local crustal structure (Arora 1969, 1971). If g_0, G_e and V_r be the output sensitivity of the geophone in volts per unit velocity of ground motion, the electronic gain of the amplifier and the recorder sensitivity in volts/cm respectively, the acutal ground displacement expressed in millimicrons corresponding to $(A)_t$ as maximum recorded zero-to-peak trace amplitude in millimetres is given by

$$(A)_g = \frac{V_r}{g_0\, G_e} \cdot \frac{10^6}{2\pi f} \cdot (A)_t \qquad (2)$$

where f is the signal frequency and the overall system response is unity. Substituting appropriate values of the instrumental constants in equation (2), the actual ground motion at 20 cps (within a few kilometres from hypocentre) can be computed with the help of the following relation—

$$(A)_g = 2.4637\ (A)_t \qquad (3)$$

The relation (3) incidentally indicates an overall system magnification of about 400,000 at the recorded frequencies around 20 cps.

3.3. *Assigning local magnitudes* — Richter (1958) laid down a procedure for assigning magnitudes to seismic events from short distances. The data used by him were a number of small earthquakes in southern California as recorded by stations equipped with identical low-magnification short-period torsion seismographs of the Wood-Anderson type. He defined magnitude M as

$$M = \log A - \log A_0 \qquad (4)$$

where A is the maximum recorded body-wave trace amplitude and A_0 is that for a standard shock called the "zero" shock.

In a practical case, such as the present one, it may be that a Standard instrument (natural period: 0.8 sec, static magnification : 2800, damping coefficient: 0.8 critical) is not deployed and instead some other type of seismometer is used. Under the circumstances, notwithstanding the type of seismographic system involved, use of actual ground motion parameter corresponding to A as well as A_0 becomes necessary. Besides, it would demand knowledge of frequency response of the Standard seismograph as well as the overall response of the actual system used.

We converted Richter's log $(A_0)_t$ values (subscript t denotes recorded trace amplitudes) into actual ground displacements $(A_0)_g$ (subscript g denotes actual ground motion) expressed in millimicrons for various distance ranges. In these conversions, flat response of the Standard instrument above 2 cps (Benioff 1955) was incorporated. Hence the modified expression for local magnitude m_L is given by

$$m_L = \log\,(A)_g - \log\,(A_0)_g \qquad (5)$$

where $(A)_g$, available under the Appendix for direct reference, represents maximum actual ground motion in millimicrons zero-to-peak for an event

TABLE 2

Observation on micro-earth tremors in the Mandya region of Mysore and the local magnitudes assigned to them

S. No.	Onset time (IST)	Δ (km)	$(A)_t$ †	Local magnitude (m_L)
		9 June 1972		
1	18·24	20	0·75	—0·6
2	18·25	25	1·50	—0·1
3	18·31	12	1·00	—0·7
4	18·39	15	1·00	—0·6
5	21·47	5	2·75	—0·3
6*	23·10	—	0·25	—
7*	23·11	—	0·25	—
8*	23·14	—	0·25	—
		10 June 1972		
1	03·20	15	1·00	—0·6
2	05·01	15	1·00	—0·6
3	07·27	12	3·25	—0·1
4*	09·03	—	0·75	—
5	09·18	25	3·50	0·3
6	13·47	5	2·75	—0·3
		11 June 1972		
1*	18·17	—	0·25	—
2*	18·25	—	0·25	—
		14 June 1972		
1	10·54	25	0·75	—0·4
2*	12·12	—	0·50	—
		15 June 1972		
1	08·09	25	1·00	—0·3
		22 June 1972		
1	14·52	20	0·75	—0·6
2	14·56	25	2·00	0·0
		13 July 1972		
1*	12·43	—	—	—
		15 July 1972		
1	10·16	10	0·75	—0·8
2*	15·33	—	—	—
3*	16·15	—	—	—
4*	16·49	—	—	—
5*	23·46	—	—	—
		16 July 1972		
1	09·01	12	0·75	—0·8
		17 July 1972		
1*	13·01	—	—	—
2*	14·55	—	—	—
		18 July 1972		
1	18·51	5	1·25	—0·7
2	23·30	10	0·75	—0·8
		19 July 1972		
1	01·52	18	1·25	—0·5

* Much smaller events whose *S-P* interval could not be read on the record.

† Maximum zero-to-peak trace amplitude in millimetres (*see* main text also).

NOTE 1. On account of low recording speed (25 mm/min), less stable timing system and high signal frequencies (about 20 to 30 cps) the entries in this table represent approximate values.

NOTE 2. All the observations listed in this table are based on 5 sets of recordings made as per the following schedule :

Set I : 9 Jun to 14 Jun (single channel)
Set II : 14 Jun to 25 Jun (single channel)
Set III : 4 Jul to 7 Jul (3-channel)
Set IV : 13 Jul to 18 Jul (3-channel)
Set V : 18 Jul to 21 Jul (3-channel)

under investigation. We have used Eq. (5) for the computation of local magnitude of all the Mandya microtremors, and it can be seen from Table 2 that almost all the values turned out to be negative implying that the events were smaller than the corresponding standard shocks.

Our formulation on local magnitude is valid except for a couple of limitations, but the errors creeping in are smaller than the limit of accuracy associated with reading of the seismograms. Firstly, the phase which appears maximum on the record traced by a Standard seismograph responding to only horizontal ground movements may not be the same phase in any other record though the difference between the logarithms of maximum ground motion associated with the two maxima may not significantly differ. Secondly, crustal structure as also the effects of lateral inhomogeneities across the wave propagation paths may give rise to different amplitude-distance relationships for different sites, and hence the basic A_0 values cannot be assumed to apply outside the southern Californian region without evidence. The amplitude attenuation laws must therefore be studied for the region using controlled underground detonations in order to supplement the A_0 data.

Acknowledgement

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APPENDIX

Standardised table for computing local magnitudes of small earth tremors from the knowledge of the actual ground motion due to the disturbances from within 600 km to recording site

S. No.	Range of epicentral distance $\triangle$(km)	—log $(A_0)_t$ †	$(A_0)_g$ ‡	S. No.	Range of epicentral distance $\triangle$(km)	—log $(A_0)_t$ †	$(A_0)_g$ ‡
1	0—9	1·4	14·2175	18	170—189	3·4	0·1422
2	10—14	1·5	11·2948	19	190—209	3·5	0·1129
3	15—19	1·6	8·9712	20	210—219	3·6	0·0897
4	20—24	1·7	7·1258	21	220—229	3·65	0·0800
5	25—29	1·9	4·4963	22	230—249	3·7	0·0713
6	30—34	2·1	2·8367	23	250—269	3·8	0·0566
7	35—39	2·3	1·7902	24	270—289	3·9	0·0450
8	40—44	2·4	1·4218	25	290—309	4·0	0·0357
9	45—49	2·5	1·1295	26	310—329	4·1	0·0284
10	50—54	2·6	0·8971	27	330—349	4·2	0·0225
11	55—59	2·7	0·7126	28	350—379	4·3	0·0179
12	60—79	2·8	0·5660	29	380—399	4·4	0·0142
13	80—89	2·9	0·4496	30	400—429	4·5	0·0113
14	90—109	3·0	0·3571	31	430—469	4·6	0·0090
15	110—129	3·1	0·2837	32	470—509	4·7	0·0071
16	130—149	3·2	0·2253	33	510—559	4·8	0·0057
17	150—169	3·3	0·1790	34	560—600	4·9	0·0045

† Common logarithm of the maximum zero-to-peak trace amplitude $(A_0)_t$ in millimetres with which a Standard Wood-Anderson Torsion Seismometer should register an event of magnitude zero. Since $(A_0)_t < 1$, its logarithm is negative and the column above shows values for—log $(A_0)_t$ as a function of $\triangle$. The set of values in this column are after Richter (1958).

‡ Actual ground displacement in millimicrons corresponding to $(A_0)_t$ associated with magnitude zero shock.