RESULTS OF A SEISMIC NETWORK IN NORTH EAST INDIA REGION

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ABSTRACT – *The North East India region is seismically one of the most active intracontinental zones where more than a dozen devastating earthquakes of magnitude $M \geq 7$ including the two $M = 8.7$ earthquakes of 1897 and 1950 took place during the last nine decades. Under a collaborative programme between the Regional Research Laboratory, Jorhat and the National Geophysical Research Institute, Hyderabad, a local seismic network consisting of 14 vertical component and 2 three component stations equipped mostly with short period instrumentation was established progressively in the region since the year 1979. With Shillong and Tura (Meghalaya) seismic stations operated by India Meteorological Department, New Delhi, the network helped to upgrade the location capabilities from magnitude 4.5 to 3. Further, the network provides scope to carry out the investigations on P-wave travel times, sub-Moho P-velocities, regional geothermal model, station factors of Khonsa (Arunachal Pradesh) and Yaongyimsen (Nagaland) seismograpy stations and seismicity associated with Indo-Burman plate margin. These studies are briefly described and the results are aptly presented. Certain interesting observations are also pointed out.*

INTRODUCTION

The North East India region is tectonically and seismically one of the most interesting regions in the world. The region demarcated by latitude 22–29°N and longitude 89–97°E is comprised of Eastern Himalayas, Mishmi Massif, Indo-Burman Ranges, Shillong-Mikir Hills Block, Assam valley and northern part of Bengal basin. Tectonic and geologic history of the region was published elsewhere[1–4] While emphasizing the poor status of earthquake location and detection capabilities, Gupta *et al.*[5] noted that till the year 1979, the Central Seismological Observatory, Shillong operated by the India Meteorological Department, New Delhi (IMD) was the only good station and only 2 to 3 earthquakes of magnitude ⩾ 4·5 were being located every month. Further, several workers[6–18] stressed that North East India is associated with high seismicity and studied to unfold the main geological features that are acting as weaknesses characterized by the occurrence of earthquakes. In view of high seismicity and with an intention to improve detection capabilities, Regional Research Laboratory, Jorhat (RRL-J) in collaboration with National Geophysical Research Institute, Hyderabad (NGRI) established progressively 14 vertical component and 2 three component seismic stations equipped with short period instrumentation. Locations of these stations and also SHL and TUR (Shillong and Tura in Meghalaya state) stations of IMD are shown in figure 1. The seismic codes, namely, AZL; BMI, INR, KOI, and TZU; DKI, NGN; GWH, HLG, HMN, JHI, KZI, TZR, and UMI; KHM and YYI represent the RRL-J/NGRI seismic stations situated at Aizawl (Mizoram); Bomdila, Itanagar, Khonsa and Tezu (Arunachal Pradesh); Dawki and Nongstoin (Meghalaya); Guwahati, Haflong, Hamren, Jorhat, Kaziranga, Tezpur and Umrongso (Assam), Kohima and Yaongyimsen (Nagaland) respectively. Particulars on seismic codes, locations, equipment etc. were given by Sitaram *et al.*[25]

While reviewing the various studies related to the crustal structure of the Indian subcontinent, Narain[19] briefly described a few interesting studies on the crustal structure in the North East India region. One such study is that of Tandon[20] who gave crustal structure as well as P and S-wave velocities. Gupta *et al.*[5] modified Tandon's crustal model and computed hypocentral parameters with least errors. Gupta's crustal model

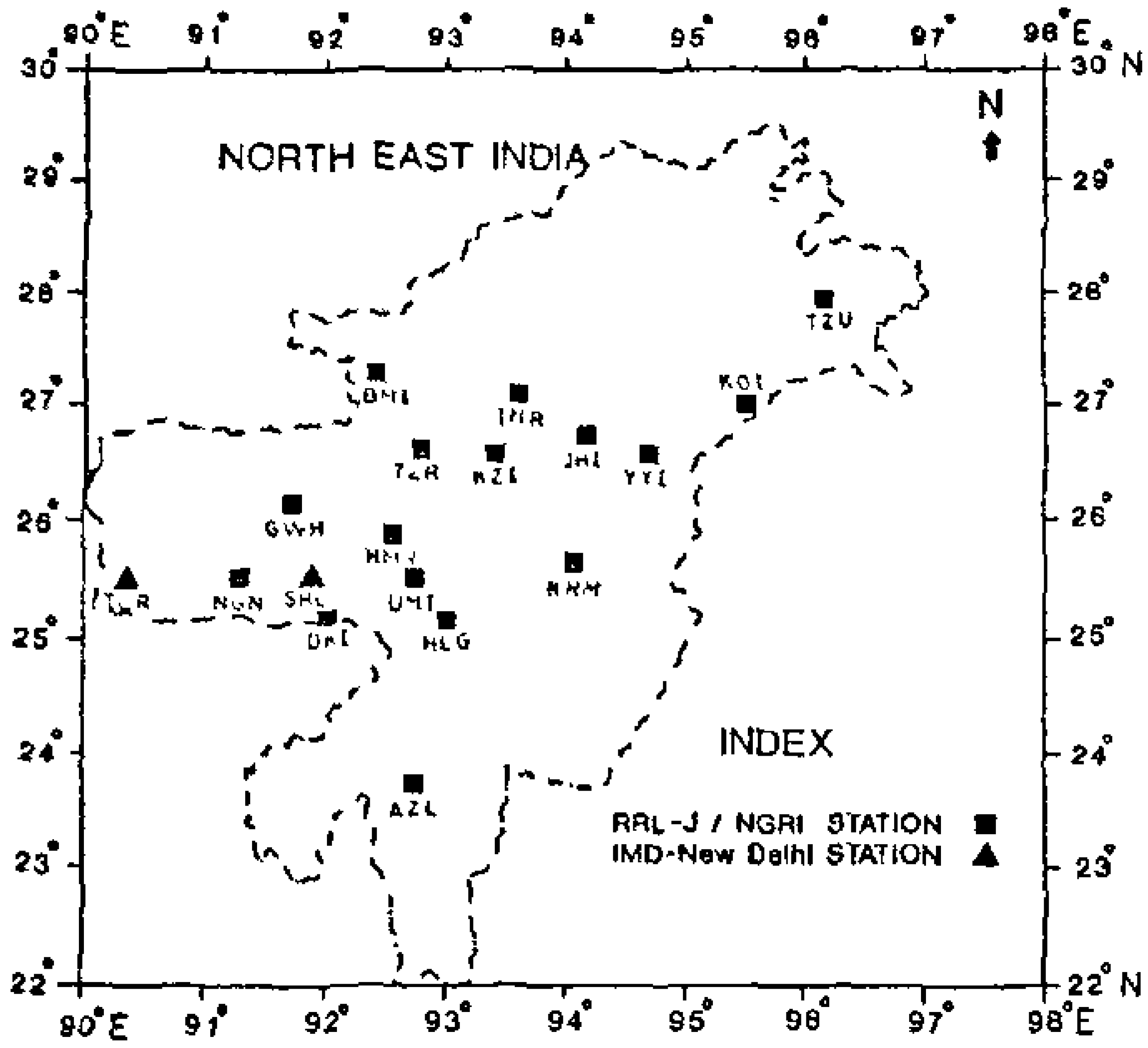

Figure 1 Locations of the seismic stations operated by RRL - J / NGRI and IMD, New Delhi in North - East India. Only locations of the stations whose data was used in the studies reported, are shown. Text may be referred for more details.

became a standard of reference for the studies reported here. Employing the seismic phase data as reported by RRL-J/NGRI seismic network and SHL and TUR stations of IMD, studies on P-wave travel times, sub-Moho P-wave velocities, regional geothermal model, station factors of Khonsa (Arunachal Pradesh) and Yaongyimsen (Nagaland) seismograph stations and the 6 August 1988 Indo-Burma border earthquake of magnitude 6·8 and its aftershocks were carried out.[21–25] Further, Saikia *et al.*[11,12] and Vanek *et al.*[17] investigated the seismicity associated with North-East India using the data reported by International Seismological Centre, Newbury, United Kingdom. These studies are briefly described. Results and salient observations are presented.

Results

Travel Times of P-waves

Based on the local crustal velocity model of Gupta *et al.*[5] Sitaram *et al.*[24] obtained relatively precise locations for 90 earthquakes in North East India Region (NER). Further, they computed the travel times of P-waves from the foci of these earthquakes at arbitrarily selected depths 5, 13, 25, 41, and 50 km to the sites of RRL-J/NGRI seismic stations located at GWH, INR, JHI, KHM, KOI, KZI and YYI and also to the sites of SHL and TUR stations of IMD (figure 1). The travel times of P-waves fit a straight line very well with the velocities of 5.97 ± 0.31, 6.18 ± 0.01, 6.41 ± 0.03, 7.82 ± 0.07 and 7.95 ± 0.01km/sec at each of the depths under study. Similar investigations of P* and P_g-waves of 16 earthquakes at a depth of 10 km gave the velocities of 6.53 ± 0.31 and 5.64 ± 0.34km/sec respectively. A simplified two-layered crustal model consisting of an average crustal thickness of 41.5 km with 22.2 and 19.3 km thick upper and lower layers was obtained. The P-wave velocity (V) appears to vary with depth (h) linearly in the depth intervals of 5–25 km and 40–50 km. In particular, the relation $V = 0.02h + 5.91$ seems to hold for the depth interval of 5–25 km.

Sub-Moho P-wave Velocities

Inter station sub-Moho P-wave velocities in NER were evaluated by Sitaram *et al.*[22] by applying the time term method of Scheidegger and Willmore[26] and Willmore and Bancroft[27] to the P_n arrival times at pairs of seismic stations to which epicentres form nearly the same azimuth. P_n times of 57 shallow earthquakes (of magnitude 4·0–5·5) were obtained from the short period vertical component seismograms for the seismic stations at JHI, KZI, INR, KOI, KHM and YYI and also for SHL and TUR stations of IMD (fig. 1). The method of analysis adopted by Haines[28] was used to determine sub-Moho P-wave velocities (km/sec) beneath the pairs of stations for

- direct paths assuming the same time terms for the stations involved in each pair,
- direct and reverse paths and
- direct and or combined paths after taking the time term differences into consideration. The velocities exhibit up to 2·5 per cent variation in the interval 7·85–8·08 km/sec. Figure 2 depicts sub-Moho P-wave velocities beneath NER wherein two dashed lines divide the region into three zones namely
- Meghalaya and western part of Assam,
- Eastern part of Assam and
- Nagaland and Tirap district of Arunachal Pradesh.

These zones are associated with mean sub-Moho P-velocities 7·94, 8·05 and 7·88 km/sec. Especially, the region of eastern part of Assam associated with relatively higher velocity of 8·05 km/sec corresponds to the Northeast directional movement of the Indian plate.

Geothermal Model

Cermak *et al.*[23] obtained the converted heat generation data for the crust beneath NER from the vertical distribution of P-wave velocities as proposed by Gupta *et al.*[5] Based on the empirical relationship between P_n-wave velocity and heat flow, they converted the individual P_n values as reported by Sitaram *et al.*[22]

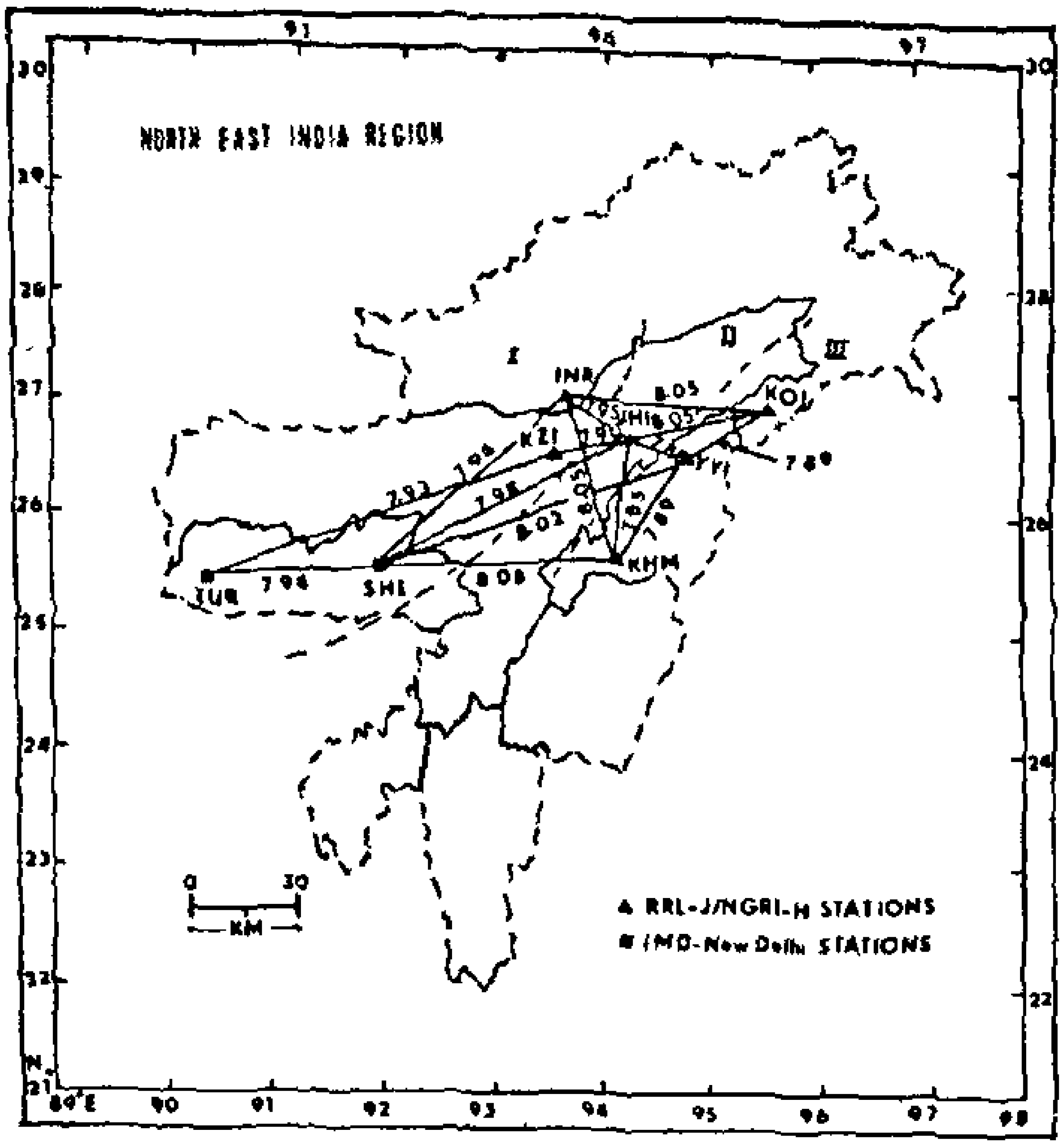

FIGURE 2 Sub-Moho P-Wave velocities (km/sec) in North-East India. Dashed lines divide the region arbitrarily in three zones I, II, III referring to the average P_n velocities 7.94, 8.05 and 7.88 km / sec respectively.

into heat flow estimates at crust-mantle boundary (figure 3) and also obtained Moho temperatures. The most probable Moho temperatures range from 800 to 900°C and the flow of heat from below the crust attains about 50 per cent of observed surface heat flow i.e., about 30 $mW.m^{-2}$. Figure 3 shows the regional geothermal regime at the crust-mantle boundary along with tectonic patterns as reported by Evans and Mathur[29] and Nandy[4]. Relatively low P_n-velocity of 7·88 km/sec beneath Naga-Patkai belt was attributed to the anomalous upper mantle structure resulting due to the subduction of Indian plate beneath the Burmese plate and it seems to correspond to relatively high heat flow estimates and Moho temperatures. Eastern part of Assam with lowest heat flows (and maximum P_n-wave velocity of 8·05 km/sec) is then linked to the Indian Shield.

Station factors of Khonsa and Yaongyimsen seismograph stations

According to Gutenberg and Richter[30] relation, Kotoky *et al.*[21] determined body-wave magnitudes of 384 and 440 teleseismic events in the distance range 9–100° using short period P-wave maximum ground amplitudes as obtained from the vertical component seismograms at KOI and YYI seismic stations. They, further, computed magnitude residuals by comparing the magnitudes reported by the stations with the corresponding body-wave magnitudes reported by the National Earthquake Information Service of the United States Geological Survey. Average residuals for KOI and YYI stations were found to be $0{\cdot}09 \pm 0{\cdot}40$ and 0.48 ± 0.42 respectively. The average residuals (ΔM) for the stations exhibited a decreasing trend with respect to epicentral distance (Δ), focal depth (h) and body-wave magnitude (m_b). Corresponding relations for KOI and YYI stations are: $\Delta M = 0.12 - 1 \times 10^{-5}\,\Delta$, $\Delta M = 0.22 - 7 \times 10^{-5} h$, $\Delta M = 1.96 - 0.356\, m_b$; $\Delta M = 0.59 - 2 \times 10^{-5}\,\Delta$, $\Delta M = 0.54 - 19 \times 10^{-5}\, h$ and $\Delta M = 2.56 - 0.391\, m_b$. The nature of variation of residuals with azimuth is nearly similar. Near similarity of residual variation with respect to distance, depth magnitude and azimuth for the two stations may be attributed to the effect of sedimentary layer beneath them.

North East India Seismicity

With the help of HCL (Hindustan Computer Limited) computer of system 4 model and based on a horizontally layered crustal model of Gupta *et al.*[5] for a P-wave to S-wave velocity ratio of 1·75, locations are obtained using HYPOCENTER algorithm software package of Lienert *et al.*[31] Duration magnitudes are determined on the basis of signal durations because the amplitudes of the strong signals are clipped. Signal duration is defined as the duration of time of earthquake signal on the seismogram from the first motion to the point at which the amplitude drops to background noise level. Further, signal duration calibration was dealt by Sitaram *et al.*[25] The errors involved in the estimation of magnitudes are of the order of $\pm 0{\cdot}2$.

Until recently, due to lack of stations only 2–3 earthquakes (magnitude $\geqslant 4.5$) per month were reported by the international agencies, whereas with the installation of RRL-J/NGRI Network 100–150 earthquakes could be recorded and about 100 events could be located with a threshold magnitude of 3.0. The errors involved in the location estimates, generally, do not exceed ± 10 km. However, recent installations of telemetry digital networks at Jorhat and Tezpur, consisting of about eight stations each, would further improve the present detection threshold and location accuracy. The annual seismological bulletins are brought out and distributed to national and international agencies.

Initially, Dutta[6,32] and Saikia *et al.*[11] divided NER in five zones namely

- Indoburmese zone,
- Mishmi Massif,

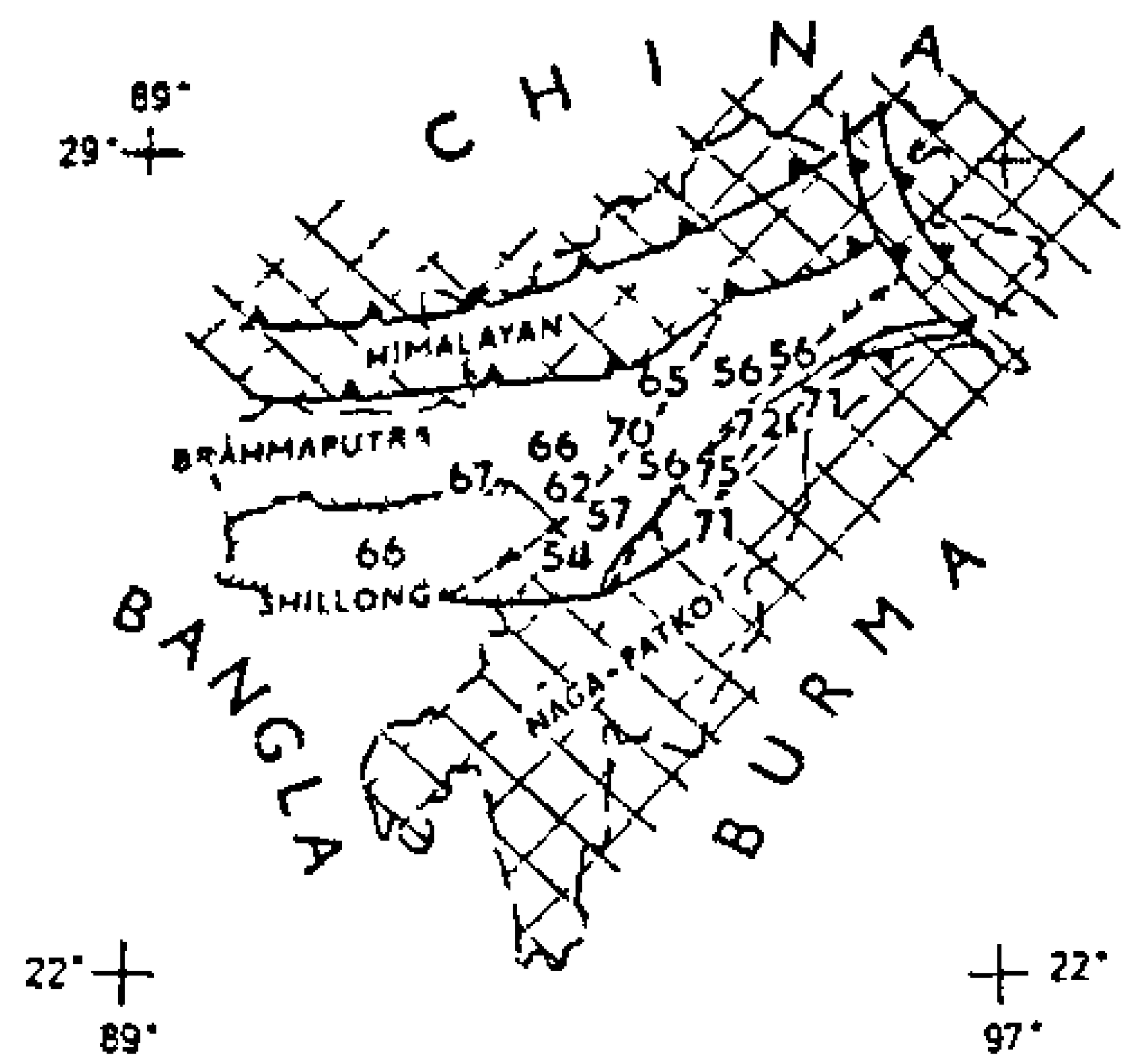

FIGURE 3 Heat flow estimates based on the empirical relationship between P_n-wave velocity and heat flow. Main tectonic patterns as reported by Evans and Mathur[27] and Nandy[4] are also shown.

- Eastern Himalayas,
- Shillong Plateau and
- Upper Assam.

Further, Saikia[33] found that a major portion of seismicity is associated with the five well defined major tectonic lineaments constituting the crustal weakness. The lineaments are

- The Himalayan thrust belts bordering the Assam valley namely the Main Boundary Thrust and the Main Central Thrust,
- the NE-SW directed Naga-Disang group of thrust faults bordering the Upper Assam Valley,
- Mishmi or Mizu thrust system,
- the E-W Dauki lineament of the southern border of the Shillong Plateau and
- the tectonic lineaments associated with the interior part of Nagaland and Manipur having genetic relationship with the Arakan-Yoma-Chin belt. However, further analysis of seismicity may lead to the demarcation of new lineaments.

Seismicity of Indo-Burman Plate Margin

The Indo-Burman Ranges comprising from north to south the Naga Hills, Chin Hills and Arakan-Yoma Hills, pass northeastwards into a belt of northwest trending structures linking them with the eastern Himalayas. Southward the ranges pass into the Andaman and Nicobar islands. The ranges represent a complex orogenic belt composed of Upper Cretaceous-Eocene strongly folded and faulted flysch strata surrounded by post-Eocene molasse strata of the central Burma molasse basin in the east and of the outer molasse basin in the west. Main folding, faulting and uplifting occurred in the Middle to Upper Pliocene. Terrestrial volcanic bodies of calc-alka-line composition post dating the Mio-Pliocene molasse occur in the central Burma molasse basin.[34,35]

The 6 August 1988 Indo-Burma border Earthquake and its Aftershocks: An Analysis

An earthquake of magnitude 6.8 occurred in the Indo-Burma border region on August 6, 1988 at the depth of about 99 km. From the analysis of network data Sitaram *et al.*[25] could find locations for about 180 aftershocks with a threshold magnitude of 3.0 during the period of about 178 days following the main earthquake. The main shock was preceded by a foreshock of magnitude 4.5 in about 91 minutes and followed by five shocks of magnitude 5.0–5.3 within six hours. The top 50 km thick crust was associated with most of the aftershocks whereas the main shock occurred at an intermediate depth. The number of aftershocks decayed following the modified Omori's formula with $p = 0.76 \pm 0.04$. The epicentres for the sequence formed a cluster of an elliptical shape covering about 4,900 sq.km. in NW-SE direction across the Eastern Boundary Thrust, a major tectonic contact between Arakan-Yoma Fold Belt and central Burma. The orientation of assigned Modified Mercalli intensity (VII and VIII) contours followed the same directional trend. The foci distribution across the Thrust revealed that about 70 km thick Indian Wadati-Benioff zone dips below the Burmese plate at about 43°. The foci distribution for the earthquakes prior to the present sequense especially during the period 1980–July 1988 showed the same observation. The whole activity appears to have resulted due to the subduction-continental collision processes.

Vanek *et al.*[17] investigated the geometry of distribution of foci across the ranges and arrived at the following observations:

- Indian Wadati-Benioff zone across the ranges dips at about 30° to the east in the Chin Hills segment and also at about 50° to the southeast in the Naga Hills segment. Its thickness ranges between 50 and 70 km. Maximum depth of earthquakes is 190 km. This observation, in general, conforms with that of Saikia *et al.*[11]
- The Wadati-Benioff zone and young calcalkaline volcanism in the central Burma molasses basin manifest the existence of a subduction, which started in the Middle Miocene along the western boundary of the Indo-Burman orogenic belt folded in the Oligocene. The average rate of this Mio-Pliocene subduction was

estimated to be between 2.0 and 2.5 cm/yr. Further, they delineated ten major seismically active fracture zones on the basis of distribution of epicentres for the shallowest earthquakes not belonging to the Wadati-Benioff zone. They also cited the parameters as reported by Chandra[9] for the disastrous earthquakes associated with these fracture zones. The zones are (1) Mandalay (2) Myingyan (3) Gangaw-Kumaon Range (4) Mingin Range-Jade Mines (5) Haka-Wuntho (6) Hailakandi (7) Bishnath (8) Assam Valley (9) Main Central Thrust and (10) Mishmis.

Salient Observations

In the absence of deep seismic sounding data, the usage of earthquake data becomes indispensable for the evaluation of wave velocities and crustal structure beneath NER. From a study on body waves of the great Assam earthquake of 15 August, 1950 and its 54 aftershocks, Tandon[20] derived P_n, P^*, P_g, S_n, S^* and S_g-velocities 7.91, 6.55, 5.58, 4.46, 3.85 and 3.43 km/sec and also obtained a Moho depth of 46.3 km along with 24.8 and 21.5 km thick upper and lower crustal layers. Sitaram *et al.*[24] derived a Moho depth of 41.5 km along with 22.2 and 19.3 km thick upper and lower crustal layers. Further, P_g and P_n-velocities 5.64 ± 0.34 and 7.82 ± 0.07 km/sec obtained by Sitaram *et al.*[24] are comparable to the corresponding velocity estimates 5.64, and 7.99 km/sec reported by Gupta *et al.*[36] for the propagation paths passing through the zone of intermediate focus earthquakes in the Arakan-Yoma fold belt. Gupta *et al.*[36] also observed P_g and P_n velocities 5.11 and 7.83 km/sec for the paths passing through Assam west of the belt. They reported P_g-velocity 5.99 km/sec also for the paths passing through Shillong Plateau and Mikir Hills. More studies need to be undertaken for knowing the lateral variation of wave velocities and crustal structure.

The presence of subduction related processes beneath the Indo-Burman plate margin was reported elsewhere.[7,10,11,16,17] Relatively low sub-Moho P-velocity of 7.88 km/sec obtained by Sitaram *et al.*[22] for the Naga-Patkai belt, and high heatflow and Moho temperatures observed by Cermak *et al.*[23] suggest the anomalous upper mantle structure resulting due to the subduction processes prevalent in the plate margin. From a study on the Indo-Burman plate margin, Saikia *et al.*[12] inferred that the region probably developed through continent-continent or continent-island arc collision. They also observed that the disposition of the zone of intermediate depth earthquakes, occurrence of ophiolitic rocks, deep sea sediments such as radiolarian cherts, pelagic limestones and pronounced negative gravity anomaly provide evidence of a suture zone between the Indian and Burmese plates. From the Wadati-Benioff zone in the plate margin and young calc-alkaline volcanism in the central Burma molasse basin, Vanek *et al.*[17] observe that the existence of a Mio-Pliocene subduction with the trench at the western boundary of the Oligocene Indo-Burman orogenic belt. The average rate of this Mio-Pliocene subduction was estimated to be between 2.0–2.5cm/yr. According to Gupta *et al.*[36] the continental collision is still taking place in the plate margin.

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ACKNOWLEDGEMENTS: We are grateful to Mr. B. P. Chaliha, Acting Director, Regional Research Laboratory, Jorhat (Assam) for according permission to publish the paper. We thank Dr. H. N. Srivastava, Director, India Meteorological Department, New Delhi for providing the necessary data. Thanks are also due to Mr. O. P. Sahu for rendering help in the preparation of diagrams.