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EARTHQUAKE ACTIVITY IN THE GREATER NEW YORK CITY AREA: MAGNITUDES, SEISMICITY, AND GEOLOGIC STRUCTURES

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Abstract

Earthquakes recorded by stations of the Lamont-Doherty seismic network in the greater New York City area are analyzed to determine magnitudes and the relationship between seismicity and geologic structures. Between 1974 and 1983, the configuration of stations in this region remained relatively stationary and the type of recording devices (visual drum recorders and 16-mm photographic recorders) did not change. This distribution of stations and recording devices allows for a uniform measurement of magnitudes and seismicity. Magnitudes of these earthquakes are determined by comparing amplitudes and signal duration measured from high-frequency (5 to 10 Hz) data recorded by the local network with m_{bLg} and M_L determined from data at frequencies near 1 Hz. During the period of time studied (nearly 10 yr), 61 earthquakes were located in this region, but none of these earthquakes exceeded 3.0 on the m_{bLg} scale. The largest event ($m_{bLg} = 3.0$) occurred in the Coastal Plain province of northern New Jersey.

The magnitude threshold for uniform detection of events throughout this region during the period of time studied is estimated to be $m_{bLg} = 1.6$. With events below this threshold removed from the catalog of network seismicity, we find that about half of the earthquakes studied occurred within 10 km of the Ramapo fault system. This fault system lies about 30 km northwest of New York City and has been interpreted by several investigators to be the most active fault system in the greater New York City area. However, earthquakes at least as large as those recorded near the Ramapo fault were located as far as 50 km from this fault (and within 20 km of New York City), in geologic structures that surround the Newark basin. While the Ramapo fault can by no means be ruled out as a possible source zone for earthquakes in the greater New York City area, the geologic structures associated with most (if not all) earthquakes in this region are still unknown. Thus, the cause of earthquakes in this region remains an enigma.

Introduction

The area within a 100-km radius of New York City has had an intermediate level of seismic activity throughout its recorded history. This region is clearly not as seismically active as some parts of the Western United States, such as California; neither has this region experienced historic earthquakes as large as the 1811 to 1812 events located near New Madrid, Missouri. Nonetheless, the greater New York City area has had its share of moderate-size historic earthquakes when compared to the rest of the northeastern United States. For example, both the historical and instrumental records of seismicity in the northeast (Figures 1 and 2) show a higher level of activity near New York City than in central New York state or western Pennsylvania. The highest intensity assigned to earthquakes in the greater New York City area in Smith's (1966) historical catalog are two events of maximum intensity VII on the Modified Mercalli (MMI) scale (1737 and 1884).

The lack of instrumental records for most of the earthquakes in the historical record makes it difficult, if not impossible, to locate these events accurately enough to relate them unambiguously to rupture along specific geologic faults. Moreover, without instrumental records of these historical earthquakes, it is difficult to compare the size of these events (i.e., magnitude or seismic moment) with earthquakes in other regions. Based on felt area and maximum intensity, however, it appears that the largest known earthquake in this region is the 1884 event. For this earthquake, Rockwood (1885) reported fallen bricks and cracked plaster from eastern Pennsylvania to central Connecticut, and the maximum intensity reported by Rockwood (1885) was at two sites in western Long Island (Jamaica, New York, 14 km east of Manhattan and Amityville, New York, 50 km east of Manhattan). The felt area of the 1884 earthquake, measured from the map published by Rockwood (1885), was about 270,000 km^2.

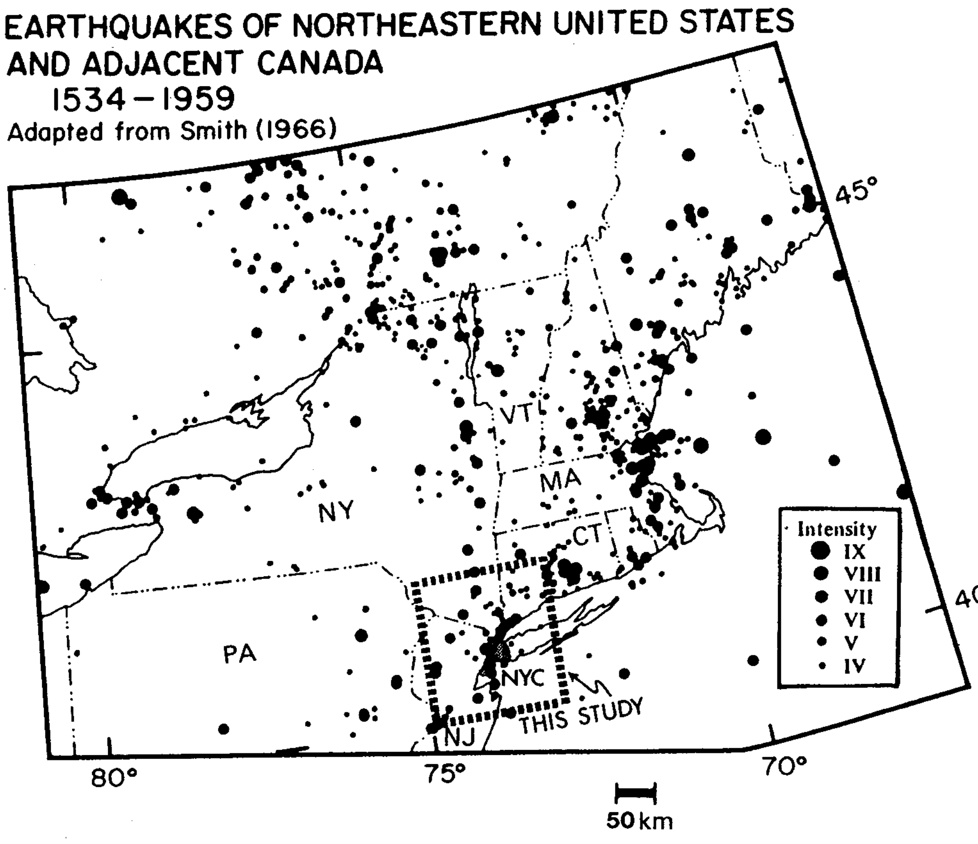

FIG. 1. Historical record of earthquakes in northeastern United States and adjacent Canada for the period 1534 to 1959 (after Smith, 1966). The *shaded region* labeled NYC is New York City.

Instrumental recording of earthquakes in the greater New York City area was quite limited before the 1970s, when numerous local and regional stations were installed. In the early 1920s, a station was installed at Fordham University. Stations were also installed at City College of New York in 1948 and Columbia University (at Palisades, New York) in 1949. Dewey and Gordon (1984) used the early instrumental record of earthquakes in the Eastern United States to relocate hypocenters of earthquakes recorded between 1925 and 1980. They found sufficient instrumental data to locate four earthquakes that occurred before 1970 in the greater New York City area. Three of these events occurred within a 4-hr period on 23 August 1938 (m_{bLg} = 3.9, 4.0, and 3.7) in New Jersey about 50 km south of New York City, and one occurred on 3 September 1951 (m_{bLg} = 3.8) about 50 km north of New York City.

More detailed information about the rate of seismic activity and the relationship between earthquakes and geologic structures can be obtained from instrumental data recorded by local seismic networks in the greater New York City area. For more than a decade, microearthquakes have been monitored in this region by seismic networks operated by Lamont-Doherty Geological Observatory (LDGO) and Woodward-Clyde Consultants (WCC). Using this record of network seismicity, we can estimate magnitudes of earthquakes with sufficient accuracy to compare the rate of seismic activity in this region during the past decade with rates of seismic activity in other parts of the world. In addition, the relationship between microearthquake seismicity and specific geologic structures in this region can be analyzed

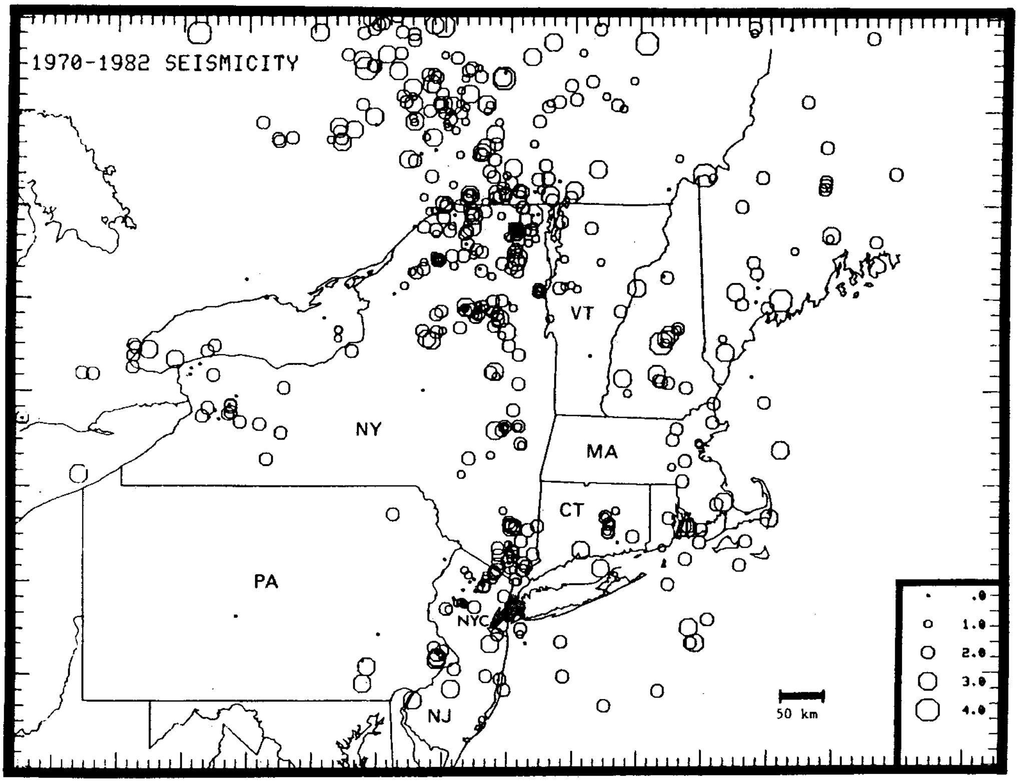

FIG. 2. Earthquakes recorded by LDGO in northeastern United States and adjacent Canada for the period 1970 to 1982. Magnitudes shown in this figure are those reported in the *Bulletins of the Lamont-Doherty Geological Observatory.*

using locations of earthquakes determined from the network data, which are generally accurate to within a few kilometers of the true epicenter.

In this paper, we analyze earthquakes recorded by the LDGO and WCC networks betwen January 1974 and September 1983; a period of time during which the configuration of stations in the region remained relatively stationary and the type of recording devices (visual drum recorders and 16-mm photographic recorders) did not change. Magnitudes of these earthquakes were determined by comparing amplitudes and signal duration measured from high-frequency (5 to 10 Hz) data recorded by the local network with m_{bLg} (Nuttli, 1973) and M_L (Richter, 1935) determined from longer period data (i.e., at frequencies near 1 Hz). These longer period data were obtained from seismograms recorded on stations of the WWSSN and two Wood-Anderson instruments operating in the Northeast. We also analyze the completeness of the catalog of network seismicity in this region. The relationship between the network seismicity and geologic structures in this region is analyzed by considering only earthquakes with large enough magnitudes to be detected without a bias due to station distribution.

LOCAL EARTHQUAKES RECORDED BY THE LDGO NETWORK

Locations. The LDGO network operating in New York State and adjacent areas (Figure 3) consists of 30 stations. Each station has a 1-Hz or 2-Hz vertical geophone, and three component stations with 1-Hz horizontal geophones are located at GPD,

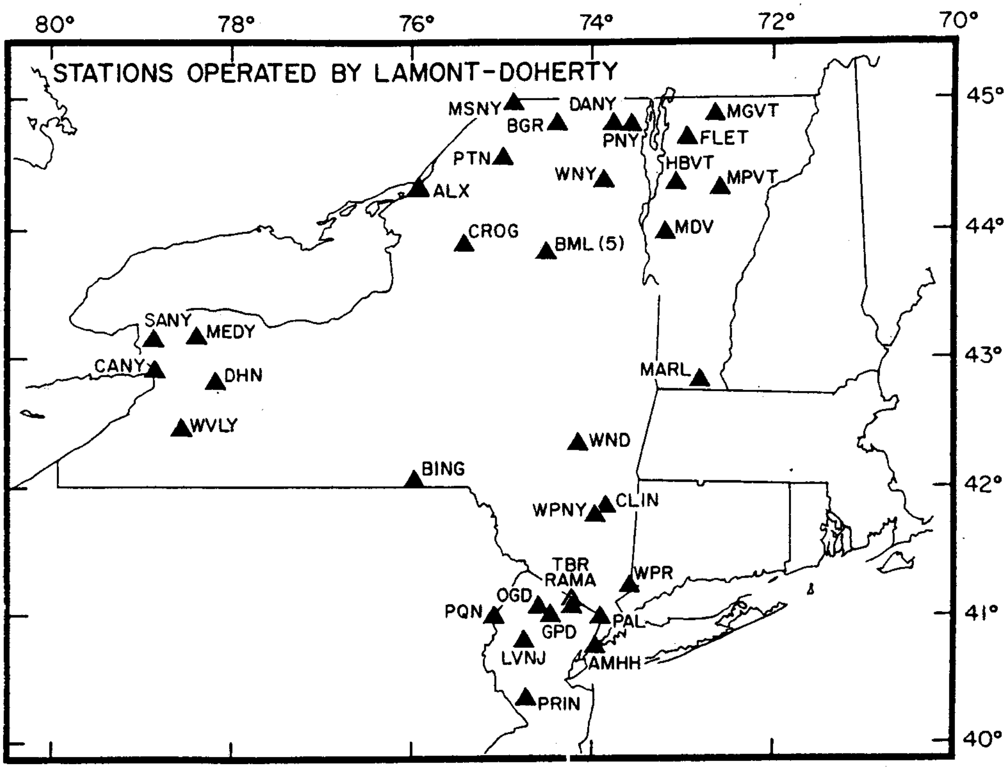

FIG. 3. Short-period seismic stations (triangles) operated by LDGO in the states of New York, New Jersey, and Vermont (1974 to 1983). Signals are telemetered to Palisades, New York, where they are recorded on a common time base.

APH, and RAMA. The peak amplitude response for these stations is at 20 Hz and ranges from about 10^6 to about 10^7. The data are telemetered to a central recording site at Palisades, New York, where they are recorded on visual drum recorders and 16-mm photographic recorders.

The earthquakes studied here were located using arrival times of P and S waves recorded on the LDGO network as well as arrival time data from the network of seismic stations operated by WCC in the vicinity of the Indian Point nuclear power plant. For the larger events, arrival times were also obtained from other local networks that are part of the Northeast United States Seismic Network (NEUSSN). Figure 4 shows the locations of network stations operated by LDGO and WCC in the greater New York City area between 1974 and 1983. Local stations have been operating in this region since about 1970, but the number of stations steadily increased between 1970 and 1974. After 1974, the configuration of the networks remained similar to that shown in Figure 4 until late 1983. Since 1970, about 100 earthquakes have been recorded in the greater New York City area, but only earthquakes that occurred between the beginning of 1974 and late 1983 (61 events) are considered in this study. This allows for a uniform measurement of the distribution and magnitudes of earthquakes.

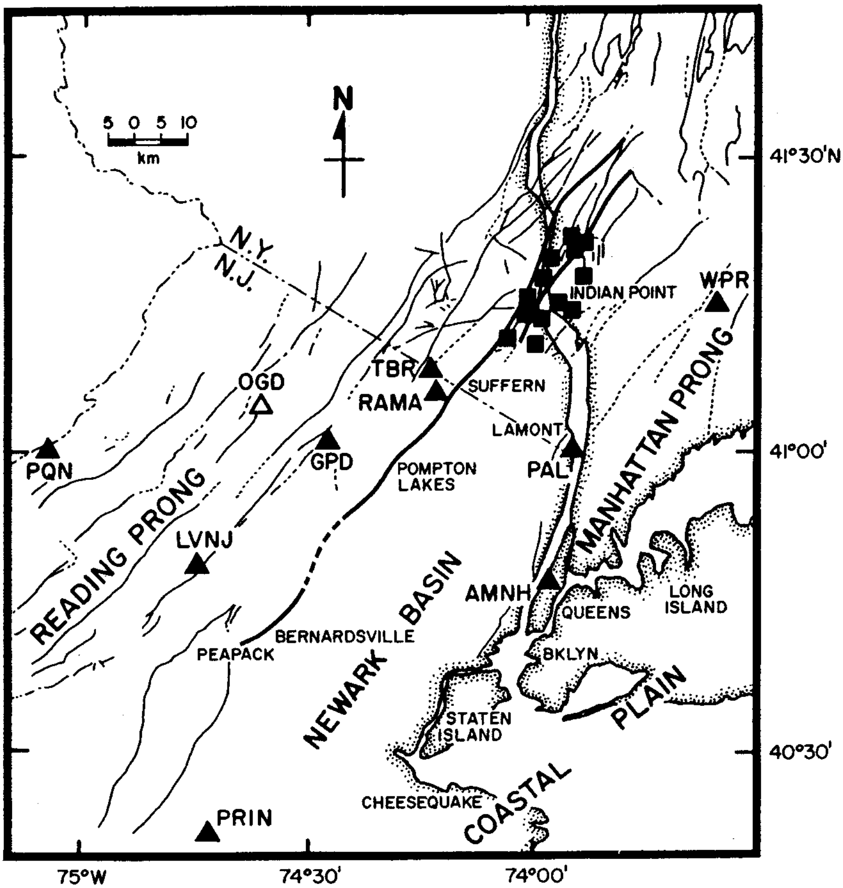

FIG. 4. Location of short-period seismic stations and geologic provinces, faults, and lineations in the greater New York City area (adapted from Yang and Aggarwal, 1981). Dark solid lines between Peapack, New Jersey and Indian Point, New York, represent the Ramapo fault system. Seismic stations operated by LDGO are shown as triangles and those operated by WCC are shown as squares. The open triangle represents a WWSSN station (OGD) located at Ogdensburg, New Jersey.

The seismic velocity model for southeastern New York and northern New Jersey determined by Yang and Aggarwal (1981) was used to locate earthquakes in this study. The uncertainty in event locations can be estimated from the ability to locate quarry blasts with known locations. Throughout New York state and adjacent areas, quarry blasts can be located to within 5 km of the true location. In the greater New York City area, where station coverage is more dense than in other parts of the region monitored, uncertainties in epicentral locations are less than 2 km. With the exception of events offshore, all earthquakes studied here are probably located to within 2 km of the actual epicenter. Events offshore may be mislocated by as much as 10 km.

Magnitudes. Most of the earthquakes that have occurred in the greater New York City area were either too small to be recorded teleseismically or occurred before the advent of instrumental seismology. Thus, it is difficult to compare directly the magnitudes of these earthquakes with those of earthquakes in other parts of the world. Nuttli (1973) developed a magnitude scale for earthquakes occurring in eastern North America and recorded at regional distances. His magnitude scale was given the symbol m_{bLg} since it utilizes the maximum sustained amplitude of the vertical component of the *Lg* wave at periods near 1 sec to determine the equivalent of teleseismic m_b.

Operators of seismic networks in the northeast United States commonly (and incorrectly) assign magnitudes to local earthquakes by applying the formulas for m_{bLg} given by Nuttli (1973) to amplitudes of high-frequency (5 to 10 Hz) L_g waves (e.g., Chiburis *et al.*, 1976–1980; Schlesinger-Miller and Barstow, 1983). In many studies of earthquakes recorded by seismic networks in the northeast, magnitudes were also assigned by applying Nuttli's (1973) formula to high-frequency *Lg* wave amplitudes. This same method of assigning magnitudes was used in two studies of earthquakes in the greater New York City area (Aggarwal and Sykes, 1978; Yang and Aggarwal, 1981). The Nuttli (1973) formulas, however, were developed for 1-Hz L_g waves, and were not intended to be used with higher frequency data.

Ebel (1982) dealt with this problem by comparing magnitudes determined from network data with seismic wave amplitudes recorded on a set of Wood-Anderson torsion seismographs operating at Weston Observatory in Weston, Massachusetts. He added an additional term to the original magnitude formula of Richter (1935) to correct for the difference in attenuation between southern California and the northeast United States. In his study, he found that substituting high-frequency L_g wave amplitudes (recorded throughout New England) into the Nuttli (1973) formula for m_{bLg}, overestimates M_L by about 0.4 magnitude units.

Herrmann and Kijko (1983) eliminated a great deal of the confusion over high-frequency versus 1-Hz L_g magnitudes by introducing an $m_{Lg}(f)$ scale. This scale uses regional, frequency-specific values of attenuation to correct for distance, and since f (frequency) is given along with the reported magnitude, e.g., $m_{Lg}(10)$, there is no confusion over what frequency was used. Herrmann and Kijko (1983) define $m_{Lg}(f)$ by the following formula

$$m_{Lg}(f) = 2.94 + 0.833\log(\Delta/10) + 0.4342\gamma\Delta + \log A \tag{1}$$

where Δ is the epicentral distance in kilometers, γ is the coefficient of anelastic attenuation, and A is the ground amplitude in microns. In this study, we determined values of m_{Lg}(f) using amplitudes read from 16-mm film records of the LDGO network and corrected for attenuation using the frequency-dependent attenuation values for the northeast given by Pulli (1984). The period range used in these calculations was 5 to 10 Hz, and the magnitudes determined in this manner are referred to as $m_{Lg}(f)$ in this paper, with the understanding that f is restricted to the period range of 5 to 10 Hz. These values of $m_{Lg}(f)$ were compared with m_{bLg} for four earthquakes and with M_L for six events.

We measured amplitudes, frequencies, and signal duration for 61 earthquakes recorded on 16-mm photographic recorders. Amplitudes of sustained ground motion were measured for the *Lg* phase (i.e., waves arriving at group velocities between 3.5

and 3.0 km/sec). Frequencies were measured from the seismograms by counting the number of zero-crossings in a given time period. Signal duration was measured as the duration of the earthquake signal in seconds from the P-wave arrival time to the point where the signal disappears into the noise.

In addition to the high-frequency data obtained from the local networks, amplitude and frequency data were also measured from records of longer period instruments for the larger earthquakes that occurred during the period of time studied. Lg waves with frequencies near 1 Hz ($0.6 < T < 1.2$ sec) were recorded by a number

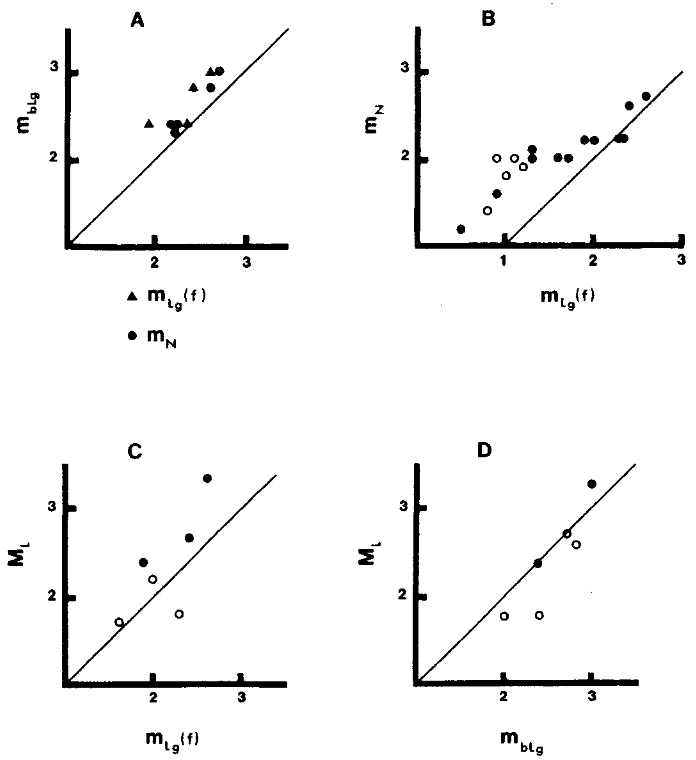

FIG. 5. Comparison of various magnitude determinations for earthquakes in the greater New York City area. Magnitudes were determined as described in the text. In each graph, a line through the origin with slope = 1 is shown. m_N represents magnitudes calculated by substituting Lg wave amplitudes from high-frequency network records into Nuttli's (1973) formula for m_{bLg}. (A) Closed triangles represent $m_{Lg}(f)$ values and closed Circles represent m_N values. (B)–(D) Closed circles represent magnitudes determined using two or more stations for both events, and open circles are shown when only one station was used to determine one or both of the magnitudes.

of stations of the WWSSN for several of the earthquakes studied here. These data were used to determine m_{bLg} magnitudes by applying the formulas developed by Nuttli (1973). A Wood-Anderson seismometer located at Palisades, New York, was used to determine M_L magnitudes by applying Richter's (1935) formula with the correction for attenuation in the northeastern United States suggested by Ebel (1982). In addition, Ebel (1982) determined M_L magnitudes for several of the earthquakes studied here from the Wood-Anderson seismometer located at Weston Obsevatory. His results are also compared with our high-frequency data.

The relationship between the various magnitude scales used in this study are shown in Figure 5. In Figure 5A, m_{bLg} determined from the WWSSN records is compared with $m_{Lg}(f)$ determined from the photographic network records. While it is clear from Figure 5A that m_{bLg} is consistently higher than $m_{Lg}(f)$, the data are too sparse to calculate a reliable best-fit line. The average difference between m_{bLg} and $m_{Lg}(f)$ is 0.4 magnitude units.

Magnitudes calculated by substituting values of *Lg* wave amplitudes from high-frequency network records into Nuttli's (1973) formula for m_{bLg} are referred to as m_N in the *Bulletins of the NEUSSN* (e.g., Chiburis *et al.*, 1976–1980) and in this paper. For the earthquakes studied here, m_N underestimates m_{bLg} by an average difference of 0.2 magnitude units (Figure 5A). Figure 5B shows that $m_{Lg}(f)$ generally underestimates m_N, particularly for the smaller events. With the exception of one M_L measurement that was based on only one station, $m_{Lg}(f)$ underestimates M_L, and m_{bLg} scatters near M_L (Figures 5, C and D).

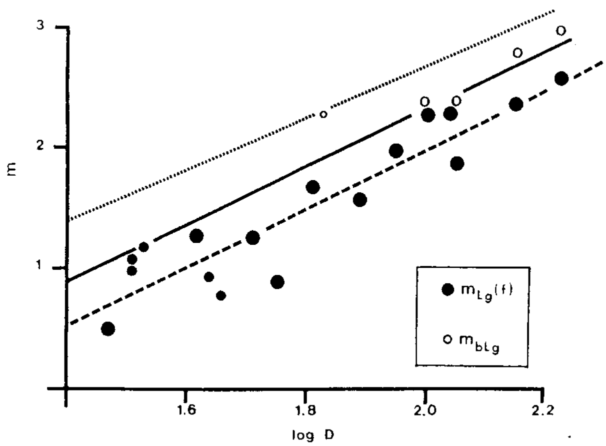

FIG. 6. Magnitudes [$m_{Lg}(f)$ and m_{bLg}] of earthquakes in the greater New York City area as a function of signal duration measured from LDGO network photographic records. Smaller symbols represent magnitudes determined from only one station. The heavy dashed line represents a least-squares fit to the $m_{Lg}(f)$ magnitudes, and the solid line represents the same least-squares fit plus 0.4 magnitude units [the average difference between $m_{Lg}(f)$ and m_{bLg} as determined in this study]. The light dashed line represents the signal duration formula of Chaplin *et al.* (1980).

The above analysis provides an upper limit of $m_{bLg} = 3.0$ for earthquakes recorded by the LDGO network in the greater New York City area during the 10-yr period studied, as well as specific values of m_{bLg} for nine earthquakes from this same period of time. We can also use the network data to estimate m_{bLg} magnitudes of the remainder of the earthquakes recorded, by comparing signal duration measured from the network photographic records with $m_{Lg}(f)$ and m_{bLg}. Figure 6 is a plot of $m_{Lg}(f)$ and m_{bLg} versus signal duration determined from the network photographic records. Also shown in Figure 6 is the relationship between signal duration and magnitude proposed by Chaplin *et al.* (1980) for earthquakes in New England (short dashes), as well as a best-fit line to the $m_{Lg}(f)$ versus signal duration data (long dashes). The Chaplin *et al.* (1980) relation which was derived from a comparison of m_N with signal duration, overestimates m_{bLg} in all but one case. For the events studied here, we prefer using signal duration to estimate magnitudes of locally recorded events and adding 0.4 magnitude units to the best-fit line for $m_{Lg}(f)$ versus signal duration to correct for the average difference observed for these two scales (solid line in Figure 6). This approach yields a catalog of earthquakes for the region studied (Table 1), in which magnitudes are related to a regional scale (m_{bLg}) and all magnitudes are determined from the same type of data (i.e., signal duration read from network photographic records). Magnitudes determined in this manner are shown in the seventh column in Table 2.

Detection threshold. The installation of regional seismic networks in the northeastern United States has significantly increased the detection of smaller earthquakes. This increased detection, if not interpreted with caution, however, can result in seismicity maps that show clusters of small earthquakes in the vicinity of clusters of stations. A question that we address in this study is, therefore: to what extent is the catalog of network seismicity in the region studied truly a measure of

TABLE 1

EARTHQUAKES RECORDED BY THE LDGO NETWORK IN THE GREATER NEW YORK CITY AREA ($m_{bLg} > 1.5$) (JANUARY 1974–SEPTEMBER 1983)

Event	Location	Latitude North (deg min)	Longitude West (deg min)	Date (mo/day/yr)
1	Stony Point, NY	41 13.12	73 59.51	04/08/74
2	Wappinger Falls, NY	41 34.27	73 56.40	06/07/74
3	West of Sandy Hook, NJ	40 20.82	73 10.62	02/20/75
4	Wappinger Falls, NY	41 34.80	73 56.63	06/15/75
5	Mahopac, NY	41 25.55	73 47.36	07/19/75
6	Wappinger Falls, NY	41 35.55	73 55.99	10/24/75
7	Riverdale, NJ	40 57.12	74 21.19	03/11/76
8	Ridgefield, NJ	40 50.10	74 02.85	04/13/76
9	Off Sandy Hook, NJ	40 29.07	73 47.74	05/11/76
10	Mt. Pleasant, NY	41 06.81	73 45.22	08/20/76
11	Suffern, NY	41 10.94	74 08.88	03/10/77
12	Hampton, NJ	40 42.22	74 56.12	07/02/77
13	Annsville, NY	41 18.78	73 55.41	09/02/77
14	North of Newburgh, NY	41 33.53	73 57.18	10/14/77
15	Off Sandy Hook, NJ	40 31.80	74 04.80	04/03/78
16	Oakland, NJ	41 04.52	74 12.10	06/30/78
17	Cheesequake, NJ	40 19.29	74 15.81	01/30/79
18	Bernardsville, NJ	40 43.34	74 30.25	03/10/79
19	Mt. Kisco, NY	41 09.38	73 42.79	12/30/79
20	Annsville, NY	41 18.53	73 55.69	01/17/80
21	Keyport, NJ	40 25.73	74 09.18	08/02/80
22	Thornwood, NY	41 06.88	73 46.70	09/04/80
23	Annsville, NY	41 18.35	73 54.72	12/12/80
24	Ramsey, NY	41 06.05	74 12.20	05/18/81
25	Suffern, NY	41 07.57	74 09.36	08/18/82
26	Oldwick, NJ	40 38.91	74 46.14	02/19/83
27	Pawling, NY	41 33.12	73 39.77	02/26/83

the distribution of earthquakes, and to what extent is it an artifact of station distribution?

We estimate that the catalog of earthquakes recorded by seismic networks in the greater New York City area between January 1974 and September 1983 is complete down to m_{bLg} at least as low as 2.1 and probably as low as 1.6. The lower limit for the detection threshold (1.6) was determined by assuming that an earthquake should have an L_g wave amplitude of at least 4 mm (peak-to-peak) and a signal duration above 50 sec at three or more stations to be detected. With the exception of points offshore in the southeast corner of the region studied, no points in that region fail to be within 100 km of at least three stations. With the station distribution shown in Figure 4, an earthquake with $m_{Lg}(f) > 1.2$ would satisfy the above conditions and therefore be detected. Since the average difference between m_{bLg} and $m_{Lg}(f)$ was found to be 0.4 magnitude units, we estimate that the detection threshold for the region studied is $m_{bLg} = 1.6$.

The more conservative estimate of the detection threshold ($m_{bLg} = 2.1$) was calculated in a similar manner; the amplitude and signal duration used for this calculation were 15 mm and 70 sec, respectively. The distribution of earthquakes

TABLE 2

MAGNITUDES OF EARTHQUAKES RECORDED BY THE LDGO NETWORK IN THE GREATER NEW YORK CITY AREA ($m_{bLg} > 1.5$) (JANUARY 1974–SEPTEMBER 1983)

Event	m_c*	m_N†	m_{bLg}‡	M_L	$m_{Lg}(f)$	m (this study)
1	2.2					1.8
2	3.0	2.6	2.8 (9)	2.6§	2.4	2.7
3	2.3	2.2				1.9
4	2.0	1.1				1.6
5	2.5	2.1				2.1
6	2.1	2.0			0.8	1.7
7	2.7		2.0 (1)			2.3
8	2.6		2.0 (2)	1.8		2.2
9	2.3			<2.5§		1.9
10	2.3	2.2	2.3 (1)			1.9
11	2.2	1.9				1.8
12	2.1	2.1			1.3	1.7
13	2.2	1.6			0.9	1.8
14	2.3	2.0			1.7	1.9
15	2.4					2.0
16	2.7	2.2	2.4 (4)	1.8	2.3	2.3
17	3.2	2.7	3.0 (8)	3.3¶	2.6	2.9
18	2.8	2.2	2.4 (8)	2.4¶	1.9	2.4
19	2.6	2.2		2.2	2.0	2.2
20	2.5	2.0		1.7	1.6	2.1
21	2.7	2.5		2.6§		2.3
22	2.8	2.2		1.5	2.3	2.4
23	2.3					1.9
24	2.5					2.1
25	2.0					1.6
26	2.7		2.4 (1)			2.1
27	3.0		2.7 (1)	2.7		2.7

* m_c is the signal-duration magnitude of Chaplin *et al.* (1980).

† m_N represents magnitudes calculated by substituting Lg amplitudes from high-frequency records into Nuttli's (1973) formula for m_{bLg}.

‡ Numbers in parentheses next to m_{bLg} values represent numbers of stations used in m_{bLg} calculations.

§ M_L magnitude determined by Ebel (1982).

¶ Average of M_L determined in this study and that of Ebel (1982).

from this study with $m_{bLg} > 1.5$ and with $m_{bLg} > 2.0$ are shown in Figures 7 and 8, respectively.

Rate of seismicity and b value. Between January 1974 and September 1983, twenty-seven earthquakes with $m_{bLg} > 1.5$ occurred in the greater New York City area. This represents an average rate of 2.8 such events per year during the period of time studied. The largest event recorded in this region during the period of time studied had an m_{bLg} of 3.0.

The cumulative frequency of occurrence of earthquakes greater than or equal to a given magnitude can be expressed in terms of the well-known relation

$$\log N_c = a - bm \tag{2}$$

where N_c is the cumulative number of events of magnitude m or greater per year, and a and b are constants. Ebel (1984) determined a b value of 0.84 for earthquakes

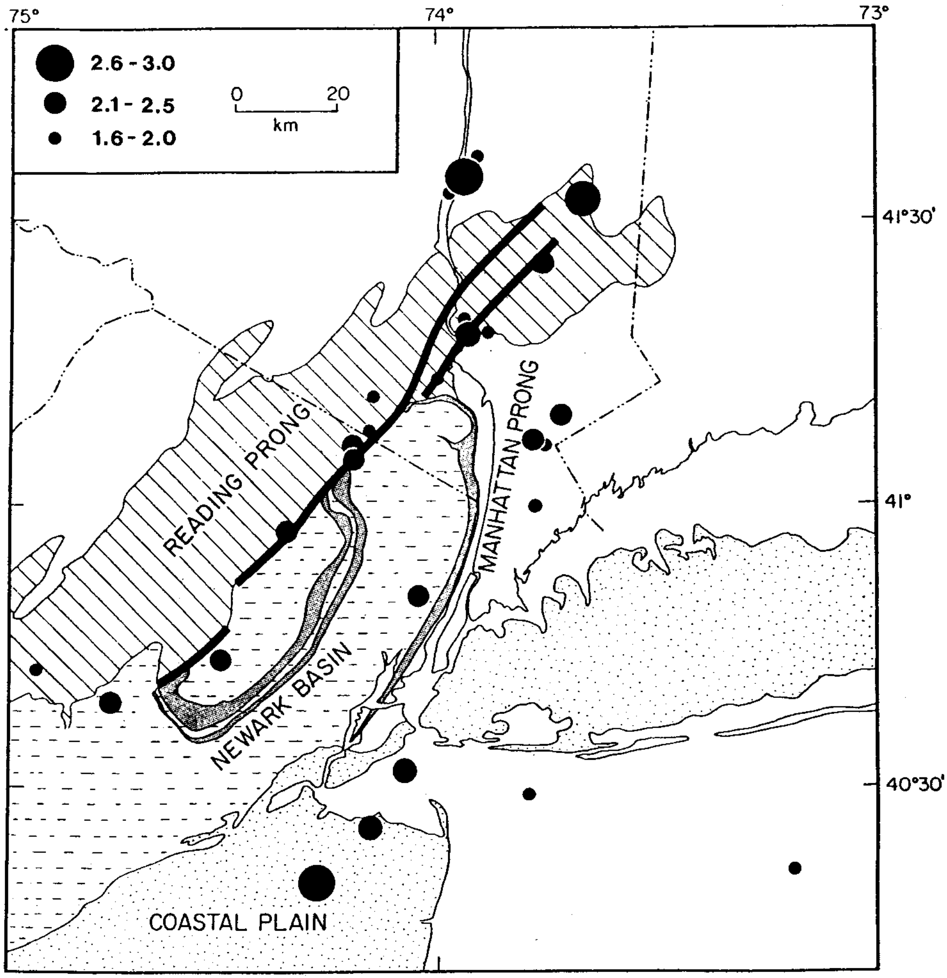

FIG. 7. Earthquakes recorded by the LDGO network in the greater New York City area (January 1974 to September 1983), plotted on a generalized map of regional geologic provinces adapted from Ratcliffe (1980). The events plotted here are also listed in Tables 1 and 2. Magnitudes shown are estimated from signal duration as described in the text. Only events with $m_{bLg} > 1.5$ are shown. The dark solid line represents the surface trace of the Ramapo fault.

recorded by local seismic networks in New England between 1975 and 1982, but he used the signal duration magnitude scale of Rosario (1979). That magnitude scale was determined by correlating m_N with signal duration and was not tied to any global or regional scale such as m_b or m_{bLg}. Aggarwal and Sykes (1978) determined a b value of 0.73 for earthquakes recorded throughout New York State, but they used m_N as a magnitude scale.

We analyzed the cumulative frequency of occurrence of the earthquakes that we studied in the greater New York City area using the magnitudes determined in this study, and we obtained a b value of 1.16 (Figure 9). We also determined the b value for a subset of this catalog, consisting of events that occurred within 10 km of the Ramapo fault. For this subset, the b value was 1.18 (Figure 9).

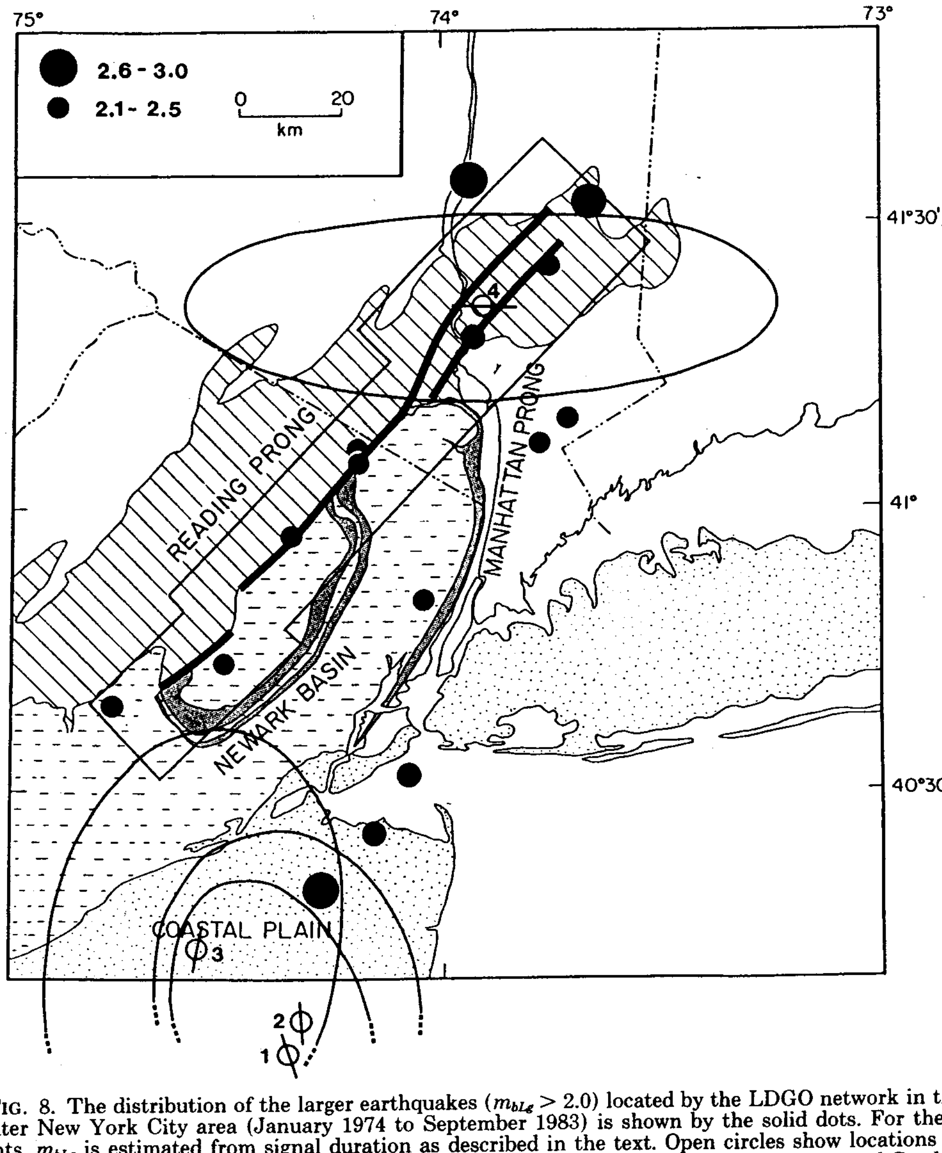

FIG. 8. The distribution of the larger earthquakes ($m_{bLg} > 2.0$) located by the LDGO network in the greater New York City area (January 1974 to September 1983) is shown by the solid dots. For these events, m_{bLg} is estimated from signal duration as described in the text. Open circles show locations of earthquakes (with error ellipses) from the early instrumental record as relocated by Dewey and Gordon (1984). Events 1, 2, and 3 occurred within a 4-hr period on 23 August 1938 and were assigned magnitudes (m_{bLg}) of 3.9, 4.0, and 3.7, respectively, by Street and Turcotte (1977). Event 4 occurred on 3 September 1951 and was assigned $m_{bLg} = 3.8$ by Street and Turcotte (1977).

Aggarwal and Sykes (1978) determined a recurrence relation for earthquakes within 10 km of the Ramapo fault that occurred from 1974 to 1977. Their recurrence relation is also shown in Figure 9 for comparison with our results. The difference between their results and ours probably reflects, more than anything else, the difficulty in estimating a and b values from such a small number of events. However, the magnitudes used in this study are probably more directly related to m_b than those used in the study of Aggarwal and Sykes (1978).

SEISMICITY AND GEOLOGIC STRUCTURES

The essential features of the surface geology in the greater New York City area are shown in Figure 7, adapted from Ratcliffe (1980). Crystalline rocks of Paleozoic to Precambrian age crop out east of the Hudson River in a "prong"-shaped geologic province that points southward toward Manhattan, i.e., the Manhattan prong. Crystalline basement rocks also crop out to the northwest of the Ramapo fault in another "prong"-shaped province of Precambrian age that points toward Reading, Pennsylvania, i.e., the Reading prong. Between these two crystalline basement

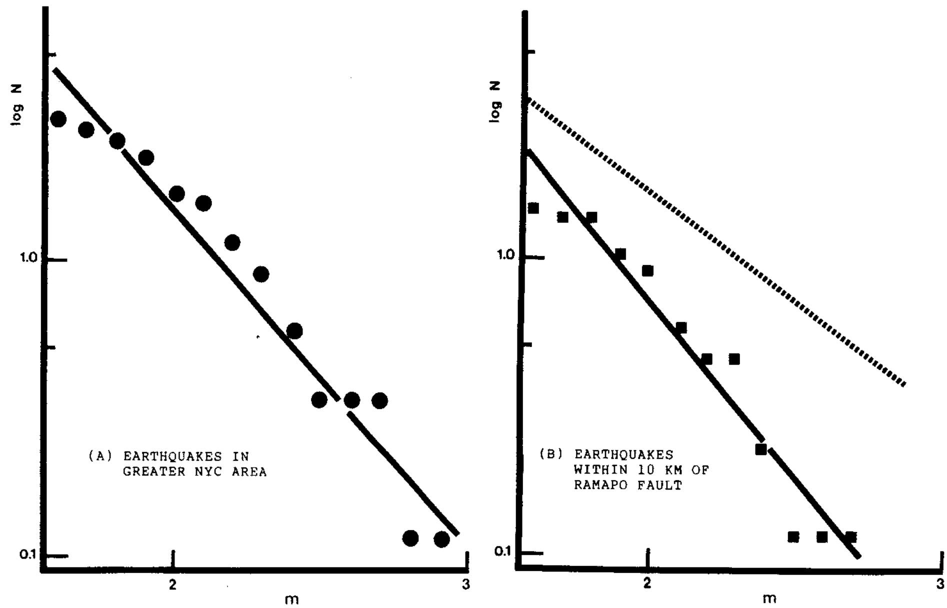

FIG. 9. Solid lines are recurrence relations for instrumentally recorded earthquakes in the greater New York City area (1974 to 1983). Magnitudes are m_{bLg} estimated as described in the text. Graph A refers to the entire region shown in Figures 7 and 8, and graph B refers to events within 10 km of the surface trace of the Ramapo fault. The dashed line in graph B is the recurrence relation determined by Aggarwal and Sykes (1978) for events within 10 km of the Ramapo fault that occurred between 1974 and 1977.

provinces lies the Newark basin, filled with Triassic sedimentary and igneous rocks. To the southeast of the Newark basin and the Manhattan prong, the basement rocks are buried beneath a thin overlap of Cretaceous and younger coastal plain sediments.

In general, the earthquakes shown in Figure 7 are located in geologic structures that surround the Newark basin, and very few events are located within the basin itself. None of the epicenters shown in that figure locate in Manhattan, Long Island, or western Connecticut, and much of the activity is concentrated within about 20 km of the edges of the Newark basin. Only three earthquakes (Wappingers Falls events) locate north of the Reading prong in the Valley and Ridge province.

In past studies of earthquake activity in this region, much emphasis was placed on the significance of the Ramapo fault and its relationship to seismicity. Early studies by Isacks and Oliver (1964) and Page *et al.* (1968) located four earthquakes in this region using arrival time data at three to four local stations. All of these events located within a few kilometers of the Ramapo fault, and Page *et al.* (1968) suggested that much of the seismic activity in southeastern New York and northern New Jersey is associated with this feature.

During the 1970s, station coverage in this region was increased, with particular emphasis on monitoring seismicity in the vicinity of the Ramapo fault. The densest station coverage is in the vicinity of the Indian Point nuclear power plant, which is located within 1 km of a major branch of the Ramapo fault system (Figures 3 and 4). With these extended capabilities, Aggarwal and Sykes (1978) and Yang and Aggarwal (1981) were able to determine locations, depths, and in a number of cases, focal mechanism solutions for the earthquakes recorded. Their results indicate that, between 1970 and 1979, about half of the earthquakes recorded by local networks in this region were concentrated in the vicinity of the Ramapo fault. In addition, each of the focal mechanism solutions from their studies showed at least one nodal plane trending north to northeast—the trend of the Ramapo and other faults in this area. Aggarwal and Sykes (1978) concluded that "seismic activity in the greater New York City area is concentrated along several northeast-trending faults of which the Ramapo fault appears to be the most active." Yang and Aggarwal (1981) concluded that the Ramapo fault system "is not the only but probably the most active fault system in the greater New York City area."

While it is hard to deny that there is a relationship between the Ramapo fault and at least *some* of the earthquake activity in this region, the association of the larger historical earthquakes with fracture along the Ramapo fault is not universally accepted. Fisher (1981), e.g., agrees that at least some of the earthquakes in this region use the Ramapo fault as a zone of weakness along which existing stresses are being released, but he argues that the distribution of earthquakes in the greater New York City area appears to indicate that other faults in the region experienced larger shocks than the Ramapo system.

Seborowski *et al.* (1982) argue that focal mechanism solutions for three earthquakes near Indian Point that they studied (events 2, 13, and 20 of this study) indicate fault planes that are transverse to the trace of the Ramapo fault. Moreover, they argue that the microearthquake seismicity (including events with m_{bLg} lower than our detection threshold) in the vicinity of Indian Point trends northwest, transverse to the trend of major geologic structures mapped at the surface.

With four more years of data than in the Yang and Aggarwal (1981) study and with a uniform detection threshold from a time period of fairly constant station configuration, we ask: How significant is the Ramapo fault as a seismic source zone when compared to other parts of the region? The most widely felt, and probably the largest known, historical earthquake in this region occurred on 10 August 1884. Did this event occur along the Ramapo fault? Is such an event (or even a larger event) possible along the Ramapo fault?

Using the lower limit of detection threshold determined above ($m_{bLg} = 1.6$), we find that 48 per cent of the events recorded during the period of time studied occurred within 10 km of the Ramapo fault (Figure 7). Using the much more conservative detection threshold of $m_{bLg} = 2.1$, we find that 53 per cent of the earthquakes occurred within 10 km of this fault (Figure 8). Three of the earthquakes from this network record shown in Figure 8 had m_{bLg} values exceeding 2.5, and one of these events occurred within 10 km of the Ramapo fault.

Also shown in Figure 8 are the locations and error ellipses of four pre-1974 earthquakes relocated by Dewey and Gordon (1984). The magnitudes (m_{bLg}) of these events were determined by Street and Turcotte (1977) and range from 3.7 to 4.0 (Figure 8). Of these four earthquakes, one occurred within 10 km of the Ramapo fault and three occurred in the New Jersey Coastal plain.

The additional data and the removal of the smaller events does not change the conclusion that about half of the microearthquakes recorded in this region occurred in the vicinity of the Ramapo fault. However, earthquakes at least as large as those located near the Ramapo fault are located 20 to 50 km from the trace of this fault in the Manhattan prong, in the Coastal Plain province, as well as offshore. Indeed, the largest event studied here occurred in the Coastal Plain province, and three of the four earthquakes relocated by Dewey and Gordon (1984) were also located in the Coastal Plain province. While the Ramapo fault cannot be ruled out as a possible source zone for earthquakes in this region, the network seismicity studied here does not necessarily indicate whether this fault is either the most active fault in the greater New York City area or if it is the fault along which the largest earthquakes will occur in the future.

Conclusions

1. We have estimated magnitudes (m_{bLg}) of earthquakes recorded by the LDGO network in the greater New York City area between January 1974 and September 1983. During this period of time, 27 earthquakes with $m_{bLg} > 1.5$ occurred in the region studied, and the largest event had an m_{bLg} of 3.0
2. We have attempted to eliminate station bias from the network seismicity in this region by estimating the appropriate magnitude threshold for detection of these events.
3. The question of whether the Ramapo fault is the cause of the larger earthquakes in the greater New York City area is still not resolved. About half of the microearthquakes recorded in this region occurred within 10 km of the fault. However, earthquakes at least as large as those recorded near the Ramapo fault are located as far as 50 km from this fault in geologic structures that surround the northern part of the Newark basin. While the Ramapo fault can by no means be ruled out as a possible source zone for earthquakes in the greater New York City area, the cause of earthquakes in this region is, in the final analysis, still unknown.
4. Since very little can be said about the actual structures associated with the larger earthquakes in this region, there is virtually no constraint on the maximum magnitude or intensity earthquake that could be experienced in this region.

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References

Aggarwal, Y. P. and L. R. Sykes (1981). Earthquakes, faults and nuclear power plants in southeastern New York and northern New Jersey, *Science* **200**, 425–429.

Chaplin, M. P., S. R. Taylor, and M. N. Toksöz (1980). A coda-length magnitude scale for New England, *Earthquake Notes* **51**, 15–22.

Chiburis, E. F., R. O. Ahner, T. Graham, P. Raica, and V. Vudler (1976–1980). *Bulletin of the Northeast United States Seismic Network*, published by Weston Observatory, Boston College, Weston, Massachusetts.

Dewey, J. W. and D. W. Gordon (1984). Map showing recomputed hypocenters of earthquakes in the eastern and central United States and adjacent Canada, 1925–1980, U.S. Geological Survey (Miscellaneous Field Studies), Map MF-1699.

Ebel, J. E. (1982). M_L measurements for northeastern United States earthquakes, *Bull. Seism. Soc. Am.* **72,** 1367–1378.

Ebel, J. E. (1984). Statistical aspects of New England seismicity from 1975 to 1982 and implications for past and future earthquake activity. *Bull. Seism. Soc. Am.* **74,** 1311–1329.

Fisher, J. A. (1981). Capability of the Ramapo fault system, in *Earthquakes and Earthquake Engineering—Eastern United States*, J. B. Beavers, Editor, 441–456.

Herrmann, R. B. and A. Kijko (1983). Short-period *Lg* magnitudes: instrument, attenuation, and source affects, *Bull. Seism. Soc. Am.* **73,** 1835–1850.

Isacks, B. and J. Oliver (1964). Seismic waves with frequencies from 1 to 100 cycles per second recorded in a deep mine in northern New Jersey, *Bull. Seism. Soc. Am.* **54,** 1941–1979.

Nuttli, O. W. (1973). Seismic wave attenuation and magnitude relations for eastern North America, *J. Geophys. Res.* **78,** 876–885.

Page, R. A., P. H. Molnar, and J. Oliver (1968). Seismicity in the vicinity of the Ramapo fault, New Jersey–New York, *Bull. Seism. Soc. Am.* **58,** 681–687.

Pulli, J.J. (1984). Attenuation of coda waves in New England, *Bull. Seism. Soc. Am.* **74,** 1149–1166.

Ratcliffe, N. M. (1980). Brittle faults (Ramapo fault) and phyllonitic ductile shear zone in the basement rocks of the Ramapo seismic zone, New York and New Jersey, and their relationship to current seismicity, in *Field Studies of New Jersey Geology and Guide to Field Trips*, 52nd Annual Meeting of the N.Y. State Geological Association, W. Manspeizer, Editor.

Richter, C. F. (1935). An instrumental earthquake magnitude scale, *Bull. Seism. Soc. Am.* **25,** 1–32.

Rockwood, C. G. (1885). Notes on American earthquakes, No. 14, *Am. J. Sci* (*Ser. 3*) **29,** 425–432.

Rosario, M. (1979). A coda duration magnitude scale for the New England seismic network, *Master's Thesis*, Boston College, Weston, Massachusetts, 82 pp.

Schlesinger-Miller, E. A. and N. L. Barstow (1983). *Bulletin of the Lamont Doherty Seismic Network in New York State and Adjacent Areas.*

Seborowski, K. D., G. Williams, J. A. Kelleher, and C. T. Statton (1982). Tectonic implications of recent earthquakes near Annsville, New York, *Bull. Seism. Soc. Am.* **72,** 1601–1609.

Smith, W. E. T. (1966). Earthquakes of eastern Canada and adjacent areas, 1928–1959, *Dom. Observ. Publ.* **26,** 271–301.

Street, R. L. and F. T. Turcotte (1977). A study of northeastern North American spectral moments, magnitudes, and intensities, *Bull. Seism. Soc. Am.* **67,** 599–614.

Yang, J. P. and Y. P. Aggarwal (1981). Seismotectonics of the northeastern United States and adjacent Canada, *J. Geophys. Res.* **86,** 4981–4998.

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