

Observations and Tectonic Setting of Historic and Instrumentally Located Earthquakes in the Greater New York City–Philadelphia Area

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Abstract A catalog of 383 earthquakes in southeastern New York, southwestern Connecticut, northern New Jersey, and eastern Pennsylvania, including metropolitan New York City and Philadelphia, is compiled from historical and instrumental data from 1677 through 2006. A magnitude-felt area relationship is used to calculate the equivalent magnitude m_bLg prior to the advent of abundant instrumental data in 1974. Revised locations are computed for a number of historic earthquakes. Most hypocenters are concentrated in older terranes bordering the Mesozoic Newark basin in the Reading, Manhattan, and Trenton prongs and in similar rocks found at a shallow depth beneath the coastal plain from south of New York City across central New Jersey. Historic shocks of m_bLg 3 and larger were most numerous in the latter zone. The largest known event, m_bLg 5.25, occurred just offshore of New York City in 1884. Many earthquakes have occurred beneath the 12-km wide Ramapo seismic zone (RSZ) in the eastern part of the Reading prong, where station coverage was the most extensive since 1974. The southeastern boundary of the RSZ, which is nearly vertical, extends from near the surface trace of the Mesozoic Ramapo fault to depths of 12–15 km. Because the Mesozoic border fault dips about 50° – 60° southeast, earthquakes of the RSZ are occurring within middle Proterozoic through early Paleozoic rocks. Which faults within the RSZ are active is unclear. Well-located activity in the Manhattan prong since 1974 extends to similar depths but cuts off abruptly at all depths along a northwest-striking boundary extending from Peekskill, New York, to Stamford, Connecticut. That boundary, which is subparallel to brittle faults farther south, is inferred to be a similar fault or fault zone. Those brittle features may have formed between the Newark, Hartford, and New York bight basins to accommodate Mesozoic extension. The Great Valley in the northwestern part of the study region is nearly devoid of known earthquakes. While few focal mechanism solutions and *in situ* stress measurements of high quality are available, the maximum compressive stress is nearly horizontal and is oriented about $N64^\circ E$, similar to that in adjacent areas. The catalog is likely complete for events of $m_bLg > 5$ since 1737, ≥ 3.5 since 1840, and ≥ 3.0 since 1928. Extrapolation of the frequency-magnitude relationship indicates that an event of $m_bLg \geq 6.0$ is expected about once per 670 yr.

Online Material: Observations and tectonic setting of historic and instrumentally located earthquakes in the greater New York City–Philadelphia area.

Introduction

The purpose of this article is twofold: to present a catalog of all known earthquakes in the greater New York City–Philadelphia area from 1677 through 2006 and then to describe those events in a context that includes geologic terranes, faults, and contemporary state of stress in this highly populated intraplate region. We describe the sizes of all events in terms of a common magnitude scale, m_bLg , and discuss the completeness of the catalog as a function of time

for various magnitudes. We end with a brief discussion of earthquake hazards and risks.

Better understanding of seismic hazard and risk in the region, especially better knowledge and calibration of the preinstrumental and limited instrumental records prior to the advent of a more extensive network in 1974, is very important. The 2° by 2° area examined (Figs. 1–3) has a relatively low earthquake hazard but high vulnerability and,

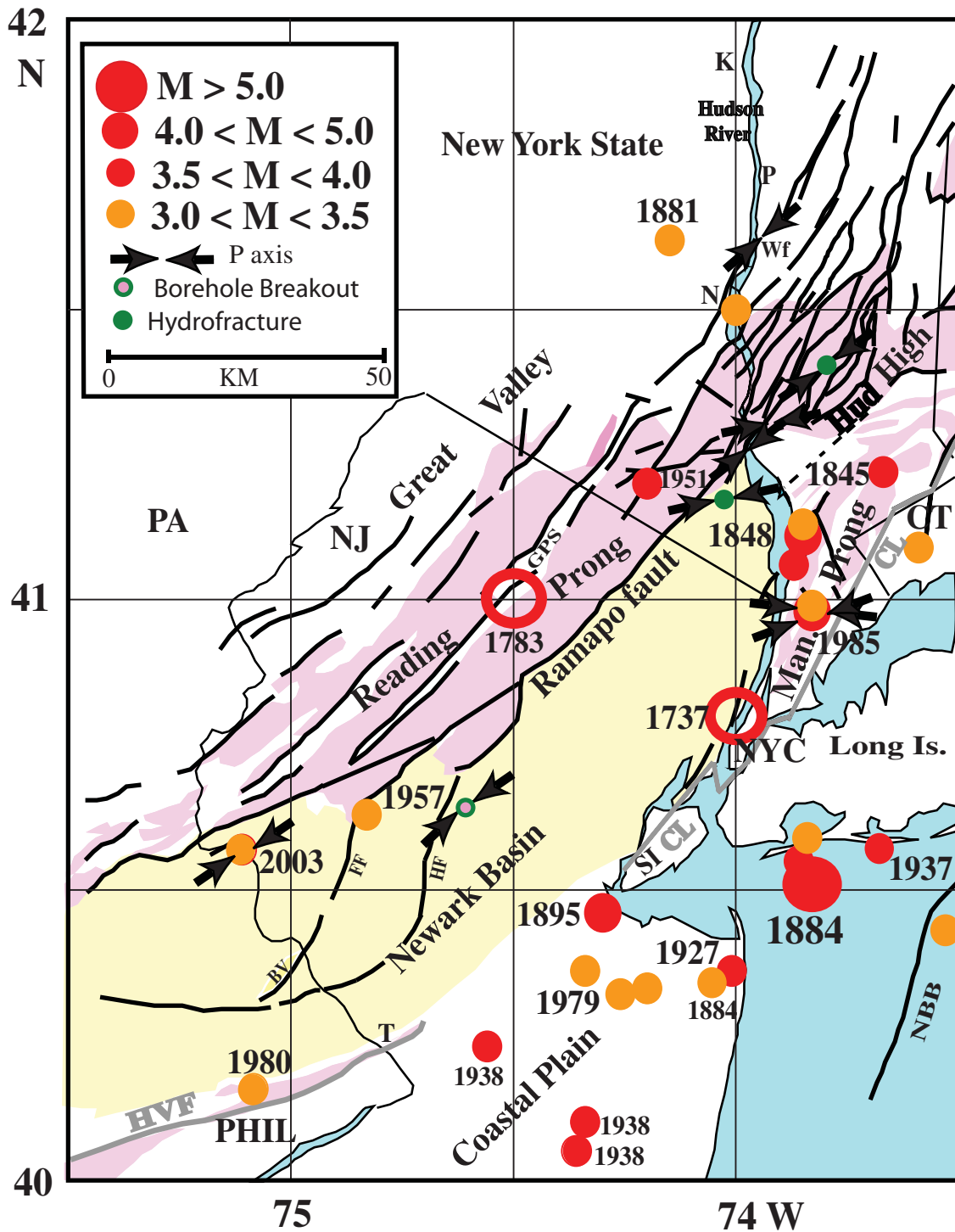


Figure 1. Known earthquakes of $M \geq 3.0$ in the greater New York City–Philadelphia area from 1677 through 2006. Magnitude, m_{best} , is $m_b Lg$ or its equivalent, as explained in text. Epicenters of large events of 1737 and 1783 maybe uncertain by 100 km and are shown as open circles. No events occurred behind the legend inset. Pink denotes Pre-Cambrian rocks; yellow denotes Mesozoic rocks of the Newark basin. Place names: Kingston, New York, K; New York City, NYC; Philadelphia, PHIL; Newburgh, New York, N; Poughkeepsie, New York, P; Staten Island, New York, SI; Trenton, New Jersey, T; and Wappinger Falls, New York, Wf. Geological features: Buckingham Valley, BV; Cameron’s Line, CL; Flemington–Furlong fault, FF; Green Pond syncline, GPS; Hopewell fault, HF; Hudson Highlands, Hud High.; Huntingdon Valley fault, HVF; Manhattan prong, Man. Prong; and New York bight basin, NBB. Major faults and geology are from Prucha *et al.* (1968), Fisher *et al.* (1970), Ratcliffe (1971, 1976), Isachsen and McKendree (1977), Ratcliffe (1980), Hall (1981), Hutchinson *et al.* (1986), Ratcliffe *et al.* (1986), Lytle and Epstein (1987), Seeber and Dawers (1989), Ratcliffe (1992), Drake *et al.* (1996), and Olsen *et al.* (1996). Horizontal projections of the P axes of better-determined focal mechanisms of earthquakes and directions of maximum horizontal compressive stress from two sets of hydrofracture (Zoback *et al.*, 1985; Rundle *et al.*, 1987) and one set of borehole breakout experiments are indicated by inward-pointing arrows (stress data in © Table 4 in the electronic edition of BSSA).

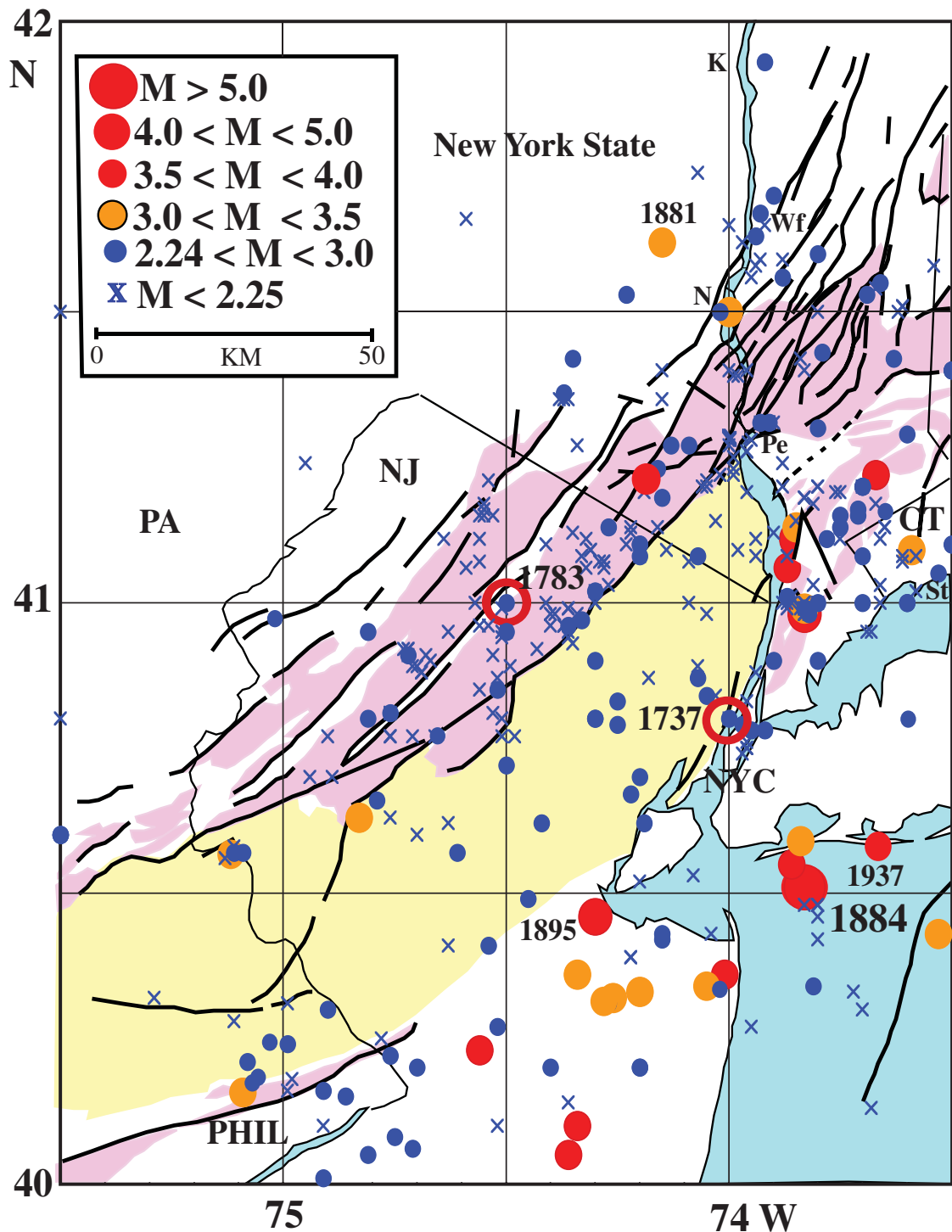


Figure 2. Entire catalog of known earthquakes in greater New York City–Philadelphia area from 1677 through 2004. No events occurred behind the legend inset. Rock units, faults, magnitudes, open circles, and place names are the same as in Figure 1. Peekskill, New York, Pe; Stamford, Connecticut, St.

hence, high seismic risk (Tantala *et al.* 2003). New York City, Newark, Trenton, and Philadelphia as well as their highly populated surrounding areas are located in the study area. The population of the area was 21.4 million in 2005 (G. Yetman, personal comm., 2007, based on U.S. census data). Better knowledge of seismic hazard is relevant to a

number of major construction projects including a replacement for the aging Tappan Zee Bridge, which carries the New York State Thruway over the Hudson River, and to the proposed extensions of the operating licenses for two nuclear power plants at Indian Point. Filled land poses greater earthquake hazards for large portions of several cities while

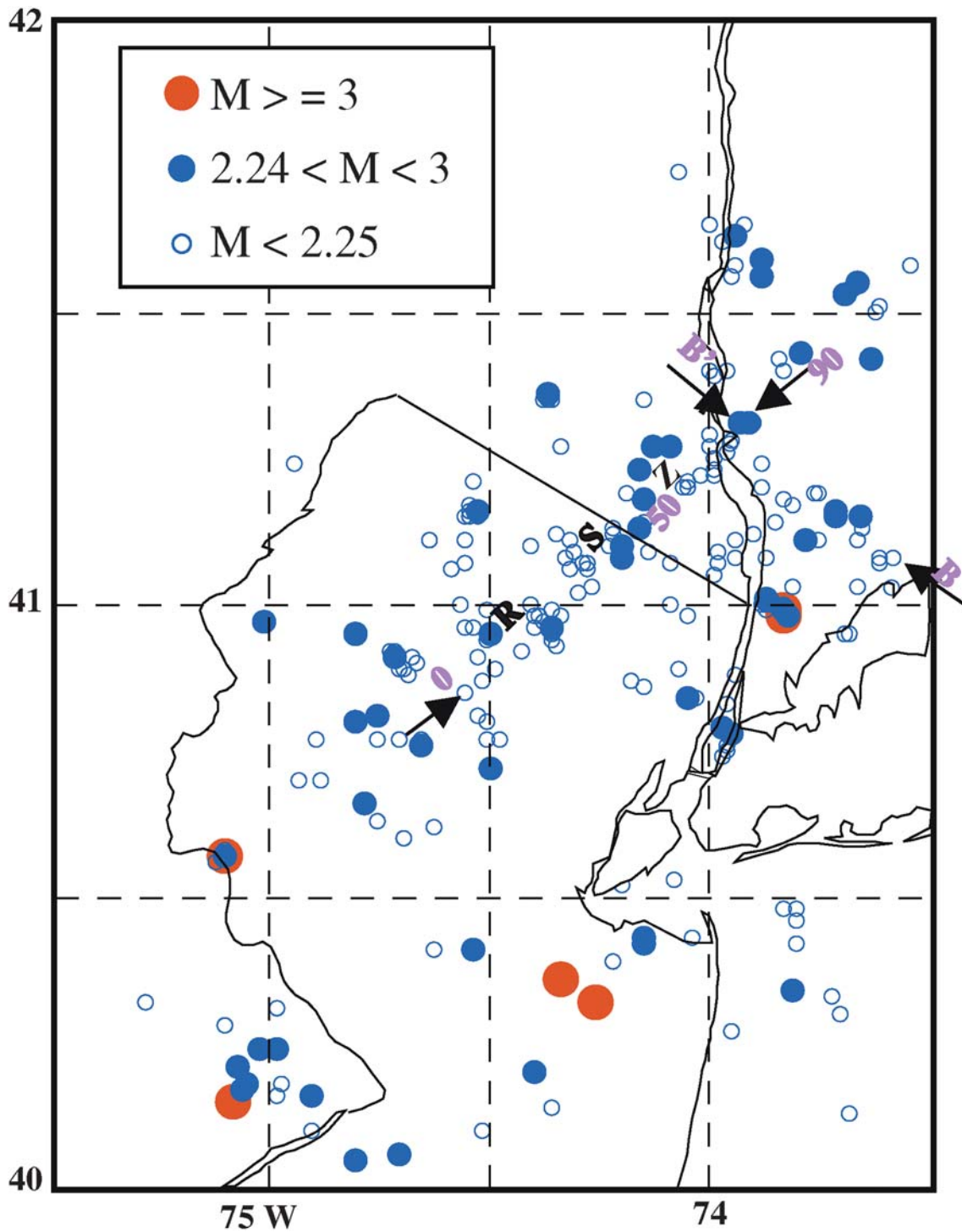


Figure 3. Instrumental locations of earthquakes from 1974.0 to 2007.0. Arrows denote approximate southeastern boundary of the RSZ and northwest-striking seismic boundary BB' between Stamford, Connecticut, and Peekskill, New York. Purple numerals denote the distance along the Ramapo zone.

other parts of the study area consist of hard rock at or close to the surface (Tantala *et al.*, 2003).

The Lamont–Doherty Earth Observatory in conjunction with several local institutions has operated a network of three or more seismograph stations in the greater New York City area since 1962 (Isacks and Oliver, 1964; Page *et al.*, 1968).

Coverage by a more extensive network extends from 1974 to the present. Limited instrumental data extend back to the 1930s. While the largest earthquakes since the 1930s were of about M 4, the historic record is much longer and includes three events larger than M 5. Historic activity of $M > 5$ has been higher in southeastern New York and northern New

Jersey than in many other areas of the central and eastern United States. It stands out as a bull's-eye on seismic hazard maps (e.g., Frankel *et al.*, 2005). While those hazard maps indicate higher values near Charleston, South Carolina, New Madrid, Missouri, and northern New York State and comparable values in southern New Hampshire, those four areas have much smaller populations and assets at risk than the study region.

Which faults are active in this intraplate region has been a subject of ongoing debate. Several faults displace Mesozoic sedimentary and igneous rocks, but evidence of faulting in younger sediments of the coastal plain and in postglacial sediments is either missing or debatable. We find that earthquakes originate from many faults rather than a few single major faults. Nevertheless, seismicity is not pervasive through our study area but is concentrated in prominent zones, such as the Ramapo seismic zone (RSZ in Fig. 3) and the Manhattan prong (Fig. 1). The southeastern side of the RSZ forms an abrupt boundary that extends to depths of 12–15 km. While that boundary intersects the surface trace of the Mesozoic Ramapo fault, its dip is generally steeper, placing it within older footwall rocks. The strikes and dips of contemporary individual faults within the RSZ are uncertain. Seismic activity within the Manhattan prong also extends from near the surface to depths of about 12–15 km. A surprising new result based on 34 yr of instrumental data is that activity in the Manhattan prong cuts off abruptly along a nearly vertical, northwest-striking boundary that extends from Stamford, Connecticut, to Peekskill, New York. This boundary is subparallel to the youngest brittle faults in the Manhattan prong, some of which are seismogenic. The eastern Hudson Highlands to the northeast, also an area of older strong rocks, is nearly aseismic.

Those who are mainly interested in the geologic and tectonic setting of earthquakes and/or hazards and risk may skip the next sections on compilation of the catalog, its completeness, magnitude determination, and rates of earthquake occurrence.

Compilation of the New Catalog

One aim in this article is to provide a common measure of earthquake size over the 330 yr for which a variety of parameters were reported or measured, such as the size of the felt area, maximum intensity, and several different seismic magnitudes. These parameters were then normalized in terms of the magnitude scale m_bLg as determined from Lg surface waves for frequencies near 1 Hz as described by Nuttli (1973). Lg waves are typically the largest seismic waves recorded at regional distances from earthquakes in the eastern and central United States and adjacent parts of Canada. Much better data on Lg are available for calibration in those regions compared to M_w , the magnitude derived from seismic moment. (E) Tables 1–4 in the electronic edition of BSSA contain the catalog, which lists various measures of size and

other pertinent data, the values of m_bLg and felt area used for magnitude calibration, and stress determinations.

Data Sources

Sykes (1986) compiled a catalog of hypocenters, origin times, magnitudes, and felt information for known earthquakes in the study area through 1983 from the literature and instrumental data from the network of stations operated by Lamont–Doherty. Only an abstract of that work was published. Seeber and Armbruster (1991) published a catalog of central and eastern U.S. earthquakes of $M \geq 3$ using improved intensity data. The present catalog updates those studies through 2006 using instrumental data from the Lamont Cooperative Seismic Network (e.g., Schnerk *et al.*, 1976; Seeber and Armbruster, 1986, 1991; Seeber *et al.*, 1993). It includes many earthquakes of $M < 3$. Locations and other parameters were obtained from detailed studies of several individual earthquakes and their aftershocks by Sbar *et al.* (1970), Pomeroy *et al.* (1976), Sykes (1976), Yang and Aggarwal (1981), Seborowski *et al.* (1982), Dewey and Gordon (1984), J. W. Dewey and D. W. Gordon (personal comm., 1984), Kafka *et al.* (1985), Quittmeyer *et al.* (1985), Thurber and Caruso (1985), Seeber and Dawers (1989), Kim (1998), and Seeber (2004).

Information on pre-1974 earthquakes, especially felt reports, were obtained from Rockwood (1872–1886), Smith (1962, 1966), the yearly series *United States Earthquakes* from 1928 to 1986 of the U.S. Department of Commerce (1968) and U.S. Geological Survey, Sykes (1976), Winkler (1979), Nottis (1983), Nottis and Mitronovas (1983), Seeber and Armbruster (1986, 1988, 1991), Stover and Coffman (1993), and Wheeler *et al.* (2005). We collected additional intensity information for several events from newspapers, which we used in revising their epicentral locations.

Revised locations for nearly all events from 1951 to 1974 of $m_bLg > 2.5$ are based on a combination of felt reports and at least some instrumental data as are those for three earthquakes in central New Jersey of m_bLg 3.8–3.9 in 1938 (Street and Turcotte, 1977; Dewey and Gordon, 1984; J. W. Dewey and D. W. Gordon, personal comm., 1984). Locations and origin times for nearly all events since 1974 are based on instrumental data from local and regional stations as reported in the series *Regional Seismicity Bulletin of the Lamont–Doherty Network* (e.g., Schnerk *et al.*, 1976), final reports on the Lamont network (e.g., Seeber *et al.*, 1993), and the series *Northeastern U.S. Seismic Network* (e.g., Chiburis and Ahner, 1976). Arrival times were reread in revising the locations and m_bLg for nine earthquakes between 1951 and 1976 (asterisks in (E) Table 1 in the electronic edition of BSSA). Other arrival times were not reread.

Quarry and other industrial explosions were eliminated by record analysts as much as possible and are not included in the new catalog. Pomeroy *et al.* (1976) identified a series of very shallow earthquakes at Wappinger Falls, New York (Fig. 1), that were triggered by the removal of a vertical load

at a large quarry. They are included in the catalog and figures. Seeber *et al.* (1998) identified a similar sequence in Pennsylvania to the west of the study area in 1994. While those events were identified as earthquakes, small triggered shocks at or near some other quarries may well have been missed by record analysts. ☉ Table 2 in the electronic edition of *BSSA* describes 15 events that were identified as either having nonearthquake sources (e.g., Nottis, 1983) or resulting from erroneous listings and duplications in previous catalogs.

Magnitude Determination

The felt area offers a measure of size for earthquakes without either any or good instrumental recordings. Earthquakes with both felt areas and good instrumental data (18 in ☉ Table 3 in the electronic edition of *BSSA*) were used for calibration of those earthquakes. Measurements of m_bLg at stations in eastern North America were made for events in the study area from 1951 to 1984 at frequencies near 1 Hz following the methodology of Nuttli (1973). We used analog records from the World-Wide Standardized Seismograph Network and similar instruments with peak magnification near 1 Hz, which, unlike many shorter-period recordings that became available in the 1970s, did not need to be filtered to measure 1 Hz Lg . We also determined felt areas for as many of these events as possible. Felt area– m_bLg values for those earthquakes along with those for the sequence in central New Jersey in 1938 (Street and Turcotte, 1977) and the 1985 Ardsley, New York, shock of m_bLg 4.2 (Kim, 1998) are shown as red circles in Figure 4. While some events as small as m_bLg 2 are felt, felt area decreases rapidly for $m_bLg < 2.5$. Also, Lg near 1 Hz often was difficult to measure even at one to a few stations for $m_bLg < 2.5$.

Because several historic events in our study area were larger than the largest instrumental event (m_bLg 4.2), it was necessary to extend the calibration using measurements from other areas with similar attenuation characteristics. Lg propagates efficiently in eastern North America, including the study area, but not so in tectonically active regions of western North America. The other values of m_bLg in Figure 4 are from Nuttli and Zollweg (1974), Sbar *et al.* (1975), Street (1976), Street and Turcotte (1977), Bollinger (1979), Street and Lacroix (1979), Kim (1998), and Seeber *et al.* (1998). The felt areas of those events are either from them, yearly issues of *United States Earthquakes*, or the Community Internet Intensity Working Group (2006). Because a linear relationship does not fit the entire data set in Figure 4, a series of solid lines were fit by eye. They were then used to derive an equivalent Lg magnitude (m_f) from felt area. In calculating m_f , however, we did not take into account population density, time of day, hypocentral depth, or site response.

The regional magnitude scale M_L , which was originally developed for southern California, was modified by Ebel (1982) and Kim (1998) to account for much smaller seismic attenuation in eastern North America. Kim (1998) finds

$m_bLg = M_L + 0.15$ for $2 \leq M_L \leq 6.5$ in eastern North America. Figure 5 shows that his relationship also holds with little scatter for earthquakes of $2 < m_bLg < 4.2$ in the study area (red circles) and for larger events reported by him in adjacent areas. $M_L + 0.15$ and m_bLg were used in determining a single best estimate of equivalent Lg magnitude, which we call m_{best} . This expands the list of reliable measures of m_bLg because only M_L was determined for nine of the events in ☉ Table 1 in the electronic edition of *BSSA*. Either m_bLg or M_L or both were determined for all but one event (an aftershock) of $m_{best} > 3.0$ in the catalog since 1937. We give either measure the highest weight in estimating m_{best} . We note that estimates of M_L made by Smith (1966) for eastern North America generally are too large because he used attenuation values appropriate to southern California.

Intensities and other magnitudes were reported for many of the earthquakes in the study area. The four subparts of Figure 6 compare magnitudes determined from Lg waves (m_bLg), M_L , felt area (m_f), coda duration (M_c), higher-frequency Lg (M_n), and maximum reported intensity on the Modified Mercalli (MM) scale. Unless stated otherwise, m_bLg was measured in a narrow frequency band near 1 Hz. Because m_f was determined from the log of the felt area using the relationship in Figure 4, it is understandable that it is nearly equivalent to m_bLg between m_f 2.2 and 4 (Fig. 6a). Hence, m_f offers a measure of size for preinstrumental events that is reasonably consistent with m_bLg .

A magnitude scale M_c based on signal (coda) duration of higher-frequency seismic waves was developed for New England by Chaplin *et al.* (1980). An advantage of M_c is that it often can be determined for very small events. Most reported magnitudes for the Lamont–Doherty network since 1982 are M_c as are some values as early as 1973. They were calculated using the parameters of Chaplin *et al.* (1980). Figure 6b indicates considerable scatter in m_bLg , m_b , and M_L as a function of M_c , especially for $M_c \leq 3$. Using data from 1974 to 1983 Kafka *et al.* (1985) found that M_c overestimates m_bLg (1 Hz) by about 0.1–0.2 units. In determining m_{best} we take $M_c = m_bLg$ but give it less weight than either m_bLg , M_L , or m_f .

Figure 6c indicates that M_n , a measure of m_bLg at higher frequency, also scatters considerably with respect to either m_f , M_L , or m_bLg (1 Hz.). Hence, we give small weight to M_n in calculating m_{best} . It is much larger than those three magnitudes for $M_n > 3$. The two largest values of M_n (Fig. 6c) were from a sequence at Abington, Pennsylvania, near Philadelphia in 1980 (Bischke *et al.*, 1980). For those two we took $m_{best} = m_f$ and then used the relative values of M_n to obtain m_{best} for four other events in that sequence. Six remaining events from 1973 to 1981 that did not have other measures of magnitude were not larger than M_n 2.2–2.8; 27 others were smaller. Only one of those was felt and it only at a single location. For those 33 events we took $m_{best} = M_n$. Inclusion of two values of M_n 2.8 and four of M_n 2.4–2.6 may contribute to the somewhat larger rates of small events

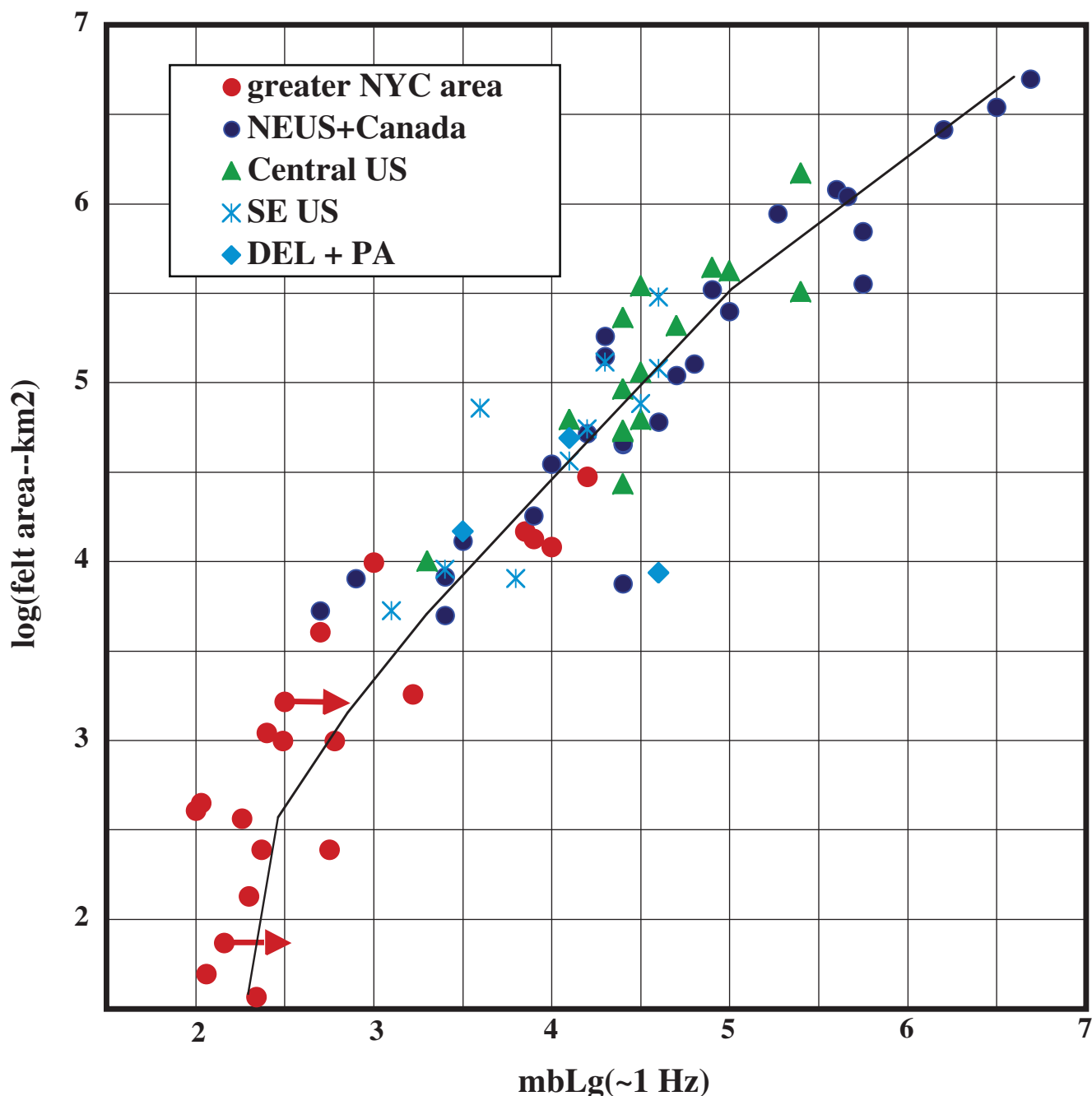


Figure 4. Log of felt area as a function of $m_b Lg$ as determined for frequencies near 1 Hz using methodology of Nuttli (1973). Black solid line is fit by eye to data. Two red arrows indicate $m_b Lg$ may be minimum values. Events and sources of data from © Table 3 in the electronic edition of *BSSA*.

between those dates in the next section on the completeness of the seismic record.

In contrast to m_f , maximum intensity, MM, scatters greatly as a function of either $m_b Lg$ or M_L and is clearly a poor measure of earthquake size (Fig. 6d versus 6a). This difference is in part a depth effect. The depth of the source has little effect on felt area but has a major effect on MM, particularly for small earthquakes that radiate rapidly attenuating high-frequency waves. Inconsistent sampling of mezo-

seismal areas is another likely factor. MM is used to determine m_{best} only when other measures were not available. Most of these events were reported from only one location and all of them have $MM \leq IV$. Many of these events were probably too small to be reported from multiple localities, but they were likely shallow and hence felt only at very short distances. Their size, therefore, will be overestimated if derived from an MM-magnitude scale developed for a representative range of depths. In our region magnitudes estimated

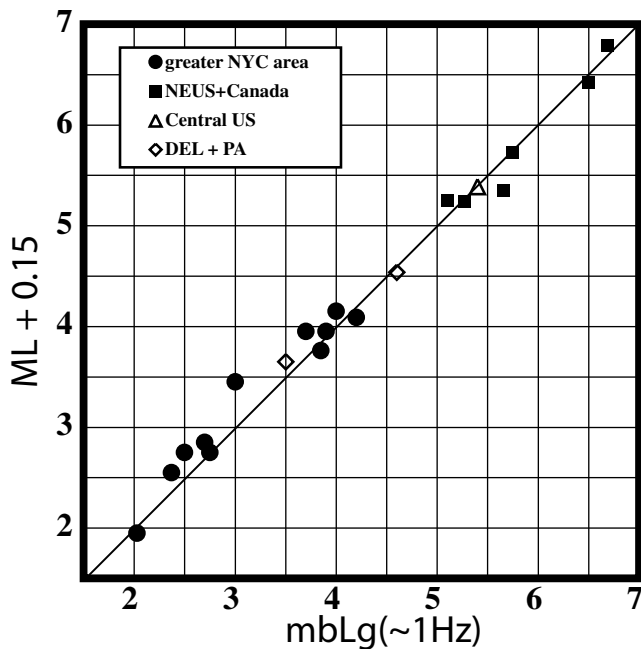


Figure 5. M_L as a function of $m_b Lg$ for events in the study area (filled circles) plus determinations from Kim (1998) for larger shocks in adjacent areas. Measurements of M_L for the study area from Ebel (1982), Ebel *et al.* (1986), and Kim (1998). Values are listed in ⑤ Tables 1 and 3 in the electronic edition of *BSSA*.

from MM using formulas in the literature typically are systematically too high. For our area we estimate the following values of m_{best} using Figure 6d: 2.7 for MM IV, 2.4 (more populated areas) to 2.6 (less populated) for MM III, and 2.2 for MM II. A few small events that were not felt (MM I) and lack any measures of size were assigned default magnitudes of 1.5–2.0.

In summary, the (best) values of magnitude, m_{best} , were obtained from all available measurements of earthquake size weighting them progressively less in the following order: $m_b Lg$ (from seismic waves near 1 Hz), M_L , m_f (from total felt area), M_c (from coda duration), M_n (from seismic waves at 2–10 Hz), and MM (from maximum intensity).

Earthquake Rates and Completeness of the Catalog

Figure 7 shows rates of occurrence of earthquakes for magnitudes greater than or equal to eight different values of $m_b Lg$. Each diagram is a log-log plot where time, T , extends back in time before 2005 on the horizontal axis as does the cumulative rate of occurrence, N/T , on the vertical axis. N is the cumulative number of events counted back in time from 2005. T and N/T are counted back in time because records typically are more complete closer to the present. A horizontal line on each diagram of constant N/T indicates that class of events is complete back to a certain date. In Figure 7a, for example, 13 events of $m_b Lg \geq 3.5$ occurred in the 165-yr period 2005–1840. Going further back in time, the decrease in cumulative values of N/T indicates the record of events of

that size is likely not complete prior to about 1840. A slope of minus one indicates a record that is totally incomplete before a certain date. The lines on each subfigure were drawn by eye, emphasizing larger values of N and, of course, deemphasizing counts of one to a few events.

The record for $m_b Lg > 3.0$ is judged complete for about the 78-yr interval 2005–1928, (since the start of the yearly series *U.S. Earthquakes*), that ≥ 4 since 1840, and that ≥ 5 for the last 270 yr, that is, since the 1737 event. The records for $m_b Lg < 3.0$ are not complete prior to about 1974, the start of the more extensive local network. From about 1975 to 1982 rates for those smaller magnitudes were about a factor of 1.5–2 times higher than those from 1983 to 2005. During that earlier interval funding for data analysis was at its peak as were the number of scientists, students, and record analysts working on the data. More marginally recorded and felt events likely were identified then. Different rates of occurrence of small events also may be attributed in part to the several different magnitude scales in use (Fig. 6), the transition from analog to digital recording, more reliance upon a triggered digital system, and a gap in the publication of felt reports by the U.S. Geological Survey.

Figure 8 shows the cumulative number of events per year greater than a given $m_b Lg$ as a function of that magnitude. The rates used for each data point were based on the values of T in Figure 7 for which the record was judged complete. A negative slope, or b -value, of 0.70 ± 0.13 was obtained by the maximum-likelihood method of Aki (1965). The a -value of 1.37 in the recurrence relationship in Figure 8 becomes $a' = -3.20$ when N/T is divided by the area of Figure 1 in square kilometers. Extrapolating the recurrence relationship to $m_b Lg$ 6.0 and 7.0 gives repeat times of about 670 and 3400 yr but with an uncertainty that increases with magnitude. Nearly all of the known earthquakes, however, are confined to about 60% of the area of Figures 1–3, where we think future large earthquakes are most likely to occur.

Distribution of Earthquakes

Figure 1 shows our study area and earthquakes in it through 2006 of $m_{best} > 3.0$. It is intended to give a perspective on the relative distribution of earthquakes in the entire area that is influenced the least by population density and distribution of seismic stations. As discussed earlier, events of $M \geq 3$ appear to be complete since 1928, those ≥ 4 appear to be complete since 1840, and those ≥ 5 appear to be complete since 1737. None of the magnitudes in Figure 1 are based on maximum intensity or the high-frequency magnitude M_n . Events of $M 3$ are typically felt over an area of about 2500 km², ensuring their likely detection well before the advent of instrumental data. Figure 2 includes all shocks in the same area from 1677 through 2004 regardless of the accuracy of their locations or completeness for various magnitude classes. Both figures indicate Mesozoic rocks of the Triassic–Jurassic Newark basin in yellow and Precambrian rocks of the Reading prong, Hudson Highlands, and the

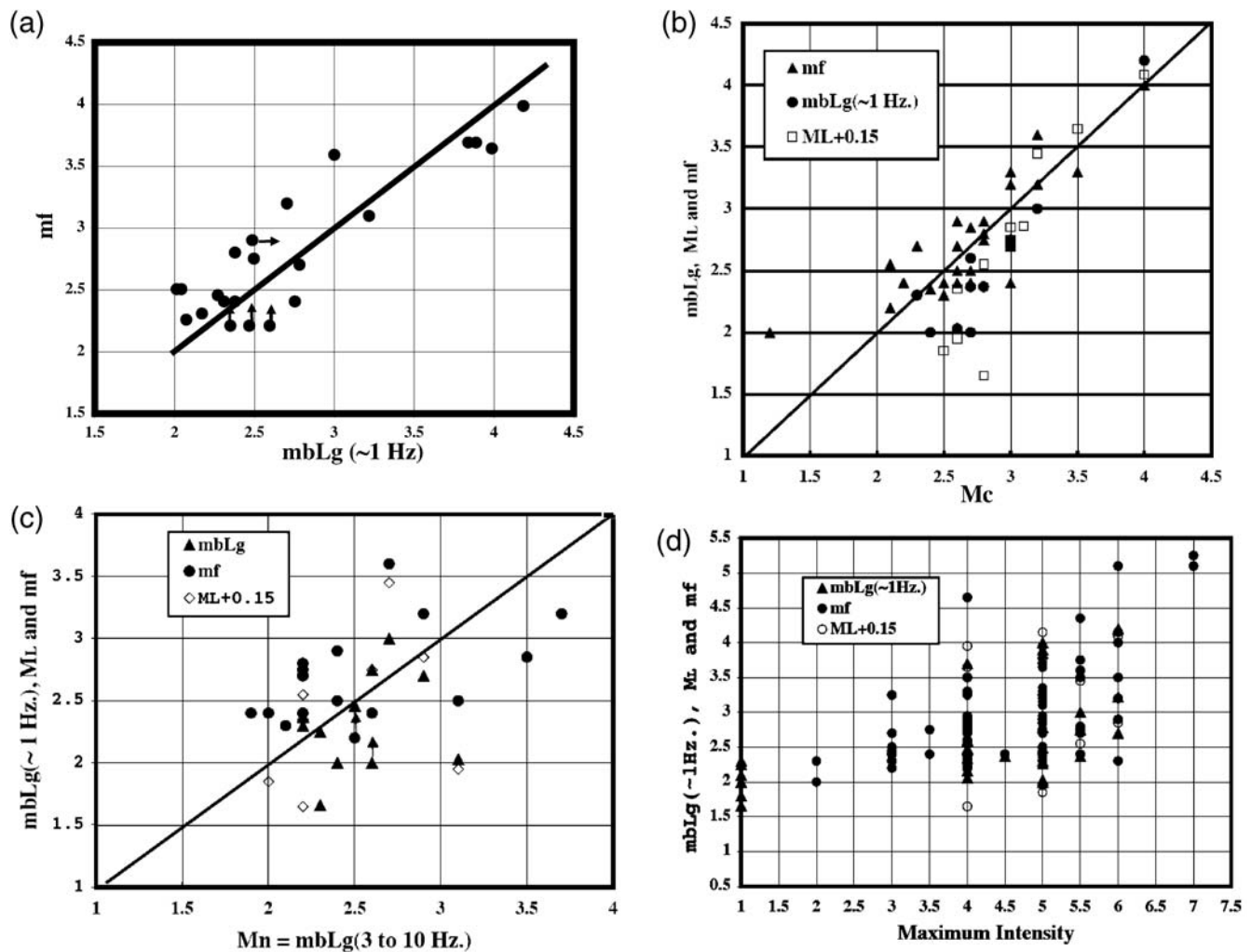


Figure 6. Relationships between various magnitudes for earthquakes in the study area. Solid lines denote equal values of quantities on two axes. (a) m_f determined from felt area using the relationship in Figure 4 as a function of m_bLg . Arrows indicate m_f or m_bLg values likely are minima based on poorly defined felt area or small seismic signals. (b) m_f , m_bLg , and M_L as a function of coda-length magnitude, M_c . (c) m_bLg measured near 1 Hz; M_L and m_f as a function of M_n (m_bLg determined between about 3 and 10 Hz). (d) m_bLg , M_L , and m_f as a function of maximum intensity on MM scale. Intensity 1 indicates not felt.

Manhattan and Trenton prongs in pink. The first five events in the catalog from 1677 to 1729 were felt at single localities in western Connecticut. Hence, for most purposes the seismic history starts with the large shock of 1737 of m_f 5.1. No events are known from the period of Dutch ownership of much of the area.

Earthquakes in the study area are not distributed uniformly in space but are concentrated in zones that generally follow the northeasterly trend of Appalachian orogenies and correlate with geologic provinces or terranes. The greatest activity in Figure 1 occurs in a belt about 35 km wide to the east and southeast of the Newark basin. The largest historic shock, m_f 5.25 in 1884, occurred along that zone. Intensity data for it and aftershocks constrain their epicenters to within 30 km of Lower New York Bay (Seeber and Armbruster, 1986). The 1884 events and the New Jersey shock of 1895 of m_f 4.15 were relocated using data based on searches

of newspapers and other documents (Armbruster and Seeber, 1986; Seeber and Armbruster, 1986, 1988). Those events and the Westchester shocks of 1845 (m_f 3.75), 1848 (m_f 4.35), and 1885 (m_{best} 4.07) were situated along the same belt as was a sequence of three shocks beneath the coastal plain of New Jersey in August 1938. The eastern belt also includes our revised location for the July 1937 earthquake of m_{best} 3.5, which moved it from west-central Long Island to near its southwestern coast.

Another belt of activity about 30 km wide (Figs. 2 and 3) is located to the northwest of the Newark basin and includes the 1881, 1951, 1957, and 2003 shocks (Fig. 1) plus numerous smaller earthquakes. Most of those events, especially those located instrumentally (Fig. 3), are concentrated in the eastern Reading prong between the Green Pond syncline and the Mesozoic Ramapo fault (Fig. 1). A special study of the 1895 event relocated it from the northwestern to the south-

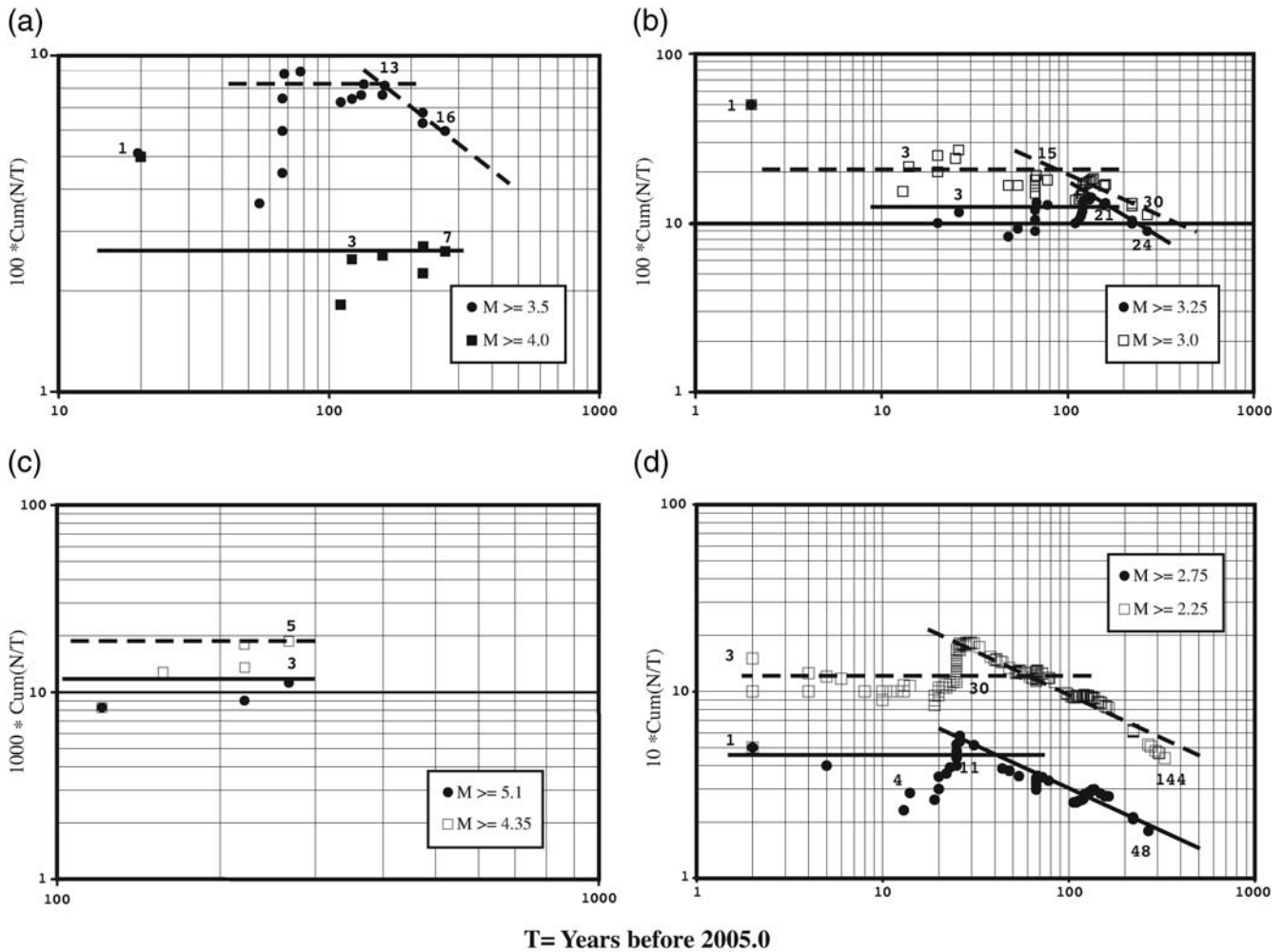


Figure 7. Cumulative number of earthquakes of a given magnitude class per year where time, T , extends backward from 2005.0. Sub-figures (a) to (d) show results for various magnitude classes. Horizontal lines indicate rates for periods of completeness. Lines of negative slope on log-log plots indicate that class is incomplete before a certain date, that is, prior to about 1840 for $M \geq 3.5$. Number beside a selected data point indicates cumulative number at various times.

eastern side of the Newark basin (Seeber and Armbruster, 1988). Unfortunately, intensity data for the early historic earthquakes of m_f 5.1 in 1737 and m_f 5.1 and 4.65 two hours apart in November 1783 do not constrain their epicenters to better than about 50–100 km. We retain the locations for them used by Stover and Coffman (1993) (open circles in Figs. 1 and 2).

While the overall distributions of earthquakes in Figures 1–3 are similar, several significant differences should be noted. Most of the smaller earthquakes in the Newark basin in Figure 2 occurred prior to 1974; many of them were felt at a single locality. The population of the basin has long been much higher than that of the mountainous Reading prong–Hudson Highlands. Hence, the record of activity in Figure 2 for the entire period since 1677 likely overportrays the rate of activity in the Newark basin relative to that of those sparsely populated regions. The distribution of events of $M > 3$ (Fig. 1) is not likely to be as biased. Epicenters of pre-1974 shocks are not as precisely located as those deter-

mined thereafter. Depths of nearly all pre-1974 events are either poorly determined or unknown.

Instrumental Locations

Instrumental epicenters from 1974 through 2006 (Fig. 3, same area as in Figs. 1 and 2) are located more precisely (about ± 2 km) than those based solely on intensities. Furthermore, instruments have detected many smaller events; about 71% of the earthquakes in the catalog occurred since 1974. Consequently, more and better data are available now than before 1974. Instrumental locations are relevant to the fine structure of the seismicity, (purposely shown free of other data in Fig. 3), but their relative numbers within the study area depend in part on station distribution. For several decades the location capability of the local seismic network was strongest for events between about 40.4° N and 41.5° N in northern New Jersey and southeastern New York State. (The distribution of stations at various times is shown in

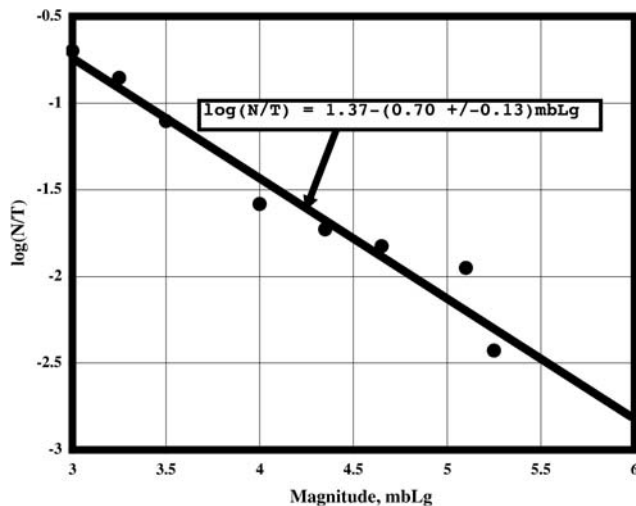


Figure 8. Cumulative frequency ($\log N/T$) as a function of magnitude, m_bLg . For each magnitude class, the number of events N is calculated for time interval T for which it is complete (Fig. 7). Slope or b -value is maximum-likelihood estimate with 68% confidence limits.

Schnerk *et al.* [1976], Kafka *et al.* [1985], and Seeber *et al.* [1993]). Coverage south and southwest of New York City in Figure 3 has not been as good as that farther north. Location capability has improved in the last decade with the installation of stations in New York City, Pennsylvania, and Delaware. No stations exist, however, in the coastal plain of New Jersey where the rate of activity is likely under represented in Figure 3.

Depths of Earthquakes

The distribution of seismic stations for much of the study area since 1974 permitted most focal depths to be estimated with a precision of a few to 5 km. It is likely poorer in the peripheries of Figure 3, especially offshore. Hence, some depths are determined with higher precision than others. Figure 9 shows a histogram of the number of instrumentally located earthquakes as a function of depth. Ninety-five percent of those events are shallower than 12.5 km. The peaks in activity at 0, 5, and 10 km are artificial and result from those being default depths of 5-km increments for some of the calculations. Similar velocity structures of two crustal layers over a mantle half-space were used in locating events since 1974.

For several years the Lamont network was augmented by a close-in array of stations surrounding the Indian Point reactors (near Pe in Fig. 2), permitting depths to be calculated with higher precision. For two events of M 1.5 and 2.1 beneath Annsville (just to the west of Pe in Fig. 2), well-determined depths of 15 km were determined independently by Lamont analysts and by Thurber and Caruso (1985). They are shown in the cross sections of Figures 10 and 11. The two depths of 16 and 19 km in the cross sections are not as well constrained. Depths for several other nearby events recorded

by the Indian Point network from 1976 to 1983 ranged from 1 to 12 km (Seborowski *et al.*, 1982; Thurber and Caruso, 1985).

Special studies using portable instruments permitted aftershocks of a few of the larger earthquakes in the region to be determined with a precision of 1 km or better. Pomeroy *et al.* (1976) report depths of only 0–1.5 km for aftershocks of the Wappinger Falls earthquake of 1974 (Fig. 2), which was triggered by the removal of a vertical load by quarrying. Seeber (2004) computed aftershock depths of 1–2 km for an M 3.5 earthquake that occurred beneath the Delaware River in 2003 (Fig. 1). Seeber and Dawers (1989) report depths of 4.5–5.5 km for aftershocks of the 1985 Ardsley, New York, earthquake (Fig. 1). Sbar *et al.* (1975) calculate depths of aftershocks of 5.0–8.4 km for the 1973 Delaware–New Jersey earthquake, which occurred just off the southwestern corner of Figure 1. Hence, well-determined depths of earthquakes in the study area range from as shallow as 1 to as deep as 12–15 km. This places all of them within the upper crust. Crustal thickness ranges from about 35 km beneath the coastal plain to 45 km in the northwestern part of the study area (Costain *et al.*, 1989).

To our knowledge all probabilistic seismic hazard calculations for the eastern United States assume that earthquakes occur at a single default depth. A shock of a given magnitude at a depth of only 0–2 km can generate higher intensities of ground motion than one at a depth of 10 km (Atkinson and Wald, 2007). Such very shallow earthquakes, which are uncommon in the western United States, occur in our area and surrounding regions where hard rock is found at or close to the surface. They present a greater hazard than deeper shocks of the same size at epicentral distances of up to several kilometers.

Tectonic Setting of Earthquakes

The study area experienced several episodes of inter-plate (plate boundary) tectonism during the last 1.3 billion

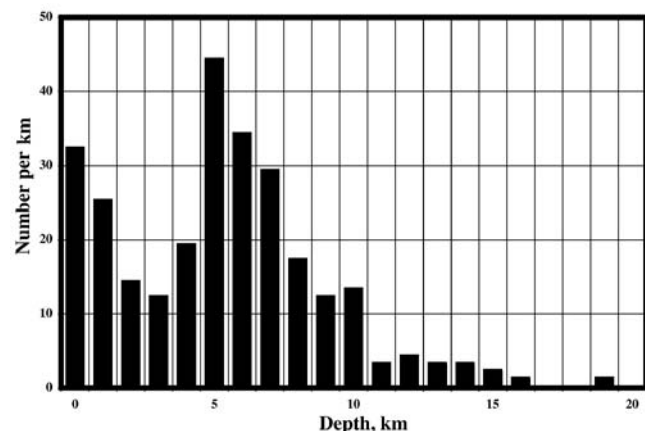


Figure 9. Histogram of the number of instrumentally located earthquakes as a function of depth.

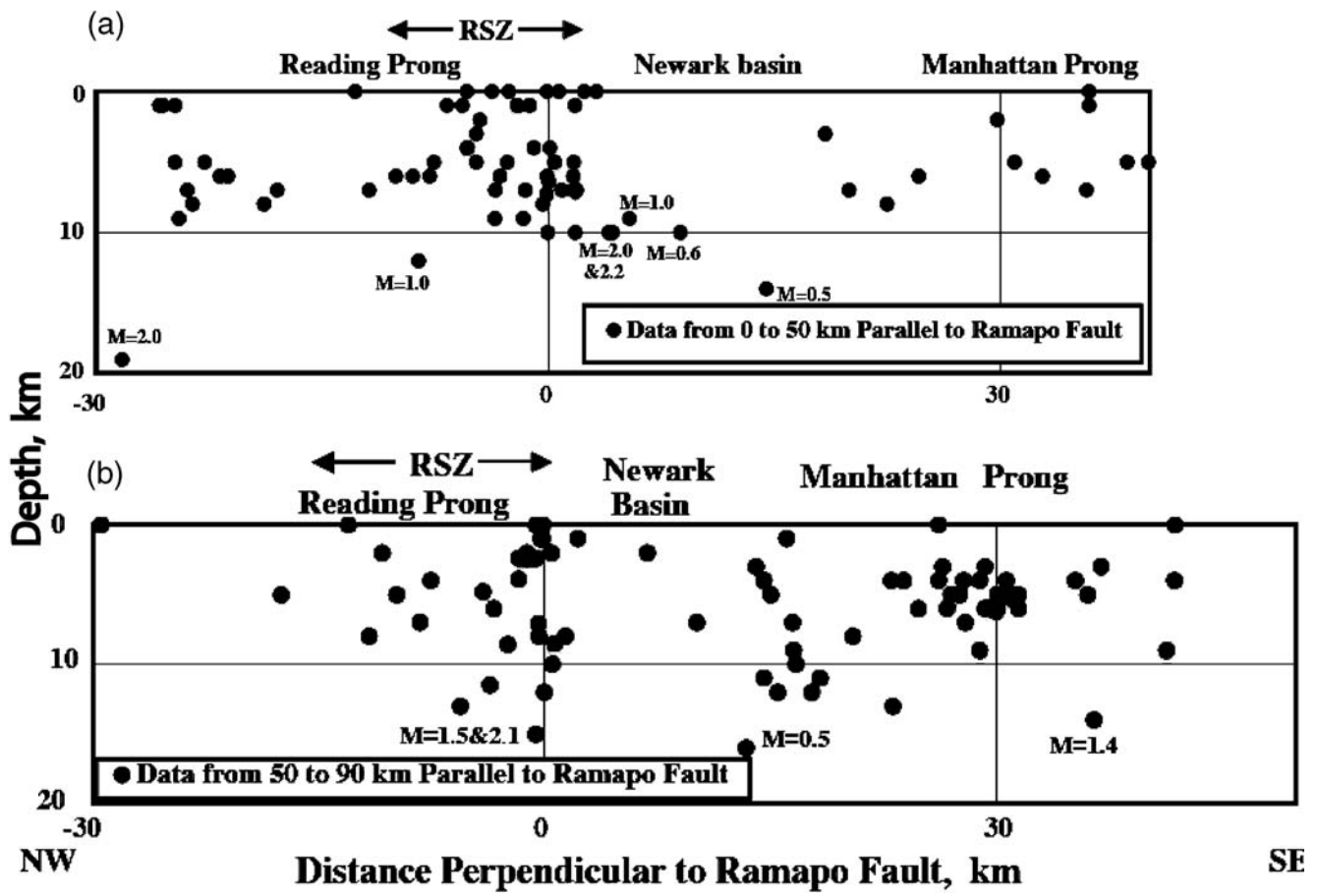


Figure 10. Vertical cross sections of instrumentally located earthquakes oriented perpendicular to the Ramapo fault. No vertical exaggeration. (a) Data from between purple kilometer markers 0 and 50 in Figure 3; (b) data from between purple markers 50 and 90.

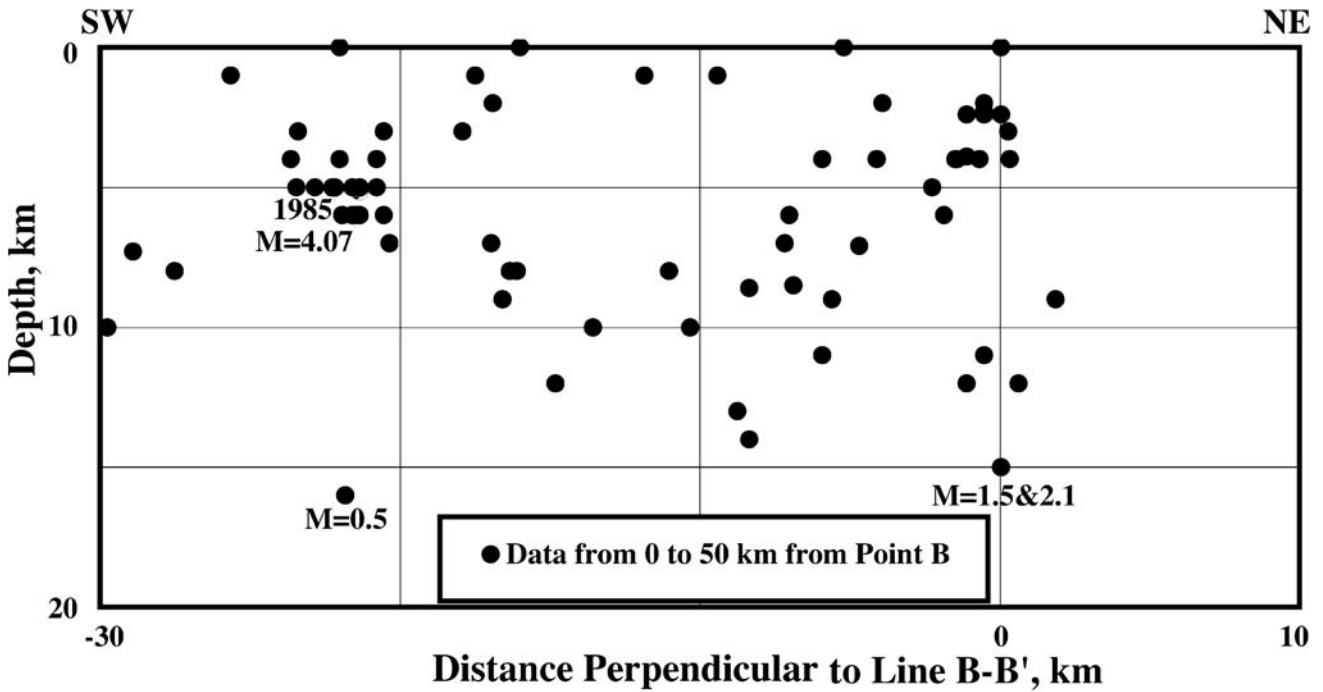


Figure 11. Vertical cross section of instrumentally located earthquakes oriented perpendicular to line BB', the Peekskill-Stamford seismic boundary, of Figure 3. No vertical exaggeration.

years (Hatcher *et al.*, 1989). The structure and lithology left by those orogenies may influence the distribution of earthquakes in what is now an intraplate region. The oldest rocks are found in the Reading prong–Hudson Highlands and the Manhattan and Trenton prongs (Fig. 1). Rifting and formation of the continental margin of the Iapetus Ocean occurred during in the late Precambrian. The New York Promontory characterized the Cambrian passive continental margin of North America in the study area. During subsequent collisional orogenies in the Paleozoic, that margin was subjected to transpressional tectonics that saw the docking of island arcs, continental fragments, and possibly smaller terranes. In contrast, the present continental margin forms an embayment south of New York City.

By the late Triassic–early Jurassic the stress regime had changed, and rifting occurred during the early separation of eastern North America from Africa (Manspeizer *et al.*, 1989). Subsequently, horizontal compression during the Jurassic resulted in some west-northwest-trending folds in the Newark basin (Schlische *et al.*, 2003). Stresses changed again to become the current contemporary intraplate regime characterized by high horizontal compressive stress oriented east-northeast but low tectonic strain (Sbar and Sykes, 1973; Zoback, 1992). Prowell (1988) estimates the onset of the present stress regime sometime between the middle Jurassic and early Cretaceous. The rift-drift transition in the Jurassic (Klitgord *et al.*, 1988) and the ensuing and continuing growth of a huge sedimentary apron are obvious reasons for changes in the stress regime near the passive Atlantic margin. During the Pleistocene, continental glaciation extended as far south as Long Island and central New Jersey and is the most recent large perturbation of the stress field. Land levels are continuing to subside (and sea level to rise) as the peripheral bulge from the last glaciation collapses.

Concentrations of Earthquakes in Older Terranes

Most of the earthquakes in Figures 1–3 are concentrated in middle-Proterozoic through Ordovician metamorphic and igneous rocks that are either exposed at the surface or present at relatively shallow depths beneath the Mesozoic and Cenozoic rocks of the New Jersey coastal plain. The coastal plain forms a southeastward-thickening prism, which is less than 1 km thick near the coast in the study area (Lyttle and Epstein, 1987; Volkert *et al.*, 1996). We think earthquakes occur in stronger basement rocks below those younger sediments. Most computed hypocenters in that area are 5–10 km deep, but are often poorly constrained. The well-located Delaware–New Jersey sequence of 1973, however, was 5.0–8.4 km deep, placing it in basement rocks.

Rates of occurrence of events of $M \geq 3$ and instrumentally located shocks (Figs. 1 and 3) are very low beneath the Mesozoic Newark basin. To first order, it is a half graben structure (Schlische, 1992; Olsen *et al.*, 1996). It is bounded on the northwest by older, crystalline rocks of the Reading prong. Ratcliffe *et al.* (1986) and Seeber (2004) deduce that

the 1957 and 2003 shocks (Fig. 1) likely occurred in those older rocks as well. Also, few instrumentally located earthquakes (Fig. 3) within the Newark basin occurred on or close to the Flemington–Furlong and Hopewell intrabasinal faults (Fig. 1). Those faults form the boundaries between three of the five major structural blocks of the Newark basin (Olsen *et al.*, 1996).

Other portions of the study area with low activity include the Great Valley to the northwest of the Reading prong, western Long Island, and the far northeastern portion of Figure 3 in the eastern Hudson Highlands. Most of the earthquakes near Wappinger Falls (Wf in Fig. 2) were triggered by quarrying (Pomeroy *et al.*, 1976).

Ramapo Seismic Zone and Other Activity in Reading Prong

The Newark basin is bordered along its northwestern margin by a series of right-stepping, southeast-dipping faults. We use the term Ramapo fault to refer to its northeastern-most Mesozoic border fault (Fig. 1). Ratcliffe (1971) refers to it and its multibranching northeastern extensions into the Hudson Highlands as the Ramapo fault zone. He concludes that Mesozoic activity extends for 25 km to the northwest of the Ramapo fault. While some instrumentally located earthquakes in Figure 3 extend that far northwest, most are concentrated in the eastern 12 km of the Reading prong. We follow Ratcliffe (1980) in calling that narrower region the RSZ; we take it to extend as far northeast as the Hudson River.

We concentrate on the RSZ first because its southeastern border forms the most conspicuous alignment of instrumentally located earthquakes in Figure 3. Figure 10 shows two cross sections of the RSZ that are oriented perpendicular to a straight line connecting points 0 and 90 in Figure 3. That line nearly coincides with the more sinuous surface trace of the Ramapo fault. In the cross section of the southwestern 0–50 km of the fault (Fig. 10a), activity is abundant in a 12-km wide zone within Precambrian rocks between the surface trace of the Ramapo fault and the Green Pond syncline of Cambrian to Devonian rocks (Fig. 1). Because that 12-km width cannot be attributed to epicentral uncertainty, more than one fault, likely many, must be involved in generating its earthquakes. The depths of events with magnitudes labeled 1.0 or smaller are not well constrained. The abrupt southeastern boundary of the zone appears to be nearly vertical. Shallower dips are obtained, however, if the hypocenters of the three small events of M 0.5, 0.6, and 1.0, in fact, were located as deep and as far southeast as shown in Figure 10a.

Other activity in the Reading prong in Figure 10a extends as far northwest as the westernmost outcrops of Precambrian rock. The depth of the event shown at 19 km is not well constrained. A band of high activity strikes about $N10^\circ W$ close to longitude $74.5^\circ W$ (Figs. 2 and 3) and extends from the Ramapo fault across the entire Reading prong.

Much of the activity in Figure 10 from 12 to 25 km northwest of the Ramapo fault occurs along that band. A left step in the Mesozoic border fault to the south and a similar step in instrumentally located events from the nearly straight southeastern boundary of the RSZ (Figs. 1–3) occur near 74.5° W and 40.8°–40.9° N. Earthquakes are nearly absent beneath the wider and deeper portion of the Newark basin that is included in Figure 10a. Some or all of the few events in Figure 3 in the eastern part of the basin may have occurred at depth within the older rocks of the Manhattan prong. We do not have enough depth resolution to ascertain if any events occurred within Mesozoic igneous bodies such as the Palisades sill.

Figure 10b shows hypocenters in a vertical section perpendicular to and across the northeastern part of the Ramapo fault (data from 50 to 90 km in Fig. 3). An abrupt increase in activity occurs across a nearly vertical boundary beneath the surface trace of the Ramapo fault. The 15-km depths of the two events of M 1.5 and 2.1 are well constrained but that for the event of M 0.5 at 16 km is not. Earthquake depths decrease systematically from a maximum of 15 km beneath the surface trace of the Ramapo fault to about 5 km at the northwestern boundary of Precambrian rocks. The lower boundary of activity beneath the Reading prong in Figure 10b appears to dip southeasterly. It may be controlled by one or more Paleozoic imbricate thrust faults that sole into a master subhorizontal detachment fault whose depth seems likely to be as great as the 12–15 km maximum depth of earthquakes beneath the Ramapo fault.

Ratcliffe (1976, 1980) and Ratcliffe and Burton (1984) examined cores from four drill holes that penetrated the Mesozoic Ramapo fault to depths of up to 0.14 km. They found no evidence of post-Jurassic displacement. Several factors could explain the apparent discrepancy between the distribution of earthquakes seen in Figure 10 and that lack of displacement for the last 150 Ma along this shallow portion of the fault. Earthquakes may be occurring either on other preexisting subparallel faults below and to the northwest of the border fault or along a series of brittle faults whose strike differs from that of the Ramapo fault. Prowell (1988) estimates fault slip rates of 0.3–1.5 mm/ka since the Early Cretaceous for several faults in the Atlantic coastal margin to the south of our study area. These are 10^{-4} smaller than rates of seafloor spreading along the Mid-Atlantic ridge today. Contemporary rates of deformation along individual faults in the RSZ may be even smaller because earthquakes occur on many faults throughout its 12-km width.

Ratcliffe (1980) and Ratcliffe and Burton (1984) report dips of 70°, >45°, and 55° southeast at three core locations along the Mesozoic Ramapo fault between the New Jersey–New York border and the Hudson River (kilometer marks 50–70 in Fig. 3). The steepest dip is found near the northeastern end of the Newark basin. The sharp southeastern boundary RSZ in Figure 10b is nearly vertical and appears to be significantly steeper than at least two of those three values. Hence, we conclude that those earthquakes occur

within older rocks of the Reading prong and not at the Mesozoic–middle-Proterozoic boundary.

Proceeding southwesterly in New Jersey, fault dips at three sites along the Ramapo fault decrease from 50° to 44° southeast and then to 34° at the Flemington–Furlong fault (FF in Fig. 1) and 25° to 35° southeast along the Mesozoic border fault at the Delaware River (Ratcliffe, 1980; Ratcliffe and Burton, 1984; Ratcliffe *et al.*, 1986). The first three sites are between purple kilometer marks 0 and 50 in Figure 3. There too, the southeastern boundary of the Ramapo seismic zone appears to dip steeper than the Mesozoic Ramapo fault.

Activity in the Manhattan Prong

The Manhattan Prong has been one of the most seismically active terranes for both historic and instrumentally located earthquakes (Figs. 1, 3, and 12). Hypocenters, like those in the Reading prong, extend from near the surface to depths of 12–15 km (Figs. 10 and 11). Nearly all of the earthquakes in those two terranes, as in the rest of our study area, appear to be confined to crystalline rocks of the upper crust. Earthquakes in New England exhibit a similar depth distribution (Ebel and Kafka, 1991) as do those in the Piedmont area and beneath the coastal plain from Virginia to Georgia (Bollinger *et al.*, 1991). The majority of earthquakes in eastern Tennessee and the adjacent Giles County seismic zone in southwestern Virginia are located at depths of about 7–25 km and occur beneath the master Appalachian decollement (Bollinger *et al.*, 1991). Earthquakes in the Canadian Shield in southern Quebec extend to depths greater than 20 km (Adams and Basham, 1991). In all of these areas earthquakes are not confined to single major through-going fault but are more distributed in map view.

Peekskill–Stamford Seismic Boundary

A surprising result based on 34 yr of instrumental data since 1974 is the existence of an aseismic area in the eastern Hudson Highlands, which contains no instrumental locations in Figure 3. It is sharply bounded on the southwest by a well-defined zone of earthquakes (between the arrows B and B' in Fig. 3) that extends northwesterly from near Stamford, Connecticut, to Peekskill, New York (St and Pe in Fig. 2, respectively). It is one of the better defined seismic boundaries in our study. Near its western end it coincides approximately with the boundary between the Manhattan prong and the Hudson Highlands, but farther east it is nearly perpendicular to and cuts across the northeasterly grain of igneous and metamorphic rocks of Precambrian to early Paleozoic age as mapped at the surface. That aseismic region is bounded on the northwest by several events to the northeast of Peekskill in the vicinity of several faults of the Ramapo fault zone on the east side of the Hudson River (Ratcliffe, 1971). That aseismic region probably extends farther east than the eastern longitude of our study.

Figure 2, which includes all earthquakes in the catalog from 1677 through 2004, gives the impression that a number

of shocks have occurred in the aseismic region of Figure 3. Accurate locations and depths of earthquakes, especially those east of the Hudson River, however, became available only after the installation of a seismic station near the New York–Connecticut border in 1971. The catalog contains no events for 1970 and 1971, one well outside that aseismic region in 1973, and one of questionable accuracy within it in 1972. The first five events in the catalog and in Figure 2 (1677–1729) are reported as being felt only at single towns in Connecticut—either Stamford or Danbury. Given the very low population density at that time, we simply do not know their locations, and some may have occurred far from either of those towns. Locations of many other pre-1974 events may be uncertain by 10–20 km, making them unreliable for defining the seismic–aseismic boundary seen in Figure 3. Many of the older less-accurately located events, such as the 1845 shock of M 3.75, may have occurred along or southwest of the Peekskill–Stamford seismic boundary.

The vertical cross section in Figure 11 is oriented perpendicular to the Peekskill–Stamford line. A sharp boundary between abundant activity to the southwest and none to the northeast extends from near the surface to a depth of 12–15 km. The two well-located events at depths of 15 km are situated at the intersection of the Ramapo seismic zone and the Peekskill–Stamford line just to the northwest of Peekskill near Annsville. Seborowski *et al.* (1982) conclude that epicenters of shallow earthquake sequences near Annsville from 1977 to 1980 are aligned northwesterly, indicating they were situated along the Peekskill–Stamford boundary.

An abrupt bend in the Hudson River to a northwesterly trend is situated near line BB' of Figure 3 and is shown on the more detailed map of Figure 12. Ratcliffe (1980) suggests that segment of the Hudson River is controlled by brittle fractures of Mesozoic age. Ratcliffe (1976) indicates a northwesterly striking, left-lateral fault on the north side of that portion of the river (the short red fault in Fig. 12). An extension of that trend to the northwest follows a major lineament that crosses the Hudson Highlands on the Preliminary Brittle Structures Map of New York (Isachsen and McKendree, 1977). It is shown as a fault (the longer red segment in Fig. 12) on the state geological map (Fisher *et al.*, 1970). The Cortland igneous complex of late Ordovician–early Silurian age (Ratcliffe, 1971) lies along or close to the Peekskill–Stamford boundary. The Croton reservoir in Westchester County (Fig. 12) consists of two northeasterly striking segments and a middle, northwesterly striking part that lies close to the seismic boundary. We do not know if that middle segment is fault controlled, but it has a geomorphic expression much like that of the 125 Street fault in Manhattan.

Seborowski *et al.* (1982) and Quittmeyer *et al.* (1985) obtained focal mechanism solutions for two small events and composite solutions for two earthquake sequences that occurred within the Indian Point seismic network near Peekskill. The solutions involve a predominance of thrust faulting along nodal planes striking northwest to north-

northwest. The strikes of nodal planes of three of those mechanisms are compatible with the orientation of the Peekskill–Stamford line.

It may be more than a coincidence that the northeastern end of the Newark basin occurs close to Peekskill and line BB' (Fig. 3). Ratcliffe (1971) finds no evidence for Triassic down-dropping at the northeastern end of the Newark basin. The boundary of the Manhattan prong seismicity along the Peekskill–Stamford line (Fig. 3) thus may coincide with the northeastern end of Mesozoic extensional tectonics. The amount of this extension, however, probably was more continuous along the former collisional orogen, but the locus of extension jumped from the Newark to the Hartford, Connecticut, basin). The New York bight basin south of Long Island (Fig. 1) may be a continuation of the Hartford basin (Hutchinson *et al.*, 1986). The series of northwest–southeast striking faults across the Manhattan prong (see next section) may have served to relay strain from the Newark basin to the other two basins where they overlap along strike. If this hypothesis is correct, all of the extension northeast of the Peekskill–Stamford boundary took place east of our study area.

Brittle Faults of Manhattan Prong

Odom and Hatcher (1980), Ratcliffe (1980), and Seeber and Dawers (1989) emphasize likely differences in the occurrence of earthquakes for two main fault types—brittle faults that are characterized by discrete breaks and cataclastic fabrics as opposed to those that exhibit semiductile and ductile behavior. The grade of Paleozoic metamorphism increases rapidly to the east of the Hudson River (e.g., Rodgers, 1970; Ratcliffe, 1971). A number of the through-going faults of Precambrian and early Paleozoic age in the Manhattan prong, the eastern Hudson Highlands, and in the Hartland terrane to the east of Cameron's Line (CL in Fig. 1) likely were healed during the Paleozoic. The grade of Paleozoic metamorphism was lower in the Reading prong and the western Hudson Highlands.

The Peekskill–Stamford seismicity boundary is nearly parallel to several brittle faults of northwesterly strike in the Manhattan prong (Fig. 12). That boundary may delineate one of the most important brittle faults in the region. Known brittle faults of the Manhattan prong include the 125 Street fault that extends across upper Manhattan to Queens, several other faults in New York City, and the Dobbs Ferry fault farther north in Westchester County (Isachsen and McKendree, 1977; Baskerville, 1982; Fluhr and Terenzio, 1984). Geologic mapping in New York City water tunnels and other field studies indicate that these northwesterly striking brittle faults are the youngest structural features in the region (Merguerian, 2004). One of them offsets a swarm of late Paleozoic (295 Ma) dikes in a tunnel (Merguerian, 2004).

The 1985 Ardsley earthquake occurred along the previously recognized west-northwest-striking Dobbs Ferry fault (Dawers and Seeber, 1991). Its mechanism involved predominantly left-lateral strike-slip motion although cumu-

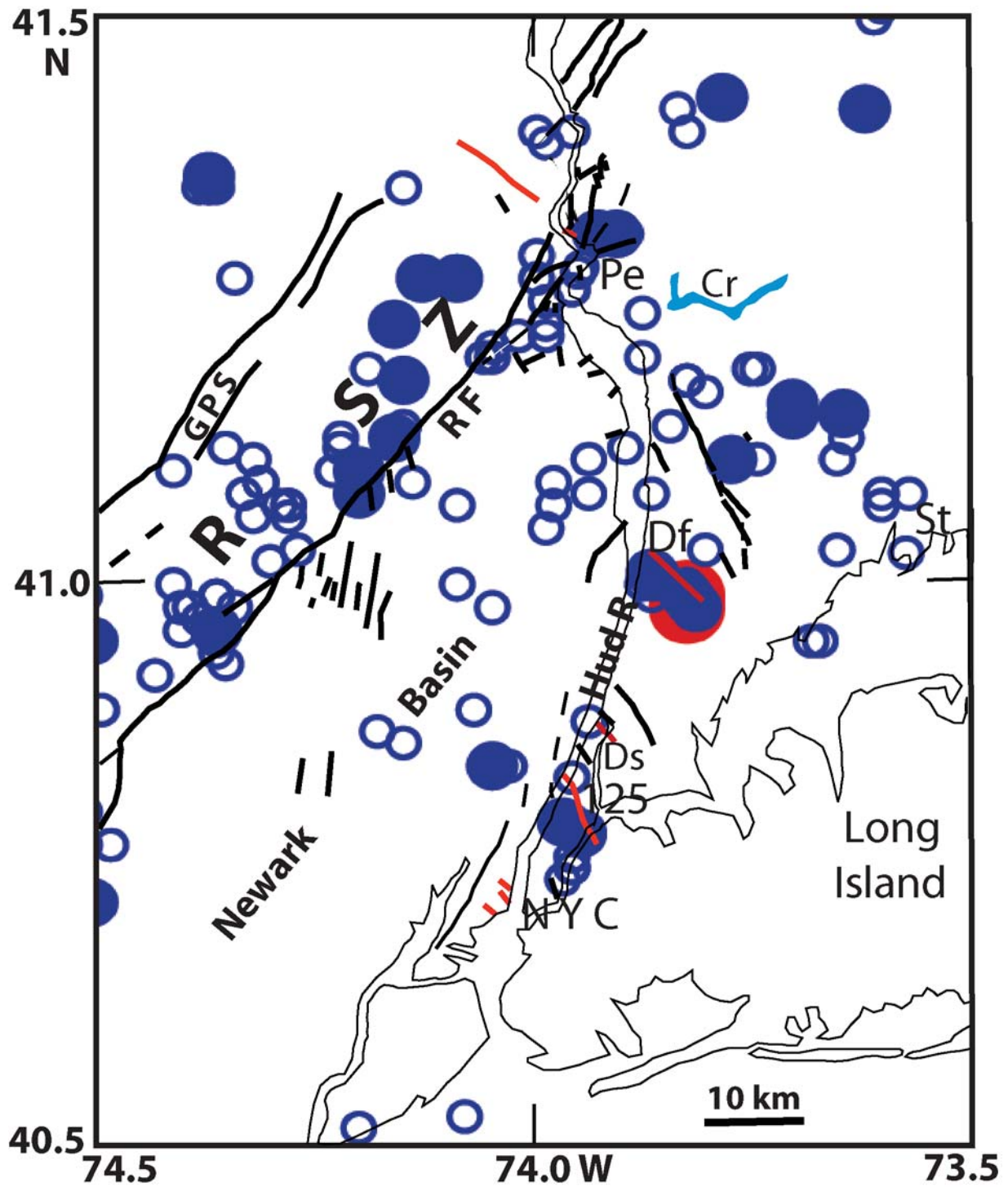


Figure 12. Brittle faults of Manhattan prong, northern Newark basin, and Reading prong from Ratcliffe (1976), Isachsen and McKendree (1977), Ratcliffe (1980), Hall (1981), Seeber and Dawers (1989), Ratcliffe (1992), Drake *et al.* (1996), Volkert *et al.* (1996), and M. N. Ratcliffe (private comm., 2007). Lineaments determined solely from aerial and satellite images are not included. Instrumentally located events 1974–2007 with symbol type the same as in Figure 3. Place names: Croton reservoir, Cr; Hudson River, HudR; New York City, NYC; Peekskill, New York, Pe; and Stamford, Connecticut, St. Tectonic features: Dyckman Street fault, Ds; Dobbs Ferry fault, Df; Green Pond syncline, GPS; Mesozoic Ramapo fault, RF; Ramapo seismic zone, RSZ; and 125 Street fault, 125. Specific faults discussed in text are shown in orange. Area is 25% of that of Figure 1.

lative offset of Paleozoic markers indicates right-lateral displacement (Seeber and Dawers, 1989). The first motions of an M 2 earthquake in 1989 along the extension of the Dyckman Street fault (Fig. 12) in northernmost Manhattan indicate a large component of left-lateral strike-slip motion along its northwesterly striking nodal plane. Merguerian and Sanders (1997) report fault breccia 90 m thick at depth where New York City water tunnel 3 crosses the 125 Street fault. A pronounced valley, which is partially filled by recent sediments, is found along that fault zone. Several small earthquakes (Figs. 3 and 12) have occurred near it. Some of the hypocenters near the west side of the Hudson River may lie along a northwesterly extension of that fault, possibly in rocks of the Manhattan prong beneath the shallow, northeastern part of the Newark basin.

Three short brittle faults (short red lines in Fig. 12) of northwesterly strike in New Jersey to the west of southern Manhattan offset rocks of both the Manhattan prong and the late-Triassic Stockton formation but are not mapped as offsetting the early Jurassic Palisades sill (Drake *et al.*, 1996; Volkert *et al.*, 1996). This suggests that they and perhaps some other brittle faults of the Manhattan prong were formed during the early development of the Newark basin in the late Triassic prior to the emplacement of the Palisades sill. Manspeizer *et al.*, (1989) and Olsen *et al.* (1996) date the emplacement of the sill about 20 Ma after clastic sedimentation in the basin was initiated. None of the other brittle faults of the Manhattan prong, however, has been shown to extend across the Hudson River and to offset Mesozoic rocks along its western side (Isachsen and McKendree, 1977; Drake *et al.*, 1996). Likewise, whether any of the other small faults that offset the Palisades sill on the west side of the River extend to its eastern side into the Manhattan prong is not known. Better mapping and dating of these various brittle faults is important in estimating seismic risk because one of them was the site of a moderate-size earthquake in 1985.

Cameron's Line and Huntington Valley Fault

A major tectonic boundary, Cameron's Line (CL in Fig. 1), separates rocks of Precambrian and early Paleozoic age of the Manhattan prong and Hudson Highlands on its northwest from those of the Hartland terrane in Connecticut and eastern New York City (Rodgers, 1970; Baskerville, 1982; Lyttle and Epstein, 1987). That zone of thrust faults of various dips is the site of several ultramafic bodies, including a large serpentine mass on Staten Island (SI in Fig. 1). Cameron's Line is usually interpreted in a plate-tectonic framework as an Ordovician (Taconic) suture zone, which involved the accretion of an island arc to the east along with intervening eugeosynclinal sediments and perhaps other terranes to the then passive continental margin of Laurentia (e.g., Drake *et al.*, 1996). Terranes on both sides of Cameron's Line were strongly metamorphosed during the Paleozoic.

The well-located 1985 shock (Fig. 1) and its aftershocks clearly occurred to the west of Cameron's Line (Seeber and Dawers, 1989; Dawers and Seeber, 1991). The Peekskill–Stamford seismic boundary (Fig. 3) cuts nearly perpendicularly across Cameron's Line. A few historic but only two instrumentally located earthquakes have occurred near its surface trace. Because it likely was healed by Paleozoic metamorphism, it is not surprising that it is no longer a zone of brittle faulting and a source of many earthquakes.

The Huntingdon Valley fault zone (Fig. 1) near Philadelphia and Trenton (T in Fig. 1) also separates contrasting older crystalline terrane—the Trenton prong to the northwest from the southern Piedmont to the southeast (Rodgers, 1970; Lyttle and Epstein, 1987). Rodgers (1970) suggests that it is an Ordovician suture zone, which may extend beneath the coastal plain of New Jersey and join Cameron's Line in Staten Island. Valentino *et al.* (1994) state that the fault and its extension to the southwest into Maryland also experienced major right-lateral strike-slip offset during the late Paleozoic. While some earthquakes such as the Abington, Pennsylvania, sequence of 1980 occurred on or close to the Huntingdon Valley fault zone, the breadth of the earthquake zone beneath the coastal plain of New Jersey (Fig. 1) suggests that most of that activity is not occurring along that fault zone.

Earthquakes beneath Coastal Plain of New Jersey

The zone of earthquakes beneath the coastal plain continues to the southwest of our study area as least as far as the New Jersey–Delaware border. The basement beneath the coastal plain of New Jersey occupies a critical position linking major Paleozoic events of the northern and central Appalachians. Seismic reflection and refraction data indicate that continental basement is present beneath the coastal plain and continental shelf of New Jersey (Klitgord *et al.*, 1988; Sheridan *et al.*, 1991; Volkert *et al.*, 1996). Most wells that reach the basement penetrate through the feather edge of the sedimentary wedge of the coastal plain. To the south and east few wells penetrate to the basement (Volkert *et al.*, 1996). Volkert *et al.* (1996) make a speculative assignment of basement rocks beneath the coastal plain to those of exposed terranes in the northern and central Appalachians based on the petrology and geochemistry of samples from wells and magnetic and gravity anomalies. They conclude that subsurface crystalline rocks beneath the coastal plain of New Jersey consist of stacks of thrust sheets representing a composite of accreted terranes.

A pronounced Bouguer gravity gradient is found along the length of the Appalachian orogen. Cook and Oliver (1981) conclude that it marks the eastern edge of the former late Precambrian-early Paleozoic continental margin, which is now buried at depth below near-surface, allochthonous crystalline rocks. A shorter-wavelength Bouguer high along that broad gravity gradient extends from Staten Island across the inner coastal plain of New Jersey (e.g., Volkert *et al.*, 1996). That high is nearly coincident with the band of seis-

mic activity beneath the coastal plain of New Jersey and with what are inferred to be crystalline rocks at depth.

Sheridan *et al.* (1991) interpret a seismic reflection profile and associated gravity anomalies in the coastal plain of southwestern New Jersey about 60 km south of our study area as indicative of a gently, southeast-dipping detachment fault that is underlain by Grenville-age continental crust. They, like Volkert *et al.* (1996), deduce that imbricate thrust faults in the upper crust above that detachment separate several allochthonous terranes of Precambrian and early Paleozoic age. They infer that several bodies representing a complex mixture of ophiolitic, metasedimentary, and metavolcanic rocks are present beneath their seismic line. They conclude that the Ordovician suture zone was transported northwestward later during the Paleozoic.

Continuity of Seismic Terranes from Manhattan Prong to Coastal Plain

Seismic velocities, magnetic anomaly patterns, and wells drilled to the basement indicate that the basement along the coast of southwestern Long Island is composed of Paleozoic or older metamorphic and crystalline rocks (Hutchinson *et al.*, 1986; Volkert *et al.*, 1996). Our location of the 1884 earthquake (Fig. 1) places it in those rocks and makes a direct association with either Cameron's Line or the offshore New York bight basin unlikely. Several small instrumentally located events (Fig. 3) appear to be aligned northwesterly just to the southeast of the 1884 epicenter and may delineate the fault along which the 1884 earthquake occurred. That alignment is roughly parallel to the strike of brittle faults farther north in the Manhattan prong. Aftershocks of the 1884 event, however, were felt more strongly either near the adjacent coast of New Jersey or near the coast of Long Island (Nottis and Mitronovas, 1983), suggesting a rupture zone striking northeasterly. Marine studies of that area and more reliable hypocenters and earthquake mechanisms are of high priority in understanding earthquake risk to New York City.

None of the known shocks of $M \geq 3$ (Fig. 1) since 1800 occurred between the Westchester earthquake of 1985 and the 1884 and 1937 events farther south. The uncertainty in the location of the 1737 event is large enough that it may or may not have occurred in New York City, but smaller instrumentally located events have (Fig. 3). The simplest interpretation is that the array of brittle cross faults that characterizes the Manhattan prong through New York City is associated with the collocated belt of seismicity and that the seismogenic potential of these faults is similar. This association has been demonstrated for the Dobbs Ferry fault. Southwest of the 1884 earthquake the seismicity continues below the coastal plain of New Jersey, and the belt of cross faults may continue as well. If this hypothesis is correct, the seismic potential of these faults can be estimated from the seismicity and from their geologic characteristics. The length of several of these faults far exceeds the inferred sizes of historic ruptures and

indicates that they are capable of generating much larger earthquakes.

Which Faults Are Active?

Cretaceous and Cenozoic faulting has been identified in the coastal plain and Piedmont from Maryland to Georgia (Prowell, 1988). Most examples involve a predominance of dip-slip reverse motion. Geological evidence of late Cenozoic fault movement is poor, largely because of the limited distribution of markers of that age. Wherever it is available, however, it points to very slow rates of fault displacement in this intraplate area. Evidence for intraplate faulting in much of our study region has been destroyed by Pleistocene glaciation, and most rocks outside of the coastal plain are Jurassic or older.

Two factors that make the field identification of Holocene faulting difficult are the small size of ruptures of earthquakes of M 4–6 in the eastern United States and the distribution of the seismicity on many faults each with very low displacement rates. The rupture area of the 1985 Ardsley shock of $m_b Lg$ 4.2 was about 0.5 km² (Seeber and Dawers, 1989); that for the 5.1 Adirondack, New York, event of 1983 was 1 km² (Seeber and Armbruster, 1986, 1988); and that for the 2002 Au Sable Forks shock of 5.1 in northeastern New York State was about 1.7 km² (Seeber *et al.*, 2002). None of those events produced surface rupture. The latter two are comparable in magnitude to the three largest historic events in our area.

Surface rupture in eastern and central North America has not been identified except for a few earthquakes of M 6–8. Shocks in that region typically have smaller dimensions and larger stress drops and are felt to larger distances than those of comparable magnitude in the western United States (Atkinson and Wald, 2007). Thus, the occurrence of earthquakes of M 5–6 in our area does not require great fault lengths, merely rupture zones about 1–10 km in length. Many faults and segments of longer faults are possible candidates for shocks of those magnitudes. Detecting Holocene fault displacement at or near the surface will continue to be very difficult. Other methods must be used to ascertain which faults are either active or potentially active.

Contemporary State of Stress and Earthquake Generation

The orientation and magnitudes of principal stresses in the crust are vital in understanding contemporary fault activity. Knowledge of *in situ* stresses in our study area, however, continues to be poor to modest. The occurrence of rock bursts during the construction of major water tunnels in hard rock (e.g., Berkey and Rice, 1919) is indicative qualitatively of high compressive stress as are results from hydrofracture experiments in our study area and elsewhere in the eastern United States (Zoback, 1992).

Combining a variety of stress measurements, Sbar and Sykes (1973) found that maximum compressive stress, σ_1 , is nearly horizontal and is oriented about east-northeast in a large area of eastern North America including northern and western New York State. Based on very limited data Sbar and Sykes (1977) concluded the state of stress differs in our study area, New England and Virginia. From studies of mechanisms of small earthquakes using only P -wave first motions Aggarwal and Sykes (1978) and Yang and Aggarwal (1981) concluded that σ_1 in our area is oriented east to east-southeast. Reliable mechanisms are rare, however, given the small shallow earthquakes and widely spaced stations typical of our area.

Subsequent work indicates that σ_1 likely is oriented northeast to east. The first 19 entries in ④ Table 4 (in the electronic edition of *BSSA*) compile P axes of selected focal mechanisms and hydrofracture and borehole breakout measurements of σ_1 for areas adjacent but outside our study region. The distribution of those stress indicators is similar to those in maps by Zoback (1992) and Du *et al.* (2003). The last nine entries in ④ Table 4 (in the electronic edition of *BSSA*) are from our study area and are shown in Figure 1. Mechanisms were selected only if they were obtained either by waveform matching (Du *et al.*, 2003) or by using data from a network of very close-in stations (i.e., either aftershocks recorded by temporary deployments of seismographs or events beneath the dense Indian Point network). Those mechanisms typically have a wider coverage of data on the focal sphere and more accurately determined depths than those derived only from first-motions of individual events. All of the P axes in ④ Table 4 (in the electronic edition of *BSSA*) are nearly horizontal. P axes, however, need not be parallel to σ_1 for preexisting faults (Michael, 1984).

An average of the azimuths of 19 values of σ_1 and P axes for the region outside our study area is $N67^\circ E \pm 16^\circ$ (± 1 standard deviation) with a standard error of the mean (SEM) of 4° . An average of nine azimuths (Fig. 1) for our area is $N64^\circ E \pm 16^\circ$, SEM = 5° . We omitted the composite mechanism for one very small earthquake and its aftershocks in 1969 that involved normal faulting (Sbar *et al.*, 1970), a completely different mechanism from that of others in the study area. Seeber *et al.* (1998) obtained an azimuth of $N68^\circ E \pm 5^\circ$ from inferred slip planes of seven focal mechanisms in our area, Pennsylvania, and Maryland using the technique of Michael (1984). The three sets of results are nearly identical. A number of breakout measurements in boreholes along the outer continental shelf to the southeast of Figure 1 also indicate that σ_1 is oriented northeasterly (Zoback, 1992). The azimuth of σ_1 may vary within our region, but we cannot as yet separate measurement uncertainties from intrinsic spatial variations.

The mechanism of the 1985 Ardsley earthquake involved left-lateral strike-slip motion along a fault striking west-northwest at a depth of 5 km. That of a large aftershock involved a combination of reverse and strike-slip motion. Mechanisms of later aftershocks abutting it displayed reverse

faulting of either north or north-northeast strike. The occurrence of those two types of faulting indicates that maximum compressive stress, at least in that immediate area, is large whereas the two other principal stresses are smaller and of comparable size. The other focal mechanisms for which P axes are shown in Figure 1 involved a predominance of reverse faulting. All but one of them was from an earthquake of very shallow depth. Likewise, the hydrofracture and borehole breakout measurements were all made within the upper 1 km. Because the least compressive stress is likely to be vertical at very shallow depths, dip-slip reverse or thrust faulting is likely.

Near-vertical strike-slip faults striking about west to west-northwest and north-northeast to northeast that have not been healed by high-grade metamorphism should be oriented favorably for rupture in a contemporary stress field oriented about $N64^\circ E$ (assuming hydrostatic pore pressure and a coefficient of friction of about 0.6). This would include west-northwest left-lateral strike-slip faulting in the Manhattan prong, as in the 1985 earthquake. Likewise, thrust faults dipping about 30° and striking nearly perpendicular to σ_1 also would be favorably oriented. Clear examples of Cretaceous and Cenozoic thrust faulting of shallow dip in the Appalachians (Odom and Hatcher, 1980), however, are lacking. Generally geologic data and focal mechanisms are inconsistent with the current reactivation of shallow-dipping Paleozoic and Precambrian thrust faults in the Appalachians, most of which are of northeasterly strike. Reverse motion on steeply dipping, preexisting faults striking north to northwest is consistent with σ_1 oriented about $N64^\circ E$ as in several of the mechanisms in Figure 1. Strike slip or oblique slip is also possible on north-northeast-striking brittle faults of steep dip in the Manhattan prong that are younger than 295 Ma (Merguerian, 2004) but older than the northwesterly striking brittle faults described earlier.

The Ramapo seismic zone and the Ramapo fault strike about $N40^\circ E$. The azimuths of σ_1 for the hydrofracture measurement (Zoback *et al.*, 1985) to the northeast of Peekskill and the one borehole breakout determination (Goldberg *et al.*, 2003) are nearly parallel to the Ramapo fault (Fig. 1). If those azimuths, rather than the average of $N64^\circ E$ obtained earlier, are indicative of the orientation of σ_1 , contemporary faulting striking subparallel to the Ramapo trend is unlikely. Another possibility is that earthquakes within the RSZ are occurring along left-lateral strike-slip faults of northwesterly strike like those in the Manhattan prong. Some faults of northwesterly strike occur within the Newark basin (Fig. 12), but those near the Ramapo fault are not mapped by Fisher *et al.* (1970), Schlische (1992), or Drake *et al.* (1996) as cutting across it into the Reading prong.

The Ramapo fault zone experienced repeated deformation from the Precambrian to the Jurassic (Ratcliffe, 1971, 1980). Significant post-middle Ordovician right-lateral strike-slip faulting has occurred along that zone (Ratcliffe, 1971, 1980; Ratcliffe and Burton, 1984). If σ_1 is oriented about $N64^\circ E$, contemporary right-lateral, strike-slip faulting

is possible within crystalline rocks of the Reading prong along preexisting, subvertical, brittle faults oriented subparallel to the Ramapo fault. The straight, nearly vertical southeastern boundary of the Ramapo seismic zone suggests that mode of contemporary deformation.

Ratcliffe (1980) states that current seismic activity along the Ramapo fault zone may be more strongly controlled by the presence of through-going crustal structures than it is by more superficial Mesozoic faults. His conclusion agrees with our finding that most, and perhaps all, earthquakes near the Mesozoic border faults appear to be occurring within crystalline rocks of the Reading prong and not at the northwestern boundary of Mesozoic rocks. Not having activity along the Ramapo and other border faults just at the boundary between Mesozoic and older rocks is not surprising because very few earthquakes occur within Mesozoic rocks of the Newark basin.

Seeber and Armbruster (1985, 1986) and Armbruster and Seeber (1987) found that aftershocks and a nodal plane of a focal mechanism solution for the Martic, Pennsylvania, earthquake sequence in 1984, about 75 km to the west of our study area, were aligned north to north-northeast, nearly parallel to the strike of a prominent Jurassic dike. The fault deduced from those data dipped steeply and involved a combination of reverse and right-lateral strike-slip motion. The P axis of the mechanism was oriented $N70^{\circ}E$, similar to the average σ_1 in our study area. Accurate isotopic dating indicates that igneous bodies in the Newark and several other Mesozoic basins were emplaced during a brief interval of about 0.6 Ma in the early Jurassic about 20 Ma after clastic sedimentation in those basins was initiated (Olsen *et al.*, 1996; Schlische *et al.*, 2003). The source of their magmas ultimately must have been the upper mantle. Their emplacement is a rather distinct and different event than the much slower development of the basins themselves. The dikes at depth that presumably fed those igneous rocks may have utilized preexisting, near-vertical faults in the crust. It is unlikely, however, that those dikes were injected along shallowly dipping faults of Mesozoic or older age. We suggest that the straight, nearly vertical southeastern boundary of the Ramapo seismic zone (Fig. 3) may coincide with one or more major dikes that fed sills, stocks, and lava flows into adjacent sections of the Newark basin.

Two mechanism solutions and orientations of aftershock zones are available for other parts of the Newark basin. The 2003 earthquake (Fig. 1) involved reverse faulting of northwest to west-northwest strike (Seeber, 2004), which is nearly perpendicular to the nearby border fault. Because that border fault strikes about $N68^{\circ}E$, it is unlikely to be active if σ_1 is oriented about $N64^{\circ}E$. To the west of our study area where border faults strike nearly east–west near Reading, Pennsylvania, the focal mechanism and aftershocks of an earthquake in 1994 indicate a combination of reverse and strike-slip motion on a fault striking northwesterly. Thus, which faults are active today in the Reading prong depends critically on their strike and dip and the orientation of σ_1 .

Low Activity in Newark Basin and Great Valley

It is important to address why so many earthquakes have occurred within crystalline terranes in our study area, so few beneath much of the Newark basin and even fewer in the Great Valley to the northwest of the Reading prong. Most of the instrumentally located earthquakes beneath the Newark basin (Fig. 3) occurred near its northeastern end where its basement shoals and crystalline rocks of the Manhattan prong are found beneath the thin Mesozoic section. In contrast, Lyttle and Epstein (1987) and Drake *et al.* (1996) indicate maximum thicknesses of Mesozoic rocks of 6–8 km for five cross sections of the Newark basin farther southwest. Schlische (1992) obtains a thickness greater than 6 km from a seismic reflection profile across the central part of the basin southwest of the Delaware River. Crystalline Proterozoic and early Paleozoic rocks are thought to be present beneath the basin (e.g., Lyttle and Epstein, 1987; Drake *et al.*, 1996). Sedimentary rocks, which constitute most of the thickness of the Newark series, either may not be strong enough to store large tectonic stresses or their rheology places them in the velocity-strengthening regime. Also, they may lack enough preexisting faults of appropriate orientation to permit rupture in response to the present stress regime.

This leaves unanswered the low level of seismicity beneath the deeper parts of the Newark basin from the base of Mesozoic rocks to depths of 12–15 km as in the RSZ and the Manhattan prong. The very few earthquakes beneath the Cambro–Ordovician rocks of the Great Valley (Fig. 1) may be a clue to the low level beneath much of the Newark basin. Major thrust and detachment faults along some of the weaker lower Paleozoic formations (e.g., Lyttle and Epstein, 1987; Costain *et al.*, 1989) may decouple them from stresses at depth. The previous three reasons also may explain the low activity. Lyttle and Epstein (1987) show about 4 km of Cambro–Ordovician rocks beneath the Newark basin on their section BB'. Thus, the combined thicknesses of Cambro–Ordovician and Mesozoic rocks along that section approaches the maximum depth of earthquakes in the RSZ. Cambro–Ordovician and Precambrian rocks outcrop within the Newark basin in Buckingham Valley (BV in Fig. 1). Lyttle and Epstein (1987) describe the former as Taconic-like rocks. Similar rocks are found on the west side of the Newark basin. Such Taconic rocks were encountered in the Paresstis exploration well in the southwestern part of the Newark basin (P. Olsen, private comm., 2007). Thus, Cambro–Ordovician rocks may be extensive beneath the deeper parts of the Newark basin and may account for the small numbers of earthquakes there.

Conclusions and Discussion

Historic and instrumental data on earthquakes in the greater New York–Philadelphia area are cataloged in terms of a common magnitude scale, $m_b Lg$. At the continental dimension, this zone is one of a few earthquake source areas

that stand out from the low intraplate seismicity background of the eastern United States. Most earthquakes within this zone are shallow (94% are ≤ 10 km deep) and occur in older crystalline basement rocks of the upper crust. They are concentrated in areas where these rocks are exposed, in the Manhattan and Reading prongs, or where they occur at shallow depth below coastal plain sediments of New Jersey and southwestern Long Island. Seismicity is much lower in the eastern Hudson Highlands where crystalline rocks are exposed at the surface and where they are more deeply buried, such as beneath the Newark basin and beneath the Cambro–Ordovician rocks of the Great Valley to the northwest of the Reading prong. Distinction needs to be made in hazard assessments between these more active and less active terranes.

Earthquake locations using 34 yr of data from a local network, delineate a number of features that were not resolvable solely with historic data. Instrumentally located activity along the Ramapo seismic zone is concentrated in a 12-km wide band in older rocks of the eastern Reading prong between the surface trace of the Mesozoic Ramapo fault and the Green Pond syncline. The southeastern boundary of that activity, which extends from near the surface to depths of 12–15 km, dips very steeply to the southeast, typically greater than the dip of the Mesozoic border fault. That activity is attributed to faults of as yet poorly determined strike within that part of the Reading prong. Seismicity is nearly absent in the Mesozoic sedimentary rocks of the Newark basin and in Cambro–Ordovician rocks inferred beneath them. This is attributed to either those rocks being relatively weak, in the velocity-strengthening rheological regime, being decoupled from the basement beneath them by thrust and/or detachment faults along weak layers, or a lack of preexisting brittle faults that are suitably oriented with respect to contemporary stresses. Because intraplate strain rates are extremely slow and geologic deformation since the Mesozoic is largely lacking, it may be useful to explore why a geologic terrane is seismically active and why another is not.

A newly identified feature—the Peekskill–Stamford seismic boundary—is nearly vertical, extends from near the surface to depths of about 12–15 km, and is subparallel to several brittle faults farther south in the Manhattan prong. Instrumentally located earthquakes are relatively abundant in the Manhattan and Reading prongs but not in the eastern Hudson Highlands to the northeast of that line. A brittle fault that crosses the Reading prong lies along the northwest extension of that line (Fig. 12). Seismic activity is low farther north in older crystalline rocks of the Green Mountains and the Berkshire and Housatonic highlands (Seeber and Armbruster, 1988). While seismicity in our study area seems to originate exclusively within old strong basement rocks of the upper crust, the presence of such rocks at or near the surface appears to be a necessary, but not a sufficient, condition for the occurrence of earthquakes. Some other factors, such as either the magnitude and/or orientation of principal stresses and the presence of brittle faults of appropriate orientation also must be important in generating varying levels of intra-

plate activity. Our sense is that the mapping of brittle faults, especially in older crystalline terranes in the study area is still in its infancy.

Earthquakes do not occur continuously along either the Appalachians or the coastal plain. Using those physiographic features by themselves for purposes of defining contemporary tectonic behavior and earthquake occurrence is not justified. The use of terranes of older stronger rocks that also have been the loci of historic seismic activity are more appropriate for calculating probabilistic seismic hazards. Future estimates of earthquake hazards and risks need to take into account the quite different levels of activity within our study area and their significance in terms of potential source terranes and faults.

Several damaging historic earthquakes and a sustained relatively high level of instrumental seismicity point to a concentration of earthquake hazard in the seismic zone that extends along the Manhattan prong to south of New York City and then beneath the coastal plain of New Jersey. Surprisingly, few modern geophysical data are available, however, considering the huge population and assets at risk. This is particularly so for the region of the 1884 earthquake just to the south of New York City and the seismic zone beneath the coastal plain of New Jersey—parts of the northeast corridor. No seismic stations are operating in that part of the coastal plain. Monitoring and knowledge of earthquakes in our intraplate area is far behind those efforts in California and Japan. Knowledge of which faults are active is in its infancy, and high-quality data on *in situ* stress are few. The accuracy of focal depths needs to be improved.

National priorities for geophysical and geological work and monitoring have been and continue to be based much more on hazard than risk. For example, the low priority of earthquake research and monitoring in New York City contrasts with high estimated annualized earthquake losses based on HAZUS calculations by the Federal Emergency Management Agency (FEMA) (2001). Their estimates rank New York City eleventh out of 40 U.S. cities. In comparison, Philadelphia and Newark are thirtieth and thirty-sixth, St. Louis is sixteenth, Boston is twenty-fifth, Memphis is twenty-ninth, and Charleston, South Carolina, is thirty-fourth.

Tantala *et al.* (2003) used HAZUS with a modified building stock for the metropolitan New York area to estimate losses for earthquakes of moment magnitude (M_w) 5, 6, and 7 at the site of the 1884 shock as well as probabilistic calculations for average return periods of 100, 500, and 2500 yr. For M_w 6 and 7 events at the site of the 1884 shock, they calculate losses from buildings and income of \$39 and \$197 billion, respectively. The addition of infrastructural losses would about double those figures (K. Jacob, private comm., 2007). Their region of investigation did not include Philadelphia and Trenton, as ours did. Extrapolated repeat times in our study area for events of $m_b L_g$ 6 and 7 are about 670 and 3400 yr, respectively. The corresponding probabilities of occurrence in a 50-yr period are about 7% and 1.5%,

respectively. The probability of an earthquake the same size as the 1884 event, $m_b L_g$ 5.25, during a 50-yr period is about 22%. Calibrating M_w in terms of $m_b L_g$ is difficult for our region because few determinations of M_w are available even for neighboring regions. Existing data indicate that M_w is typically somewhat smaller than $m_b L_g$.

Two nuclear power plants at Indian Point (near Peekskill in Fig. 2) are located closer to more people at any given distance than any other similar facilities in the United States. Entergy, their owner, recently applied for 20-yr extensions of their existing 40-yr licenses. Much new seismological information is available since their initial approvals in 1973 and 1975. Nevertheless, the U.S. Nuclear Regulatory Commission so far has not permitted any new information to be used or old information on which the original licenses were based to be contested in considering extensions of licenses. Indian Point is situated at the intersection of the two most striking linear features marking the seismicity (Fig. 3) and also in the midst of a large population that is at risk in case of an accident to the plants. This is clearly one of the least favorable sites in our study area from an earthquake hazard and risk perspective.

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