NUREG/CR-3079

Earthquake Hazard Studies in New York State and Adjacent Areas

Final Report
April 1976 - June 1982

Prepared by A. L. Kufka

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GPO Printed copy price: $5.00

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Manuscript Completed: November 1982
Date Published: January 1983

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Prepared for
Division of Health, Siting and Waste Management
Office of Nuclear Regulatory Research
U.S. Nuclear Regulatory Commission
Washington, D.C. 20555
NRC FIN B6096
Under Contract No. NRC-04-76-369

ABSTRACT

Lamont-Doherty Geological Observatory (LDGO) currently operates a network of 38 short period seismic stations in the states of New York, New Jersey and Vermont. It is part of the larger Northeastern United States Seismic Network (NEUSSN) operated by several university groups in New York, New Jersey, Pennsylvania and New England. These networks provide a wealth of data to study seismicity, earthquake hazards, earthquake source properties, tectonic processes, and crustal and upper mantle structure in the northeastern United States and adjacent parts of Canada. The LDGO network provides data for more specific studies of earthquake processes in New York State and adjacent areas.

The operation and maintenance of the LDGO network has been supported primarily by funds from the United States Geological Survey (USGS), the United States Nuclear Regulatory Commission (NRC), and the New York State Energy Research and Development Authority (NYSERDA). This report discusses results of research related to the operation of the network during Phase I through Phase VII of our contract with NYSERDA, and also introduces current directions of research for future studies.

SUMMARY

The past decade of Lamont-Doherty research associated with the New York State seismic network includes the following specific results: (1) a greater general understanding of the relationship between seismicity and geologic structures in the northeastern United States; (2) identification of seismic provinces based on seismicity, earthquake fault plane solutions, state of stress, and other geological and geophysical data; (3) a greater understanding of the origin of the intraplate stress field in the northeastern United States; (4) studies of crustal and upper mantle velocity structures; (5) studies of strong-motion and spectral content of earthquakes in the northeastern United States; and (6) recent field studies of large earthquakes in 1982 in New Brunswick and New Hampshire.

We are currently pursuing a number of research efforts that were initiated during the period of time covered by this report and will continue in our future studies. These research efforts include: (1) obtaining more fault plane solutions of earthquakes to better delineate the intraplate stress field and the configuration of seismic provinces; (2) continued studies of strong-motion and spectral content of earthquakes in the northeastern United States coupled with studies of modified Mercalli intensity of recent and historical earthquakes in this region; (3) further studies of crustal and upper mantle velocity structure in the northeastern United States; (4) an attempt to develop depth diagnostics for local earthquakes from digital seismograms recorded by the network; (5) calculations of the maximum intensity, acceleration, and velocity that can be expected at given locations; (6) a comparison of the Nuttli (1973) magnitude scale for frequencies near 1 Hz with magnitudes calculated from frequencies more typical of those recorded by the northeastern network (3 to 10 Hz); and (7) determination of magnitudes for the past 250 years of historic earthquakes in New York, New Jersey and Vermont from felt areas and attempts to determine the probability of exceeding a given level of ground acceleration and velocity for specific sites within that region.

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INTRODUCTION

As the trends of urbanization and industrialization spread throughout the world and the density of the built environment grows ever greater, the problem of understanding earthquake hazards in regions of high population density becomes increasingly serious. The northeastern United States is a region of very high population density, and the density of the built environment as well as the concentration of critical facilities, such as nuclear power plants, in this region is among the highest in the nation. From a seismic risk point of view it is, therefore, important to understand earthquake phenomena in the northeastern United States. It is, for example, important to know if large earthquakes, such as the Massena, New York earthquake of 1944 or the recent sequence of large earthquakes in New Brunswick, Canada (January 1982) can occur anywhere in this region or if such large events are confined to particular tectonic zones or faults. In addition, it is important to know if very destructive earthquakes such as those which occurred in other portions of the eastern United States (e.g., the 1811-1812 earthquakes of New Madrid, Missouri or the 1886 earthquake of Charleston, South Carolina) could occur in the Northeast. In this report we describe our research on northeastern United States seismicity, earthquake hazards, and tectonic processes. The long term goal of this research program is to arrive at an understanding of earthquake phenomena in the Northeast in general and in New York State and adjacent areas in particular.

Local seismic networks in the northeastern United States operated by Lamont-Doherty and several other institutions during the past decade (Figure 1) have elucidated the general features of seismicity in this region. Yang and Aggarwal (1981) have indicated that the record of instrumentally located earthquakes in this region reveals the same general features as the longer term (several hundred years) historical record. Those areas that have had little or no seismicity historically were relatively aseismic for small shocks recorded by the networks during the past decade, whereas the historically active areas are also active today. This similarity suggests that seismic activity in the Northeast is fairly stationary in space over a several hundred year period, and allows us to make some general statements about which areas are prone to seismic activity and which are not.

The operation of local seismic networks in the Northeast has enabled us to make these general statements about the seismicity and has helped to correct erroneous interpretations of the historical seismic record caused by biases which result from the uneven distribution of population. In addition, the earthquake locations which have been determined from network data are accurate to within a few kms, whereas the historical data are only acurate to within a few tens of kms. This accuracy of locations makes it possible to begin to associate the details of particular geologic structures with seismicity, and this has been a major part of the Lamont-Doherty research effort during the past several years (e.g., Fletcher and Sykes, 1977; Aggarwal and Sykes, 1978; Sykes, 1978; Yang and Aggarwal, 1981; Kafka et al., in preparation, Appendix C).

Our more recent research efforts constitute an integration of several approaches to understanding earthquake phenomena involving detailed analysis of both the network data and the historical record

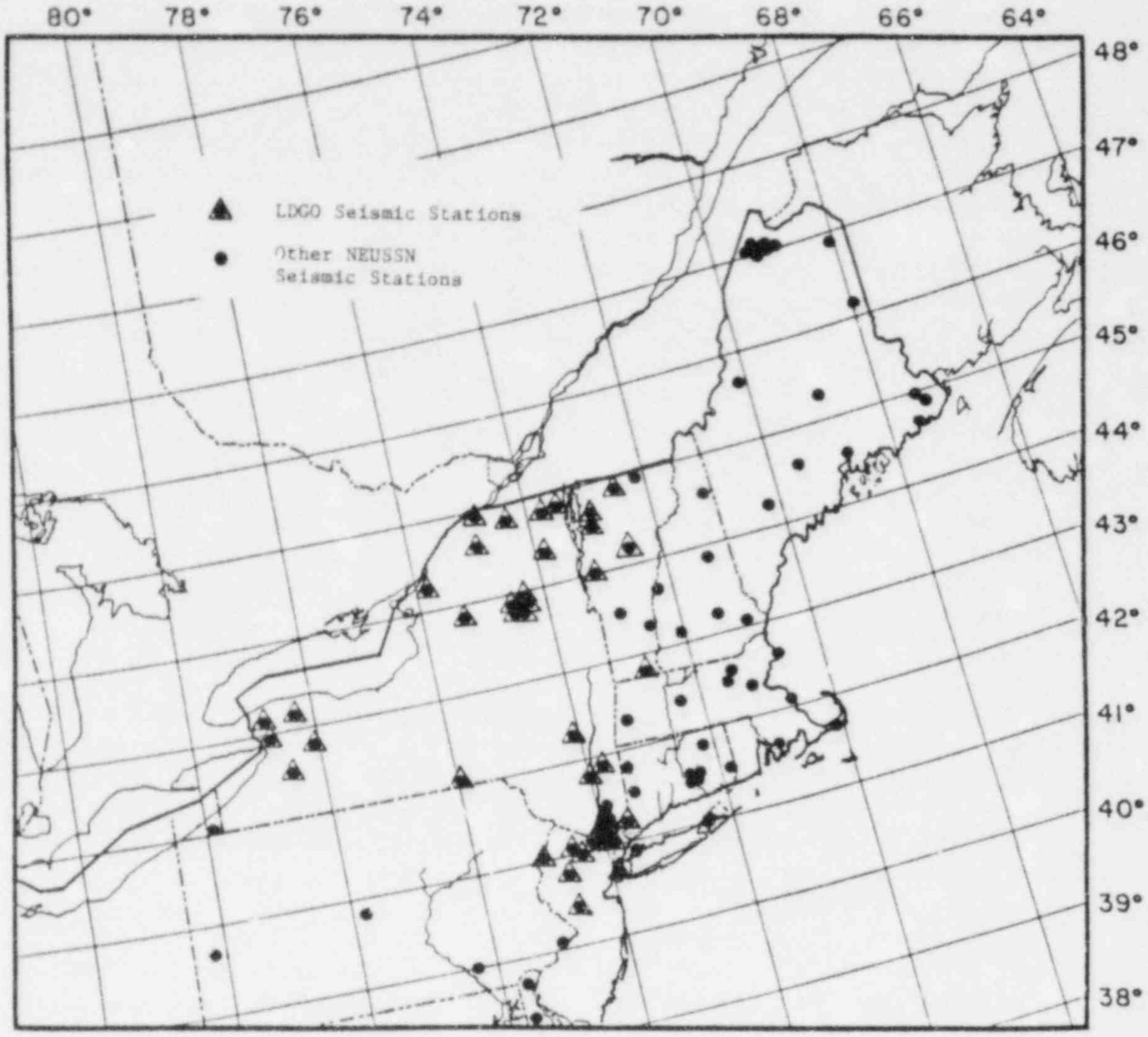

Figure 1: Present configuration of the Northeastern United States Seismic Network (NEUSSN) including the LDGO network in New York, New Jersey, ..d Vermont.

with a corresponding study of the details of the earth structure and seismic wave propagation in the New York State region. These studies address several interrelated problems. More detailed determination of the three dimensional velocity structure is essential in constraining the depths, locations, and focal mechanisms of earthquakes very accurately, and therefore, essential in determining what faults or structures are seismically active. Studies of the attenuation of intensity with distance in the region are aimed at constraining the magnitudes of older earthquakes; and integration of these studies with results of strong motion investigations will help to determine what intensity of ground shaking can be expected at a particular site. In addition, we have been searching through historic accounts of older earthquakes to help constrain their locations and magnitudes. This integrated approach should greatly elucidate our understanding of earthquake hazards in New York State and adjacent areas.

TECHNICAL DISCUSSION

Present Status of the Lamont-Doherty Network

Figure 1 shows the configuration of the various seismic networks operating in northeastern United States and adjacent parts of Canada. The Lamont-Doherty network in New York State, Vermont and New Jersey currently consists of 38 stations, 36 of these stations have single-component vertical seismometers and two (Ramapo Mountain, NJ and Palisades, NY) are 3-component sites. The signals are telemetered by telephone line and radio to a central recording site at Palisades, New York, and recorded on a common time base. Fourteen channels are recorded on a develocorder and all are recorded on an analog magnetic tape recorder. Ten helicorders are used to monitor activity in real time, enabling rapid detection of earthquakes. The entire recording system at Palisades is powered by an uninterruptable power supply system allowing for continuous operation in the event of an emergency failure of public power. The analog magnetic tapes are digitized for detailed analysis of the wave forms of particular events. In addition to these short period seismometers, three SMA-1 strong-motion accelerographs are deployed in the field; one in each of the three areas of relatively high activity in the New York State region (as described below).

In the near future we expect to acquire digital recording capability by connecting the network directly to the PDP 11/34 computer presently at LDGO. This computer will be dedicated to the network as an on-line, real time system for event detection and recording. The develocorder will be shut off as soon as reliable operation of the PDP 11/34 system has been demonstrated.

Routine Data Analysis

The data recorded by the network are routinely analyzed on a daily basis to identify and locate earthquakes. Since quarry blasts are almost a daily occurrence, care is taken to discriminate the smaller shocks from blasts. Regular contact is maintained with several quarries to facilitate and ensure proper discrimination. Particular events are analyzed for determination of accurate locations, depths, focal mechanisms, and other source parameters whenever possible.

Portable instrumentation is deployed by LDGO personnel in the epicentral region of the larger earthquakes as soon after the event as possible to record aftershocks and strong ground motion. Such surveys were recently conducted for the two large earthquakes in New Brunswick and New Hampshire as well as for the Long Island Sound earthquake of October 1981 and the Abington, Pa. earthquakes of March 198[illegible].

The network data are exchanged with other participating institutions in northeastern United States and adjacent parts of Canada to obtain more accurate earthquake locations. A bulletin, listing all earthquakes in and around New York State, is published quarterly to provide preliminary locations and other earthquake parameters (such as magnitude and origin time). A final yearly bulletin is published at the end of each year.

RESULTS OF ONGOING RESEARCH PROGRAM

The ultimate goal of our ongoing research program is to arrive at an understanding of earthquake phenomena and earthquake hazards in the northeastern United States with specific emphasis on New York State and adjacent areas. The program is also involved in research related to the problem of understanding intraplate earthquakes in general. Important contributions to these long term goals have emerged from previous studies involving data recorded by the Lamont-Doherty network. Here, we discuss our results to date, the present status of our research efforts, and suggested directions for future investigations.

Seismicity

Since the inception of the network in 1970, more than 800 earthquakes ($1 \leq m_b \leq 5$) have been recorded and located in the northeastern U.S. and adjacent Canada. Figure 2 shows the distribution of the larger earthquakes ($m_b \geq 2$) recorded by our network from its inception in 1970 through 1980. In this figure we chose the magnitude threshold so as to reduce the bias introduced by non-uniform coverage in space and time. Figure 2 reveals some important features of the seismic activity in this area and it is instructive to compare it with Figure 3 which shows historical seismicity for the period 1534-1959.

The features of interest common to both Figures 2 and 3 are: (1) a NNW-trending zone of seismicity extending from northern New York to southern Quebec, (2) concentration of seismicity in western New York, and in the region near New York City, (3) relative absence of activity in the central part of New York State, southern Vermont, and western Massachusetts, and (4) the general preponderance of earthquake activity along coastal New England as well as offshore.

The strong similarities in the distribution of seismicity between Figure 2 (1 decade of instrumental data) and Figure 3 (over 400 years of historical data) suggest that the spatial distribution of earthquakes in this area has been relatively stationary over the last few hundred years. The most important difference between the historical data and that from our network is, however, in the quality and in the amount of meaningful information that can be derived from them. Whereas the locations of most historical events suffer from larger

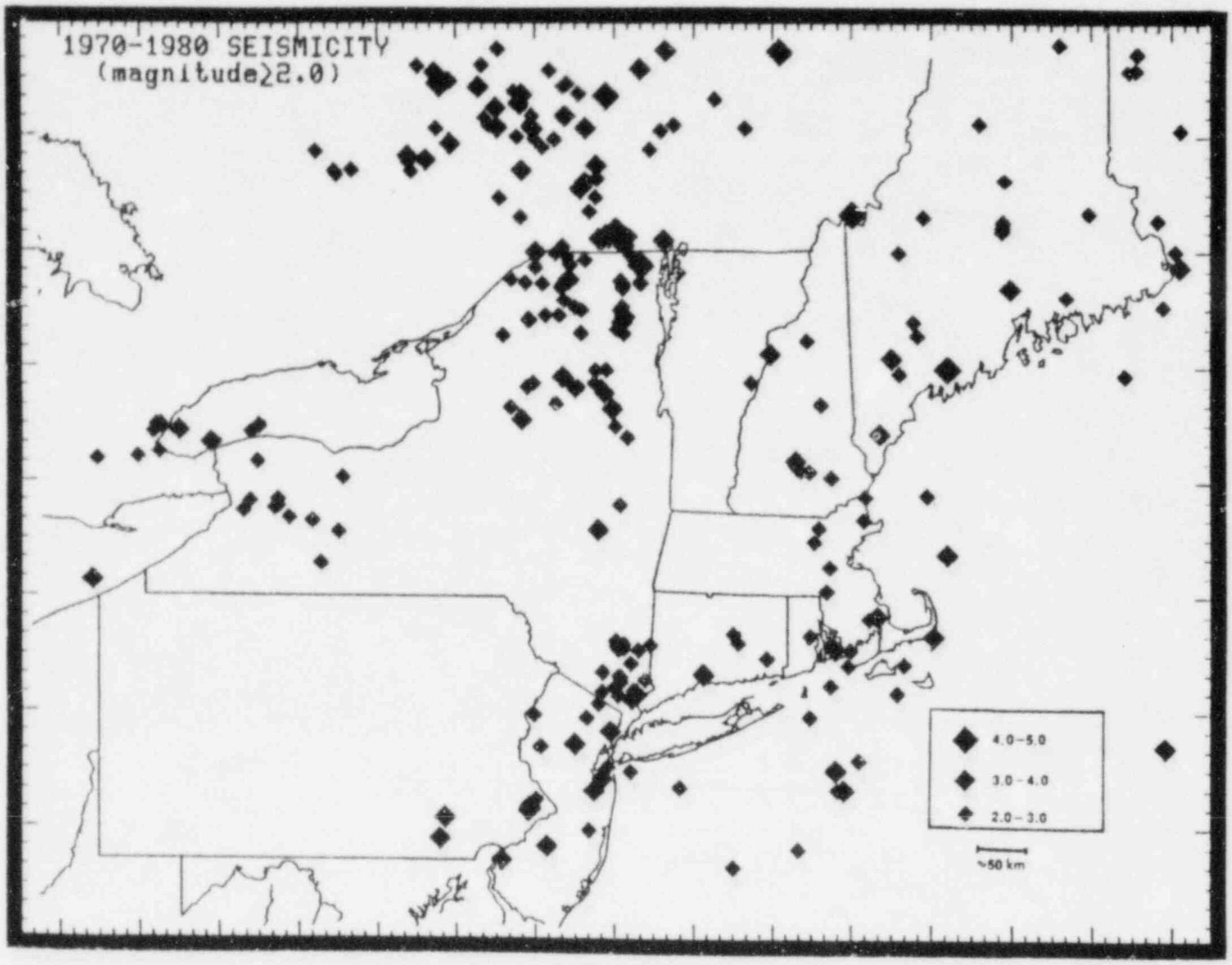

Figure 2: Earthquakes (magnitude 2.0 and greater) recorded by the LDGO seismic network in New York State and adjacent areas from 1970 through 1980.

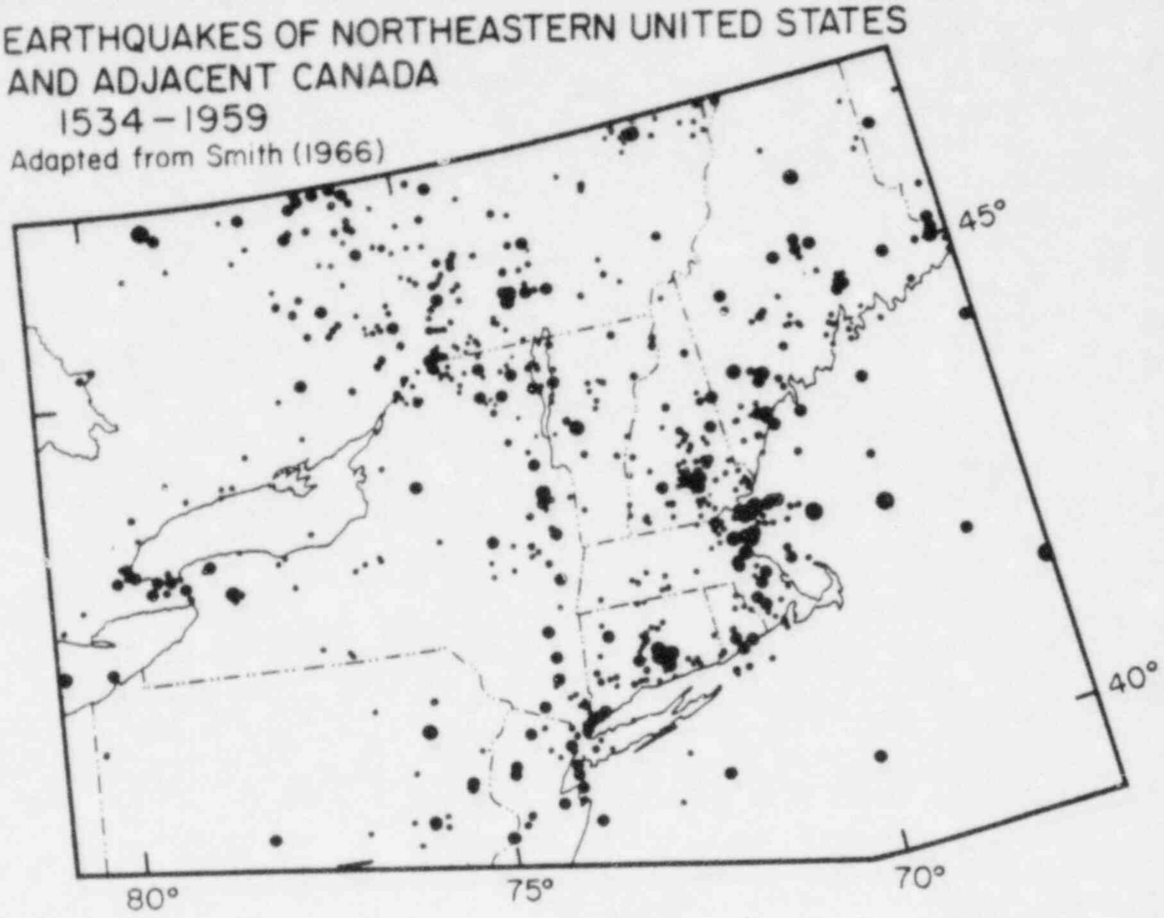

Figure 3: Earthquake locations (solid circles) in northeastern United States and adjacent Canada for the period 1534-1959 (adapted from Smith, 1966). The data shown include both instrumentally located epicenters and locations determined from intensity data.

uncertainties (a few tens of kilometers), the instrumental locations are now usually determined to better than a few kilometers. Focal depth determinations are only available from the network data. Furthermore, the details of hypocentral distribution and their possible association with faults or other structural features can only be meaningfully attempted with the instrumental data.

Focal Depth Determination

Earthquakes located by the LDGO network in the New York State region are, most likely, confined to the crust (depth <40 km), but in general focal depths of these events are not very well constrained. At present, accurate focal depth determination is only possible when at least one station is located close enough to the event that the epicentral distance is less than about twice the depth of the event. Unfortunately, the density of stations in the New York State array is not sufficient to assure us that a station will be close to every event. To overcome this problem we have, in the past, deployed portable seismograph stations to supplement the permanent network whenever a significant sized event occurred in the New York State region. These studies have enabled us to obtain accurate depths of aftershocks. Better depth control is also available in those parts of northern and southern New York where the distribution of stations is more dense.

With the inclusion of the aftershock data the following results have been obtained. In the southeastern New York - northern New Jersey area well determined focal depths range from near surface (about 0.5 km) to at least 10 km. Within the Adirondack massif, in northern New York, focal depths are usually less than 5 km, although events as deep as 18 km are known to have occurred to the north of the massif in northern New York and western Quebec. In western New York well determined focal depths are quite shallow and confined to the upper crustal layers (<5.0 km).

To obtain a more complete picture of the distribution of focal depths within the New York State region we are investigating additional methods of determining focal depths of local earthquakes. We are currently digitizing our analog magnetic tape data from various events and looking for characteristic depth phases in the entire wave form. In particular we are attempting to determine if short-period surface waves can be used as depth diagnostics. Longer period surface waves (20-50 sec) have been used in various studies as depth diagnostics for earthquakes with $m_b > 4.5$ (e.g., Kafka and Weidner, 1979). To assess the utility of surface waves in the shorter period range (as low as 1 sec) it will be necessary to study the effect of the propagation path on the observed seismic signal, and this will be one of our future research efforts.

The seismograms shown in Figure 4 were recorded by our network from two earthquakes which occurred within a dense seismic network (operated by Woodward-Clyde Consultants) in the vicinity of the Indian Point nuclear power plant near Peekskill, N.Y. Because of the large number of stations surrounding the epicenters, the depths of these two events are quite well constrained. Also, since the events are very near each other, the effect of the path on the observed seismograms is nearly identical, and differences in the observed waveforms should

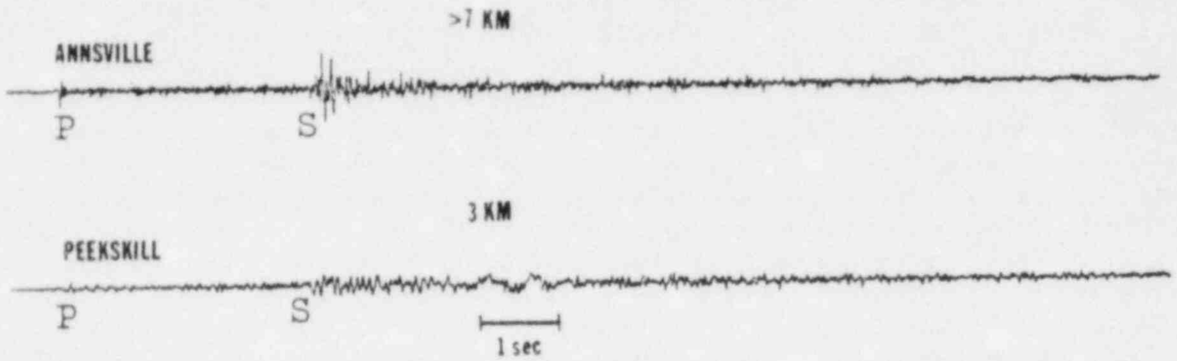

Figure 4: Seismograms recorded at station WPR from two earthquakes located within 2 km of each other (about 30 km from WPR) and at different depths. Note how the seismogram of the shallower earthquake shows a long period ($\sim$ 1 sec) wave beginning about 2 sec after the arrival of the S-wave, i.e. above the bar marked '1 sec'.

result mainly from source characteristics. The seismogram of the shallower event shows a long period (≈1 sec) wave which may be a fundamental mode Rayleigh wave. Similar long period waves have been observed from quarry blasts. We are investigating the utility of using these waves as depth diagnostics.

This method could be very valuable in differentiating faults along which activity extends from the surface to depths of 10 to 15 km from those along which activity is very shallow. The former could be regarded as having a greater probability of being sites of large shocks ($5 < m_b < 7$) whereas the latter would have a low probability of having large shocks. This could also be a way of studying the configuration of faults within a detached (allochthonous) block or sheet.

Relationship of Seismicity to Structural Features

An important element of the ongoing and future research is to elucidate the relationship between earthquake occurrence and structural features. Although that relationship remains a mystery for much of the east coast, our efforts have proven successful in this context for parts of the Northeast. For example, the study by Aggarwal and Sykes (1978) has demonstrated that earthquakes in the greater New York City area occur predominantly along northeast trending faults. Their study also suggests that the Ramapo fault zone which bounds the Newark basin on its NW side is an active feature. More recent results (Kafka et al., in preparation, Appendix C) show that about half of the earthquakes located by the network in the New York City metropolitan area occur along other geologic structures which surround the Newark basin (Figure 5). Fletcher and Sykes (1977) have shown a correlation between earthquake activity near Attica, N.Y. and the Clarendon-Linden fault zone.

For other parts of the Northeast, however, this relationship remains unclear. For example, an examination of Figure 2 reveals that the NNW trend in seismicity extending from northern New York into western Quebec, Canada, is transverse to the St. Lawrence River Valley and bears no apparent relationship to known structures. Also focal mechanism solutions indicate that earthquakes in this region occur predominantly along NNW trending fault planes. In contrast, the predominant trend of the mapped faults in northern New York is northeasterly, although some faults are known that are nearly parallel to (and appear to be related to) the NNW trend of the seismic zone and the fault plane solutions (Yang and Aggarwal, 1981). Thus, the question arises as to whether some of the earthquakes in northern New York and western Quebec represent new faulting, or occur along pre-existing faults not readily indentifiable through geologic mapping. We believe that the combined use of seismic data from the network, LANDSAT imagery, as well as available gravity and magnetic data should help resolve such problems. Combining these data Yang and Aggarwal (1981) tentatively identified a few NNW trending lineaments in northern New York that can be spatially correlated with earthquake locations and fault plane solutions. We will continue our efforts in this direction.

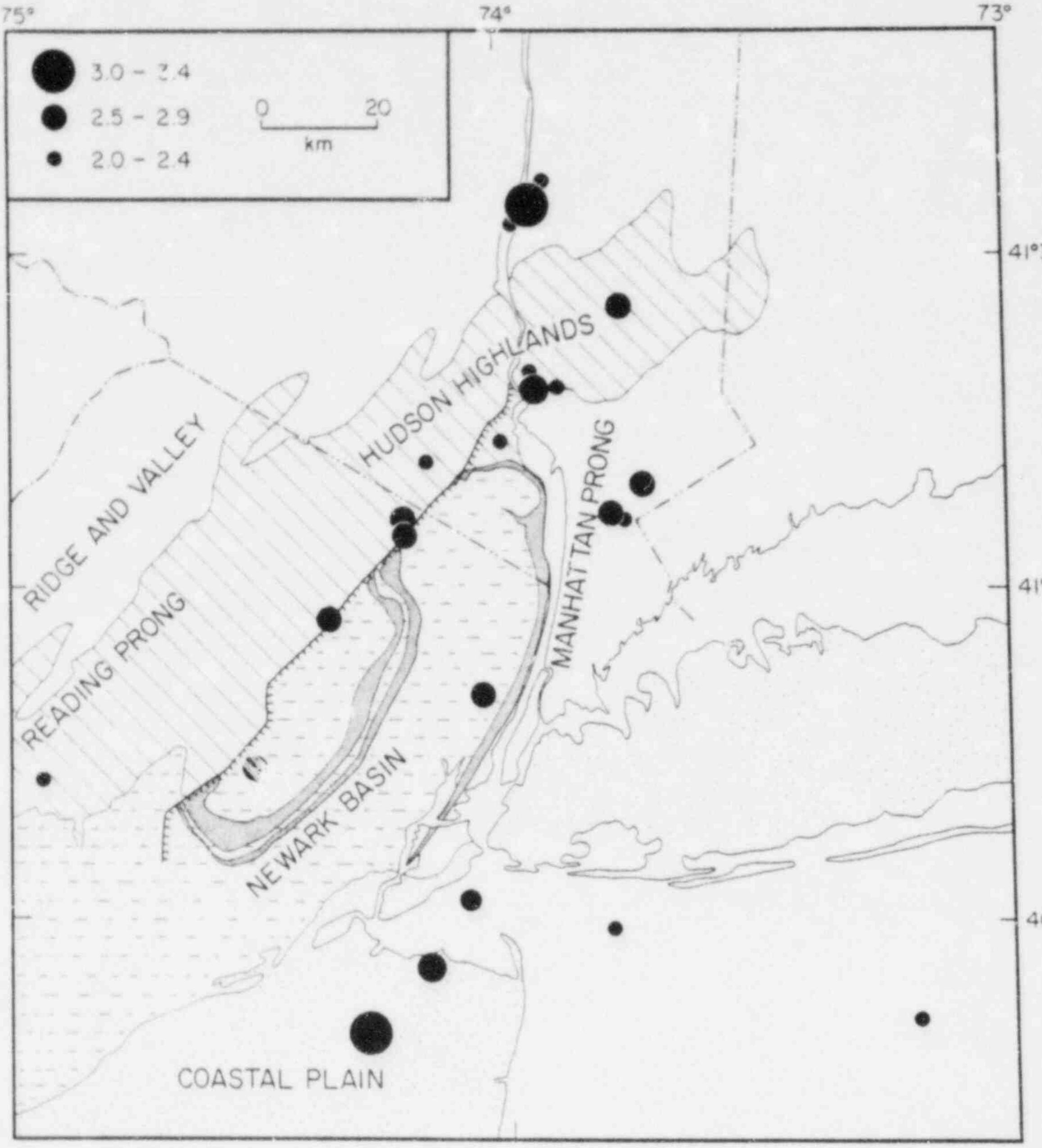

Figure 5: Earthquakes recorded by the LDGO network in the New York City metropolitan area, and a schematic representation of regional geologic structures adapted from Ratcliffe (1980). Coverage is complete for area shown for $m_b \geq 2.0$.

Seismic Provinces

One of the important long-term tasks of this project is to define seismic provinces in the northeastern United States. A significant portion of our research, therefore, involves identifying and isolating parameters that influence earthquake processes in particular regions within the Northeast. We think that a combination of parameters rather than a single parameter (for example seismicity alone) should be used in defining seismic provinces.

Our recent research in this context (Yang and Aggarwal, 1981) indicates that the parameters most useful for such purposes are: (1) the observed spatial distribution of earthquakes, (2) similarities in fault plane solutions, (3) relative uniformity in the inferred directions of principal stress, (4) correlations of earthquake locations with mapped or inferred faults, and/or agreement between regional fault patterns and fault trends inferred from fault plane solutions.

On the basis of such factors we have tentatively identified two seismic provinces: (1) The Adirondack-Southern Quebec Province. A NNW trending zone of seismic activity, about 200 km wide and at least 500 km long, extending from the southeast Adirondacks into Quebec, Canada. In this zone thrust faulting on NNW trending planes appears to predominate and the inferred axis of maximum horizontal compression (σ_1) is relatively uniform and trends ENE (Figure 6). (2) The Piedmont-Appalachian Province. A northeast trending zone of seismic activity along coastal northeastern U.S., extending from northern Virginia to southeastern New Hampshire and possibly continuing farther northeast. The recent large earthquake (m_b = 4.5) near Laconia, New Hampshire (January 18, 1982) appears to have been located in this province, and the recent large earthquake (m_b = 5.8) in New Brunswick, Canada (January 19, 1982) may have also been located in this province. In this province, either high angle reverse faulting or thrust faulting on northeasterly trending planes appears to predominate. The inferred direction of maximum compressive stress is relatively uniform and trends ESE (Figure 6). Where data are sufficient, earthquake locations and focal mechanism solutions usually correlate well with the surface traces and trends of mapped faults.

We think that such efforts will lead to a more rational definition of seismic provinces in the East than has been possible in the past. This in turn would facilitate the decision making process in siting critical facilities and evaluating earthquake hazard in the East.

Fault Plane Solutions - Intraplate State of Stress

A key element in understanding the earthquake phenomenon near the east coast and in other intraplate regions is the state of stress and its origin. The network has thus far produced about 30 fault plane solutions for the Northeast. The maximum compressive stress axes (σ_1) inferred from these fault plane solutions are plotted in Figure 6 (solid symbols) along with the orientations of σ_1 inferred from hydrofracturing experiments (Haimson, 1977; open squares). The inferred σ_1 delineate two distinct stress provinces, and show

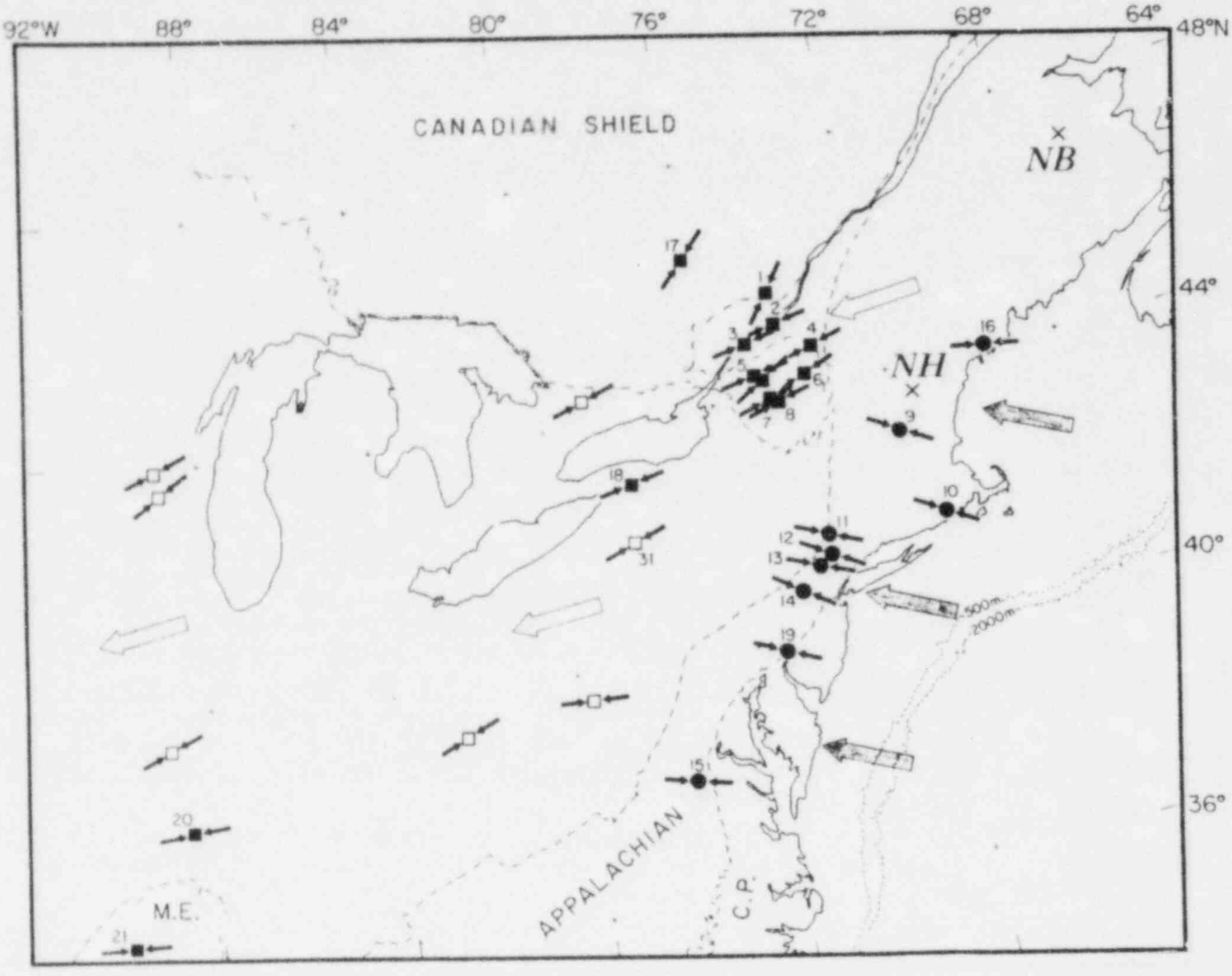

Figure 6: Stress map of northeastern North America (from Yang and Aggarwal, 1981) showing (converging arrows) the orientation of maximum horizontal compression (σ_1) inferred from fault plane solutions (solid symbols) and hydrofracture data (open squares). Major tectonic provinces are outlined; C.P. = Coastal Plain and M.E. = Mississippi Embayment. Crosses are locations of two recent earthquakes; NB = New Brunswick (January 9, 1982; m_b = 5.8), NH = New Hampshire (January 18, 1982; m_b = 4.5).

a remarkable internal uniformity. Within the Appalachians, near the Atlantic coast, σ_1 trends ESE; west of the Appalachians, in the interior of the continent, σ_1 trends ENE in agreement with the findings of Sbar and Sykes (1973). The change in the maximum compressive stress direction from ESE to ENE appears to be fairly abrupt. The zone of transition coincides to some extent with regions that nearly lack seismicity (Figures 2 and 3), and also coincides with a zone of transition in crustal structure (Figure 7) determined from travel time residuals of teleseismic P-waves by Peseckis and Sykes (1982, Appendix B).

Important contributions to the understanding of the interrelationship between seismicity, stress, and geologic features in the northeastern United States have recently emerged and are discussed by Yang and Aggarwal (1981). The ESE orientations of the σ_1 axes within the Appalachians are observed to be almost parallel to the direction of motion of North America relative to Africa (Chase, 1978; shaded arrows in Figure 6). In contrast, the ENE orientations of the σ_1 axes in the interior of the continent are observed to be nearly parallel to the directions of absolute motion of the North American plate as inferred from hot spots (Chase, 1978; open arrows in Figure 6).

In the context of intraplate stress it is obviously very important to know the fault plane solutions and depths of the recent large earthquakes (Figure 6) which occurred during January 1982 in New Brunswick, Canada (m_b = 5.8) and in New Hampshire (m_b = 4.5). The New Brunswick event occurred in what would appear to be the northern extension of the Appalachian stress province (discussed above) and the New Hampshire earthquake occurred within this province. These events were well recorded on many different types of seismic instruments over a very large portion of the United States and Canada. Many WWSSN stations recorded these events, and at Lamont-Doherty they were also recorded on a Wood-Anderson seismometer as well as being recorded by the local network. These events have generated sufficient data to be studied in great detail. Our future research will include an attempt to determine the fault plane solutions and depths of these events (and perhaps some of the larger aftershocks) by a combined use of both P-wave first motions and radiation patterns of teleseismic surface waves. The techniques described in Kafka and Weidner (1979) will be used for the surface wave study. We are, of course, anxious to know whether or not those mechanisms are consistent with the stress provinces inferred thus far.

Crustal and Upper Mantle Velocity Structures

An adequate knowledge of the crustal and upper mantle velocity structure is a prerequisite for accurate earthquake locations and meaningful interpretation of the seismic data. For much of New York State and areas adjacent to it, such information was either nonexistent or very limited in scope prior to the installation of local seismic networks. A better determination of the seismic velocities in the Northeast has, therefore, been and remains one of our primary objectives.

The crust and the upper mantle is being probed on a local and regional scale, using quarry blasts as well as earthquakes. Detailed

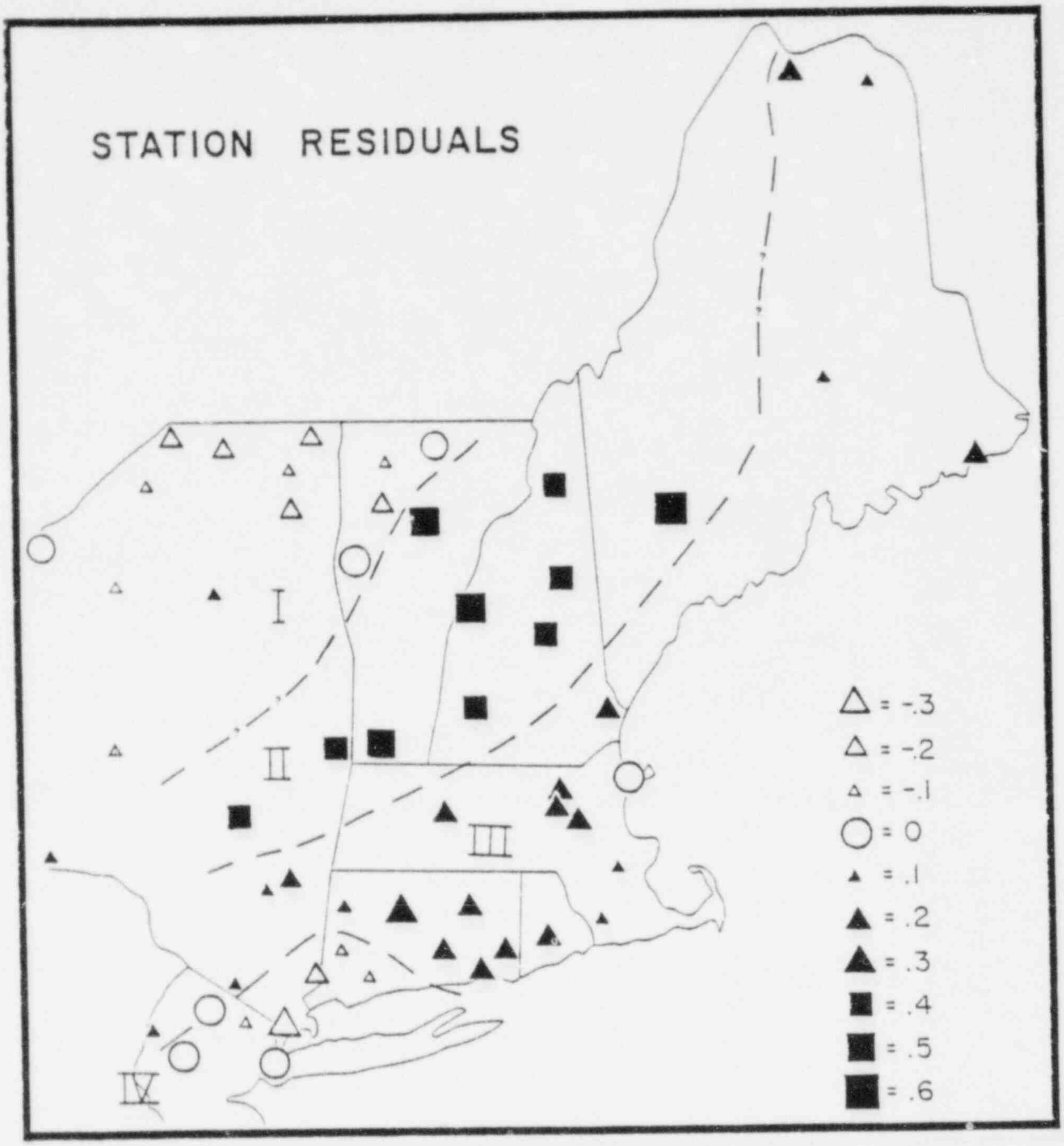

Figure 7: Travel time residuals for P-waves at stations of Northeastern United States Seismic Network. Residuals shown are taken from Peseckis and Sykes (1982) and Taylor and Toksöz (1979). Several subregions are defined (shown by dashed lines) in which residuals are constant to within 0.15 sec.

investigations of P and S velocities have been carried out in parts of the Adirondacks (for example, Aggarwal et al., 1975) using road construction and quarry blasts. In addition P arrivals from distant earthquakes and nuclear explosions have been studied to scan vertical sections of the crust and upper mantle beneath the seismic stations (Fletcher et al., 1978; Pesechkis and Sykes, 1982, Appendix B).

Thus far, tentative crustal and upper mantle velocities have been deduced for parts of New York State and adjacent areas, and these models are used in our earthquake location procedures. Our most recent investigations indicate that lateral variations in seismic velocities occur locally as well as on a more regional scale. Peseckis and Sykes (1982, Appendix B) examined the crust and upper mantle of the northeastern United States using relative residuals of P-waves from distant earthquakes and underground nuclear explosions recorded by the New York State seismic network and other seismic stations in New England and adjacent parts of Canada. Relative residuals (Figure 7) vary from -0.3 to +0.6 sec in the area, but they are constant to within 0.15 sec in broad subregions. Station residuals vary markedly between subregions. Since the transition zones between subregions are no more than 50 to 100 km wide, most differences in velocity appear to be situated in the upper 100 km. This result appears to conflict with existing refraction data which indicate that very little of the differences in travel time can be associated with the crust or the uppermost mantle. All of the stations with negative residuals (early arrivals and higher velocities) are situated in New Jersey and New York on basement of Grenville age. Most of the positive residuals occur in New England and appear to be associated with terranes that were sutured onto North America during the Paleozoic. In New England lines of constant residual appear to be rotated about 30° with respect to the pattern of regional geology (Figure 7). Thus, the structure of the upper crust in parts of New England may not correspond to that of deeper units and may be allochthonous as preliminary results from COCORP lines also indicate. Large azimuthal variations in residuals are found for most stations and appear to reflect short-wavelength changes in velocity within the outer 50 km of the earth. The pattern of azimuthal variations is similar at nearby stations in many cases. We will continue such studies of velocity structure and wave propagation in New York State and adjacent areas using local earthquakes, quarry blasts, and teleseismic earthquakes and explosions.

Historical and Pre-Network Data

Valuable instrumental data are available for some of the events that occurred prior to the inception of the present network in 1970. Several seismic stations were in operation in the Northeast (both in the U.S. and Canada) for various periods of time. The Canadian seismic records, which go as far back in time as 1907, are available at Lamont-Doherty. We have recently transferred seismograms from Fordham University, which go back to the 1920's, to Lamont-Doherty.

With the progressive increase in our knowledge of the velocity structure, we have relocated pre-network events in the New York City region for which sufficient instrumental data exist (Aggarwal and Sykes, 1978). We will continue this effort for other areas of the

Northeast. We are in the process of determining magnitudes for as many past events as possible using the available data. We are also examining intensity data for larger historical events. Recent results from studies of the attenuation of intensity with distance in the New York State area (Schlesinger-Miller et al., 1980) will be combined with future studies of the attenuation of seismic waves recorded by e network. This research will not only help to estimate magnitudes of historic earthquakes but will also help to predict the intensity of ground shaking to be expected at a given site. Particular emphasis has been recently directed towards this type of analysis for earthquake activity in the New York City metropolitan area (Kafka et al., in preparation, Appendix C), and these studies are discussed in detail below. It is hoped that such improved relocations and magnitudes combined with better knowledge of attenuation will provide a more accurate data set to understand the seismotectonics of parts of the east coast and to evaluate seismic risk.

Seeber and Armbruster (1981) have reevaluated historical records of earthquakes in the southeastern United States. On the basis of these data they have challenged long-held concepts about the 1886 earthquake in Charleston, South Carolina and intraplate earthquakes along the Atlantic seaboard in general. Their model of the faulting process of the 1886 event involves back-slip on Paleozoic detachments, and they propose that this is a fundamental tectonic process in the crust of this region. Seeber and Armbruster's research on the southeastern United States is continuing and expanding to other areas including the Northeast (discussed below). We consider their efforts and our efforts as part of an integrated program to study seismicity and tectonics along the Atlantic seaboard for the purpose of estimating the seismic hazard in this region.

Strong-Motion Studies

Very little is known about the strong ground motion and spectral content of eastern earthquakes. We have deployed a few strong motion accelerometers in the field. To date, several accelerograms have been recovered: four for Blue Mountain Lake earthquakes (m_b = 2.2 to 2.7), two for the Raquette Lake, New York earthquake (m_b = 3.9), one for an earthquake near Dale, New York (m_b = 1.5) and three for aftershocks of the New Brunswick earthquakes of 1982. The results of several Lamont-Doherty studies of these records (Fletcher and Anderson, 1974; Anderson and Fletcher, 1976; Boatwright, 1978) indicate that the earthquakes at Blue Mountain Lake and Raquette Lake had much higher stress drops ($\approx$100 bars) and higher corner frequencies than the earthquake at Attica ($\approx$10 bars). The mean stress and the frictional regimes of the regions where the earthquakes occurred appear to differ considerably (Boatwright, 1978), indicating that the eastern United States regionalization proposed by Street et al. (1975) is inaccurate and probably inapplicable to the Northeast. If the Blue Mountain Lake and Raquette Lake events are more typical of earthquakes in the Northeast, then the design parameters of many structures in this region may be in error because they are derived from records of western United States earthquakes.

At the present time, three SMA-1 strong-motion accelerographs are deployed in the field as part of our permanent network; one in each of the three seismically active areas in the New York State region. One SMA-1 is deployed in northern New Jersey (Ramapo Mountain, New Jersey), one in the northern New York (Massena, New York), and a third in western New York (near Attica).

Lamont-Doherty participated along with various other institutions in field studies of the epicentral region of the recent large earthquakes in New Brunswick and New Hampshire. We deployed a strong-motion accelerograph in the vicinity of the New Brunswick epicenters, and this instrument was triggered by several aftershocks. In addition several strong-motion records have been recorded from the New Hampshire earthquake at permanent sites (J. Ebel, personal communication). These records will be analyzed and should greatly enhance our knowledge of strong ground motion in the Northeast.

Historic Earthquakes and Present-Day Seismicity in the Greater New York City Area

The instrumental record of earthquakes for the greater New York City area and for much of the eastern United States is quite poor for the period of time prior to the installation of the Northeast Seismic Network about 10 years ago. Only one station, Fordham, was in operation in a large area near New York City prior to the installation of the stations at City College and Palisades in 1948 and 1949 respectively. The accuracy of epicentral locations for much of the area was still very poor until Ogdensburg, New Jersey and Sterling Forest, New York were installed in 1962. Hence, the detection of events smaller than about magnitude 3.5 and the accurate location of earthquakes of all sizes is effectively limited to the period since 1962. It is only during the last 20 years for which epicentral locations became accurate enough that they can be meaningfully correlated with specific geologic structures.

Nevertheless, the historic record of felt earthquakes near New York City is much longer and extends back about 250 years. Sykes (in preparation) has been studying data from about 50 older events within a 2° x 2° area centered near New York City in an attempt to extract more information from the pre-instrumental history of felt earthquakes. Kafka et al. (in preparation, Appendix C) determined magnitudes on the Nuttli scale for periods near 1 sec for 7 moderate sized earthquakes in that area for which instrumental data are available. Reported maximum intensities scatter greatly when plotted against either magnitude or the size of the felt area. Magnitudes scatter very little, however, when they are plotted as a function of the felt area. Nuttli, Street, and others find a similar behavior for data from the central United States and New England. The magnitude-felt area relationships for the greater New York City area is very similar to that for other areas of the central and eastern U.S.

Sykes has obtained magnitudes from this magnitude-felt area relationship for historic earthquakes in the New York City area extending back to about 1730. It is clear that large errors in magnitudes arise if they are calculated instead from maximum intensities. Thus, more accurate magnitudes can now be assigned to events of the past 250 years using the size of felt areas. Maximum intensities can then be used as a rough indication of depth of focus. Seismic moment may also be related to felt area, and such relationships will be investigated.

Calculated magnitudes of some of the larger earthquakes in the area are listed in Table 1 along with the maximum reported intensity. The largest event is that of 1884 ($m_N \approx 4.9$) while three other shocks for which the maximum reported intensity was VI or VII are assigned magnitudes of 4.7. The magnitudes of the shocks of 1783 and 1895 are larger than what would be inferred from most magnitude-intensity relationships. For shocks in New Jersey in 1927 and 1957 the magnitudes computed from their felt areas are much smaller than those computed from their maximum intensities. Those two events, like the Moodus, Connecticut, earthquake of 1791 or some events reported by Nuttli, appear to have very large maximum intensities for a given magnitude. The high intensities appear to be real and are probably related to very shallow focal depths. Realistic calculations of seismic risk (e.g., estimates of the probability of shaking exceeding a given intensity) must take into account the huge scatter in magnitude-intensity relationships and the occurrence of unusual shocks with high intensities but small felt areas and magnitudes.

Recently there has been considerable debate about the location of the August 10, 1884 earthquake which occurred in the vicinity of New York City (e.g., Aggarwal and Sykes, 1978; Fisher, 1981). Accurate location of this event is very important in estimating seismic risk in the New York City area. We have begun an examination of original data sources on this event (Armbruster, unpublished data) in an attempt to resolve some of the unanswered questions. This work (Figure 8) indicates that intensities of the 1884 earthquake were remarkably uniform from western Connecticut to the vicinity of Philadelphia, Pennsylvania without the concentrated center of higher intensities that is used to define the location of a historical earthquake. The most reasonable preliminary interpretation is that this earthquake was centered offshore some distance southeast of New York City (Armbruster, unpublished results). If the earthquake was offshore, then the highest intensities observed may have been smaller than the intensity near the epicenter. Thus, the 1884 earthquake could have had an energy release much greater than that of other known events in the vicinity of New York City.

TABLE 1

Magnitudes of Nuttli Scale (m_N) Inferred from Felt Areas for Some Larger Earthquakes in a 2° x 2° Area Centered Near New York City

Date	Magnitude	Maximum Report Intensity
1884	4.9	VII
1737	4.7	VII
1783	4.7	VI
1895	4.7	VI
Lower magnitudes were obtained for the following shocks of intensity VI or VII:		
1927	3.8	VI-VII
1957	3.2	VI

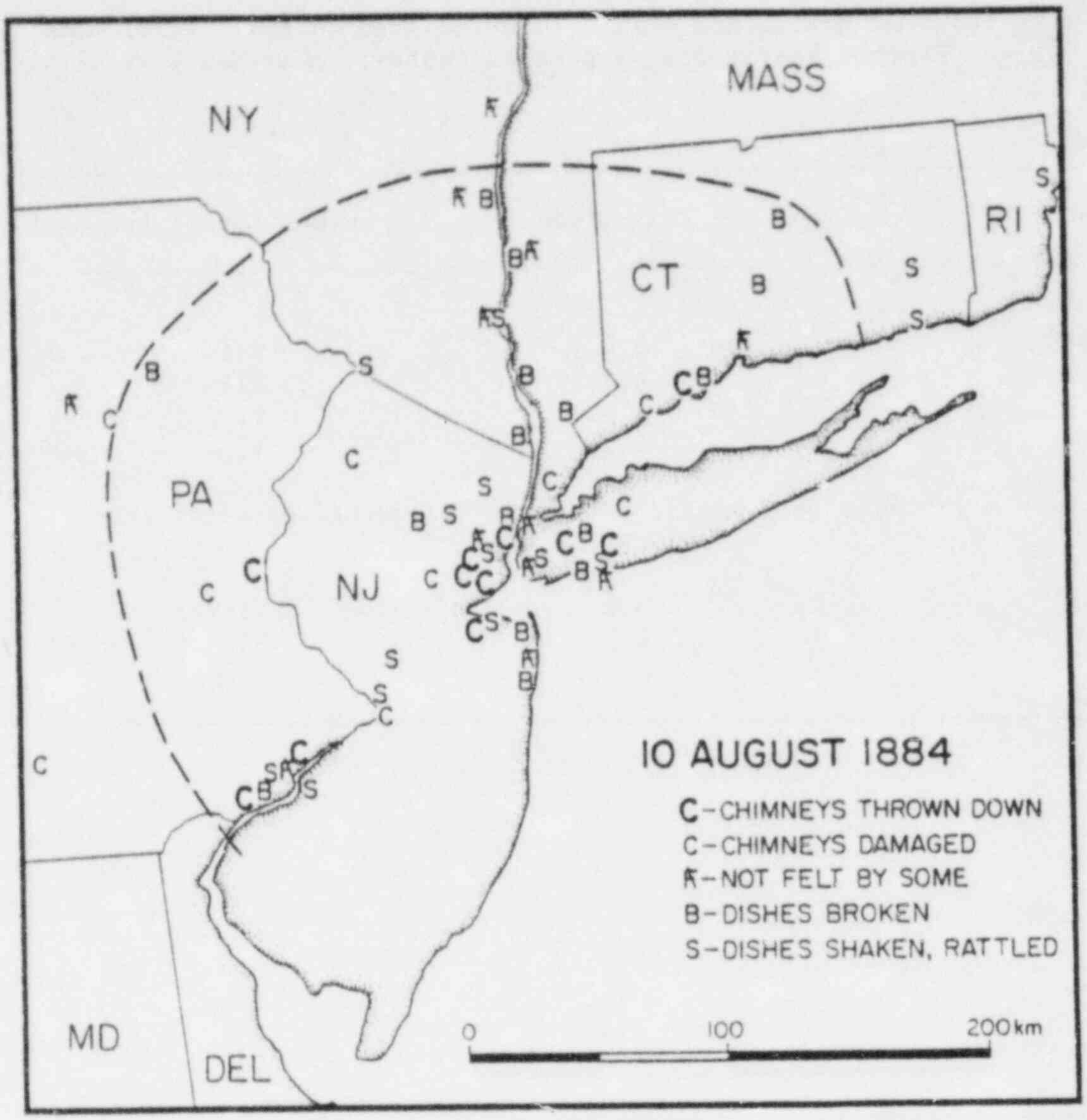

Figure 8: Specific effects of the August 10, 1884 earthquake (Armbruster, unpublished data, 1982). High intensity effects occur in a wide area centered on New York City, but there is no concentrated center of higher intensity on land as indicated by places where some did not feel the earthquake. Data obtained from the following newspapers published in New York City: Times, World, Herald, Tribune, Sun and Post. These high intensity effects appear to be limited to the region enclosed by the dashed line.

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APPENDIX A

LAMONT-DOHERTY PUBLICATIONS THAT HAVE RESULTED FROM RESEARCH STIMULATED BY NYSERDA*, NRC*, AND USGS GRANTS FOR STUDIES OF SEISMICITY AND TECTONICS OF THE EASTERN UNITED STATES

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APPENDIX B

P-WAVE RESIDUALS IN THE NORTHEASTERN UNITED STATES AND THEIR RELATIONSHIP TO MAJOR STRUCTURAL FEATURES

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Abstract. The crust and upper mantle of the northeastern United States are examined using relative arrival times of P-waves from underground nuclear explosions and distant earthquakes recorded by the Lamont-Doherty seismic network and by other stations in the Northeast. Average station residuals vary by as much as one second in the northeastern United States. Residuals are constant to within 0.15 s, however, in broad sub-regions. Five such areas are defined based on the near constancy of residuals. Station residuals vary markedly between these sub-regions. One transition zone between sub-regions is 100 km wide indicating velocity differences in this area extend to depths of 100-200 km. However, the transition zone between the sub-regions with the largest residual difference is 55 km wide, indicating that these differences in velocity are situated in the upper 50-100 km. This result conflicts with existing refraction data which indicate that very little of the differences in travel time can be associated with the crust or the uppermost part of the mantle. All of the stations with negative residuals (i.e., early arrivals and fast velocity) are situated in New Jersey and New York on basement of Grenville age. Most of the positive residuals occur in New England and appear to be associated with terranes that were sutured onto North America during the Paleozoic. In New England lines of constant residual appear to be rotated with respect to the pattern of regional geology. Thus, the structure of the upper crust in parts of New England may not correspond to that of deeper units and may be allochthonous. Large azimuthal variations in residuals are found for most stations and appear to reflect short-wavelength changes in velocity within the outer 50 km of the earth. The pattern of azimuthal variations is similar at nearby stations in many cases.

Introduction

Laterally varying structures with anomalous velocities are found near active plate margins. Some of these are temperature dependent effects which decay in time once the area is no longer situated near an active plate boundary. Some structural features that are formed near plate boundaries, such as those related to the juxtaposition of units of contrasting seismic velocity, are not temperature dependent. Hence, differences of those types may be expected to persist long after a region has been situated near a plate boundary, probably for hundreds of millions of years. Relatively few attempts have been made to look for geophysical signatures of past plate tectonic events, such as the suturing or overthrusting of terranes of contrasting properties. Few attempts have been made to examine old plate boundaries in three dimensions. In this paper we use travel-time residuals of P waves to study differences in velocity structure for a former plate boundary.

The northeastern United States has been situated at or near a plate margin several times during the past 1 billion years. At least two collisional episodes (Grenville and early to mid-Paleozoic) and two of rifting (late Precambrian-Cambrian and Triassic-Jurassic) are evident [Wilson, 1966; Dewey and Burke, 1973; King and Zietz, 1978; Osberg, 1978; McLelland and Isachsen, 1980; Seyfert, 1980]. The area has been located well inside the North American plate for the last 150 Ma. Parts of New England appear to have been sutured onto the rest of North America during the Paleozoic. Hence, one or more former plate boundaries pass through the region, and their deep structure can be studied by various geophysical techniques such as gravity and magnetic measurements, seismic reflection and refraction studies, and travel-time residuals.

Past studies to determine the velocity structure of the Northeast include refraction work [Katz, 1955; Schnerk et al., 1976; Chiburis and Graham, 1978; Taylor et al., 1980], surface wave analysis [Dorman and Ewing, 1962; Taylor, 1980], as well as teleseismic P wave residual studies [Fletcher et al., 1978; Taylor and Toksöz, 1979]. COCORP has made some multichannel reflection lines in the area [Brown et al., 1981] as part of a project to complete a profile from the Grenville craton eastward through the New England Appalachians. Earlier reflection work in the southeastern United States and in Quebec [Cook et al., 1979] indicates that large portions of the Appalachian orogen, especially the eastern parts, may be allochthonous. The preliminary COCORP results in the Northeast indicate similar thin-skinned overthrusting as farther south. If large parts of New England are, in fact, allochthonous, care must be used in comparing travel-time residuals and refraction data which may sample rather different materials and structures in the lower crust and uppermost mantle. One of our major findings is that while travel-time residuals do correlate in a general way with Paleozoic tectonic provinces, the locations and strikes of boundaries between sub-areas in which residuals are nearly constant do not always coincide with those of major geologic provinces as mapped at the surface.

Our investigation has several similarities to a study of P-wave residuals that Taylor and Toksöz [1979] performed for New England and eastern New York. Several of the stations used in our study and theirs overlap in eastern New York and western Vermont. We also examined data from four stations in western New York, an area not included in their study as well as data from nine other stations in New Jersey and eastern New York that either were installed since their work or were not analyzed by them. Since the set of earthquakes we used differs from theirs, our results provide independent evidence on travel time for the stations that are common to our two studies.

Taylor and Toksöz [1979] infer higher velocities beneath the Precambrian Grenville province than for that part of New England that was affected by Paleozoic orogeny. They account for some of these differences in terms of changes in crustal structure; they infer that major differences in velocity extend to depths of about 200 km and can be correlated with surficial geologic structures. Taylor et al. [1980] find crustal structure differences and slight differences in upper mantle velocity between the two provinces.

We also find that travel time residuals differ by 0.5 to 1.0 s between the Grenville province and portions of New England. Major

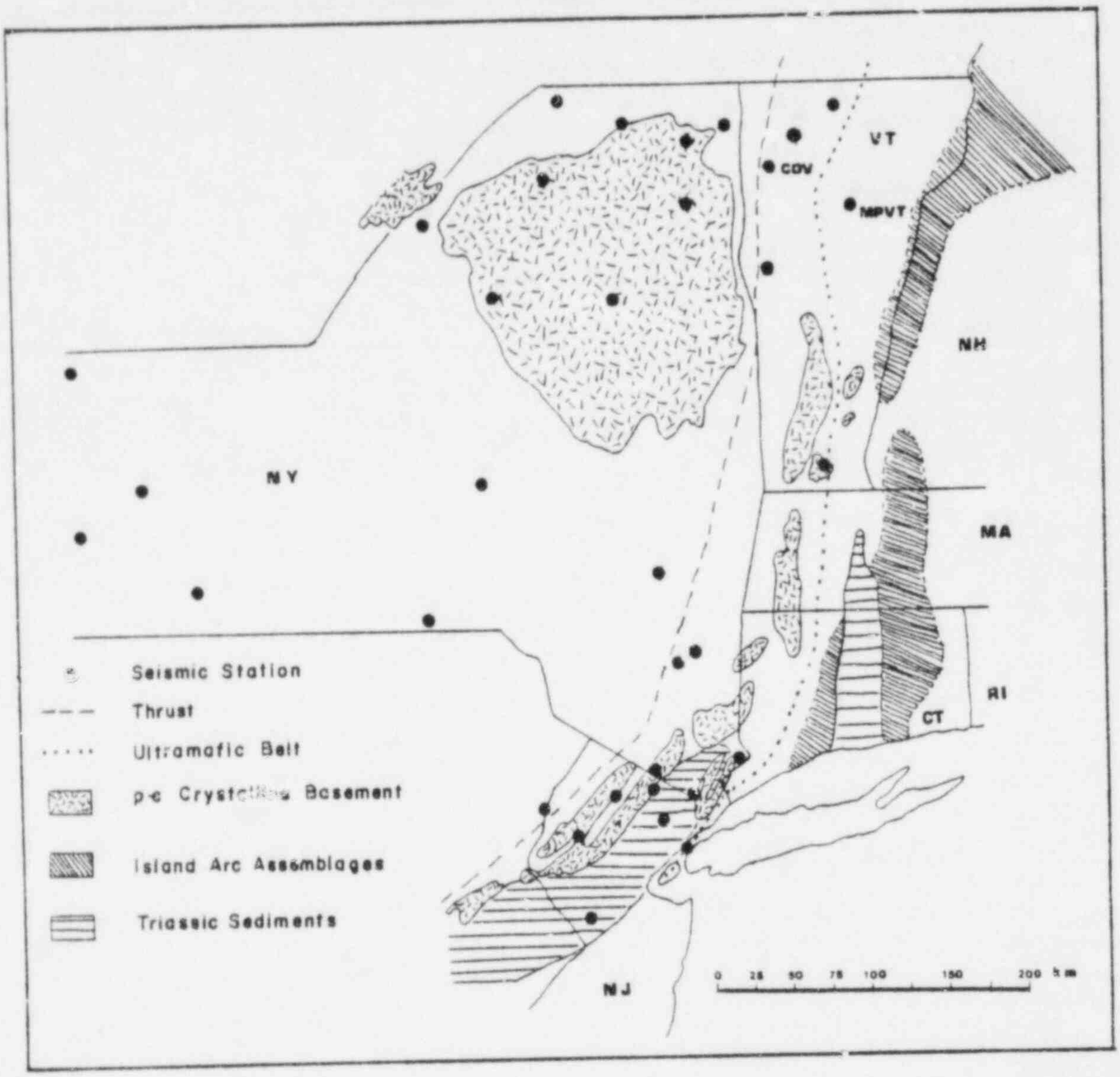

Fig. B1. Seismic stations of Lamont-Doherty network and some major geological features of the Mid-Atlantic states and New England. Outcrops of Grenville-age (1.1 by) basement are shown as crystalline rock. Dashed line represents western limit of Paleozoic thrust faulting. Dotted line denotes belt of ultramafic rocks of Ordovician age [Osberg, 1978]. These latter two features are thought to be associated with the closing of the proto-Atlantic Ocean.

differences in residual, however, are also found between northern and southern New England, both of which were affected by Paleozoic deformation. In fact, the Northeast can be divided into five sub-regions in each of which the residuals are nearly the same. Since transitions between these sub-regions occur rapidly over short distances, i.e., within 50 to 100 km, we argue that major changes in velocity are largely confined to the upper 50-100 and 100-200 km of the earth, respectively. In several places these transition zones do not coincide with mapped boundaries between major geologic provinces. Also, lines of constant residual are rotated with respect to the directions of major geologic elements at the surface in parts of New England. Hence, the structural grain of the lower crust and uppermost mantle in those areas may differ from that of the upper crust, suggesting that the latter is allochthonous in large regions.

Data and Analysis

The Lamont-Doherty seismic network (Figure B1), part of the larger northeastern United States network, consists of short period seismic stations located throughout New York, northern New Jersey and Vermont (Figure B1). Between 1977 and 1979 the network included 36 stations operating on a common time base and has since been expanded to 38 stations. Signals from these stations are telemetered to Lamont-Doherty where all of them are recorded on magnetic tape, 28 on develocorder film and seven are displayed on helicorder records.

P-wave residuals were calculated for selected earthquakes and underground nuclear explosions recorded by the network. From 1977 to 1979, 42 events were chosen for their impulsive first arrivals. To obtain an azimuthally well-distributed data set, however, this criterion was relaxed for events from azimuths with poor data coverage. All of the events chosen for analysis were above magnitude 5.5 and in the distance range 35° to 95°. In addition, all events were well recorded by most of the network. At a particular station, however, the number of events is a subset of the original 42 because of station down-time.

Arrival times of the first prominent peak or trough of P-waves were picked from develocorder film to 0.05 s. Residuals, R_{ij}, for the i th station and j th event were calculated with respect to the Jeffreys-Bullen travel time tables where

$$R_{ij} = T_{ij}^{(obs)} - T_{ij}^{(JB)}. \tag{1}$$

T_{ij} is the observed arrival time at the i th station of the j th event and T_{ij} is the arrival time predicted by the tables.

To find travel time differences between stations in the network, relative residuals, R_{ij}, were calculated for each station with respect to the average residual

$$\overline{R}_{ij} = R_{ij} - \frac{1}{n} \sum_{i=1}^{n} R_{ij} \qquad (2)$$

where n is the number of stations recording event j so $\frac{1}{n} \sum R_{ij}$ is the average residual for event j. Calculating relative residuals in this way eliminates effects of the source and travel path common to all rays entering the array. Thus, mislocations of the event in time and space should not significantly affect the relative residuals. Also eliminated is the effect of choosing the first prominent peak or trough rather than the first motion. Positive (negative) relative residuals indicate that the rays arrived later (earlier) than average.

To see how relative travel times vary across the network, average station residuals, r_i, were calculated by averaging the relative residuals of all the events recorded at each station i

$$r_i = \frac{1}{m} \sum_{j=1}^{m} \overline{R}_{ij} \qquad (3)$$

where m is the number of events. The station residuals, r_i, are accurate to ±0.2 s which includes some systematic error and will be discussed further in a later section. We neglected elevation corrections which are smaller than 0.1 s and corrections for difference in sedimentary thickness, which do not appear to be appreciable unless thick sequences of low velocity materials are present at depth.

Station Residuals

Station residuals for the Lamont network are shown in Figure B2; these data are combined with those of Taylor and Toksöz [1979] for other stations of the Northeastern seismic network in Figure B3. To bring results from the two studies into agreement 0.1 s was added to their data. The size of the correction was found by inspection. This removes the effect of calculating the residuals relative to different averages. When that was done, the average residuals at all but one of

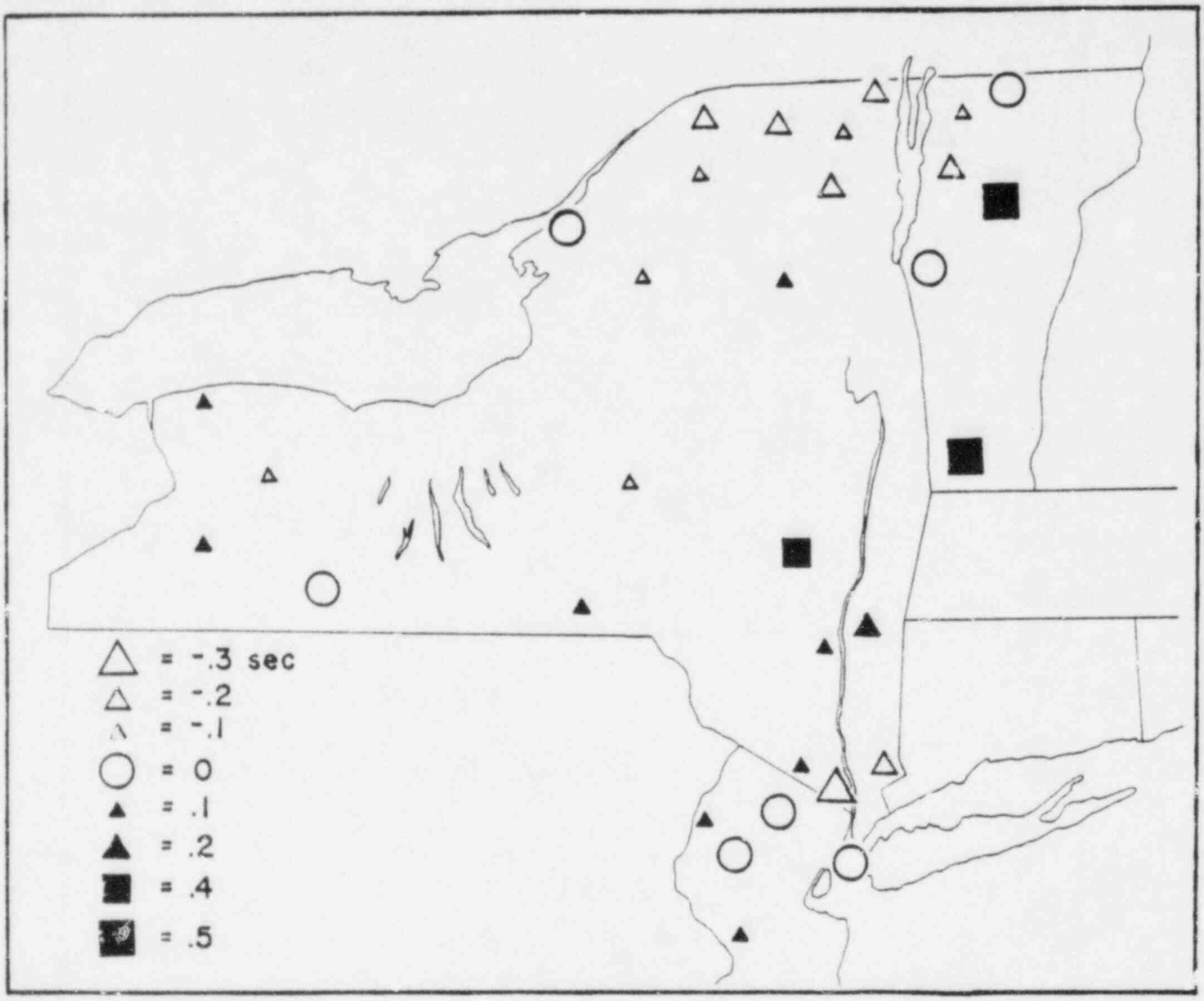

Fig. B2. Travel time residuals for P-waves at stations of Lamont-Doherty seismic network. Data shown represent average of relative residuals of all events recorded at a station. Open symbols are negative residuals (early arrivals) and solid symbols are positive (late arrivals). The size and shape of symbol indicate residual size. Zero residuals are shown as open circles. Residuals are accurate to ±0.2 s.

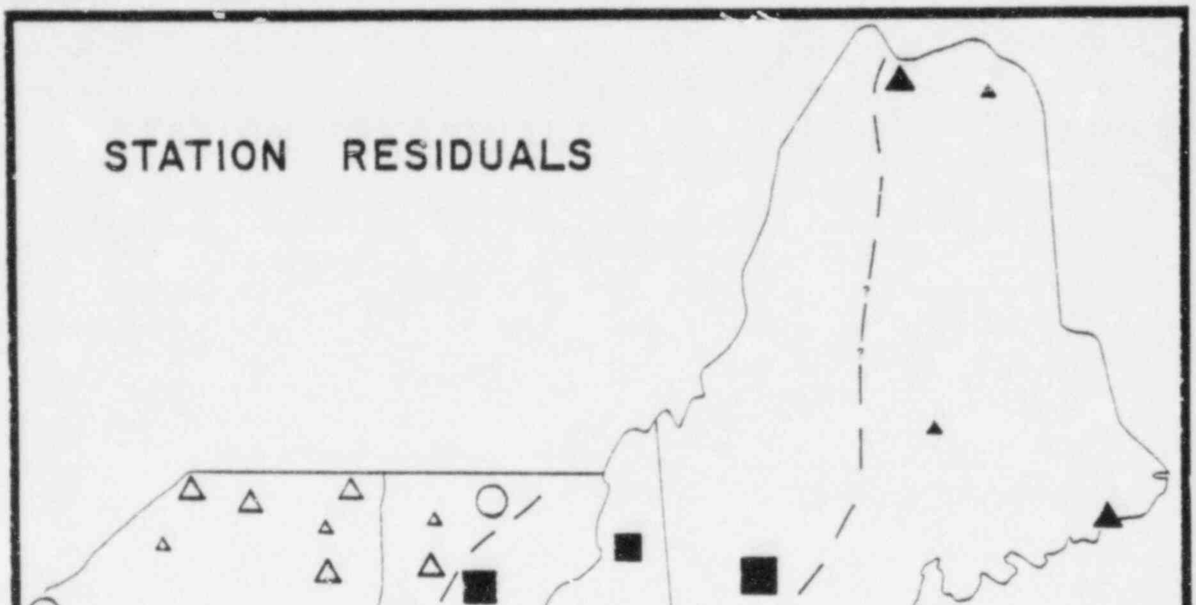

Fig. B3. Travel time residuals for P-waves at stations of northeastern seismic network. Residuals from this study are combined with those of Taylor and Toksöz [1979]. Several sub-regions are defined (shown by dashed lines) in which residuals are constant to within 0.15 s. Note the abrupt change in residuals between regions. The largest change (0.6 s) occurs in 55 km between stations COV in Region I and MPVT in Region II.

the stations that were common to the two studies agreed to within 0.1 s.

An interesting pattern of station residuals is seen in Figure 3 in that residuals at many stations within a given sub-area are equal to within 0.15 s. Between sub-regions, however, the residuals exhibit large differences. This near constancy within sub-regions is surprising since station residuals vary up to 1.0 s in the Northeast as they do in many other networks of similar size. Since residuals at large numbers of nearby stations are nearly identical, it is unlikely that the pattern in Figure 3 results from random occurrence.

We divided the area of Figures 2 and 3 into five sub-regions solely on the basis of the near constancy of residuals. Regions I through IV are shown in Figure 3. The small positive residuals in northern Maine may define either a separate region or an extension of Region III. The data are too sparse to make this distinction. Region V includes the zone of approximately zero (±0.1 s) residuals in central and western New York (Figure 2). Regions I through IV differ from V in that their average residuals are not zero (Table 1). Within each region the standard deviation from the average residual is about 0.1 s.

Between these sub-regions the residuals change significantly. The largest such change (0.63 s) occurs in Vermont over a distance of 55 km between station COV in Region I and MPVT in II (Figure 1). A 0.3 s difference in residuals exists between Regions II and III as well as between III and IV. Stations near the boundary between II and III are separated by at least 100 km so the transition zone is not well defined. Nevertheless, this places an upper limit on the width of the transition zone. The transition zone between Regions III and IV is located in western Connecticut where residuals gradually change in about 100 km from +0.3 s to -0.2 s. That transitions between regions occur over short distances argues that the sources of these differences are located at relatively shallow depths, i.e., approximately the upper 50 to 100 km between regions I and II and the upper 100-200 km in the other cases.

Figure 4 schematically helps illustrate this point. On the Earth's surface are two stations (COV and MPVT) in different residual regions. Four rays (a,b,c, d) are shown entering the stations. The depth (z) of the cross-over point can be determined from the separation distance (x) between the stations and the angles of incidence (i) of rays as

$$z = \frac{x}{\tan i_{COV} + \tan i_{MPVT}} .$$

Velocity differences causing the difference in residuals must lie above the cross-over point when the following four conditions are satisfied:

1) the residual difference between rays (a) and (d) equals the average residual difference between the two regions,

Table 1. Average Residuals for Various Regions in the Northeast.

Region	Average (sec)
I - Northern New York and northwest Vermont	-0.1
II - Central New York, New Hampshire into Maine	+0.5
III - Massachusetts, Rhode Island, Connecticut	+0.2
IV - Southern New York, New Jersey and adjacent Connecticut	-0.1
V - Central and western New York	0.0

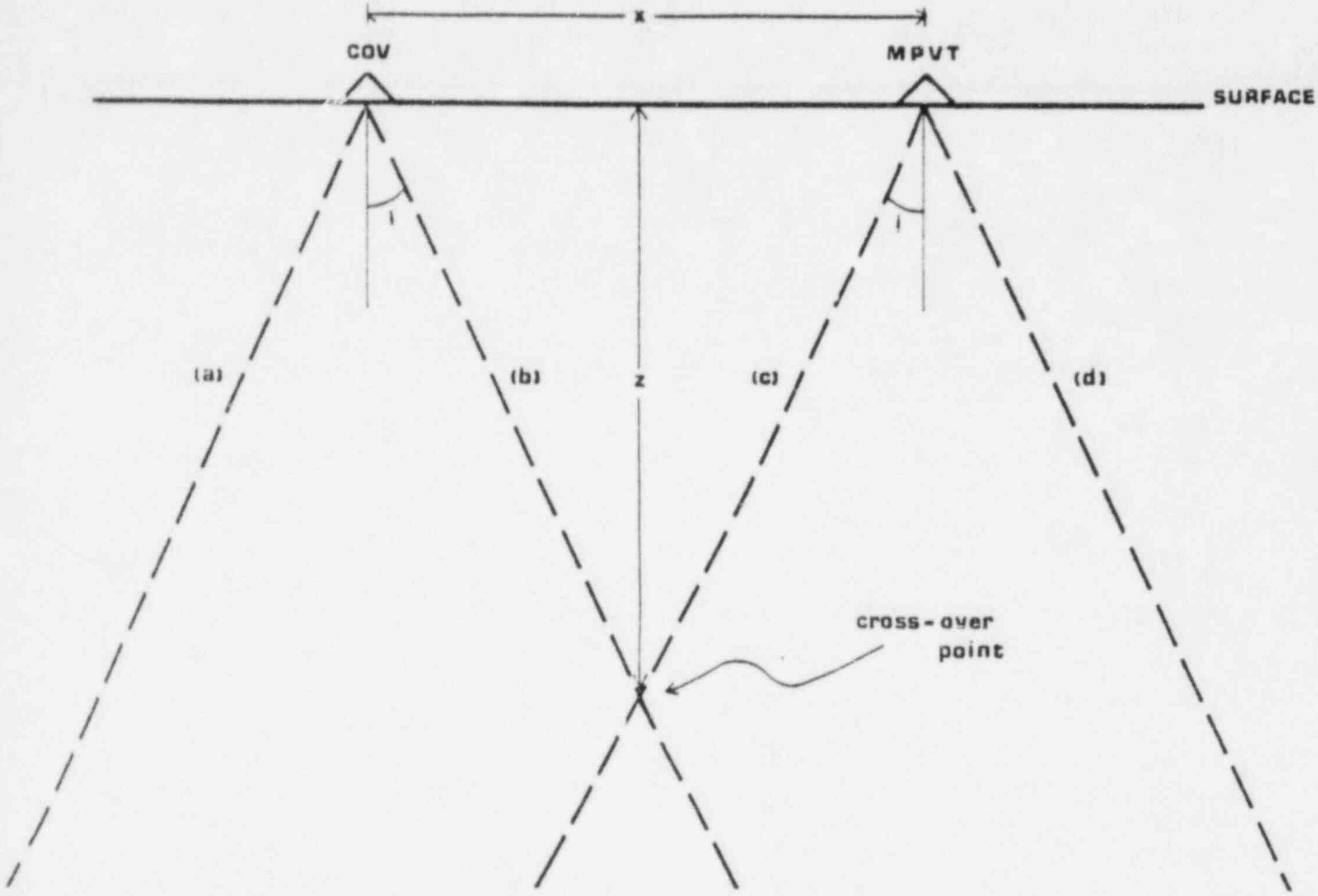

Fig. B4. Teleseismic rays (a), (b), (c), (d) incident at angles i at two stations separated by distance x in different residual regions. Conditions on the differences in arrival time of these rays determine whether velocity differences causing the residuals lie above the cross-over point.

2) the residual difference between rays (b) and (c) equals the residual difference between rays (a) and (d) (i.e., equals the average residual difference between the two regions),
3) the residual difference between rays (b) and (d) equals the average difference between residuals in the two regions for rays incident from that direction, and
4) same as (3) but for rays (a) and (c).

Residuals at stations COV and MPVT in regions I and II satisfy these conditions. Rather than comparing individual rays, however, averages were calculated over all the rays entering the stations from the directions (a), (b), (c) and (d). Given the relative positions of COV and MPVT (MPVT lies at 113° from north with respect to COV) the residuals for the two stations from events in the quadrant with back azimuths of 270°-360° were averaged to determine residuals for rays (a) and (c), respectively. Residuals from events in the quadrant 90°-180° were averaged to find residuals for (b) and (d).

Averages corresponding to residuals for rays (a) and (d) have a difference of 0.7 sec which is very close to the average difference between regions I and II (0.6 sec). Averages corresponding to residuals for rays (b) and (c) differ by 0.6 sec. Thus, conditions (1) and (2) are satisfied to within 0.1 sec. Similarly, conditions (3) and (4) are shown to be satisfied to within 0.1 sec. The residual difference between (a) and (c) ((b) and (d)) is 0.9 sec (0.5 sec) which compares with 0.9 sec (0.4 sec) as the value of the average residual difference between the two regions for events from that quadrant.

Since all the conditions are satisfied to within the accuracy of the residuals, the lateral velocity change that causes the residual difference between regions I and II lies above the cross-over point. Angles of incidence in this study range from 15° to 30°. This establishes a ange of depths (z) of the cross-over point: the maximum for $i = 15°$ and the minimum for $i = 30°$. For stations COV and MPVT, $x = 55$ km so $48 \text{ km} \leq z \leq 103$ km.

For the other transition zones this data set is not complete enough to determine whether the four conditions are satisfied. Because of the large separation distance between stations in regions II and III, the depth cannot be constrained to better than the upper 100 to 200 km. The transition between regions III and IV gradually occurs (Figure 3) over about 100 km, indicating that the differences between these two regions extend to depths of 100-200 km. The lateral velocity differences between regions III and IV are likely deeper than those between I and II.

The size of residual differences and the transition zone width between regions III and IV can be explained by small upper mantle velocity differences (about 2% which is similar to the differences found by Taylor et al. (1980)). However, the explanation of the residual differences between regions I and II which are the largest in size and occur over the shortest distance is not so clear.

For typical crustal and upper mantle velocities, a 20% velocity contrast over a travel path of 20 km can cause a residual difference of 0.6 s. Velocity contrasts of this size are not uncommon in the crust where P-wave velocities can range from 5.0 to 7.0 km/s. Consistent with confining much of the residual differences between regions I and II to the crust is the rapid transition between these regions which indicate that depths of lateral variation are no greater than about 50-100 km. This observation suggests that even with observed differences in the depth to the M-discontinuity much of the velocity contrast between Regions I and II must be accounted for in the crust unless either differences extend far deeper than 50-100 km or sub-M-discontinuity velocity differences are extreme.

This conclusion seems straightforward, however, it disagrees with results from refraction data in the area and thus with the results of Taylor and Toksöz [1979] and Taylor et al. [1980]. Taylor and Toksöz [1979] claim that velocity differences in the entire region that can be correlated with surficial geology extend to depths of 200 km or more. In the refraction study of Taylor et al. [1980] crustal models for the Grenville province in New York State and the Paleozoic New England Appalachians were obtained by inverting travel times from local earthquakes and quarry blasts. Resulting velocity profiles account for very little (less than 0.1 s) of the residual differences. If these profiles accurately describe the vertical velocity structure of the crust in this region, the residual difference between regions I and II would have to be accounted for mainly by differences in velocity structure below the M-discontinuity.

P_n velocities in the entire region are normal, i.e., about 8.1 km/s [Taylor et al., 1980; W. Menke, personal communication] indicating no large, dramatic differences in velocity at the top of the mantle. For example, if the velocity of the upper mantle is 8.1 km/s in Region II and 8.3 km/s in Region I, these differences would have to extend to depths of about 240 km to cause the 0.6 s difference in residuals between the two regions. This depth is clearly inconsistent with the rapidity of the change between Regions I and II.

Problems exist, however, in comparing residuals determined from teleseisms with refraction data insofar as teleseismic rays sample the crust nearly vertically (15° to 30°) while rays from local events sample the crust more nearly horizontally. Horizontal rays average the velocity structure over large lateral distances in the crust and upper mantle whereas near-vertical rays sample a very small horizontal area. Some of the few refraction profiles in the area cut across and therefore average over the residual regions shown in Figure 3. Thus, until much denser refraction and reflection data become available for the region teleseismic residual studies probably will continue to give a better indication of the rapidity of lateral velocity changes and of the depths to which these variations persist. Also, if P velocity in the uppermost mantle is strongly anisotropic, a comparison of rays travelling horizontally and those travelling nearly vertically is complicated further.

Azimuthal Variation

Fletcher et al. [1978] observed that relative residuals at some stations in the northeastern United States and in adjacent parts of Canada vary rapidly with back azimuth, i.e., direction to the source. Taylor and Toksöz [1979] noted that a similar variation exists at most stations of the Northeastern network as did Raikes [1980] for stations of the southern California network. To show this variation with back azimuth at stations in the Lamont network relative residuals at each station are averaged by quadrant (Figure 5B). The data are averaged

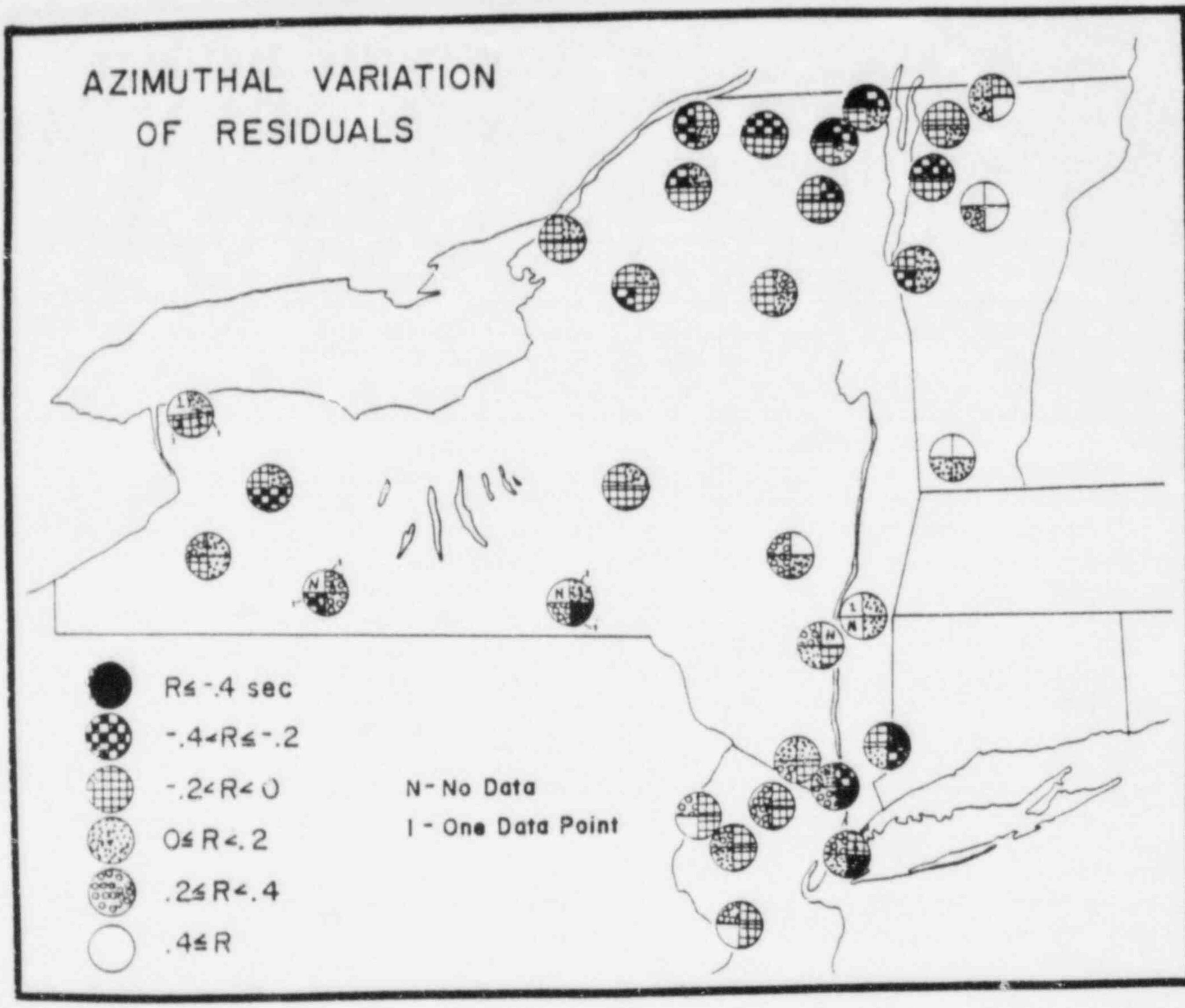

Fig. B5. Variation in relative residuals as a function of back azimuth. Relative residuals are averaged by quadrants. Symbols range from light (positive residuals) to dark (negative residuals). Largest variation at a single station between two quadrants ranges from 0.18 to 0.81 s and averages 0.45 s. A simple pattern of variations is not evident but nearby quadrants tend to be alike.

in this way since the sampling in azimuth was not sufficient to find the Fourier components of the azimuthal variation.

The largest residual difference between quadrants for a single station ranges from 0.15 to 0.81 s and averages 0.45 s. This variation contributes to the calculated uncertainty in the station residuals. A pattern of azimuthal variation is not easily discernible but similar quadrants at nearby stations often have similar residuals. For example, arrivals at the seven stations in northern New Jersey and southern New York are consistently early for easterly and southeasterly azimuths and late for southwesterly azimuths whereas arrivals from the south and southwest are early relative to other azimuths at the four stations in western New York.

Azimuthal variation is often intepreted as indictive of either dipping layer(s) or lateral heterogeneity in the crust and upper mantle. Dipping layers produce a regular pattern of azimuthal variation [Nuttli and Bolt, 1969] which are not observed here. Hence, lateral heterogeneity in the crust and upper mantle is the most likely cause of the azimuthal variation in the northeastern United States.

Rays from teleseismic distances arriving at one station from opposite directions are separated by less than 50 km at the base of the crust and by about 115 km at a depth of 100 km. Since travel time differences of up to 0.8 s are associated with rays arriving at some stations, lateral variations in velocity associated with them must be quite large and of short wavelength. This type of lateral heterogeneity seems to be different from the broad regional differences depicted in Figure B3. Hence, it seems that at least two different and probably independent scales of lateral heterogeneity coexist in the crust an upper mantle: small-scale lateral variations that cause the observed azimuthal variation are superimposed upon large-scale lateral variations that are reflected in the pattern of average residuals. Since azimuthal variations are typically large, it is interesting that residuals averaged over all azimuths (Figures B2 and B3) are nearly identical at large numbers of nearby stations. This is not too surprising since azimuthal variations tend to be similar at nearby stations.

Since the station spacing is sparse for much of the Northeast, we decided to obtain averages over azimuth at each station in our discussion of the relationship between residuals and tectonic structure rather than to use an inversion scheme such as that of Aki et al. [1977], which relies upon arrivals from differing azimuths. For the present distribution of stations that method cannot resolve velocity structure finer than about 150 km in depth [Taylor and Toksöz, 1979] for most of the Northeast. We were also concerned that rapid variations of arrival times in azimuth that result from velocity changes of short wavelength may lead to aliased estimates of larger wavelength variations when information from individual rays is used directly in the inversion of velocity structure. Note that large azimuthal variations are found (Figure B4) not only near the boundaries of the five sub-regions of Figures B2 and B3 but also within the interiors of those sub-regions. This indicates that the azimuthal variations are not a product of the transitions alone.

Correlation with Tectonic Setting

The belt of ultramafic rocks shown in Figure B1 is considered to be the approximate location of a suture zone associated with the closing of an ocean basin or a back-arc basin during the Ordovician [Osberg, 1978]. Basement of Grenville age (1100 Ma) is found to the west of the suture. East of the suture lies what is thought to be an island arc that collided with the rest of North America during the Ordovician. The island arc terrane is separated on the east from the Avalon zone by the Lake Char and Clinton-Newbury fault zones [Osberg, 1978]. All of the residuals at stations east of the ultramafic belt are positive while most station residuals to the west are near zero or negative. This change in residual from west to east must represent a profound difference in the velocity structure between New England and most of the area to the west.

While this change agrees in a gross way with changes in regional geology, there are many ways in which the pattern of residuals does not match the regional geologic pattern. The boundaries of sub-regions in Figure B3 (or lines of constant residual) generally strike northeasterly in New England and eastern New York (and perhaps in northern New Jersey). Thus, they tend to follow the overall strike of the Appalachian orogen rather than the local strike, which is northerly in southern and western New England between New York City and Montreal.

A series of negative and postive gravity anomalies (both free air and Bouguer), which are among some of the largest anomalies in eastern North America, strike northeasterly in eastern Pennsylvania and southeastern Quebec and northerly along the eastern border of New York [Simpson et al., 1981]. These anomalies generally follow the edge of the Cambrian continental margin of North America [Rodgers, 1970] and are situated between the suture zone and western limits of Paleozoic thrust faulting (Figure B1). Maps of Bouguer anomalies for the northeastern United States and adjacent parts of Canada filtered to pass wavelengths shorter than 100 and 250 km [Simpson et al., 1981] indicate that the sources of these prominent anomalies must be located at depths shallower than 80 km and that at least some of the sources must be shallower than 33 km. These short wavelength gravity anomalies tend to follow the local strike of the geology better than the average residuals do.

Three stations near the New York-Massachusetts border, which are situated on basement of Grenville age, have residuals that are 0.3 to 0.6 s more positive than those of other stations that are located to the west of the suture zone shown in Figure B1. Their residuals are like those of stations in New Hampshire and most of Vermont, and hence were assigned to the same sub-region (II) in Figure B3. While those stations are situated to the west of the suture, they are still located near the Cambrian continental margin of North America [Rodgers, 1970]. Those positive residuals must, of course, reflect anomalous velocities at depth when they are compared to residuals for stations on crust of Grenville age. The residuals are not sufficiently dense in number, however, to place useful limits on the depths of the causative anomalies.

One explanation could be that the upper mantle and perhaps the lower crust beneath those stations are similar to that beneath New Hampshire and parts of Vermont and were thrust beneath that area during the Paleozoic. Another possibility is that a thick sedimentary section of low velocities may be present beneath the Cambro-Devonian sequence of rocks that are exposed at the surface near those stations. It is possible that a thick section of late Precambrian sediments could have been deposited along that portion of the former continental margin that existed from the Precambrian through Cambrian time. The anomalies in travel time are large enough that sedimentary thicknesses of 2 to 7 km are required to produce them. Explaining the three anomalous residuals as a result of a thick sedimentary sequence requires, however, that the near equality of the three residuals and those in other parts of sub-region II be attributed to coincidence.

While residuals at stations in southern New England are 0.2 to 0.3 s more positive than those of sub-regions I, IV and V (Figures B2 and B3), they are still about 0.3 s smaller than the residuals at stations in New Hampshire and most of Vermont. Hence, sizeable differences in velocity must be present at depth beneath northern and southern New England as well as between them and the area to the west. The basement in at least part of sub-region III is of latest Precambrian (Avalonian) age [Rodgers, 1979]. Parts of southern New England are known to have been affected by late Paleozoic orogeny whereas New Hampshire and Vermont were not.

The way that lines of constant travel-time residuals cut across the suture zone and other major geologic boundaries in Figure B1 is reminiscent of similar patterns in southern California that is described by Hadley and Kanamori [1977] and Raikes [1980]. They, like we, find that velocity variations do correlate with general tectonic provinces. Nevertheless they find the Transverse Ranges of southern California are underlain at a depth of about 40 km by a refractor with a P-velocity of 8.3 km/s. Early P arrivals from teleseisms indicate that this anomaly extends to a depth of 100 km. This high-velocity, ridge-like structure is coincident with much of the areal extent of the geomorphic Transverse Ranges and is not offset by the San Andreas fault. They suggest that the plate boundary at depth is displaced from the location of the San Andreas at the surface. Hence a zone of decoupling in the lower crust or uppermost mantle is necessary to accommodate the horizontal shear that must result from the divergence of the plate boundary at different depths.

One interpretation of the divergence between lines of constant residual and major tectonic boundaries in Figure B1 is that the deep structure sampled by P-waves at nearly vertical incidence also differs from that near the surface. In this view the suture zone of Figure B1 and large parts of New England would be allochthonous. Brown et al. [1981] reach a similar conclusion using COCORP data along a profile across Vermont.

Since the Cambrian continential margin of North America did not have the same strike throughout the Mid-Atlantic states, New England and southern Quebec, the island arc that is inferred to have collided with it during the Ordovician must have interacted obliquely with at least some parts of that passive margin. Hence, the suturing of the arc onto North America is likely to have resulted in the rotation of individual terranes within the arc both with respect to one another and with respect to North America. The upper mantle and perhaps the lower crust of the arc may not have deformed in the same manner as more surficial terranes. Obviously, more studies of the distribution of seismic velocities are needed to unravel the relative movements of various units of the upper mantle and lower crust relative to one another and relative to the upper crust.

Conclusion

Travel-time residuals for P-waves from distant sources vary by up to 1.0 s throughout the Mid-Atlantic states and New England. Average station residuals at many nearby stations, however, are identical to within ±0.15 s. Five sub-regions are defined based upon the near constancy of the residuals at stations in their interiors. These sub-regions are bounded by transition zones only 50 to 100 km wide over which rapid changes in residuals occur.

One transition zone is clearly about 100 km wide suggesting lateral variations in velocity causing these residual differences might extend to depths of 100-200 km. The transition width along with the size of the residual difference can be explained by small (2%) upper mantle velocity differences. The transition between regions with the largest residual change occurs within 55 km indicating that the lateral velocity changes between these regions only extend to depths of 50-100 km. Such shallow sources, however, appear to conflict with existing refraction data which do not indicate obvious or major changes in crustal thickness or P_n velocity. The latter only provide averages over large regions, however, whereas rays from distant sources arriving at a given station from various azimuths sample a much smaller horizontal area. Hence, we suspect that more detailed investigations in this particular area, such as multichannel reflection profiling, will detect major lateral changes in velocity structure within either the crust or the uppermost 60 km of the mantle. Preliminary results from a COCORP line across Vermont [Brown et al., 1981] suggest that this is the case.

Variations in P velocity in the mantle portion of the lithosphere from 7.8 to 8.7 km/s have been reported from long refraction profiles that were designed to study the uppermost mantle [Hadley and Kanamori, 1977; Nagumo et al., 1981] in tectonically active regions. With this size velocity difference it would be possible to account for all of the variations in P residuals by changes in the uppermost 45 km of the mantle. Variations at those depths would be compatible with the rapid transitions in residuals between sub-regions of Figures B2 and B3. It seems more reasonable to us, however, to place some of the contrast in the crust and perhaps a small amount at depths greater than 100 km. Obviously, more data from residual and other studies are needed to better resolve changes in velocity as a function of depth and location.

Briden et al. [1981] report differences in travel-time residual for P-waves of up to 1.0 s along a profile extending from the 250 Ma Mauritanide orogenic belt to the 2000 Ma West African craton. That difference in residual as well as wavelength and polarity of the gravity anomalies across that boundary are similar to those between eastern New York and New Hampshire. Residuals at stations in N

Hampshire and in the Mauritanide belt, which are more positive than those at stations on the adjacent cratons, are still negative compared with those for most stations in the western United States and other regions of Cenozoic tectonism.

While residuals in the northeastern United States show a general correlation with the age of the basement and with major tectonic provinces, lines of constant residual cut across some major tectonic boundaries such as the suture zone of Ordovician age in western New England. Hence, the structure of the uppermost mantle and perhaps the lower crust as inferred from P residuals appears to follow the overall northeasterly strike of the Appalachian orogen rather than the regional strike which is more northerly in southern and western New England. This difference can be reconciled if the upper crust is allochthonous in those parts of New England and if the suture zone is itself allochthonous.

Major differences in velocity structure also must be present between northern and southern New England. It seems clear that lateral heterogeneity in the crust and uppermost mantle as reflected in travel-time residuals and velocity differences probably has persisted in the Northeast for hundreds of millions of years.

We also find large variations in travel time with azimuth at most stations. These variations in azimuth tend to be similar at nearby stations and to be present both within as well as near the boundaries of regions in which station residuals are nearly constant. Hence, they are not a property of the transition regions per se. They could result from a major anisotropy in velocity in the uppermost mantle. Nevertheless, since the ray paths we studied were inclined no more than 30° from the vertical, our experiment was not well suited to measure anisotropy in the mantle.

Acknowledgments. This research was partially supported by the Advanced Research Projects Agency of the Department of Defense and was monitored by the Air Force Office of Scientific Research under contract number F49620-C-79-0021, by the National Science Foundation under contract EAR 7822770 and by the U.S. Geological Survey under contract USGS-14-08-0001-19750. We thank Drs. Alan Kafka and David Simpson for their critical reviews. Lamont-Doherty Geological Observatory Contribution Number 0000.

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APPENDIX C

EARTHQUAKE MAGNITUDES AND SEISMICITY IN THE NEW YORK CITY METROPOLITAN AREA

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PREPRINT
To be Submitted for publication to BSSA

Seventh Draft - 11 June 1982

ABSTRACT

Several magnitude scales are analyzed to determine an estimate of the size of earthquakes recorded by microearthquake networks in the New York City metropolitan area. Signal duration is found to be a very useful measure of the size of earthquakes in this region. Seismic waves recorded on instruments with peak response near 1 Hz are used to compare duration measured from the higher frequency (≈10 Hz) data recorded by the local networks with m_b and M_L. During the period of operation of dense microearthquake networks in the New York City region (1970-present) the largest earthquake in the study area occurred near Cheesequake, New Jersey on January 30, 1979. The magnitude of this event (m_b) is estimated to be 3.0±0.2. A map of the distribution of the larger events in this study suggests that earthquakes in the New York City metropolitan area are primarily concentrated in geologic structures that surround the Newark (Triassic-Jurassic) basin. We find no evidence from the record of microearthquakes recorded from 1970 to 1981 to suggest that NE trending faults that lie to the northwest of the Newark basin (such as the Ramapo fault) are any more active than faults that lie to the north and east of the basin. Only about half of the earthquakes above magnitude 2.0 occurred along the Ramapo fault zone.

INTRODUCTION

During the past 250 years maximum intensities as high as VII on the modified Mercalli scale have been reported for three earthquakes in the New York City metropolitan area (1737, 1884, and 1927). These events having occurred long before the installation of local seismic networks in the region, cannot be located accurately enough to be unambiguously related to specific geologic features. Using seismic data recorded by local seismic networks operated by Lamont-Doherty Geological Observatory (LDGO) and Woodward-Clyde Associates, the Ramapo fault zone (Figure C1) has been documented as a seismiclly active feature, and many earthquakes recorded by the local networks are concentrated along this fault zone (Aggarwal and Sykes, 1978; Yang and Aggarwal, 1981). Several relatively large earthquakes, however, have occurred to the north and east of the Ramapo fault zone during the past decade of microearthquake monitoring, emphasizing that the Ramapo fault zone is not the only seismically active feature in the New York City metropolitan area. Indeed, the largest event (based on maximum intensity, felt area, signal amplitude, and signal duration) recorded by the networks (m_b ≈ 3.0; January 30, 1979) was located to the southeast of the Newark basin (Figures C1, C4 and C6) within (or beneath) the Atlantic coastal plain sediments. In this study we have reevaluated the past decade of seismicity in the New York city metropolitan area, and we conclude that the Ramapo fault zone is one of several zones in the New York City metropolitan area that is seismically active, and that earthquakes are also likely to occur in geologic structures that lie to the north and east of the Newark basin.

Studies of the spatial and temporal distribution of seismicity in a given region often depend upon a scale for comparing earthquakes in terms of the energy released at the source. The frequently quoted

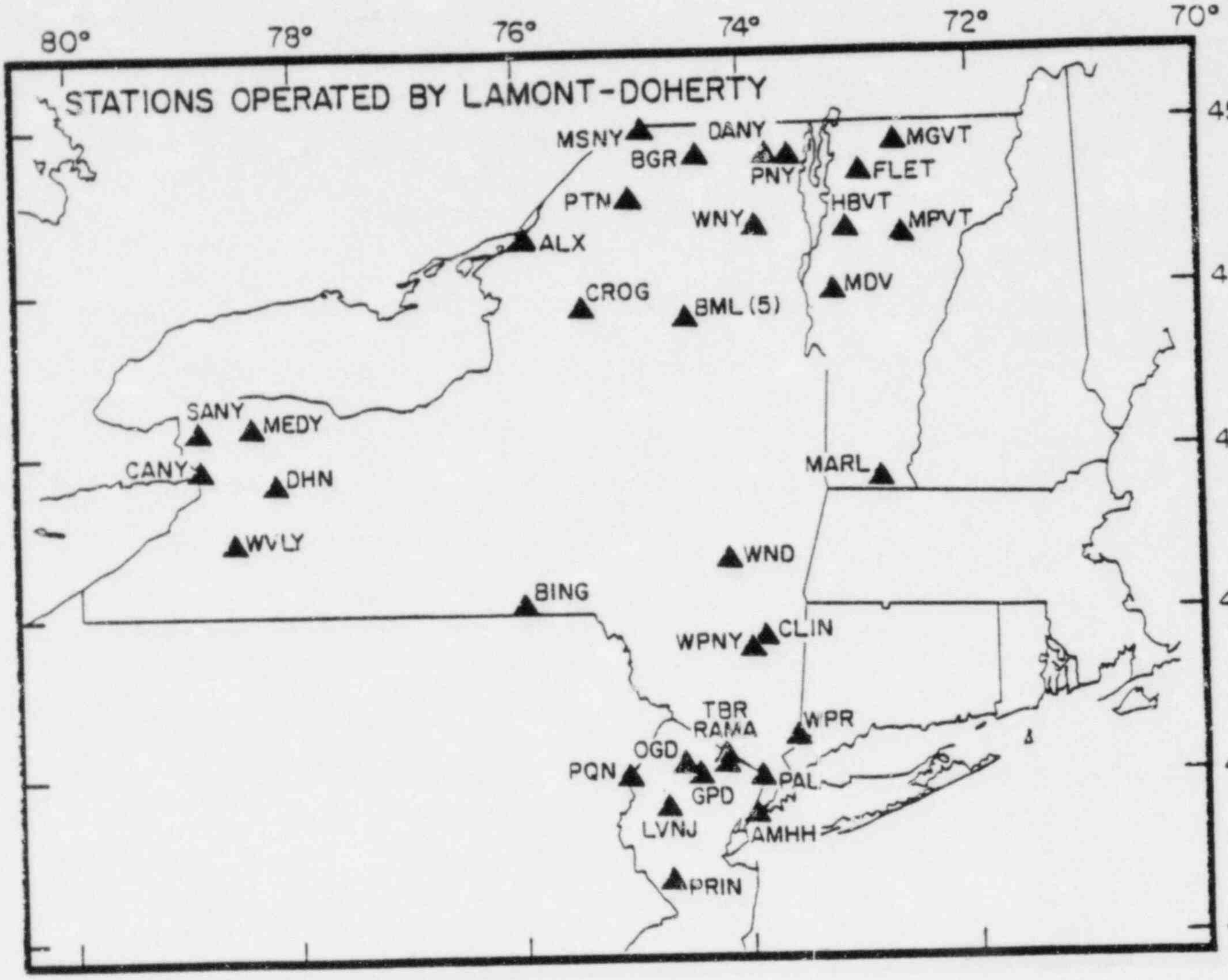

Figure C1: a) Present configuration of short-period seismic stations (triangles) operated by Lamont-Doherty Geological Observatory (LDGO).

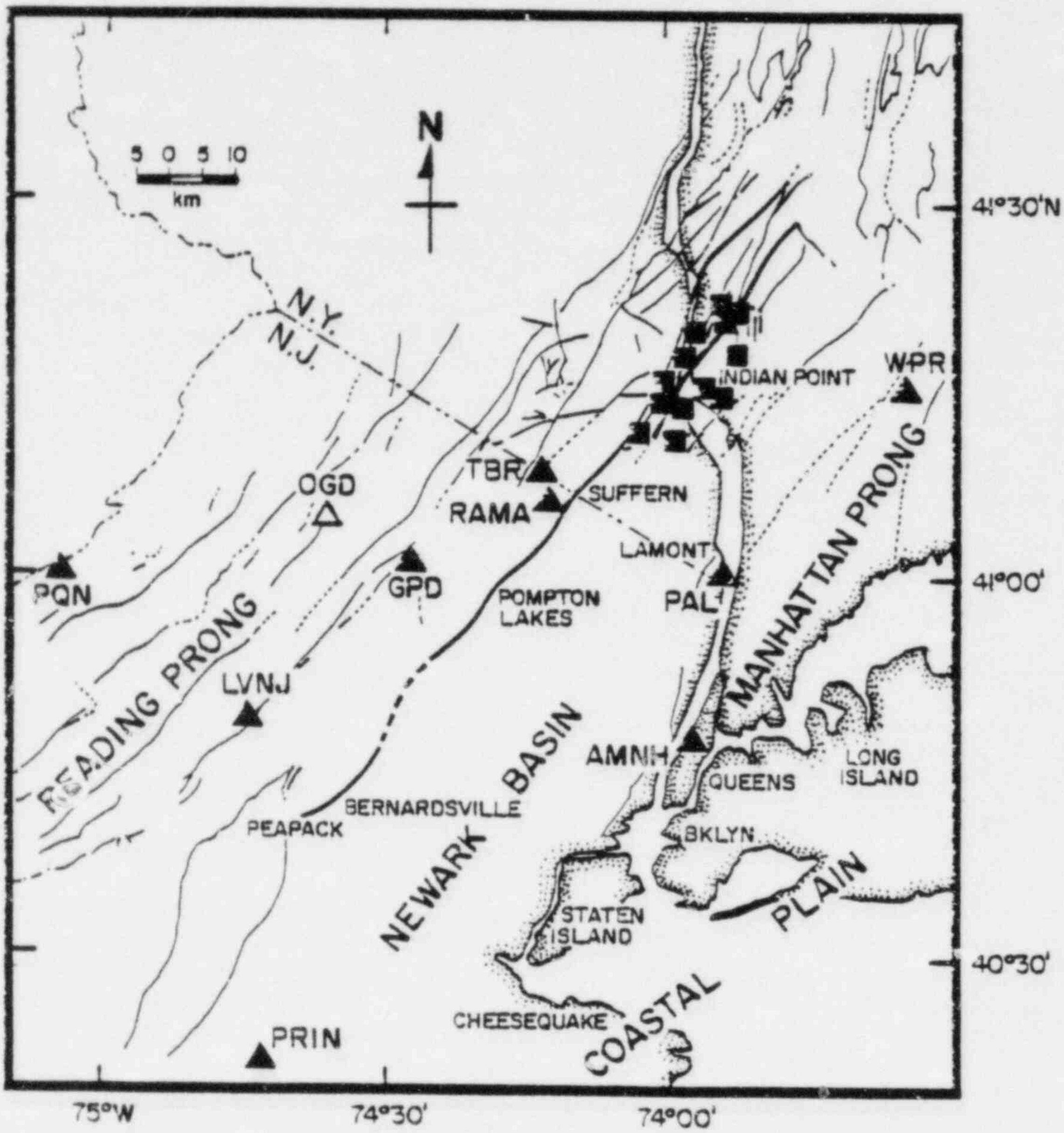

Figure C1: b) Location of short period seismic stations superimposed on a map of geologic provinces, faults and lineations (adapted from Yang and Aggarwal, 1981) in the New York City metropolitan area. Dark solid lines between Peapack, N.J. and Indian Point, N.Y. represent the main trace of the Ramapo fault zone. Seismic stations operated by LDGO are shown as triangles and those operated by Woodward-Clyde are shown as squares. The closed triangles represent short period seismic stations and the open triangle represents the WWSSN station.

magnitude scale (M_L) of Richter (1935) attempts to quantify the notion of energy released at the source of earthquakes by comparing the amplitude of ground motion generated by the events. Other magnitude scales, such as m_b and M_s, have been developed that relate the amplitude of ground motion to the size of earthquakes, and these scales are usually constructed such that they coincide with M_L over some range of magnitudes. In addition to amplitude of ground motion, signal duration has been proposed as a measure of the size of an earthquake (e.g., Brisztricsany, 1958; Lee et al., 1972). In this paper we analyze the applicability of various magnitude scales for earthquakes recorded by local seismic networks in the New York City metropolitan area, and we estimate the magnitudes of earthquakes which occurred in this region during the past decade. This analysis results in a new catalog of instrumentally located earthquakes with well defined magnitudes. A cut-off threshold for magnitude is determined below which events detected should not be entered into the regional catalog since they introduce a bias in the distribution of earthquakes resulting from an uneven distribution of stations. This new catalog of earthquakes in the New York City region suggests that seismic activity is primarily concentrated along geologic structures that surround the Newark (Triassic-Jurassic) basin.

The Nuttli (1973) m_{bLg} scale was developed to estimate m_b from L_g waves (near 1 sec period), and this scale has been used to determine magnitudes from short period seismic data recorded by local networks in the northeastern United States (Chiburis et al., 1976-1981). The m_{bLg} scale is not necessarily the most appropriate scale to use for the data recorded by these networks for a number of reasons. One problem is that the m_{bLg} scale was developed from observations of much longer period Lg waves (~1 sec) than the Lg waves recorded by the local networks (~.5 to ~.15 sec). In this paper we examine the applicability of the m_{bLg} scale for earthquakes recorded by local networks operating in the New York City metropolitan area using two complementary approaches. One approach is to calculate magnitudes of larger earthquakes ($m_{bLg} > 2$) that were not only recorded by the local network stations but were also recorded on 1 Hz WWSSN stations and on a Wood-Anderson instrument located at Palisades, NY. In these cases mangitudes determined from several types of seismograms using a number of different scales can be compared. The results of this analysis suggest that amplitudes measured from develocorder records of the local network stations can be used to estimate m_{bLg} if the amplitudes are divided by the period of the waves recorded. The other approach is to analyze spectra of earthquakes recorded by the local networks, and to compare the level of excitation of the 5 to 15 Hz Lg waves with that of the 1 Hz Lg waves.

A second problem with using the m_{bLg} scale (or any magnitude scale that depends on signal amplitude) for earthquakes recorded by the local networks is that for earthquakes of magnitude greater than about 3.0 many of the develocorder records are off scale. In an effort to record the smaller earthquakes during the past decade, station gains for the local networks have been set very high (~10^6 to ~10^7) resulting in a lower magnitude threshold for events detected, but with a limited number of unclipped records of the larger events. Thus, magnitudes of the larger events are often very poorly determined.

This problem of records being clipped for the larger events introduces a serious difficulty in determining the relative sizes of events in an earthquake sequence. Specifically, if a sequence of larger (magnitude > 2) and smaller (magnitude < 2) earthquakes occurs, the smaller events are recorded only by the close stations (epicentral distance less than 100 km), while the larger events are clipped at the close stations and only well recorded at the more distant stations. In some instances this mixed sampling of stations would yield nearly the same magnitudes for two earthquakes when the two events intuitively appeared to be very different in size based on other observations (such as the number of stations recording each event, differences in maximum intensity and felt area and relative amplitudes for the 2 events when recorded at the same station).

Magnitudes determined from measurements of signal duration are not as dependent on the problems discussed above as magnitudes determined from amplitudes. A signal duration scale (see Table 1) has been developed for small (m_b < about 4.5) earthquakes in New England (Chaplin et al., 1980). In this study, signal duration and Lg wave amplitudes were measured for earthquakes recorded by microearthquake networks in the New York City metropolitan area and magnitudes determined from signal duration and amplitudes recorded on a number of instruments of different frequency response are compared.

THE DATA BASE

Short period data for this study were recorded by a local seismic network operated by Lamont-Doherty Geological Observatory (LDGO) in the states of New York, New Jersey and Vermont. Figure 1 shows the distribution of short period stations operated by LDGO in this region. These local stations have been operating since about 1970, although the configuration of the network has changed over the past decade. Since 1970 about 100 earthquakes ranging in magnitude from approximately 1.0 to approximately 3.0 have been recorded in the New York City metropolitan area. Only earthquakes that occurred since 1974 have been considered in this study because the station distribution of the Lamont-Doherty network has been similar to that shown in Figure 1 since about that time.

Amplitudes and frequencies of Lg waves, signal duration and amplitude of background noise were measured from develocorder films for earthquakes recorded by the LDGO network between 1974 and 1981. A total of 194 amplitudes, 192 frequency, and 233 signal duration measurements were made from 37 events. Maximum amplitudes were measured for the Lg phase (i.e., the waves arriving at group velocities between 3.5 and 3.0 km/sec) for most cases. Although in some cases one or two inordinately large peaks in the wave train of the Lg phase were ignored and the "average maximum" amplitude was measured. Frequencies were determined for Lg waves by counting the number of peaks in a one second interval. If the develocorder film was not in sharp focus, however, frequencies greater than about 10 Hz were difficult to resolve and are not very reliable observations. Signal duration was measured as the duration of the earthquake signal in seconds from the P arrival time to the point where the signal disappears into the noise. Since signal duration was measured relative to the amplitude and frequency of the background noise level, the instrument and site

TABLE 1

Magnitude Formuals Used in This Study

	Formula	Reference
1	$m_{bLg} = 3.75+0.90 \log\Delta+\log(A/T)$ $0.5^\circ<\Delta<4^\circ$ $m_{bLg} = 3.30+1.66 \log\Delta+\log(A/T)$ $4^\circ<\Delta<30^\circ$ Δ = epicentral distance (km) A = amplitude of Lg wave (microns) T = period (sec)	Nuttli (1973)
2	$M_L = \log A+\log A_o-\delta \log A_o$ A = maximum amplitude on Wood-Anderson seismogram (mm) $\log A_o$ = correction factor for southern California - Richter (1935) $\delta \log A_o$ = additional correction factor for northeastern United States - Ebel (1982)	Ebel (1982)
3	$m_c = 2.21 \log D+1.70$ D = signal duration (sec)	Chaplin et al. (1980)

magnification effects are in part compensated for. Thus, although the station gains range from ~10^6 to ~10^7, the variation in gain does not affect signal duration because station gains have been adjusted such that each station has approximately 20 millivolts of background noise at the field site.

We also obtained amplitude and frequency data from records of longer period instruments for the larger earthquakes that occurred between 1974 and 1981. A Wood-Anderson seismometer located at Palisades, New York was used to determine local magnitudes (M_L) by applying Richter's (1935) formula with the correction for northeastern United States attenuation suggested by Ebel (1982). In addition, Lg waves (~1 Hz) recorded on the WWSSN station OGD (Ogdensburg, New Jersey) were used to determine m_{bLg} magnitudes by applying the formula of Nuttli (1973). These M_L and m_{bLg} magnitudes are compared to measurements of signal duration and to m_{bLg} magnitudes obtained from amplitudes of high frequency (~10 Hz) Lg waves recorded by the local network data.

MAGNITUDES AND SPECTRAL CONTENT OF EARTHQUAKES RECORDED BY THE LOCAL NETWORK

A major portion of this study involved assessing various magnitude scales that have been applied to microearthquake network data in the northeastern United States and adjacent Canada and deciding upon a scale that is most appropriate for data recorded by the microearthquake network operated by Lamont-Doherty in the New York City metropolitan area. Once we decided upon an appropriate magnitude scale a cut-off threshold for magnitude was determined below which events detected should not be entered into the regional catalog since they introduce a bias in the distribution of earthquakes resulting from an uneven distribution of stations. The criteria we used in searching for a magnitude scale were the following:

(1) The magnitudes obtained from a given formula should not be very dependent upon site response, the number of stations recording the event, and/or which stations recorded a given event.

(2) The magnitude scale should, in general, give larger magnitudes for earthquakes that generated greater amplitudes at seismic stations. Although this criterion may appear to be obvious in principle, it can be quite difficult to quantify and can often only be applied qualitatively since for a given earthquake many parameters (e.g., focal mechanisms, depth, and epicenter to station path) can affect the amplitudes of different seismic phases at different azimuths. Nonetheless, an attempt was made in this study to qualitatively evaluate magnitudes determined for different events against this criterion.

(3) On average events with larger magnitudes should be recorded at greater distances, and have a larger felt area and greater intensities than events with smaller magnitudes. As we discuss below this criterion (like criterion number 2 above) can only be applied qualitatively.

(4) The magnitude scale should be made to coincide at least to some extent with global estimates of magnitude (such as m_b) so that studies from one region can be compared to the next.

(5) The magnitude scale chosen should enable us to determine a magnitude threshold above which earthquakes can be detected and located by the network regardless of where the events occurred within the region being studied.

Chaplin et al. (1980) developed a magnitude scale based on signal duration (referred to as m_c below) for New England earthquakes by relating mean signal duration to m_{bLg} as reported in the bulletins of the NEUSSN and the formula given in their study is shown in Table 1. Such signal duration formulas have two advantages over formulas based on signal amplitudes.

(1) Signal duration can be measured for events which are relatively large and have amplitude peaks that are clipped.

(2) Signal duration does not appear to be as dependent on station gain, distance and site response as amplitude measurements are. Hence, one or two records are sufficient to estimate the magnitude; even if the event is poorly located.

In Figure 2 magnitudes determined by several methods are plotted against mean signal duration measured from develocorder records of the LDGO network. Closed triangles are m_{bLg} magnitudes also determined from the develocorder records using formula (1) of Table 1. When applying the Nuttli (1973) formula to these high frequency measurements we divided by the period (T) as suggested by the A/T term. Open circles and open squares are estimates of Richter (1935) magnitudes determined from a Wood-Anderson instrument located at Palisades, New York. These magnitudes were calculated by applying the corrections to the M_L scale suggested for the Northeast by Ebel (1982) as shown in Table 1. In his study magnitudes of northeastern U.S. earthquakes where calculated from Wood-Anderson seismograms recorded at stations operating in the northeast, and Richter's M_L formula was rewritten to account for differences between attenuation for the eastern vs. western United States. Open circles represent events within the study area shown in Figure 6, and open squares represent other events in New York State and adjacent areas that were outside the study area. Closed stars represent m_{bLg} magnitudes determined from a short period vertical component of a number of WWSSN seismograms recorded from several of the larger earthquakes in this study. In the case of these magnitudes the formula given by Nuttli (1973) was applied to the frequency of waves for which it was developed.

The best fit (least-squares) line to the high frequency magnitudes (closed triangles) is shown by the dashed line in Figure C3. The solid line represents the coda-length scale of Chaplin et al. (1980) which was determined from the same type of data for earthquakes in New England. The magntiudes calculated from the Wood-Anderson instruments show a significant amount of scatter, and differences between these magnitudes and those obtained from the network data were as great as 1 magnitude unit for certain events. Such differences might be expected, however, since the Wood-Anderson magnitudes were determined from only one station. We note, however, that the amount of scatter shown here is greater than that observed by Ebel (1981) for events studied in New England.

The m_{bLg} magnitudes determined from short-period WWSSN stations (i.e., using the instrument and frequency band for which the m_{bLg} formula was developed) correlate fairly well with the magnitudes determined from the high frequency network data Figure C2, and Table

Magnitudes of Earthquakes in N.Y. City Region

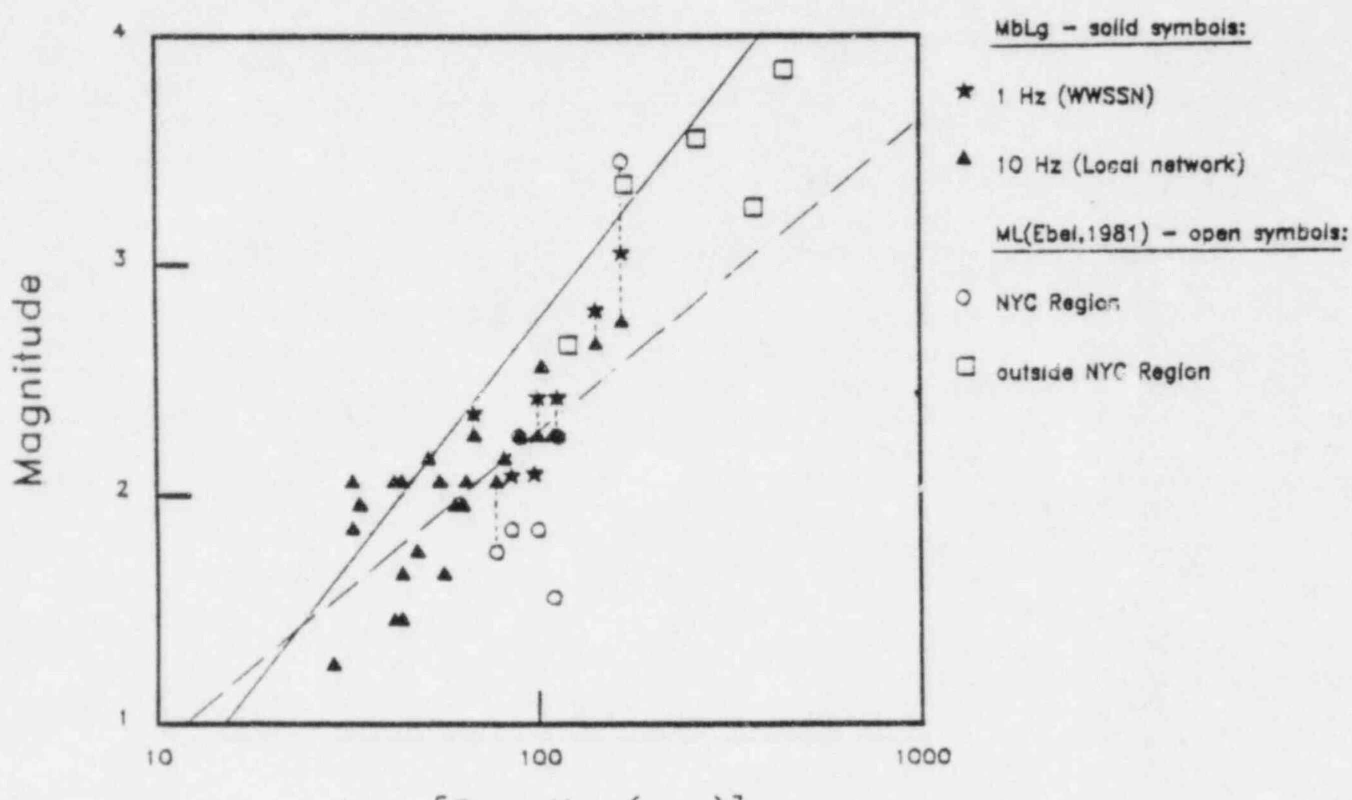

Figure C2: Magnitudes of earthquakes in the New York City metropolitan area and nearby regions as a function of signal duration measured from LDGO develocorder records. Closed triangles are estimates of m_{bLg} magnitudes calculated from the local network stations as described in the text. Open circles and open squares are estimates of Richter (1935) magnitudes determined from a Wood-Anderson seismometer. Open circles represent events within the study area shown in Figure 6, and open squares represent other regional events. Closed triangles represent m_{bLg} magnitudes determined from short-period ($\approx$1 Hz) compontents of WWSSN stations. Formulas used for these magnitude calculations are given in Table 1. The dashed line represents a best fit (least squares) to the high frequency magnitudes and the solid line represents the coda-length scale of Chaplin et al. (1980). Dotted lines connect different magnitude determinations for the same earthquakes.

TABLE 2

LIST OF EARTHQUAKES IN NEW YORK METROPOLITAN AREA: 1974-1981

*SIGNAL DURATION MAGNITUDES $\geq$ 2.0

Date	Location	Latitude	Longitude	Signal Duration Magnitude
4/8/74	Stony Point	41°13.12'	73°59.51'	2.2
6/7/74	Wappingers Falls	41°34.27'	73°56.40'	3.0
2/20/75	West of Sandy Hook	40°20.82'	73°10.62'	2.3
6/15/75	Wappingers Falls	41°34.80'	73°56.63'	2.0
7/19/75	Mahopac	41°25.55'	73°47.36'	2.5
10/24/75	Wappingers Falls	41°35.55'	73°55.99'	2.1
3/11/76	Riverdale	40°57.12'	74°21.19'	2.7
4/13/76	Ridgefield	40°50.10'	74°02.85'	2.6
5/11/76	Off Sandy Hook	40°29.07'	73°47.74'	2.3
8/20/76	Mt. Pleasant	41°06.81'	73°45.22'	2.3
3/10/77	Suffern	41°10.94'	74°08.88'	2.2
7/2/77	Hampton	40°42.22'	74°56.12'	2.1
9/2/77	Peekskill	41°18.78'	73°55.41'	2.2
10/14/77	North of Newburgh	41°33.53'	73°57.18'	2.3
4/3/78	Off Sandy Hook	40°31.80'	74°04.80'	2.4
6/30/78	Oakland	41°04.52'	74°12.10'	2.7
1/30/79	Cheesequake	40°19.29'	74°15.81'	3.2
3/10/79	Bernardsville	40°43.34'	74°30.25'	2.8
12/30/79	Mt. Kisco	41°09.38'	73°42.79'	2.6
1/17/80	Peekskill	41°18.53'	73°55.69'	2.5
8/2/80	Keyport	40°25.73'	74°09.18'	2.7
9/4/80	Thornwood	41°06.88'	73°46.70'	2.8
12/12/80	Annsville	41°13.35'	73°54.72'	2.3
5/18/81	Ramsey	41°06.05'	74°12.20'	2.5

* These magnitudes were calculated using the formula of Chaplin et al. (1980), and the applicability of this formula for earthquakes in the New York City metropolitan area is discussed in the text.

TABLE 3

Earthquakes in New York City Metropolitan Area That Were Recorded on WWSSN Instruments: 1974-1981

Date	Location	Latitude	Longitude	Number of Stations	m_{bLg}
6/7/74	Wappingers Falls	41°34.27'	73°56.40'	9	2.75
3/11/76	Riverdale	40°57.12'	74°21.19'	1	2.04
4/13/76	Ridgefield	40°50.10'	74°02.85'	2	2.03
8/20/76	Mt. Pleasant	41°06.81'	73°45.22'	1	2.30
6/30/78	Oakland	41°04.52'	74°12.10'	4	2.37
1/30/79	Cheesequake	40°19.29'	74°15.81'	8	3.00
3/10/79	Bernardsville	40°43.34'	74°30.25'	8	2.37

3). If, however, we did not divide by T (in the A/T term of the m_{bLg} formula) we would have underestimated magnitudes by about 1 magnitude unit. Thus, we conclude that a magnitude scale similar to that of Chaplin et al. (1980) that relates signal duration to m_{bLg} determined from amplitudes of high frequency network data yields a reasonable estimate of m_{bLg} determined at 1 Hz provided that the amplitude is divided by T. Since the m_{bLg} scale was designed to estimate m_b from observations of Lg waves we have satisfied the above mentioned criterion that the magnitudes reported here coincide, to some extent, with global estimates of magnitude.

Due to the small number of WWSSN stations and only one Wood-Anderson instrument operating in the vicinity of the region studied some of the magnitudes presented in Figure C2 were determined from only one station. Figures C3 and C4 show an example of the caution necessary in interpreting magnitudes which are determined from only one station and this example also illustrates the caution necessary in estimating magnitudes from intensity data. The seismograms from WWSSN station OGD for the Cheesequake, New Jersey earthquake of January 30, 1979 and for the Wappingers Falls, New York earthquake of June 7, 1974 are shown in Figure 3. Note that the amplitude of the Lg wave on the Wappingers Falls seismogram is much larger than that of the Cheesequake seismogram. Signal duration, 10 Hz m_{bLg} magnitude and the S-wave at OGD, however, were slightly higher for Cheesequake than for Wappingers Falls. Also the area within the intensity IV isoseismal is much smaller for Wappingers Falls than for Cheesequake as can be seen in Figure 4. The distance to OGD is nearly the same for both events. Hence, the amplitudes shown in Figure C4 should be comparable.

Several factors can account for the lack of correlation between the various estimates of the size of these earthquakes. First, in the case of Wappingers Falls the Lg waves travelled parallel to the structural gain (which trends NE) and the entire path consists of competent bedrock, whereas for the Cheesequake event the Lg waves crossed many structural boundaries and traversed the sediments of the Newark basin and the sediments of the Atlantic coastal plain. Second, there may be radiation pattern effects in addition to these propagation effects. Third, the Wappingers Falls earthquake was extremely shallow (≈1 km) and was probably caused by crustal unloading associated with quarrying operations (Pomeroy et al., 1976). The depth of the Cheesequake earthquake, on the other hand, was poorly determined, and this event may have been deeper than the Wappingers Falls earthquake. Such a difference in depth is consistent with the much larger amplitudes of the surface waves in the Wappingers Falls seismogram than in that of the Cheesequake event (Figure C3).

In an effort to compare the amplitude of ground motion of Lg waves at different frequencies the Lg wave portion of seismograms recorded by the local network on analog magnetic tape were digitized and Fourier analyzed, and the resulting spectra were corrected for instrument response. Samples of Lg wave spectra determined from these seismograms are shown in Figure C5. Note that corner frequencies in these samples range from about 3 Hz to about 20 Hz, although in the large majority of cases corner frequencies were in the range 7 to 12 Hz. Amplitudes and frequencies for magnitude calculations were measured in the time domain from develocorder records. In nearly every case where spectra were determined the dominant frequency

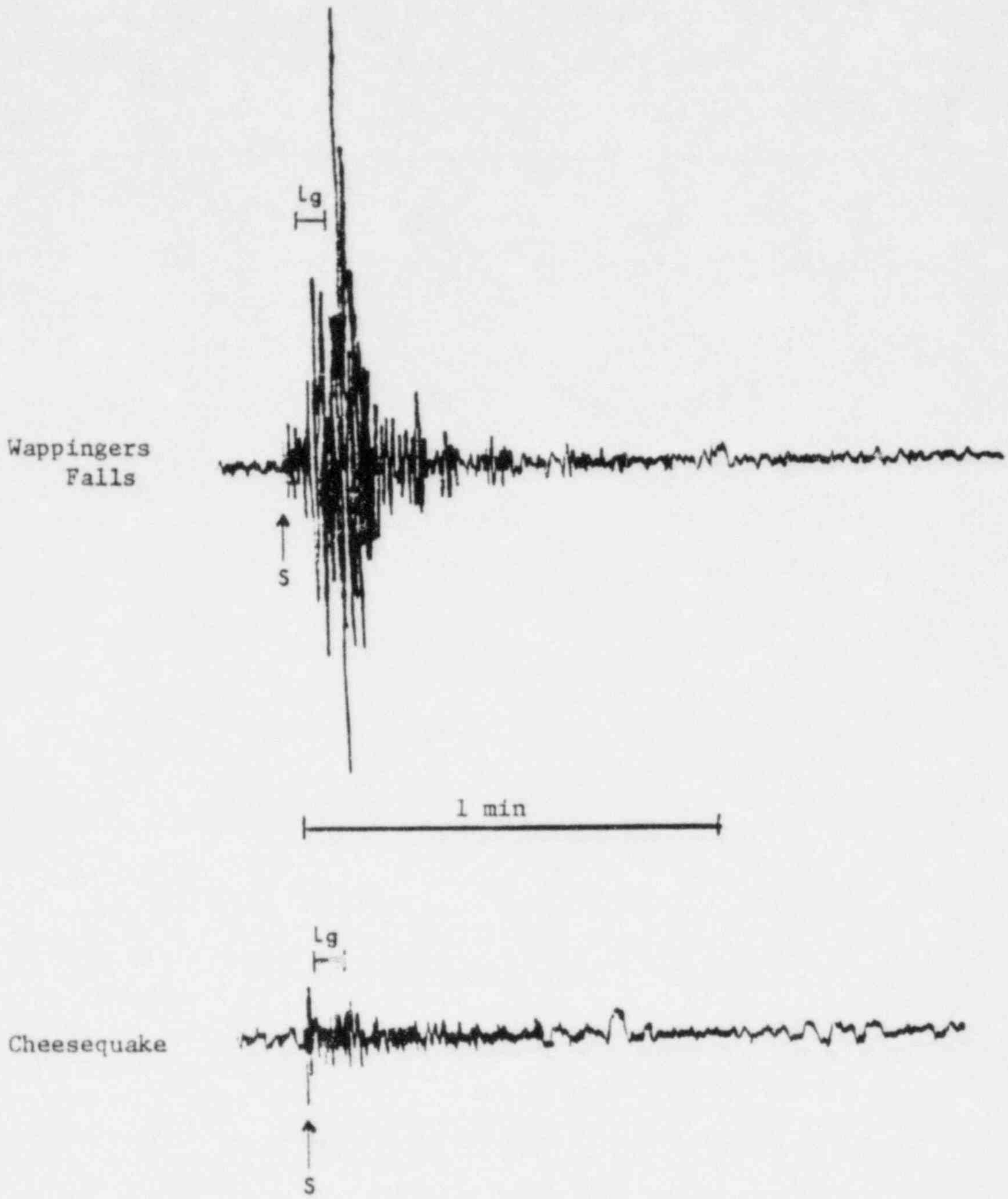

Figure C3: Seismograms recorded on the short-period vertical component of WWSSN station OGD for the Wappingers Falls, New York earthquake of June 7, 1974 (top) and the Cheesequake, New Jersey earthquake of January 30, 1979 (bottom). Arrows marked 'S' represent arrival times of the S-waves, and bars marked 'Lg' represent approximate arrival times of Lg-waves (3.5-3.0 km/sec).

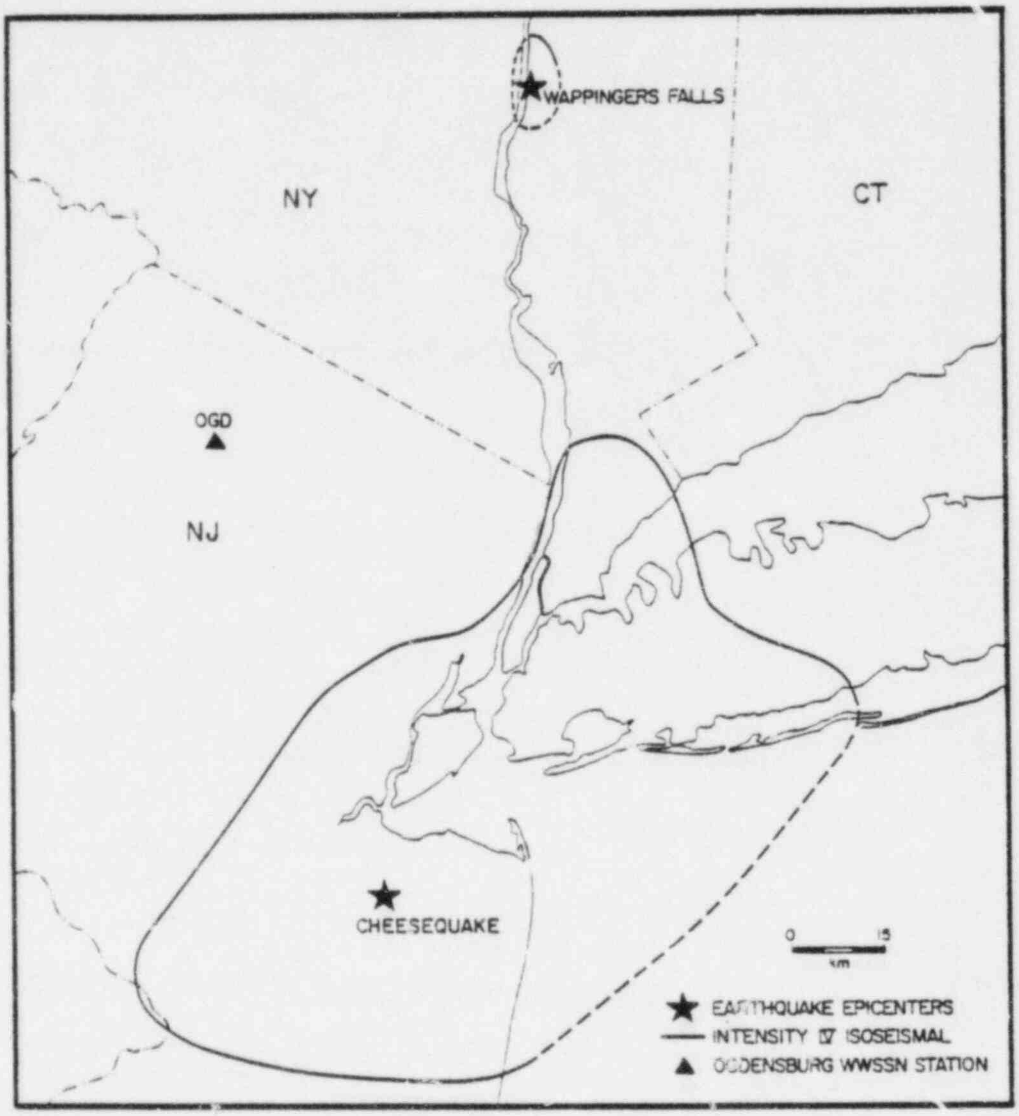

Figure C4: Isoseismals (for modified Mercalli intensity IV) of the Wappingers Falls and Cheesequake earthquakes. Stars are epicenters determined from the LDGO and Woodward-Clyde stations. Triangle represents the WWSSN station at Ogdensburg, New Jersey. The intensity IV isoseismal for Wappingers Falls is taken from Pomeroy et al. (1976), and the intensity IV isoseismal for Cheesequake was determined by Schlesinger-Miller et al. (1980).

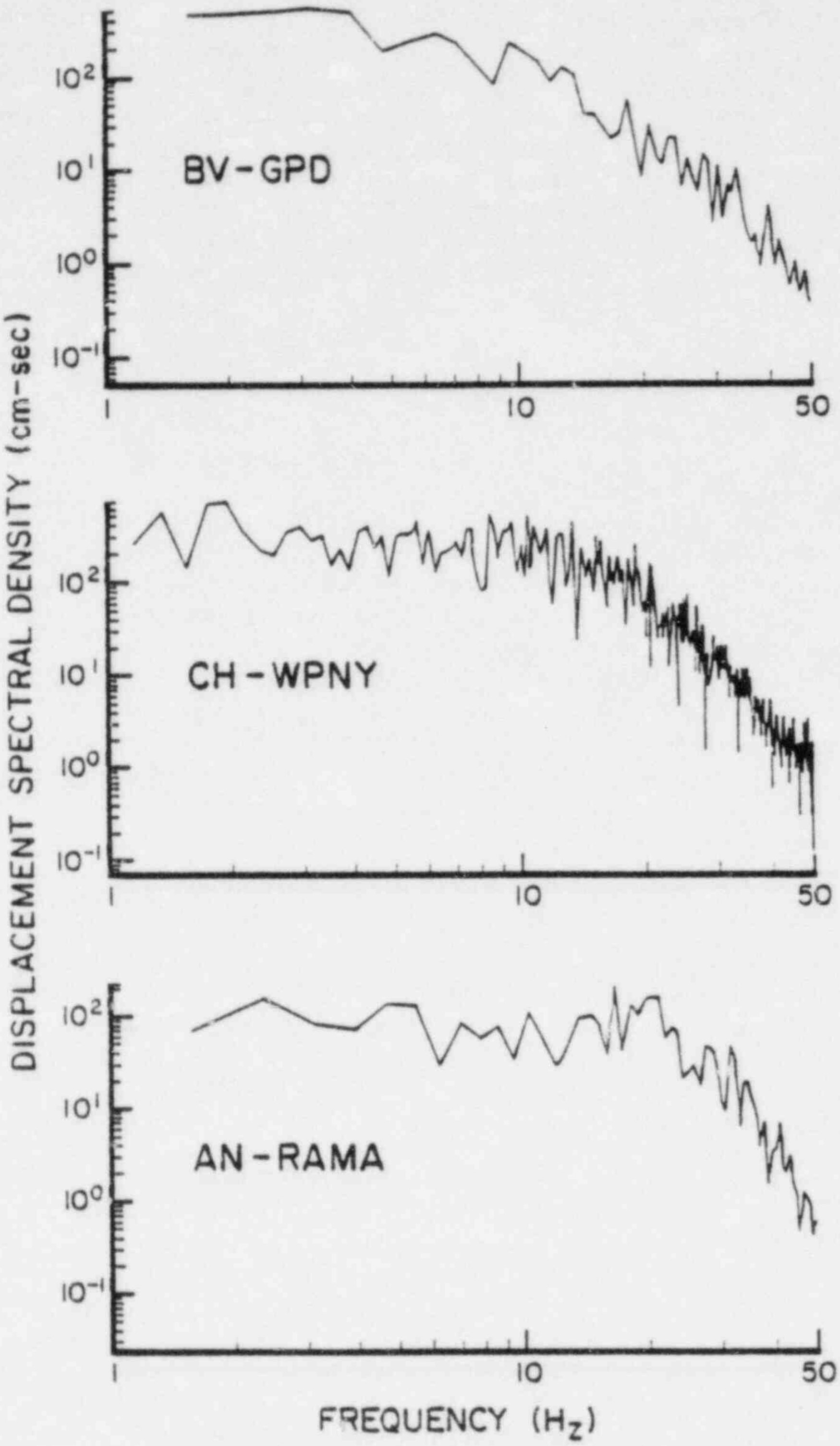

Figure C5: Samples of Lg wave spectra from seismograms of local earthquakes recorded by the LDGO network in the New York City metropolitan area. These spectra represent ground displacement (cm-sec) and are corrected for instrument response only. BV-GPD = Bernardsville (3/10/79) to station GPD; CH-WPNY = Cheesequake (1/30/79) to station WPNY; AN-RAMA = Annsville (12/12/80) to station RAMA.

content of the develocorder records was close to the corner frequency of the displacement spectrum. A comparison of frequencies read off the develocorders and corner frequencies determined from the spectra suggests that the develocorder frequencies tend to be slightly higher than the corner frequencies. Amplitudes read from develocorder records at these high frequencies, therefore, appear to be representative of the high frequency falloff portion of the displacement spectrum, and might be expected to be lower than amplitudes read from records of instruments with peak response close to 1 sec.

Since the m_{bLg} magnitudes calculated using develocorder amplitudes were found to be in reasonable agreement with those calculated from data near 1 Hz (dividing by T in both cases), dividing by T appears to, on average, correct for the fact that we are using amplitudes measured from data recorded in the high frequency falloff portion of the spectrum of ground displacement. The 1 Hz measurements, on the other hand, are measured from data recorded in the flat portion of the spectrum.

We conclude from this study that a magnitude scale based on signal duration is the most appropriate scale to use for comparing the size of earthquakes recorded by the mi roearthquake networks in the New York City region. The large amount of scatter in the various magnitude calculations shown in Figure C2 precludes any definitive statement on the signficance of differences between signal duration measurements in this study and those for New England used to develop the m_c scale. We, therefore, did not think that it was appropriate to introduce a new magnitude scale into the literature, and we use m_c below in our discussion of magnitudes and geologic structures in the New York City region.

SEISMICITY AND GEOLOGICAL STRUCTURES IN THE NEW YORK CITY METROPOLITAN AREA

Having concluded that the m_c scale is appropriate for this study, we investigated the magnitude cutoff threshold on the m_c scale that should be applied to the catalogue of earthquakes studied here to remove station bias from the distribution of seismicity. This cutoff threshold was chosen by determining the detection threshold within the study area. We first calculated the average background noise levels at several stations. We then assumed that any earthquake whose P wave amplitude (on short period network stations) was not above the background noise level would not be suitable for this study because it could not be located. Next, we measured the distance from the edge of a map of the study area (Figure C6) to the stations. Applying these distances and assuming ground motion amplitudes that are slightly greater than background noise levels, we calculated an average m_{bLg} (10 Hz) of 1.8 as the smallest magnitude earthquake that would not go undetected. Because station gains varied to some extent between 1974 and 1981, however, it is possible that some magnitude 1.8 earthquakes would not have been detected by enough stations to obtain good locations. So, to be conservative, we chose a cutoff threshold of 2.0.

The resulting distribution of the larger earthquakes ($m_c \geq 2$) recorded by the local network in the New York City region between 1974 and 1981 is shown in Figure C6 on a map of geologic provinces. These

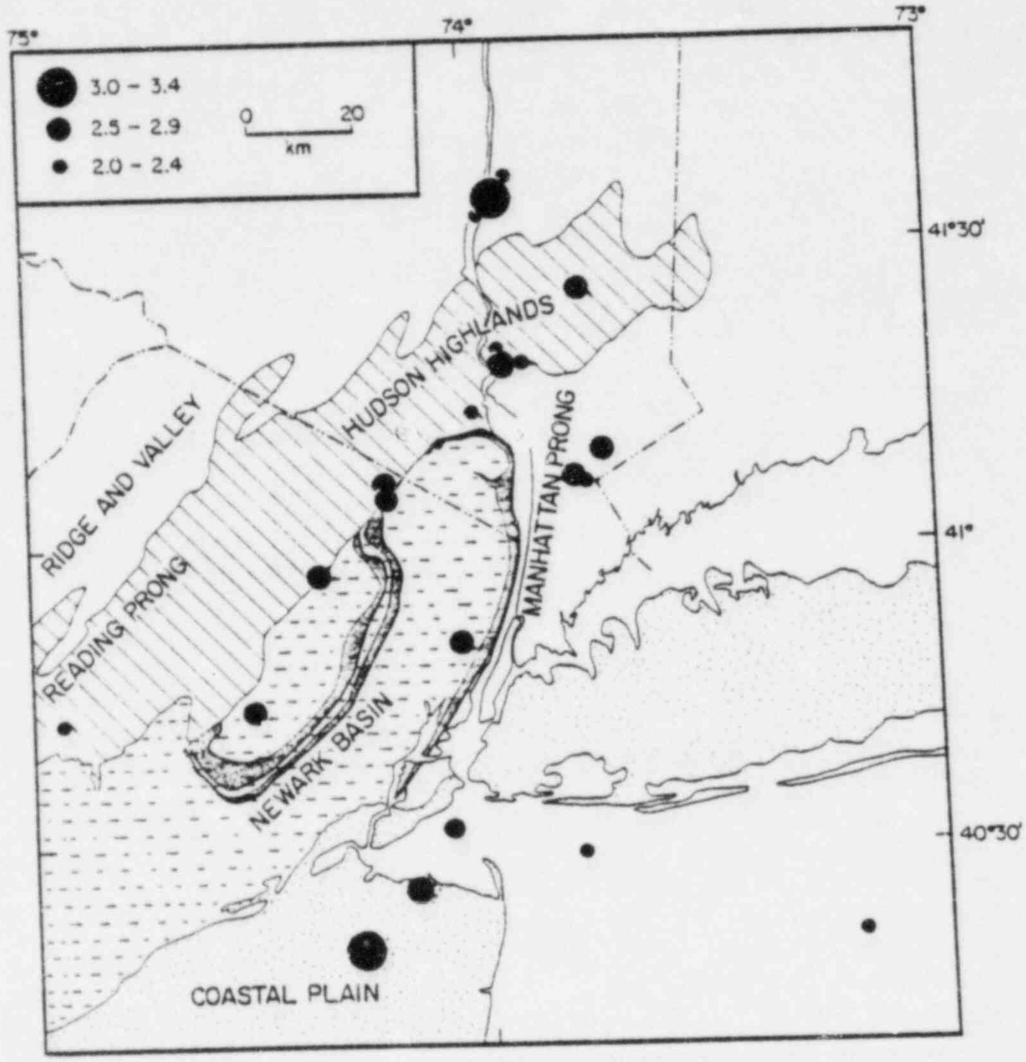

Figure C6: Earthquakes recorded by the LDGO network in the New York City metropolitan area (1974-1981), plotted on a generalized map of regional geologic provinces adapted from Ratcliffe (1980). Larger symbols are larger magnitudes as shown in upper left of the figure. The lower bound for magnitudes in this sample was chosen such that no event of magnitude greater than or equal to this threshold would fail to be detected if it occurred in the region shown.

events are also listed in Table 2. The pattern of seismicity shown should not be biased by station distribution because of the cutoff threshold chosen. In general, the earthquakes surround the Newark basin, and very few events are located within this sedimentary basin. This basin was formed by stretching and block faulting of continental crust just prior to the opening of the Atlantic Ocean in Triassic time (e.g., Dewey and Bird, 1973). On the northwest side of the basin earthquakes are occurring in the vicinity of a Triassic border fault, the Ramapo fault, as was suggested by Aggarwal and Sykes (1978), but to the north and east, earthquakes tend to occur within the older rocks of the Manhattan and Reading prongs as well as beneath (or within) the coastal plain sediments. Northeast trending faults and lineaments have been mapped in these older structures (e.g., Ratcliffe, 1980), and these features may be zones of weakness in crust that was highly fractured as a result of the stretching of the crust in this region during Triassic and Jurassic time.

Focal mechanisms of earthquakes in the structures which surround the Newark basin are generally high-angle reverse faulting with northeast trending nodal planes (Aggarwal and Sykes, 1978; Yang and Aggarwal, 1981). This correlation between mapped faults and nodal planes of focal mechanisms has been noted by Aggarwal and Sykes (1978) and Yang and Aggarwal (1981). These studies suggest that the Ramapo fault system is a zone of weakness which is being reactivated by the present day stress field. Yang and Aggarwal (1981) claim that the Ramapo fault system is probably the most active fault system in the New York City region.

The high concentration of earthquakes located by the microearthquake networks to the northwest of the basin noted by Aggarwal and Sykes (1978) and Yang and Aggarwal (1981) appears to be enhanced by the uneven distribution of stations (Figure C1b) and the shorter period of time used in their studies (1974-1979). The pattern of seismicity shown in Figure C6 should not have as great a detection bias, and, therefore, this figure is probably more representative of the longer-term distribution of earthquakes in the New York City metropolitan area than the Aggarwal and Sykes (1978) and the Yang and Aggarwal (1981) studies. Only about half of the earthquakes studied here are concentrated along the Ramapo fault zone, and faults that lie to the north and east of the Newark basin (possibly including faults which may lie beneath the Coastal Plain sediments) may be just as active as the Ramapo and other faults to the northwest of the basin. The larger earthquakes in this study are to the east of the basin (Cheesequake, New Jersey), and to the north of the basin (Wappingers Falls, New York) rather than being concentrated near the Ramapo fault zone.

The often debated questions about the long-term predominance of earthquake activity along the Ramapo fault zone relative to earthquake activity in other zones in this region (e.g., Aggarwal and Sykes, 1978; Ratcliffe, 1980, Fischer, 1981) are not yet resolved. The crustal material in the New York City metropolitan area appears to be under a relatively uniform stress field with maximum compression trending approximately W-NW (as inferred from focal mechanisms by Aggarwal and Sykes, 1978 and Yang and Aggarwal, 1981). This stress field, however, is being applied to an inhomogeneous medium, and stress concentrations are likely to be distributed in a very complex manner throughout the region. In addition, zones of weakness are also likely to be distributed in a very complex manner. Earthquakes will occur in those areas where the local stress exceeds the breaking strength of the crustal material. It is, therefore, this combintion of both high concentrations of stress and zones of weakness which causes earthquakes. An earthquake may occur in a region of relatively high strength because of a very high stress concentration. Conversely, an earthquake may occur in a relatively weak zone without a very high concentration of stress. The linear trend of earthquakes which parallels the Ramapo fault in Figure C6 may be the surface expression of a long, continuous, and very weak zone, but the lack of larger earthquakes ($m_c > 3$) along this zone between 1974 and 1981 may be the result of a relatively low concentration of stress during that period of time. Higher concentrations of stress may be responsible for the larger earthquakes which occurred away from the Ramapo fault in areas where there is no obvious evidence from surface features that would suggest a long, continuous zone of weakness.

ACKNOWLEDGEMENTS

We wish to thank A. Frankel for his help with the use of his computer programs for spectral analysis of short-period seismic data. Helpful discussions with W. R. McCann, R. E. Habermann, L. Peseckis, W. Mitronovas, G. Nottis, L. Seeber, J. G. Armbruster, J. Ebel, S. Nishenko, and D. W. Simpson are appreciated. P. G. Richards and L. Seeber critically reviewed the manuscript and their helpful suggestions are also appreciated. We thank L. Murphy who typed the manuscript and K. Nagao who drafted the figures. This research was supported by the United States Geological Survey under grant USGS-14-08-0001-19750 and the United States Nuclear Regulatory Commission under grant NRC-81-179. Lamont-Doherty Contribution Number 0000.

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NRC FORM 335 (11-81)

U.S. NUCLEAR REGULATORY COMMISSION

BIBLIOGRAPHIC DATA SHEET

1. REPORT NUMBER (Assigned by DDC): NUREG/CR-3079

2. (Leave blank)

3. RECIPIENT'S ACCESSION NO.

4. TITLE AND SUBTITLE (Add Volume No., if appropriate)

Earthquake Hazard Studies in New York State and Adjacent Areas--Final Report (April 1976 to June 1982)

5. DATE REPORT COMPLETED

MONTH	YEAR
Nov.	1982

7. AUTHOR(S)

Alan L. Kufka

DATE REPORT ISSUED

MONTH	YEAR
Jan.	1983

9. PERFORMING ORGANIZATION NAME AND MAILING ADDRESS (Include Zip Code)

Columbia University
Lamont-Doherty Geological Observatory
Palisades, NY 10964

6. (Leave blank)

8. (Leave blank)

12. SPONSORING ORGANIZATION NAME AND MAILING ADDRESS (Include Zip Code)

Division of Health, Siting and Waste Management
Office of Nuclear Regulatory Research
U. S. Nuclear Regulatory Commission
Washington, DC 20555

10. PROJECT/TASK/WORK UNIT NO.

11. FIN NO.: B6096

13. TYPE OF REPORT: Final Technical Report

PERIOD COVERED (Inclusive dates): April 1976 - June 1982

15. SUPPLEMENTARY NOTES

14. (Leave blank)

16. ABSTRACT (200 words or less)

Lamont-Doherty Geological Observatory (LDGO) currently operates a network of 38 short period seismic stations in the states of New York, New Jersey and Vermont. It is part of the larger Northeastern United States Seismic Network (NEUSSN) operated by several university groups in New York, New Jersey, Pennsylvania and New England. These networks provide a wealth of data to study seismicity, earthquake hazards, earthquake source properties, tectonic processes, and crustal and upper mantle structure in the northeastern United States and adjacent parts of Canada. The LDGO network provides data for more specific studies of earthquake processes in New York State and adjacent areas.

The operation and maintenance of the LDGO network has been supported primarily by funds from the United States Geological Survey (USGS), the United States Nuclear Regulatory Commission (NRC), and the New York State Energy Research and Development Authority (NYSERDA). This report discusses results of research related to the operation of the network during Phase I through Phase VII of our contract with NYSERDA, and also introduces current directions of research for future studies.

17. KEY WORDS AND DOCUMENT ANALYSIS

17a. DESCRIPTORS

Earthquakes
New York
New Jersey
Seismographic Networks

17b. IDENTIFIERS/OPEN-ENDED TERMS

18. AVAILABILITY STATEMENT

19. SECURITY CLASS (This report): Unclassified

20. SECURITY CLASS (This page): Unclassified

21. NO. OF PAGES

22. PRICE: $

NRC FORM 335 (11-81)

UNITED STATES
NUCLEAR REGULATORY COMMISSION
WASHINGTON, D.C. 20555

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EARTHQUAKE HAZARD STUDIES IN NEW YORK STATE AND ADJACENT AREAS

JANUARY 1983