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A Study of Seismicity and Tectonics in New England

Final Report

Prepared by J. E. Ebel

Weston Observatory
Boston College

Prepared for
U.S. Nuclear Regulatory Commission

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ABSTRACT

The operation by Weston Observatory of a seismic network in New England from 1974 to 1985 is described, and the results of the seismic monitoring are summarized. The network coverage of Weston Observatory increased from 2 operating stations in 1974 to 36 stations in 1979 and was stabilized at 30 stations in the early 1980's.

The network was used to find the locations and magnitudes of all earthquake activity detected during the study period. Most earthquakes from 1974 to 1985 were found to occur in the same places as those which have been documented historically, although the activity appears to be random both in space and time. Studies of aftershocks and detailed monitoring in selected areas did not show any strong correlations between the earthquake locations and mapped geologic structures. It is concluded that the relationships among earthquakes, cectonic or structural zones and faults exposed on the surface are not well understood. The causes of the earthquake activity in the northeast are not clearly established with the seismic data which was gathered and analyzed.

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EXECUTIVE SUMMARY

This is the final report for U.S. Nuclear Regulatory Commission contract No. NRC-04-74-238 with Weston Observatory of Boston College entitled "A Study of Seismicity and Tectonics in New England". The contract period was from April 1, 1974 to March 31, 1985. During the contract period a seismic network of 30 stations was established and operated in New England, and the data gathered by the network was analyzed to find the causes of the seismicity and to assess the earthquake hazard of the region. This report describes the operation of the seismic network and summarizes the major research accomplishments during the contract period.

Section 1 presents a brief history of the establishment of the network and an overview of the operational hardware at both the transmitting sites and at the recording facility at Weston Observatory. The response curves of the network system are given, as well as the magnitude thresholds of the network for detecting and locating earthquakes in New England.

Section 2 presents a discussion of both the historic and recent seismicity of the northeast regions and compares the regional patterns of earthquake activity. Those places which were seismically active in historic times were found also to be active during the recent time period.

In Section 3 there is an analysis of the earthquake magnitude scales which were adopted and used in New England. It is shown that the code and L_g magnitude scales overestimate the local magnitude scale by about 0.4 units.

Section 4 presents the results of a study of the temporal and spatial patterns of the earthquake activity. The earthquakes in New England appear to be both spatially and temporally random. However, the time period during which there was the highest rate of earthquake occurrence throughout the region (the early 1980's) was also the time when the largest earthquakes in the region took place. Earthquake recurrence times for New England as a whole were also computed from the recent data with the return times for magnitude 5.0, 6.0 and 7.0 being 21 years, 144 years and 994 years respectively.

Descriptions of the detailed studies of several local areas of earthquake activity are described in Section 5. These include aftershock studies (Bath, Maine; Sanbornton, New Hampshire, and Dixfield, Maine) as well as anlyses of the seismicity in more active areas (Moodus, Connecticut; Passamaquoddy Bay, Maine; and the greater New York City area). In all of these cases highly accurate locations and focal depths were obtained. In two cases (Bath, Maine and the New York City area) some earthquakes were found to occur on mapped faults.

In three cases (Sanbornton, New Hampshire; Dixfield, Maine and Passamaquoddy Bay, Maine) the seismicity did not occur on mapped faults but geologic structures in the area were noted which could indicate the unrecognized existence of local faults. In two cases (Moodus, Connecticut and some of the New York City area earthquakes) the seismicity is clearly not occuring on any mapped faults or other geologic structures.

Section 6 presents summaries of studies to better define the New England seismic crustal structure. The structure is found from analyses of seismic refraction studies, from analyses of teleseismic travel times and waveforms, and from studies of the group dispersion of R_g waves.

Section 7 summarizes the results of the research conducted during the contract period. The improved documentation of the seismicity in the region due to the expanded station coverage is noted. However, the causes of the earthquakes remained unexplained since neither the relationship between the seismicity and the geology nor the sources of the stress field triggering the earthquakes could be clearly established.

1. THE NEW ENGLAND SEISMIC NETWORK OF WESTON OBSERVATORY OF BOSTON COLLEGE.

1.1 <u>Network Configuration and Analog Recording</u>

Seismic monitoring in New England by Weston Observatory began in under this contract with the operation of stations WES amd BNH in 1974 and the installation of 8 additional stations in 1975. During the first five years of the contract period, the number of seismic stations operated by Weston Observatory was steadily increased and the configuration of the network stations was adjusted in order to provide the optimum station coverage for earthquake detection and location. Notable events in the development of the network were the merger of the Connecticut network run by the University of Connecticut with the northern New England network of Weston Observatory in June 1977 and the installation of the Moodus subnetwork in 1979. The New England network reached its maximum extent in 1979 when thirty-six stations were in operation, and that same station coverage was effectively maintained throughout the duration of the contract period. The only reduction in the seismic network since 1979 has been the dismantling of the six-station Dickey microearthquake subnetwork in northern Maine in 1982. A map of the seismic network from 1979-1982 is shown in Figure 1.1, and a listing of information on all stations operated by Weston Observatory during the contract period is given in Table 1.1. The seismic network since 1982 is identical to that in Figure 1.1 except that the coverage in northernmost Maine has been reduced from seven stations (D1A, D2A, D3A, D4A, D5A, D6A and AGM) to one station (AGM).

There was an evolution of the network hardware at both the field sites and at the Weston Observatory recording laboratory between 1974 and 1975. The first stations installed in 1975 monitored three components of ground motion using 1 sec period Benioff geophones. Telephone telemetry with Weston Observatory was accomplished with used, commercially-made amplifier/VCO/discriminator units. Later stations were installed with vertical component, 1 sec period HS-10 geophones and the telemetry was through new or used, commercially-made amplifier/VCO/discriminator units or with equipment which had been fabricated in-house. In 1976 the three-component stations were converted to single vertical-component stations with HS-10 seismometers. The signals from all of the stations except WES were sent via FM telemetry over voice grade telephone lines to Weston Observatory. Several stations were typically multiplexed on one telephone circuit. The six Dickey stations D1A, D2A, D3A, D5A and D6A were radio-linked to the AGM site before they were multiplexed for telephone telemetry to Weston Observatory. At Weston Observatory, the data were sorted using homemade or commercial

TABLE 1.1

WESTERN OBSERVATORY NEW ENGLAND SEISMIC NETWORK

1974-1985

STA ID	LATITUDE degrees	LONGITUDE degrees	ELEVATION meters	LOCATION	OPENED	CLOSED
CONNECTICUT						
BCT	41.4933N	73.3839W	69	BROOKFIELD, CT	6/75	
ECT	41.8346N	73.4113W	342	ELLSWORTH, CT	1/76	
HDM	41.4857N	72.5232W	24	HADDAM, CT	12/74	
MD1	41.5529N	72.4667W	113	MOODUS (COMSTOCK BRIDGE), CT	3/79	
MD2	41.5314N	72.4337W	61	MOODUS (PICKEREL LAKE), CT	3/79	
MD3	41.5066N	72.4715W	152	MOODUS (CAVE HILL), CT	3/79	
MD4	41.5023N	72.5121W	106	MOODUS (HADDEM NECK), CT	12/78	
MD5	41.4551N	72.4950W	101	MOODUS (SHAILERVILLE), CT	10/79	
NSC	41.4807N	71.8516W	110	N STONINGTON, CT	6/77	
UCT	41.8317N	72.2505W	149	STORRS, CT	11/74	
MAINE						
AGM	47.0817N	69.0233W	240	ALLACASH, ME	11/75	
BPM	44.6317N	68.7893N	80	BUCKSPORT, ME	6/78	
CBM	46.9325N	68.1208E	250	CARIBOU, ME	7/75	
D1A	47.0586N	69,0989W	304	DICKEY, ME	10/76	8/82
D2A	47.1304N	69.1524W	402	DICKEY (KELLY MTN), ME	10/76	8/82
D3A	47.0876N	69.1687W	259	DICKEY (CARTER BROOK), ME	10/76	8/82
D4A	47.1881N	69.2767W	490	DICKEY (ROCKY MTN), ME	10/76	8/82
D5A	47.0113N	69.2650W	365	DICKEY (BROWNS BROOK), ME	10/76	8/82
D6A	47.0890N	69.4957W	430	DICKEY (TWO MILE STREAM), ME	10/76	8/82
EMM	44.7392N	67.4894W	20	EAST MACHIAS, ME	7/75	
HKM	44.6564N	69.6408W	79	HINCKLEY, ME	3/78	
HNME	46.1599N	67.9867W	209	HOULTON, ME	12/75	
JKM	45.6555N	70.2426W	378	JACKMAN, ME	6/78	
MIM	45.2436N	69.0403W	140	MILO, ME	7/75	
PQ0	44.9863N	67.4674W	219	COOPER HILL, ME	1/79	
PQ1	44.9035N	67.3271W	93	EAST RIDGE, ME	1/79	
TRM	44.2597N	70.2551W	113	TURNER, ME	9/76	
MASSACHUSETTS						
FLR	41.7167N	71.1215W	52	FALL RIVER, MA	3/75	
LNX	42.3389N	73.2724W	345	LENOX, MA	6/78	
QUA	42.4566N	72.3738W	201	QUABBIN, MA	5/77	
WES	42.3847N	71.3221W	60	WESTON, MA	12/30	

NEW HAMPSHIRE

BNH	44.5906N	71.2564W	472	BERLIN, NH	1/63
HNH	43.7053N	72.2856W	180	HANOVER, NH	8/75

VERMONT

BVT	43.3488N	72.5853W	300	BALTIMORE, VT	10/78
DVT	44.9620N	72.1709W	370	DERBY, VT	6/78
IVT	43.5221N	73.0533W	295	IRA, VT	10/78

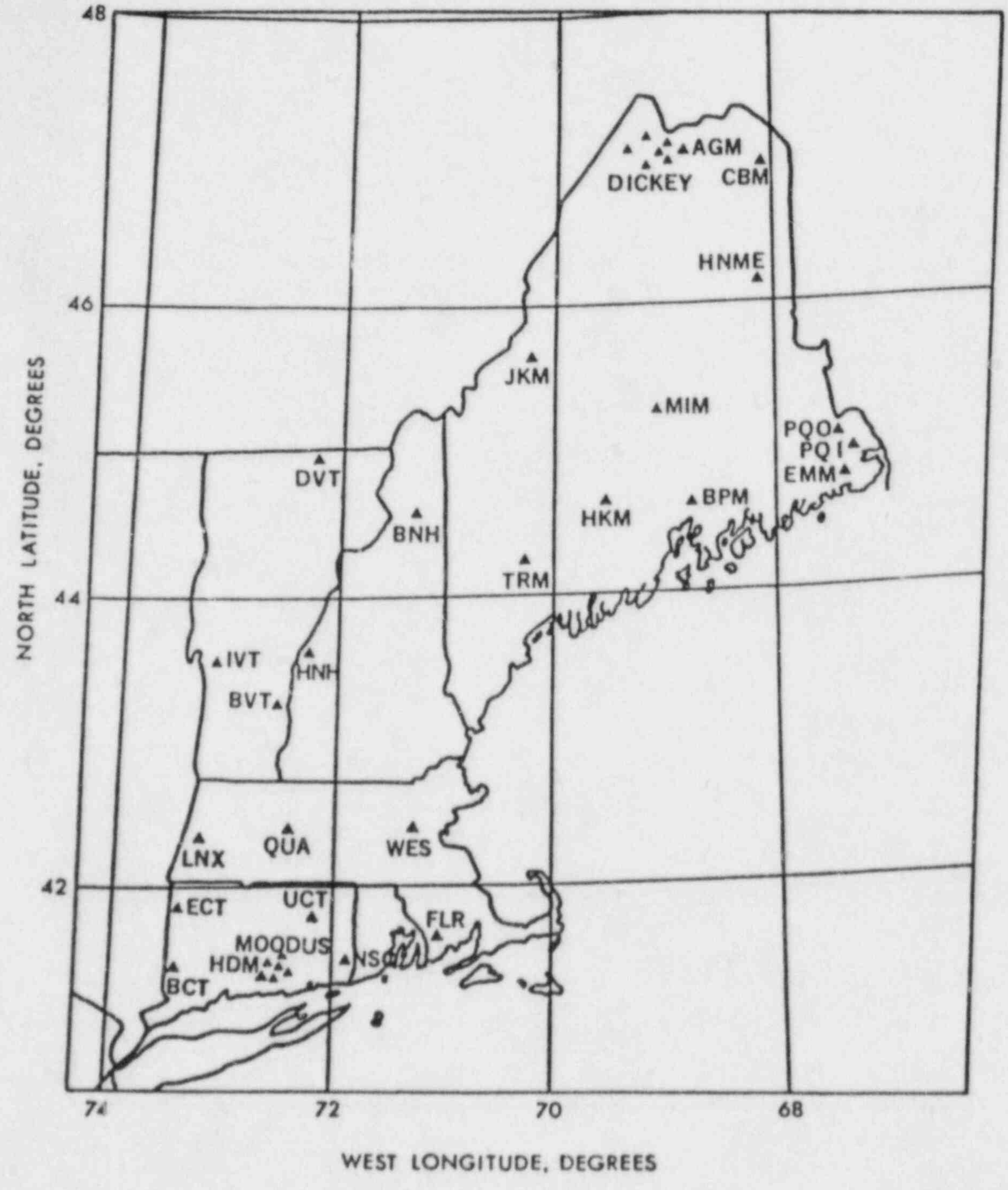

Figure 1.1 Map of the maximum extent of the Weston Observatory seismic network in New England from 1979 to 1983. Stations D1A, D2A, D3A, D4A, D5A, and D6A are indicated by Dickey in northern Maine, and stations MD1, MD2, MD3, MD4, and MD5 by Moodus in southern Connecticut. After 1983 the configuration was the same as that shown less the Dickey stations.

discriminators and then recorded in analog fashion on conventional paper drum recorders and 16mm film develocorders. Data were continuously recorded twenty-four hours a day, seven days a week. Beginning in 1978, new sites were installed with new, commercially made (Teledyne-Geotech) field electronics. The telemetry links were all by telephone with the exception of that fromPQ1 to PQO, which was by radio. The discriminator/attenuator circuitry for the stations after 1978 at Weston Observatory was also commercially made.

Beginning in 1980, there began a program to replace the homemade and used field and telemetry equipment with new commercial hardware. The purposes for this were twofold: first, this lead to a standardization of the equipment and therefore station responses, and second, it improved the reliability of the network as a whole. The commercial equipment was purchased between 1980 and 1983 and the equipment standardization was finished in 1983. The average amplitude response curve calculated from the Develocorders for the network with the standard equipment is shown in Figure 1.2.

1.2 Digital Recording

Beginning in 1982 a microcomputer digital recording system was also developed to improve the quality and utility of the data being gathered. The desire to use low-cost, state-of-the-art microcomputers for this system meant that all hardware configuration plans and software design had to be done in-house. During the first phase of the system development (1982 to 1984), the work consisted of designing, acquiring and building the hardware system, designing and writing the data acquisition software, and testing each of these designs. The first digital earthquake seismogram was taken on May 27, 1983 and the entire seismic network came on-line into the digital system in March, 1984. Writing and testing the event analysis software was the second phase of development. This work was begun in 1984 and continued to the end of the contract period. While a workable data analysis package was in use in late 1984, refinements and additions to the basic software package necessitated the long time for development of such a system. By the end of the contract period, however, both the digital data acquisition and analysis sytems were in regular use at Weston Observatory. Table 1.2 lists the earthquakes recorded by the digital system through March, 1985. Figures 1.3 and 1.4 show the configurations of the entire network hardware system and of the microcomputer-based digital recording and analysis sytem respectively at the end of the contract period.

Calibrations have also been performed for the response of the digital system. Both the impulse response to ground displacement (Figure 1.5) and the amplitude response of the system (Figure 1.6) have been calculated. Figure 1.5 was computed by taking three time

Table 1.2

Earthquakes Recorded Digitally by the Weston Observatory
New England Seismic Network

Date	Origin Time	Magnitude	Location
May 27, 1983	23:04Z	3.4	Northern Maine
May 29, 1983	05:46Z	4.2	Rumford, Maine
June 2, 1983	04:15Z	2.1	Machias, Maine
June 2, 1983	06:30Z	3.7	La Pocatiere, Quebec
August 10, 1983	23:53Z	2.3	Portsmouth, New Hampshire
August 13, 1983	17:56Z	2.3	Central New Brunswick
December 4, 1983	06:32Z	2.3	Machias, Maine
December 4, 1983	10:49Z	3.7	Dover-Foxcroft, Maine
January 19, 1984	05:27Z	3.8	Calais, Maine
February 10, 1984	07:43Z	2.2	Harvard, Massachusetts
February 22, 1984	22:36Z	2.6	S. of Long Island, New York
February 23, 1984	19:12Z	1.9	Central New Brunswick
February 24, 1984	03:17Z	3.9	Central New Brunswick
February 24, 1984	10:06Z	2.4	Central New Brunswick
May 11, 1984	15:46Z	1.7	Central New Brunswick
June 14, 1984	20:56Z	2.4	Central Massachusetts
June 28, 1984	17:00Z	2.5	Central New Brunswick
August 20, 1984	10:58Z	2.8	Hero, Vermont
August 21, 1984	19:09Z	2.5	Central New Brunswick
August 22, 1984	14:02Z	2.6	Newton, New Jersey
August 25, 1984	10:27Z	2.4	Lake Champlain, Vermont
October 13, 1984	01:45Z	3.1	Plaster Rock, New Brunswick
November 30, 1984	05:51Z	3.6	Central New Brunswick
December 30, 1984	18:08Z	1.7	Central New Brunswick
January 8, 1985	04:35Z	2.1	Kingsley Falls, Quebec
January 15, 1985	06:46Z	2.1	Taymouth, New Brunswick
February 2, 1985	17:46	2.3	Kinderlook Lake, New York
February 23, 1985	19:12Z	2.5	Central New Brunswick

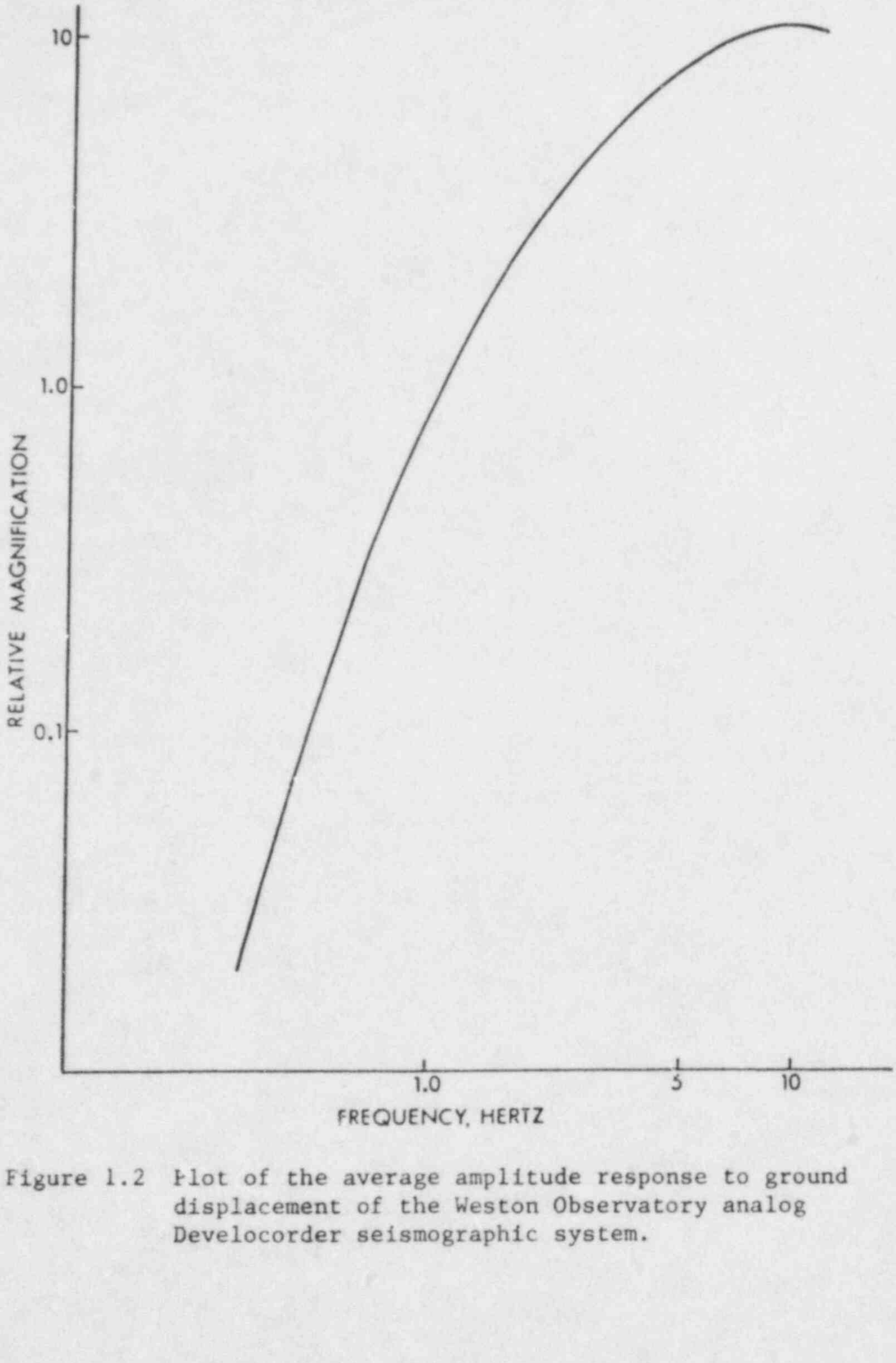

Figure 1.2 Plot of the average amplitude response to ground displacement of the Weston Observatory analog Develocorder seismographic system.

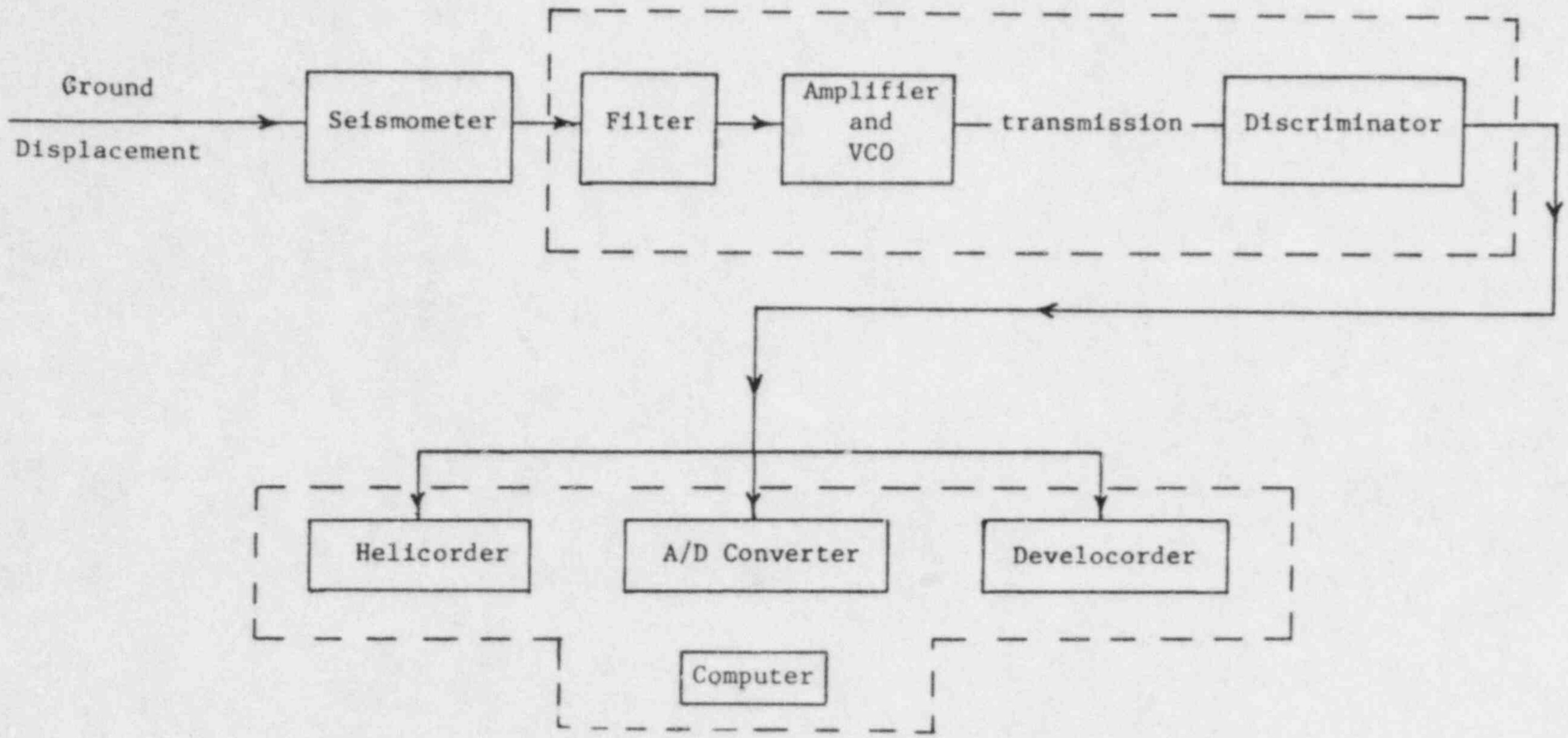

Figure 1.3 Component diagram of the Weston Observatory seismic network hardware configuration for the New England Seismic Network.

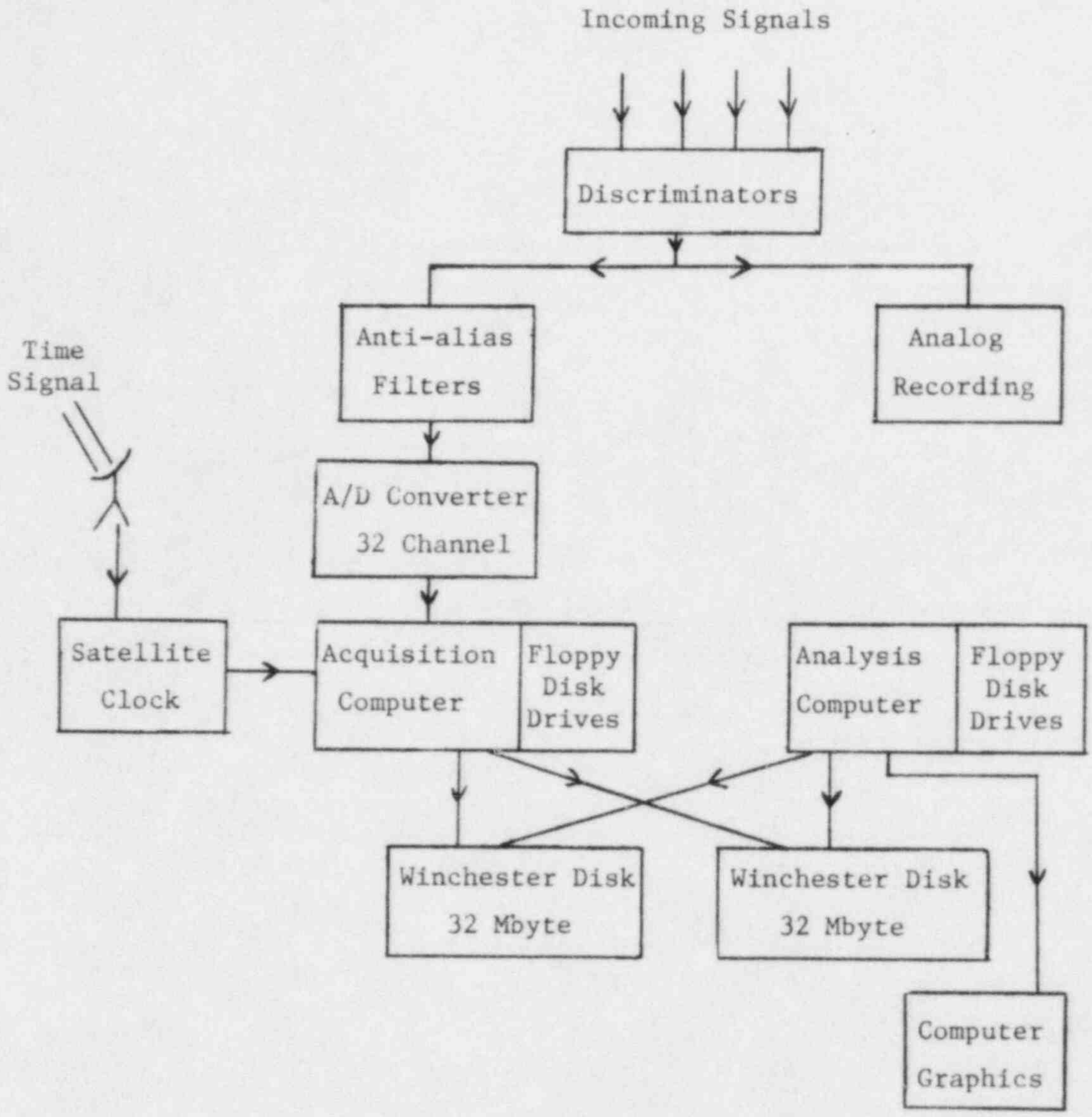

Figure 1.4 Hardware configuration of the Weston Observatory digital recording and analysis system.

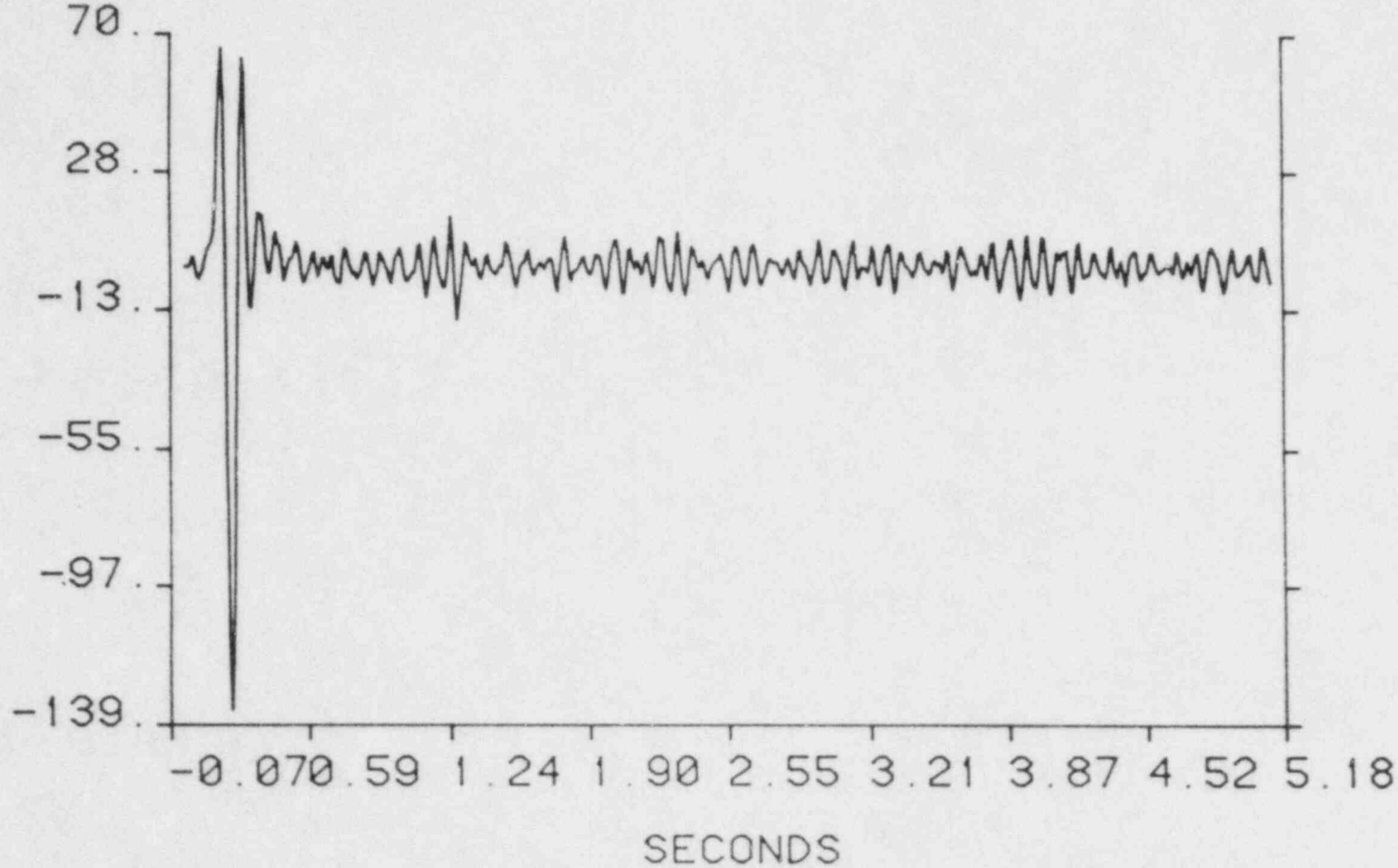

Figure 1.5 Measured impulse response to ground displacement of the Weston Observatory digital seismic system. The low amplitude energy of approximately 10 Hz apparent throughout the entire record is the background noise measured simultaneously by the seismometer as the calibration signal was being taken.

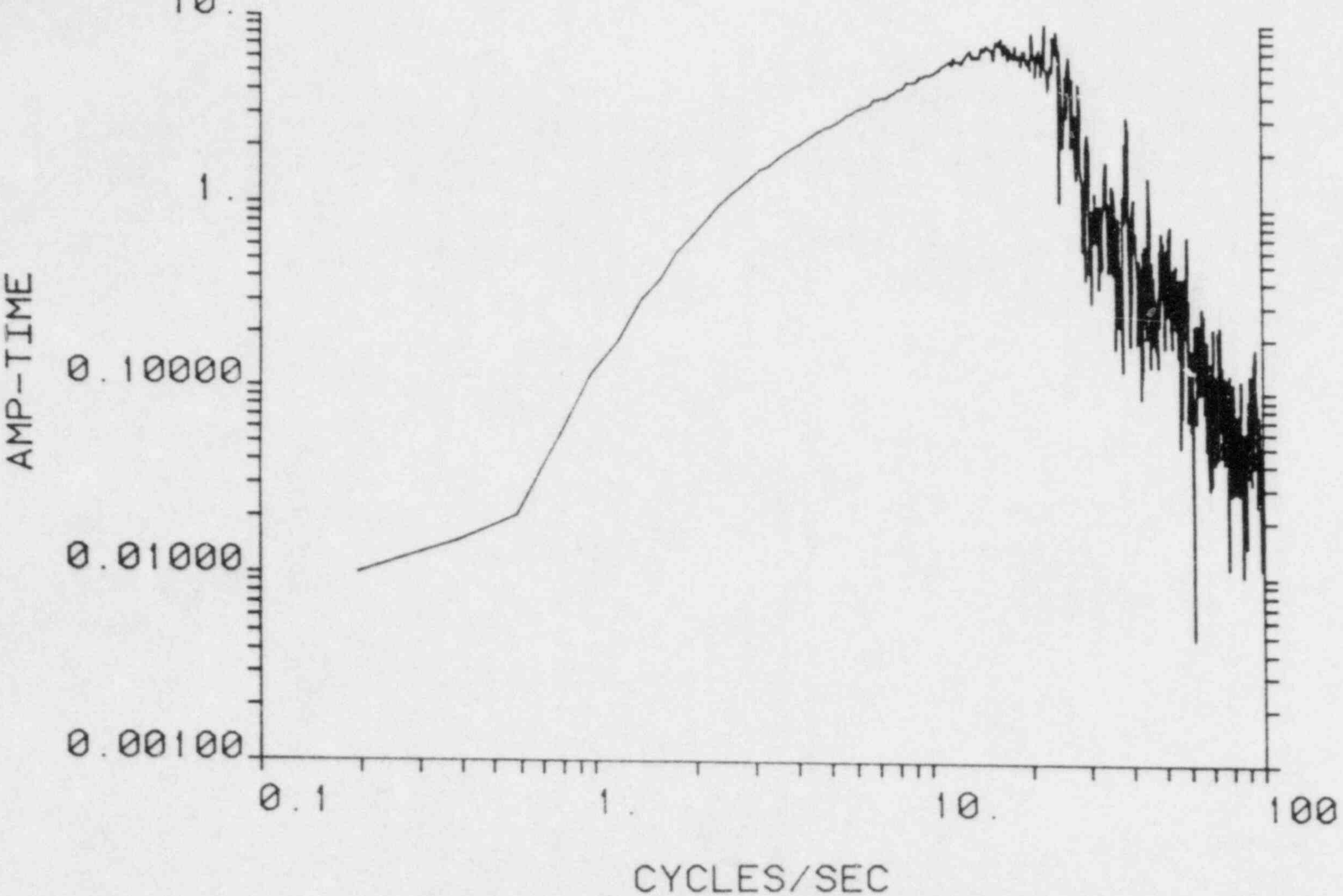

Figure 1.6 Fourier transform of the signal in Figure 1.5. This curve represents the measured displacement amplitude response of the Weston Observatory digital seismographic system.

derivatives of the step response of the system, and Figure 1.6 is the Fourier transform of the curve in Figure 1.5. With the common telemetry equipment having been installed at all the network stations before the development of the digital acquisition system, the responses of all stations are very close to those shown in Figures 1.5 and 1.6.

1.3 Strong Motion Network

Prompted by the occurrences of several strong earthquakes in 1982, a number of strong ground motion accelerographs were purchased in 1982 and 1983. These instruments, Kinemetrics SMA-1 accelerometers which record in analog form on photographic film, were deployed in 1983 and 1984 at a number of sites around New England (Figure 1.7). The trigger threshold for all the instruments is 0.01 g on the vertical component. Several of the instruments were located on bedrock sites, and those located in the area near Boston, Massachusetts were sited with the specific idea of having them easily accessible so that they can be quickly collected and moved to the epicentral zone of a large earthquake after it has occurred somewhere in the region. Through the end of the contract period, none of these instruments had been triggered by an earthquake. One instruments was installed near Gaza, New Hampshire in February, 1982, but did not trigger on any of the small aftershocks which took place there after the M_c = 4.7 earthquake of January 19, 1982. Three strong-motion instruments were deployed in May, 1983, near Dixfield, Maine to monitor aftershocks of the M_c = 4.4 earthquake which occurred on May 29, 1983. No significant aftershocks occurred during the six months the instruments were deployed. Several of the instruments were lent to Lamont-Doherty Geological Observatory in October, 1983 for aftershock recording near the epicenter of the M_c = 5.1 Goodnow, New York, earthquake. Again, no aftershock was of a sufficient size to trigger the instruments. Two instruments were also installed in eastern Maine on January, 1984 during a period of moderate earthquake activity there. Both instruments were located about 20 km away from an M_c = 3.8 earthquake which took place on January 19, 1984. At both sites the strongest vertical ground motions apparently did not exceed 0.01 g since neither instrument triggered. Thus in all cases where strong-motion data were sought, there were no earthquakes of sufficient size to allow for the collection of any strong motion data.

1.4 Network Locatability and Detectability

Ebel (1984) published a theoretical analysis of the capabilities of the seismic network for detecting and locating earthquakes in New England. Included in the analysis were all seismic stations in the area which were operating during 1983. Figure 1.8 shows the approximate thresholds for observing some sort of an earthquake signal on two different stations, while Figure 1.9 shows the approximate threshold contours for locatability of earthquakes. Figure 1.9 was computed

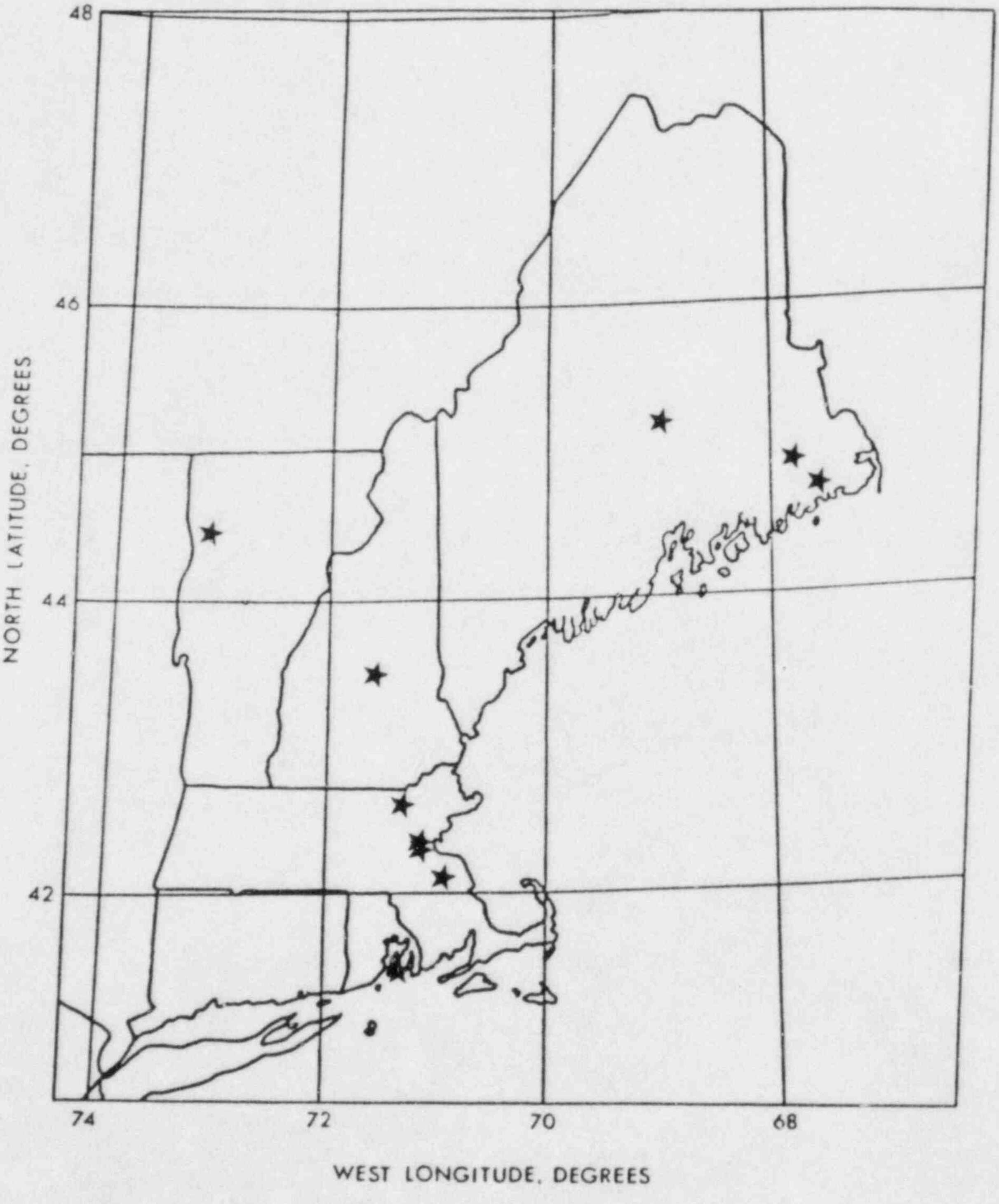

Figure 1.7 Map of the strong motion accelerograph (SMA) sites in New England maintained by Weston Observatory during 1984 and 1985.

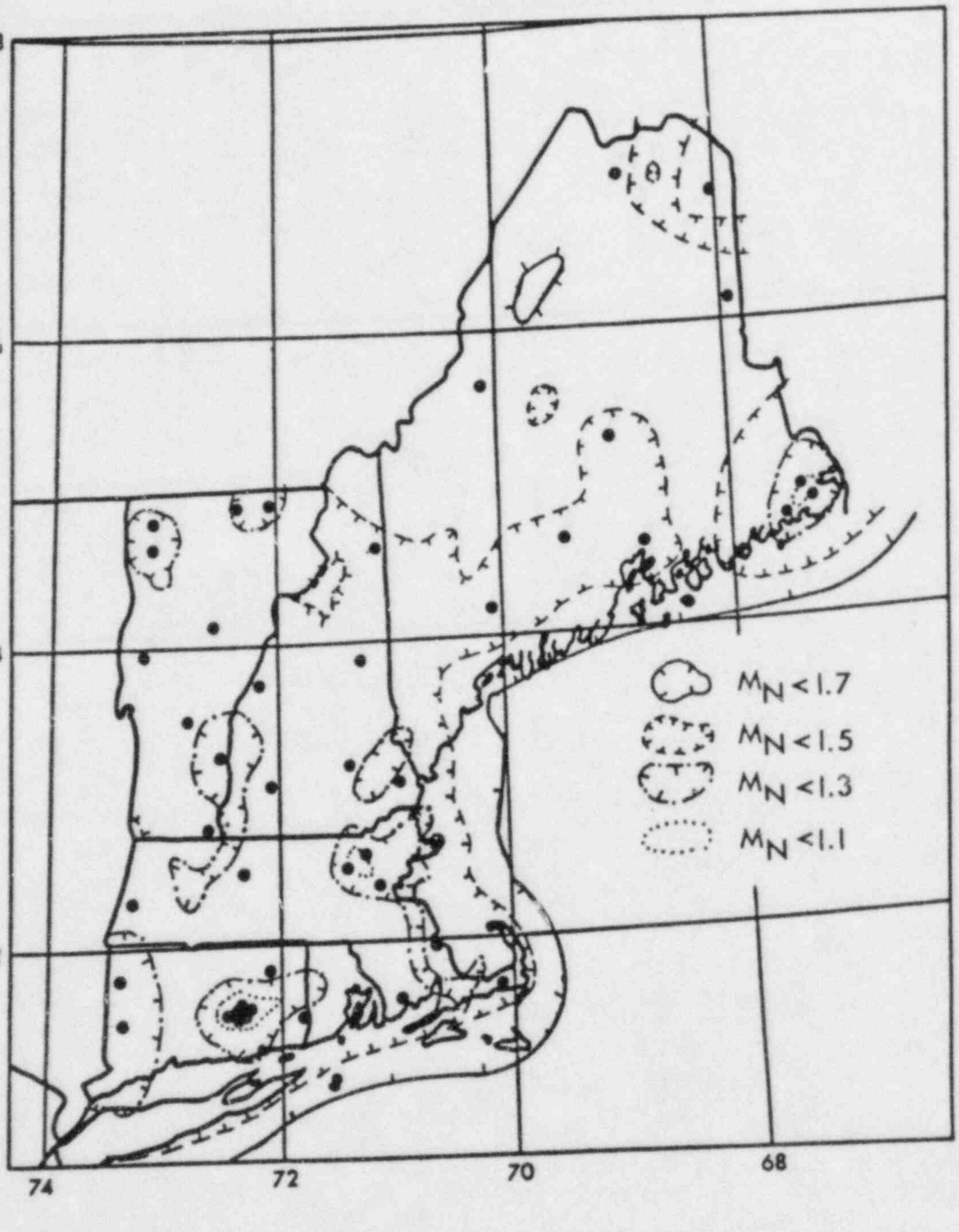

Figure 1.8 Detectibility contours for New England. The contours were calculated assuming that an Lg wave could be observed on two different stations. The solid dots show the locations of seismic stations in New England during 1983.

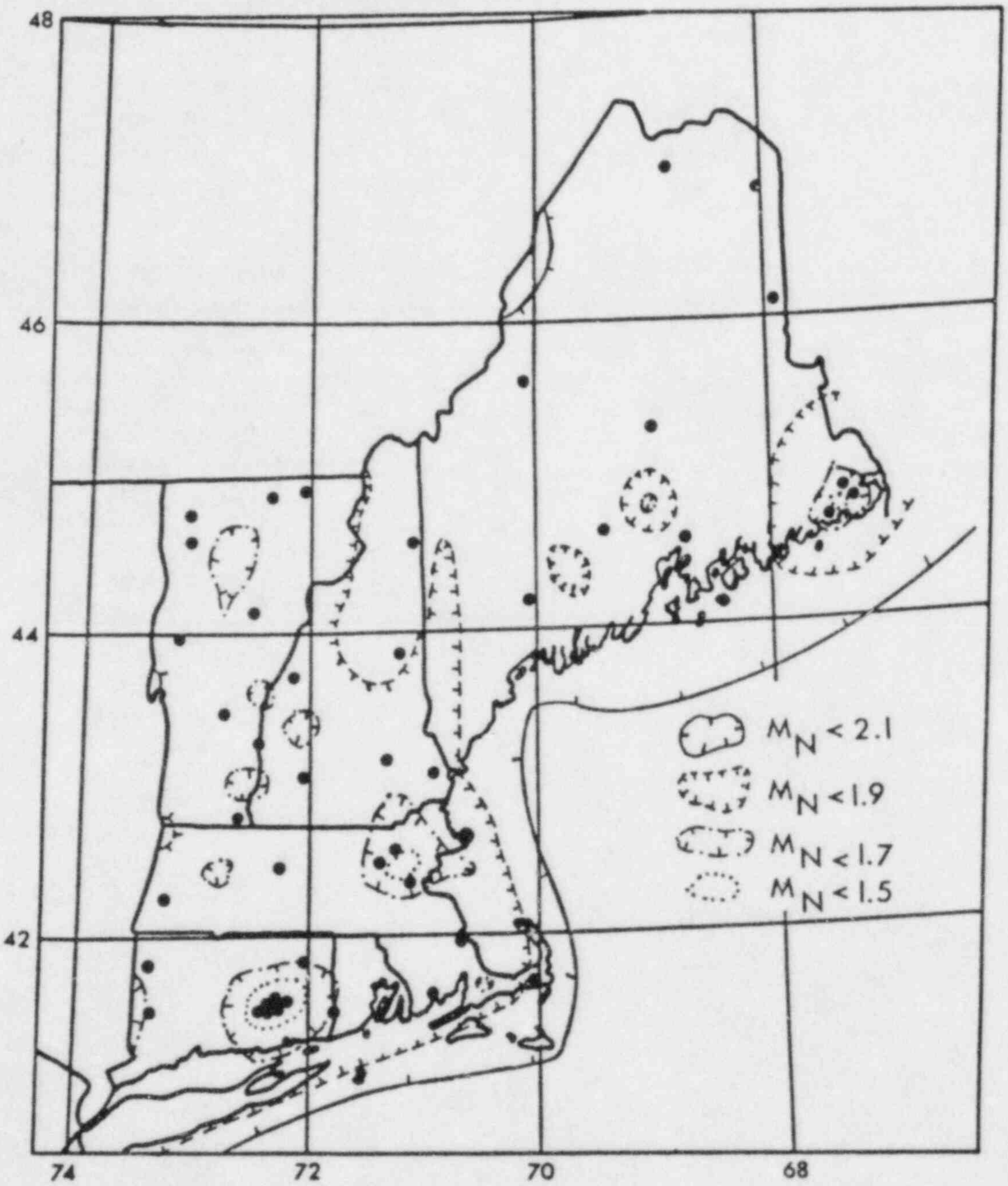

Figure 1.9 Locatibility contours for New England. The contours were calculated assuming that a P wave could be observed on three different stations. The solid dots show the locations of seismic stations in New England during 1983.

assuming that a P wave had to be observed on three separate stations in order to allow for the event to be located with some reasonable degree of accuracy (10 km or less). The contours in Figures 1.8 and 1.9 are average values and are supported by experience; it has been found that under optimum conditions the detectability and locatability of the network are better than indicated by the theoretical calculations. It was also argued by Ebel (1984) that the typical epicentral error of events recorded by the network is 5 to 10 km, which is reduced to about 1 km when tight-knit arrays of stations near epicenters are used (in cases like aftershock surveys). Depth determinations are more poorly constrained, but probably are about ±6 km or less.

1.5 Portable Seismograph and Special Studies

Following the occurence of the Bath, Maine, earthquake in 1979, Weston Observatory acquired five MEQ-800 portable seismographs to be used for earthquake monitoring. These instruments were used frequently for aftershock recording and crustal velocity studies subsequent to their acquisition. In 1984, two portable digital DR-200 recording units and force-balance accelerometers were purchased for digital field recording of ground motions, especially accelerations. The DR-200 units were used for travel time studies in Maine in 1984, but no opportunity to use them for aftershock monitoring arose before the end of the contract period. A portable field clock to provide master timing for all the portable instruments was purchased in 1983.

2. EARTHQUAKE ACTIVITY IN NEW ENGLAND

2.1 Historic Activity

An extensive effort to compile a catalog of the historical earthquake activity of the northeast was carried out during the contract period and was partially supported by this contract. The results, published by Chiburis (1981), indicated that over 2500 earthquakes could be documented in New England and adjacent regions from earliest colonial times through 1977. Of these earthquakes, ones which were widely felt or caused damage include those from 1727 and 1755 off Cape Ann, Massachusetts; 1904 near Passamaquodday, Maine; 1940 near Ossipee, New Hampshire; 1638, 1663, 1860, 1870 and 1925 near La Malbaie, Quebec; 1884 near New York City; 1935 near Timiskaming, Quebec; and 1944 at Cornwall, Ontario and Massena, New York. Figure 2.1 is a map of all seismicity contained in the Chiburis catalog and augmented with the recent seismicity through March, 1985.

2.2 Recent Seismicity

Starting with the second quarter of 1976, Weston Observatory began compiling and publishing a quarterly Bulletin of the seismicity of the northeastern United States and adjacent Canada (the Northeastern United States Seismic Network [NEUSSN] Bulletins). The first two bulletins, covering the time period from October 1, 1975 to March 31, 1976, had been published by the University of Connecticut. Bulletins through the third quarter of 1983 were completed during the contract period. Approximately 989 located earthquakes were reported in those bulletins, and several thousand additional aftershocks from the large earthquakes in the region and microearthquakes from several localities were also listed. Earthquakes of note because of their large size, location, or applicability for special study include: 1979 near La Malbaie, Quebec; 1979 near Bath, Maine; 1982 in central New Brunswick; 1982 at Gaza, New Hampshire; 1981 and 1982 at Moodus, Connecticut; 1983 near Dixfield, Maine; and 1983 near Goodnow, New York. A map of the seismicity for the time period from October 1, 1975 to March 31, 1985, taken from the quarterly bulletins of the NEUSSN and from the quarterly reports of Weston Observatory, is shown in Figure 2.2.

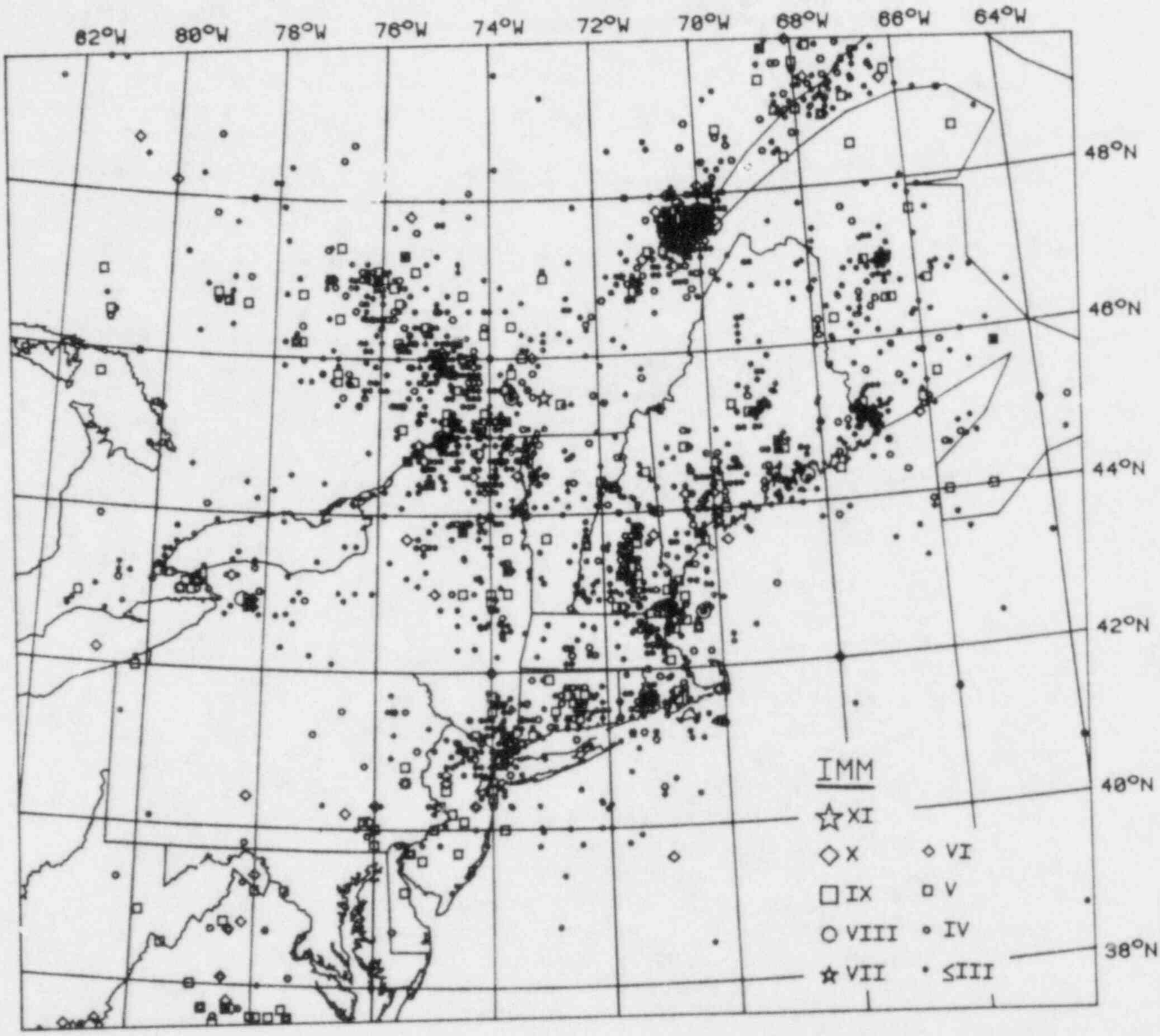

Figure 2.1 Map of the historic and recent seismicity of the northeastern United States. The historic earthquakes are from Chiburis (1981) while the recent seismicity is from the NEUSSN Bulletins or from the Weston Observatory quarterly reports of northeastern seismicity. The sizes of all earthquakes with only magnitude determinations have been converted to Modified Mercalli intensity using the formula of Chiburis (1981) for plotting on a common scale.

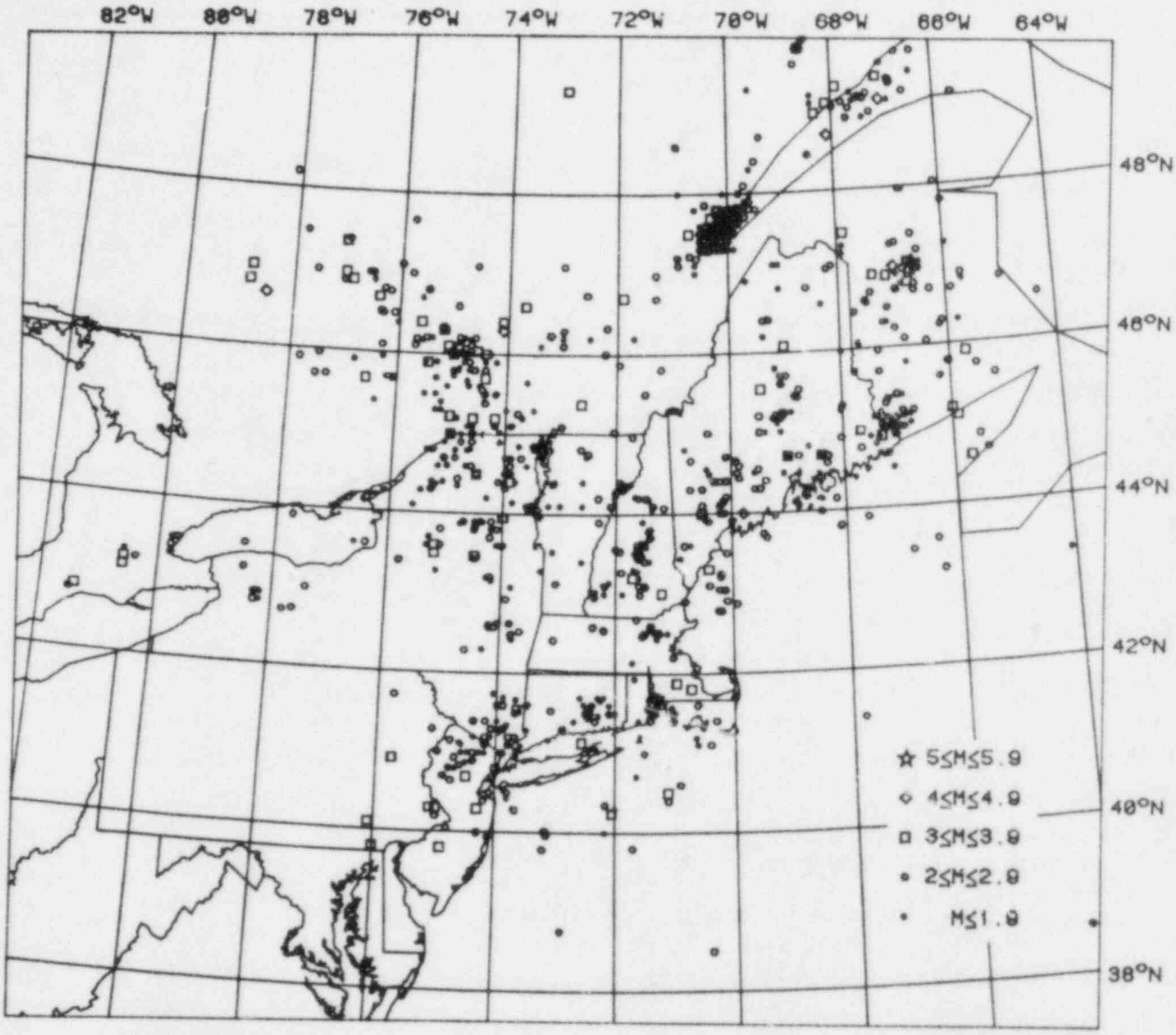

Figure 2.2 Map of the recent seismicity of the northeastern United States and adjacent Canada for the time period from October 1, 1975 to March 31, 1985. The data are from the NEUSSN Bulletins (through September 30, 1983) and the Weston Observatory quarterly reports of northeastern seismicity (October 1, 1983 to March 31, 1985).

3. EARTHQUAKE MAGNITUDES

3.1 M_N

When the seismic network became active in 1974, one of the first questions that had to be addressed was that of how to assign magnitudes to the earthquakes which were recorded. The major problem was one of finding a magnitude formula that would be accurately calibrated against theM_L scale of southern California and the m_b scale used world-wide. Thefirst magnitude scale to be used for earthquakes of the region was the M_{blg} scale of Nuttli (1973). Because the Nuttli (1973) formulae were meant to be applied to L_g waves of 0.8 to 1.3 sec period and such long periods were not generally observed on the records of the L_g waves recorded by the instruments in the region, the practice of measuring the largest sustained amplitude of the L_g wave, dividing that amplitude by the period of the signal and then applying the Nuttli (1973) formulae was adopted. These magnitudes, as reported M_N values, were computed using the described procedure throughout the entire contract period.

3.2 M_c

Because it was not possible to clearly measure L_g wave amplitudes for all events due to recording limitations, a magnitude scale based upon the maximum duration of coda waves was adopted (Rosario 1979). The formula, calibrated against the M_N scale described in Section 3.2, is

$$M_c = 2.23 \log(D) + 0.12 \log(K) - 2.36$$

where D is the coda duration (end of the coda wave train minues the origin time) in seconds and K is the epicentral distance in kilometers. The major advantage of this formula is that it can be applied to all earthquakes where the coda duration can be measured, regardless of whether or not any of the individual body or surface wave phases were recorded on scale. A disadvantage of the formula is that it may break down at large magnitudes since the m_b = 5.7 Central New Brunswick earthquake in 1982 had a measured M_c of only 5.1. Lack of data have precluded further research into this question. This possible problem notwithstanding, the M_c magnitudes for individual events agree generally within ±0.2 magnitude units of the M_N values, indicating that these two magnitude scales are well calibrated against each other.

3.3 M_L

A study designed to directly calibrate northeastern magnitudes with those from southern California was published by Ebel (1982). The data set was comprised of over 50 local and regional earthquakes which were registered between 1967 and 1981 on a set of Wood-Anderson seismographs which have been operating at Weston Observatory since 1967. The records from these events were analyzed in order to determine M_L magnitudes based upon the suggestion of Richter (1935). The Richter (1935) formula was revised to take into account the attenuation of seismic waves in New England which is different from that in southern California. The M_L magnitudes computed using the revised Richter (1935) formula were then compared to the M_N or M_c values for the events of the study. As is shown in Figure 3.1, the M_L values are consistently smaller than the M_N measurements, with the median difference being 0.4 units. No dependence of the difference between M_N and M_L with magnitude is seen. The appropriateness of the Ebel (1982) corrections to the Richter (1935) magnitude scale is illustrated in Figures 3.2 and 3.3. In Figure 3.2, the M_L values calculated using the original Richter (1935) formula with no additional correction for New England show a clear distance dependence, whereas the corrected M_L measurements shown in Figure 3.3 show no distance-related bias. The data in Figures 3.1 and 3.3 demonstrate that the M_N values overestimate the corrected M_L magnitudes, and hence the size of the equivalent earthquake in southern California, by about 0.4 magnitude units. This overestimation was also noted by Street and Herrmann (1976) in their study of M_{blg} (really M_N) and m_b magnitudes.

All of the M_N and M_c magnitudes reported by Weston Observatory during the contract period were those calculated as described in Sections 3.1 and 3.2 above. While these magnitudes overestimate the true event magnitudes by about 0.4 units, it was decided not to change the formulae used to calculate M_N and M_c in the quarterly reports of seismicity by Weston Observatory or in the NEUSSN Bulletins. There were two reasons for this. The first is that the magnitude independance of the difference between M_N or M_c and M_L means that the relative sizes of large and small events is the same for M_N, M_c and M_L. Thus b values calculated for events with magnitudes found from any of the scales should be the same. The second reason is simply to avoid the confusion that would arise when M_N and M_c are suddenly recalibrated and therefore have different values from those previously used.

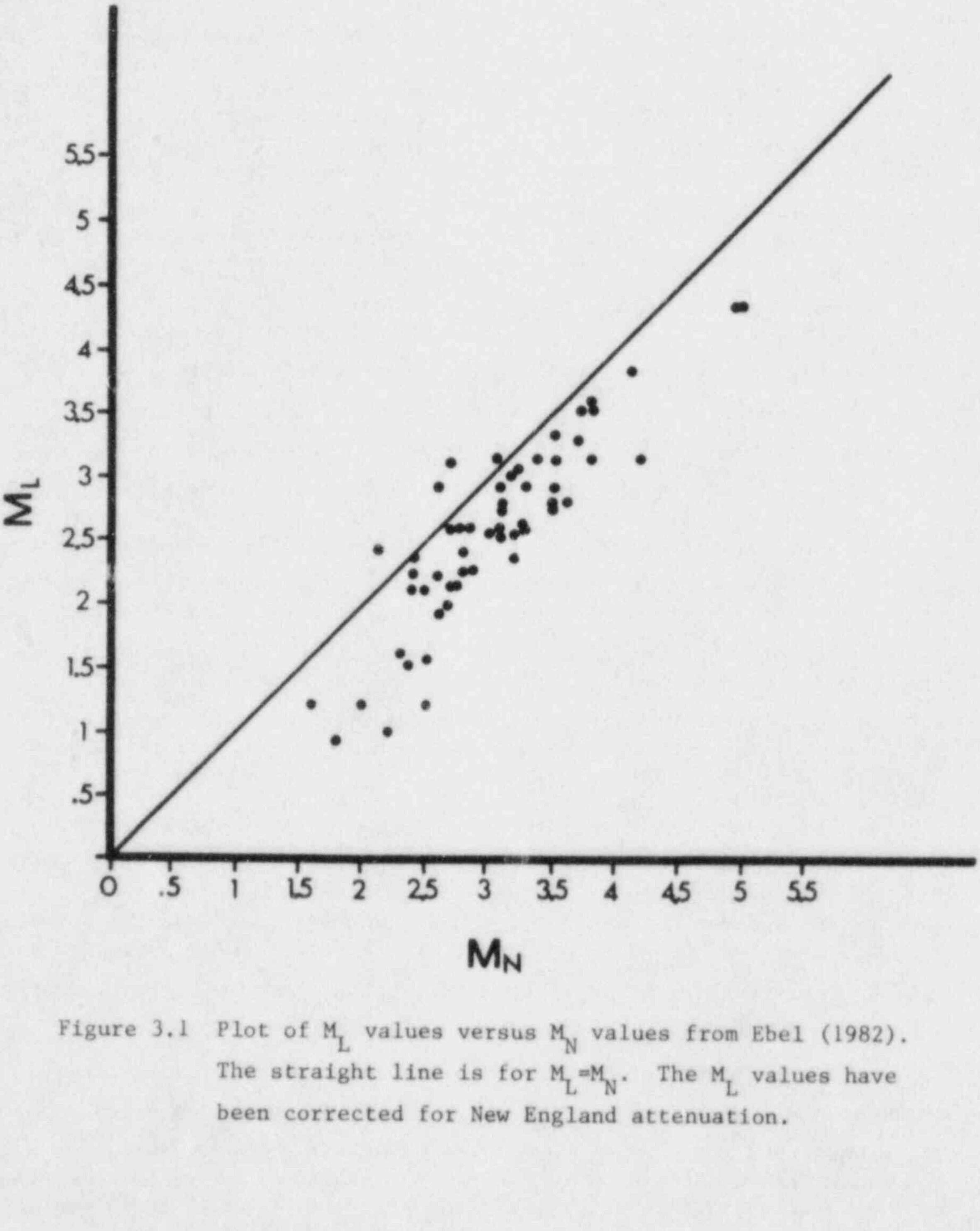

Figure 3.1 Plot of M_L values versus M_N values from Ebel (1982). The straight line is for $M_L = M_N$. The M_L values have been corrected for New England attenuation.

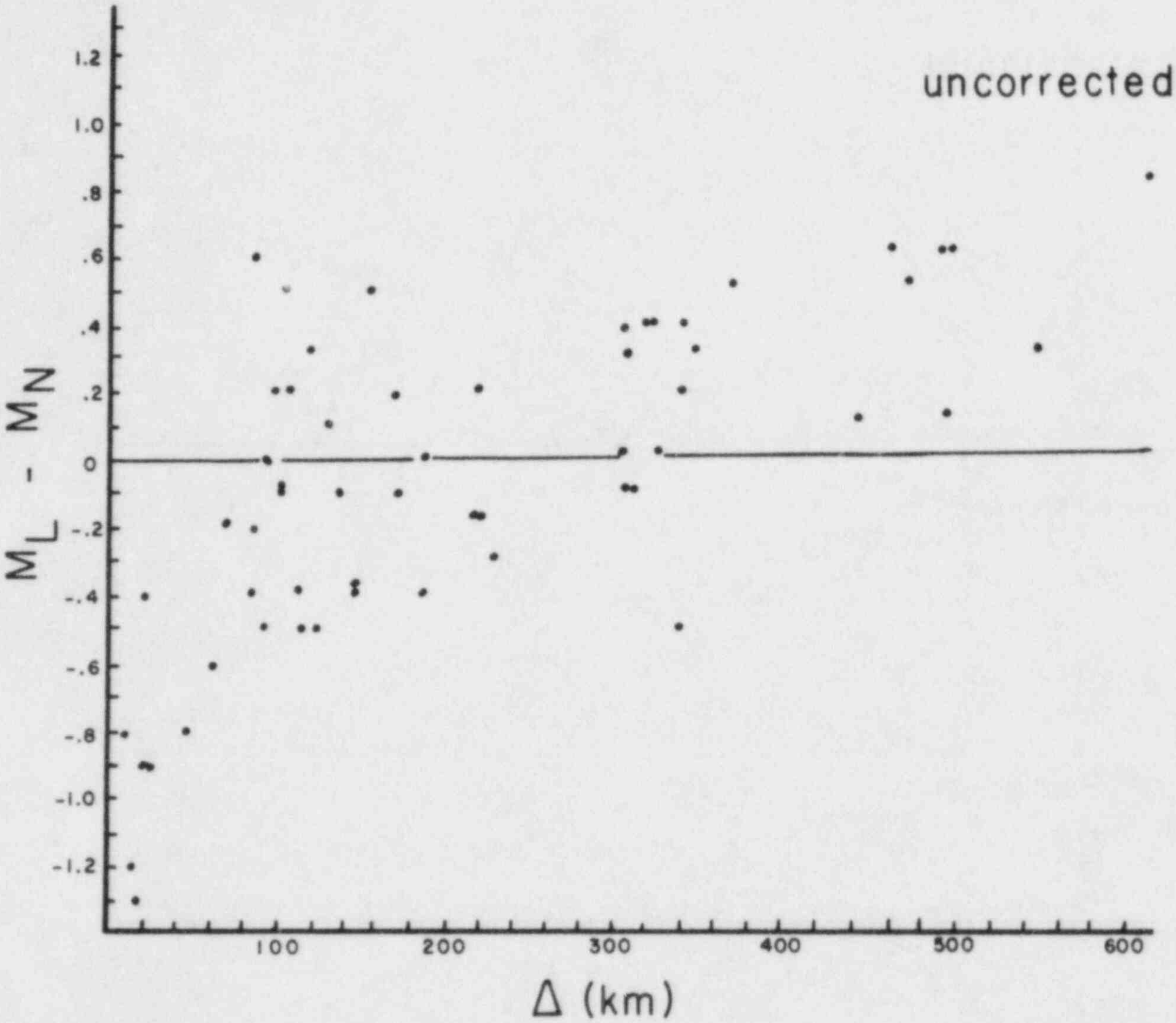

Figure 3.2 Plot of the difference M_L-M_N for the events from Figure 3.1. In this plot the M_L values have been calculated from the formula of Richter (1935) and have not been corrected for New England attenuation.

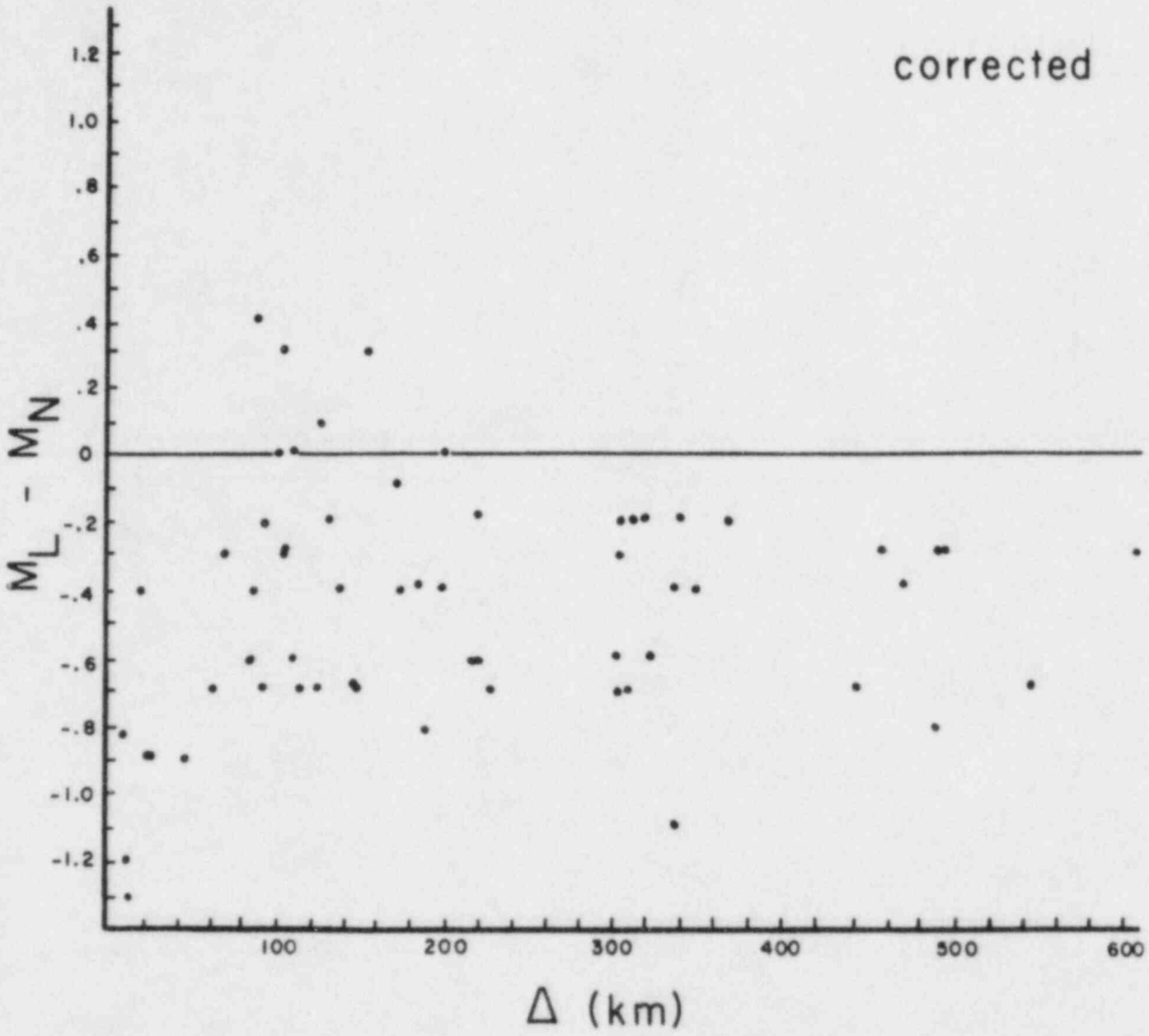

Figure 3.3 Plot of the difference M_L-M_N for the events from Figure 3.1. In this plot the M_L values have been calculated with the correction for New England attenuation (Ebel, 1982).

4. EARTHQUAKE STATISTICS AND RETURN TIMES

4.1 Variations of Seismicity with Time

The compilation of the historic seismicity by Chiburis (1981) and the careful instrumental monitoring during recent times means that the variations of seismicity with time can be examined in a meaningful way. Chiburis (1981) showed that the number of earthquakes reported per year has increased from early colonial times due to the spread of the population over the northeast and to the improved documentation of natural events by newspapers and other sources during more recent times. Ebel (1984) showed that the seismic network coverage operated since 1975 has been sufficient to record all earthquake activity above magnitude 2.0 in New England. He went on to examine the statistics of earthquake occurrences in New England between October 1975 and November 1982 since this is a complete data set. In this report the time range for the temporal statistical analysis has been expanded to encompass the time to the end of the contract period of March 31, 1985, also a complete data set for all earthquakes above magnitude 2.0 in New England.

A plot of the number of events per month for the contract time period is shown in Figure 4.1. All foreshocks and aftershocks have been removed from the seismicity data set before this figure was put together (Ebel 1984). There is obviously quite a range of seismicity from month to month. In order to remove some of the high-frequency fluctuations, Figure 4.1 was smooth with a six-month moving average boxcar window (Figure 4.2). Figure 4.2 shows the longer term fluctuations in the seismicity with time. The middle and late 1970's were relatively quiet times while the time period from 1980 to 1984 showed a trend to increased seismic activity. From early 1984 to March, 1985 there was a decrease in the rate of earthquake activity.

The temporal history of earthquake sizes is shown in Figures 4.3 and 4.4. The equivalent magnitudes were computed by summing the seismic energy release in each month using the relation $\log E = 1.9\ M_c + 9.9$ (see Bath, 1979; here M_c was used rather than M_L) and then converting that total energy back to magnitude. Practically speaking, in most months the equivalent magnitude is the same as the magnitude of the largest eathquake which occurred in those months. Figure 4.3 is a plot of the raw data and Figure 4.4 is the data filtered in the same way as in Figure 4.2. Again, the late 1970's had relatively low energy releases while the early 1980's had larger energy releases. The largest earthquakes, included in Figures 4.3 and 4.4 which occurred during the most active time in this contract period, were in January, 1982, at Gaza, New Hampshire (M_c = 4.7) and in May, 1983 near Dixfield, Maine (M_c = 4.4). Thus, the filtered data demonstrate that the larger eventsduring the contract period have occurred primarily during times of higher seismic activity throughout the region as a whole.

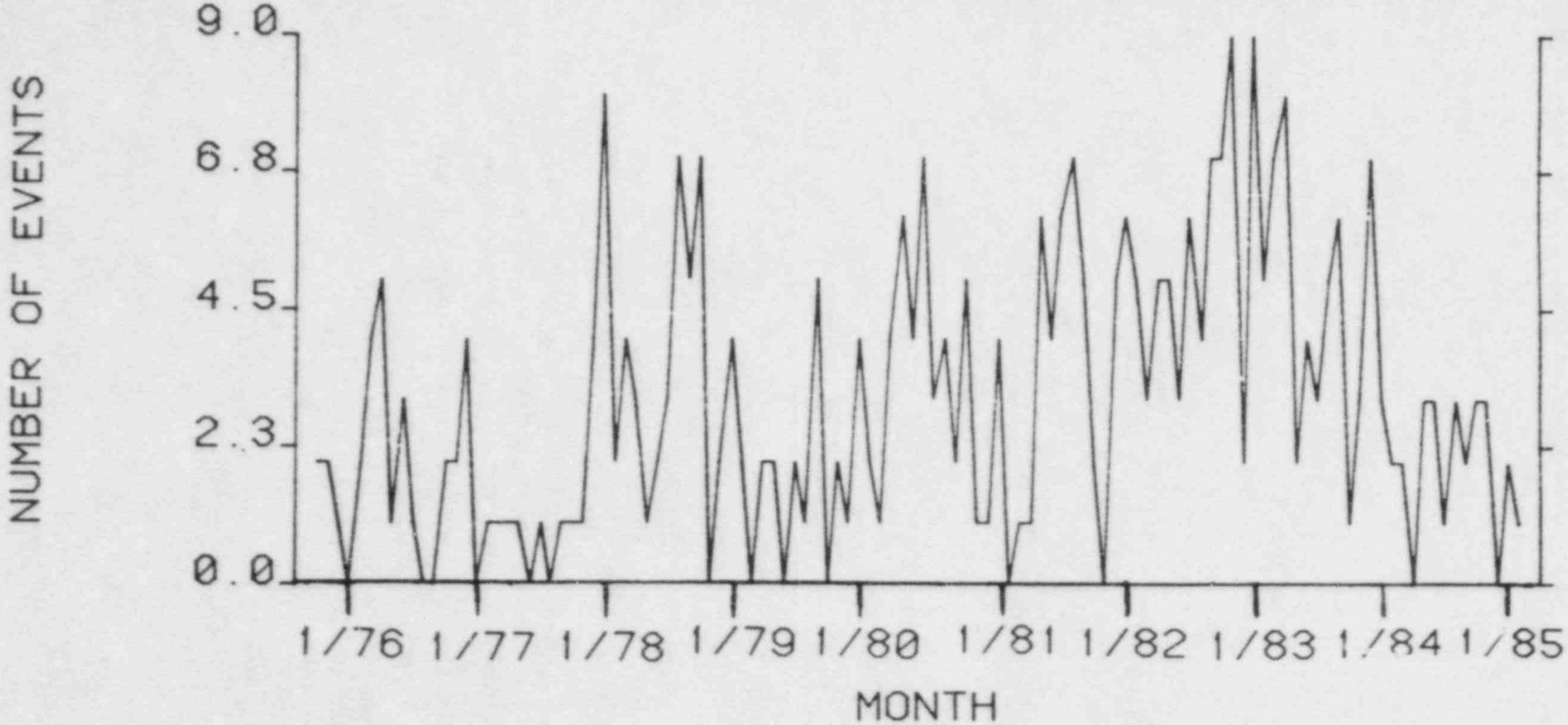

Figure 4.1 Number of earthquakes per month for the time period from October, 1975 to March, 1985 for New England.

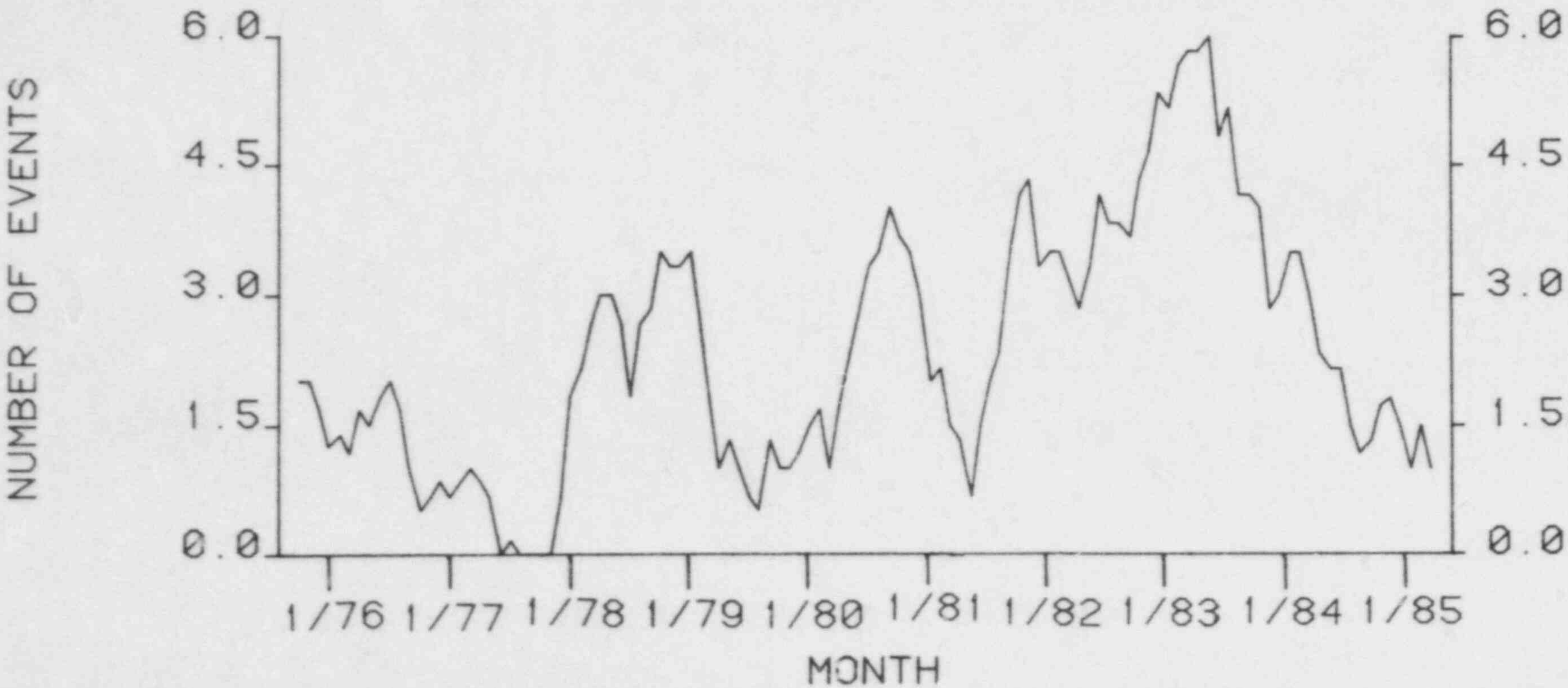

Figure 4.2 The data of Figure 4.1 filtered with a 6 point, moving-average boxcar filter.

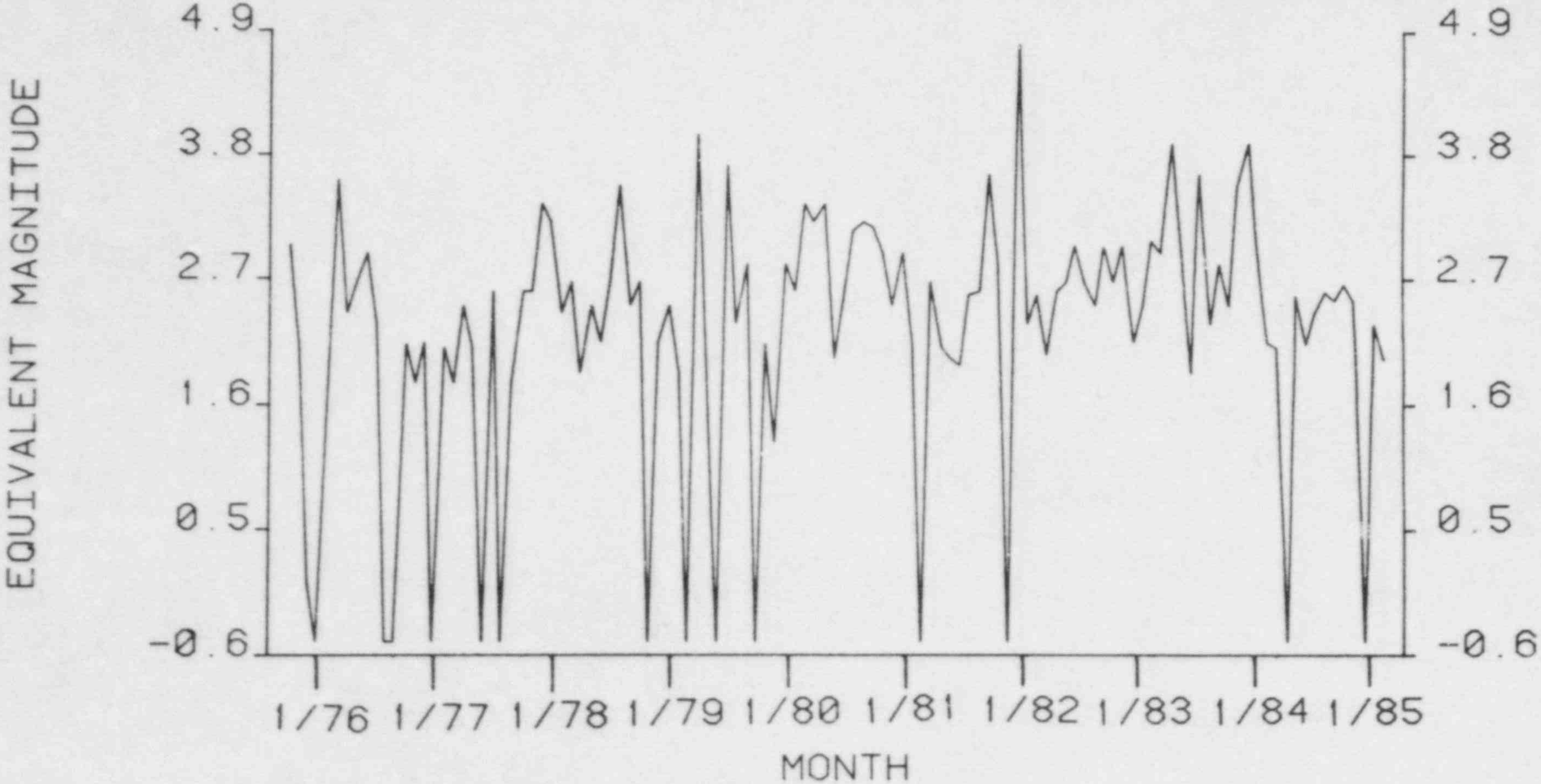

Figure 4.3 Equivalent magnitude per month for the time period from October, 1975 to March, 1985 for New England. The equivalent magnitude was computed by finding the total earthquake energy release per month and then converting that energy back to magnitude.

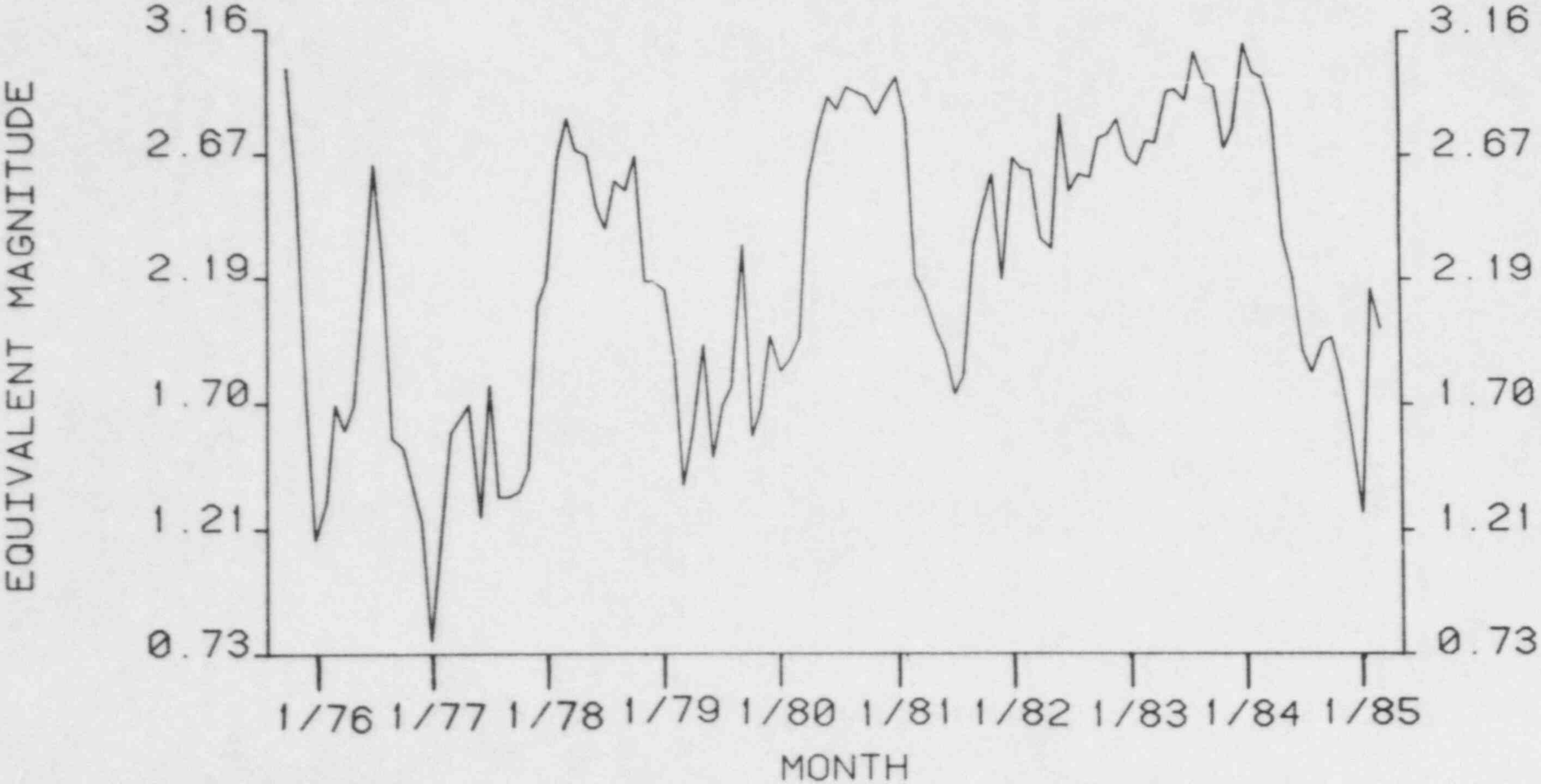

Figure 4.4 The data of Figure 4.3 filtered with a 6 point, moving-average boxcar filter.

4.2 Temporal Randomness of the Seismicity

A search for periodicities in the time and equivalent magnitude data of Figures 4.1 and 4.3 was made by autocorrelating each sequence. These autocorrelations, each of which was normalized by the variance of the time series so that the maximum value of the autocorrelation is 1, are shown in Figures 4.5 and 4.6. If the sequences are random, as has been proposed for earthquakes in other areas (Gardner and Knopoff, 1974; Shimazaki, 1973), the autocorrelation coefficients should be 1 for zero relative shift of the sequences and should be 0 for all other shifts. Any large coefficients indicate a periodicity in the data with a period equal to the corresponding shift in the autocorrelated sequences. The autocorrelations of both number of events (Figure 4.5) and equivalent magnitude (Figure 4.6) appear to be very close to random sequences. In neither case are the autocorrelation values significant at any relative shift other than 0. These data show even more temporal randomness than that reported by Ebel (1984), as would be expected when a longer, finite, random sequence is being autocorrelated.

The normalized number of events and equivalent magnitude sequences were also crosscorrelated (Figure 4.7). Once again there is quite a bit of scatter in the correlation coefficients, and the only strong peak in the curve (0.60) is for a zero lag between the two sequences. The significance of this strong correlation is less than it might appear since it arises in part due to the fact that there is very little energy released in months when there are no earthquakes detected. The crosscorrelation clearly indicates that there is no recognizable relationship between the number of earthquakes in a month and the energy release in any month preceding or following it. Statistically speaking, this means that one cannot use either a temporal seismic quiescence or a period of increased activity in New England to forecast an impending large energy release somewhere within the region. However, there is reason to expect an increased chance of having a larger earthquake during periods of high seismicity throughout the region.

4.3 Spatial Randomness of the Seismicity

Ebel (1984) published an analysis indicating that the spatial pattern of New England seismicity with time is approximately random. Figure 4.8 is a map of the seismicity used in that analysis, and Figure 4.9 shows two time lines where the earthquakes from Figure 4.8 have been projected onto NW-SE and NE-SW lines. Only earthquakes with magnitudes greater than 2 have been plotted in Figure 4.9 both to insure that a uniform data set is used and to emphasize any migrations of the more significant events. The seismicity appears to be approximately spatially random although there may also be some broad migrations of the epicenters. For instance, in 1980 the earthquakes clustered predominantly in central Maine while in 1981 they took place in southern and coastal New England as well as in northern Maine. In general, however, this data set does not show any strong spatial migrations of the earthquake activity with time.

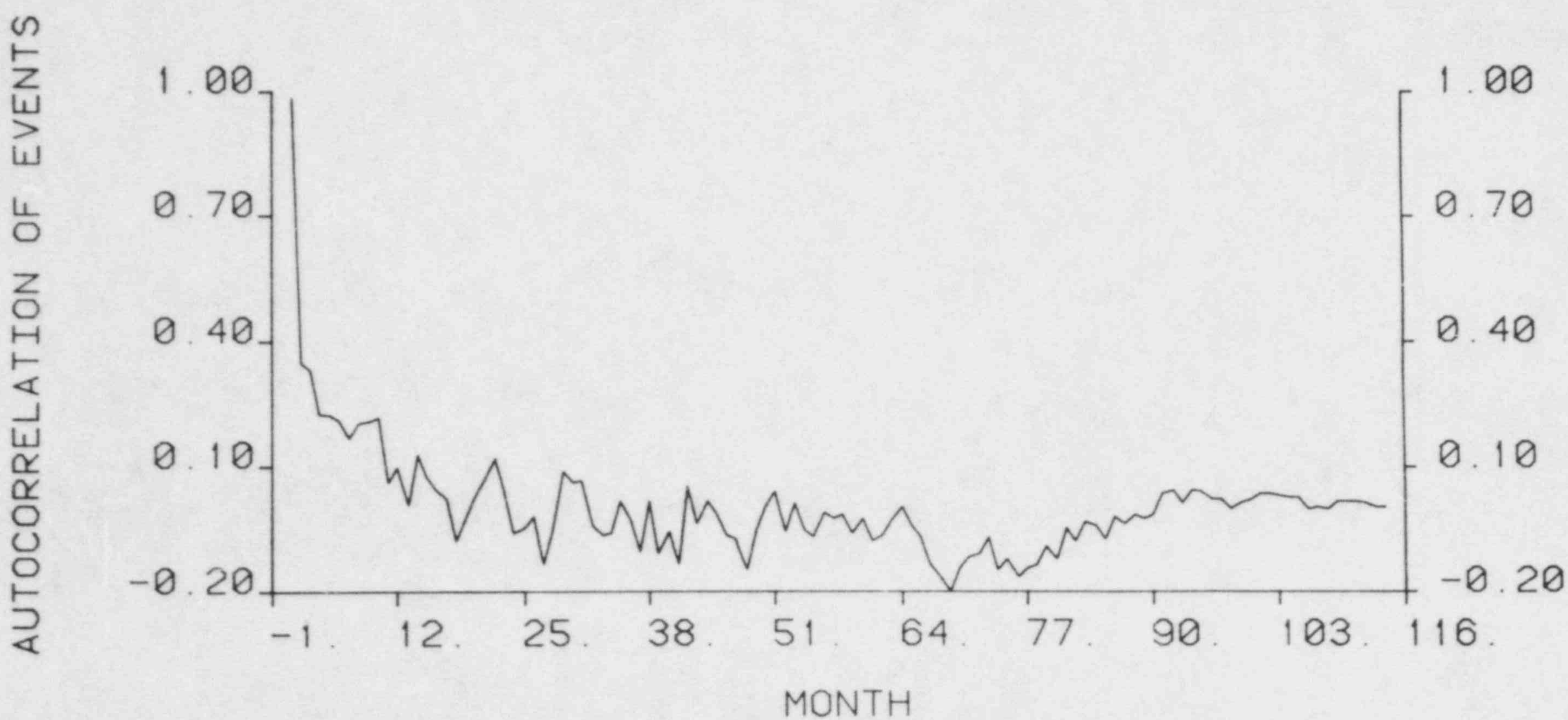

Figure 4.5 Autocorrelation of the time series of Figure 4.1.

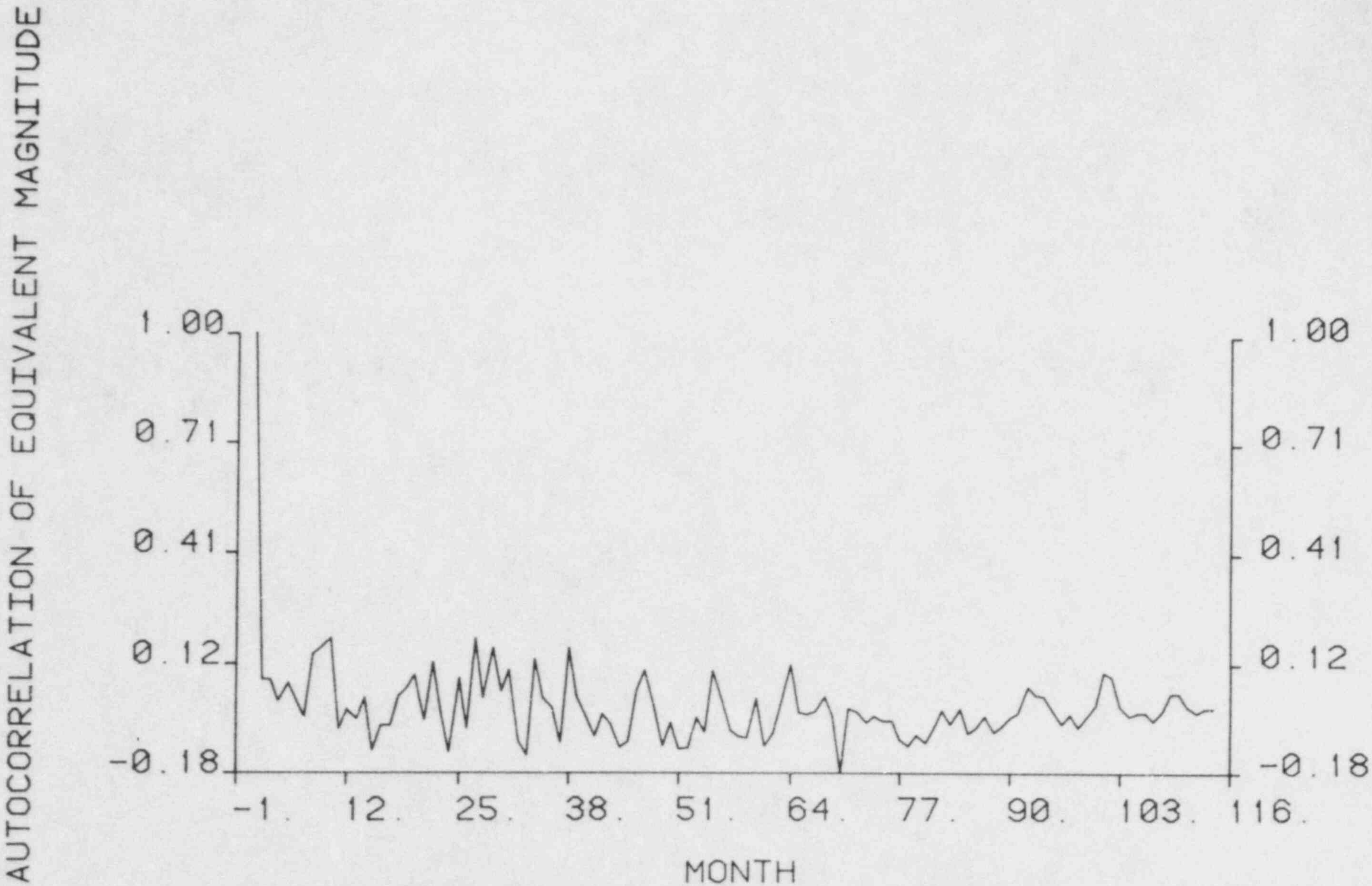

Figure 4.6 Autocorrelation of the time series of Figure 4.3.

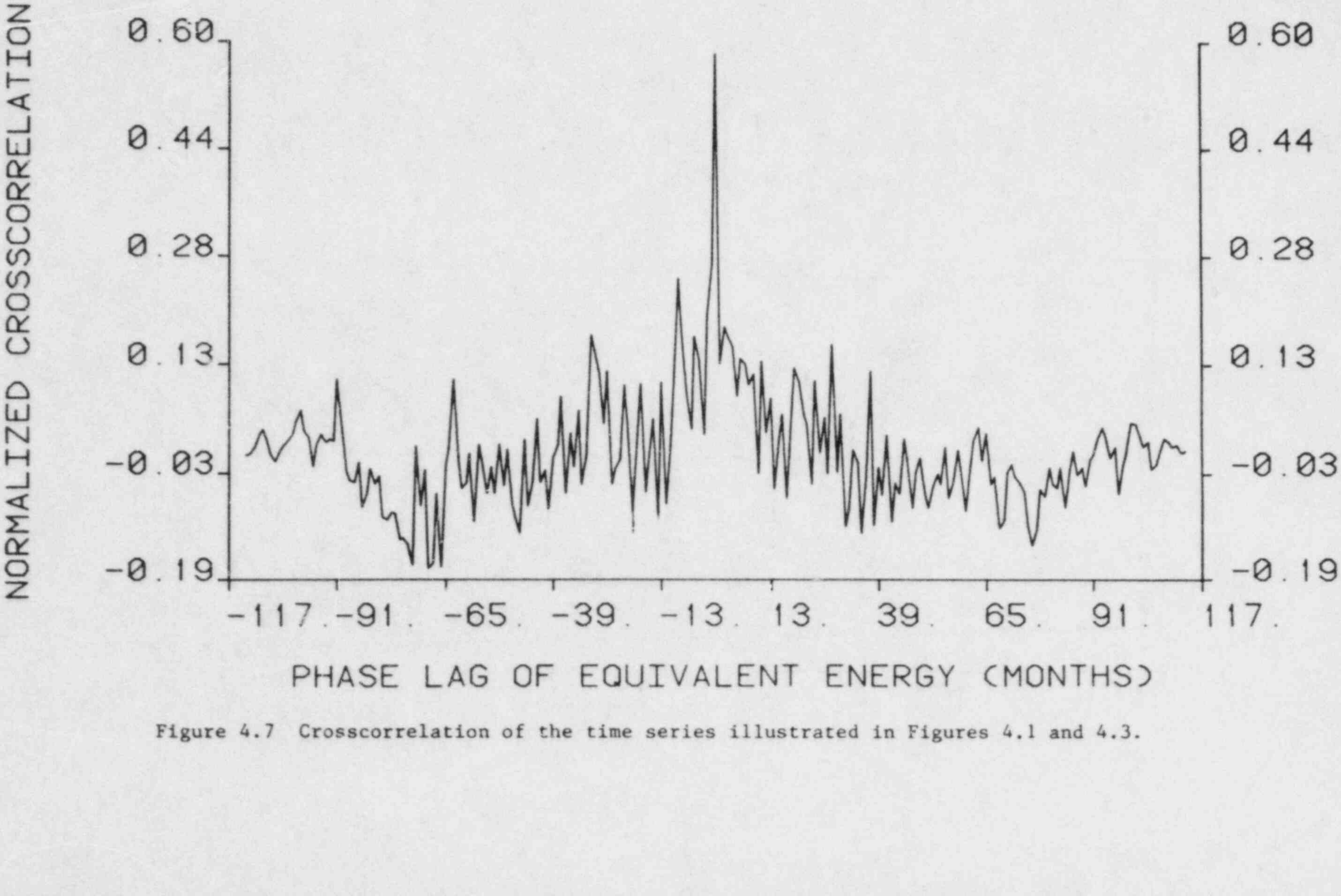

Figure 4.7 Crosscorrelation of the time series illustrated in Figures 4.1 and 4.3.

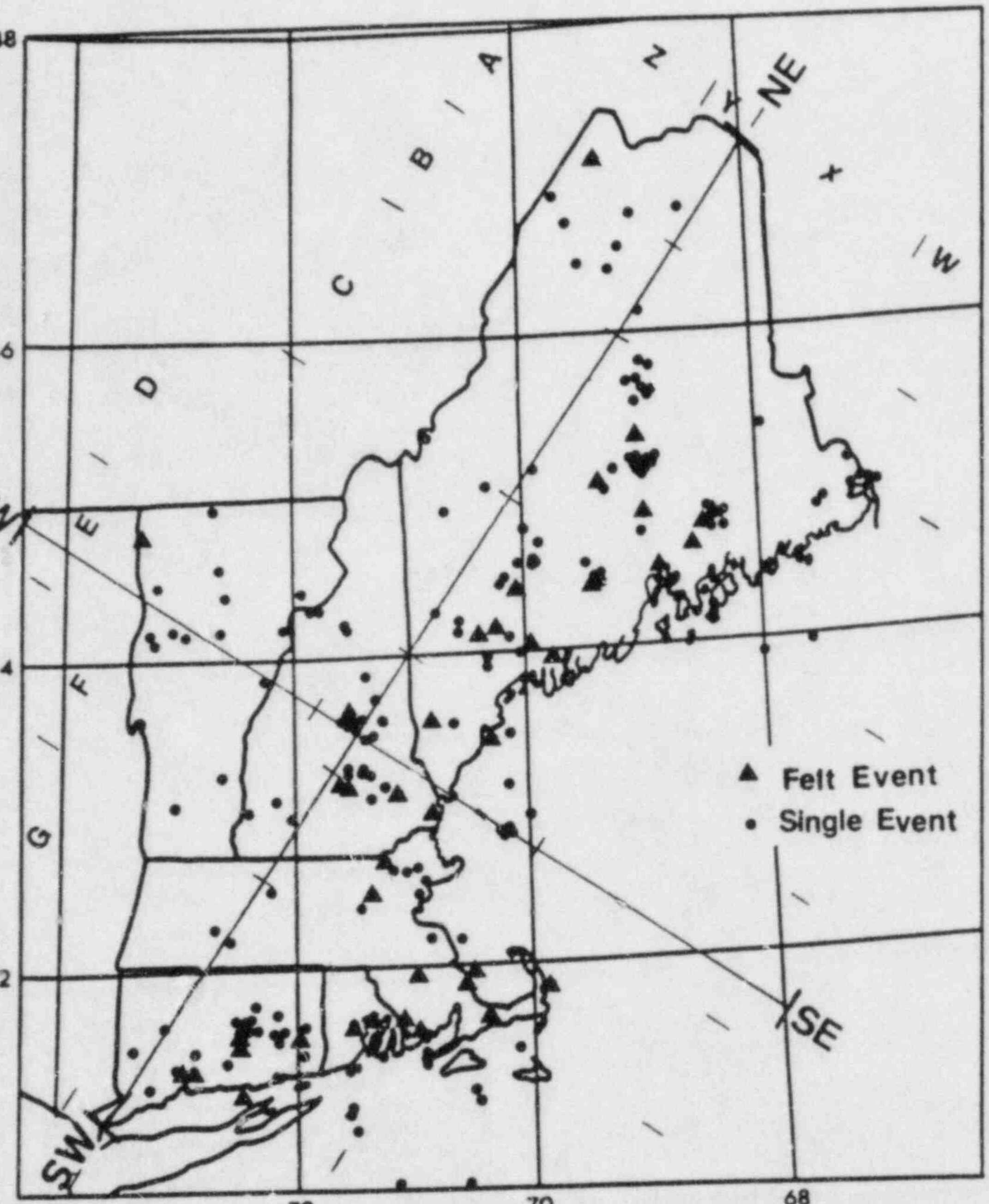

Figure 4.8 Map of the New England seismicity from October 1, 1975 to November 30, 1982.

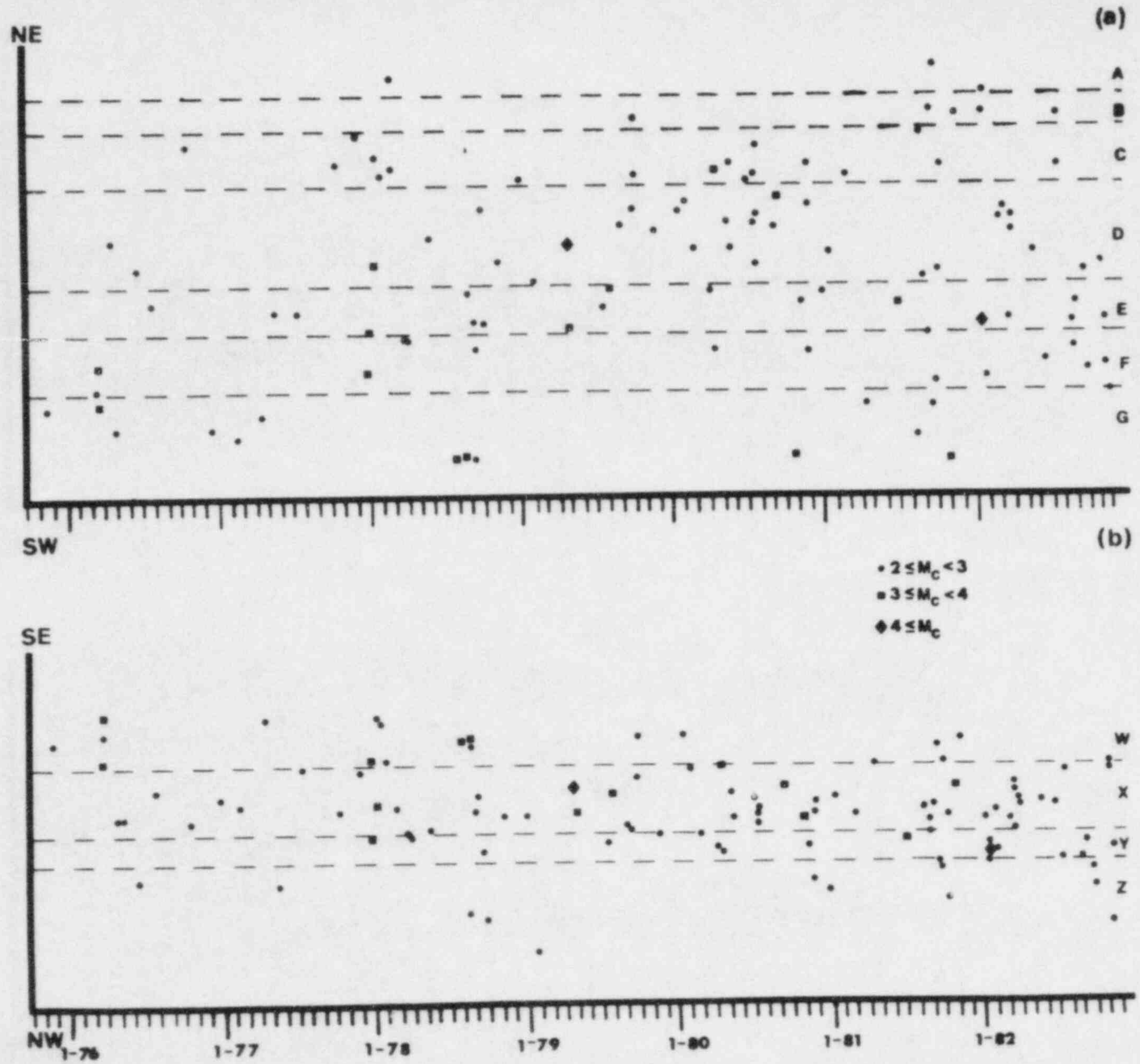

Figure 4.9 Space-time diagram of the seismicity shown in Figure 4.8. The earthquakes in (a) were projected onto the NE-SW line on Figure 4.8 while those in (b) were projected onto the NW-SE line. The regions A,B,C,D,E,F,G,W,X,Y and Z are indicated on Figure 3. This figure is from Ebel (1984).

4.4 Earthquake Return Times and Probabilities of Occurence

Chiburis (1981), using his historic catalog, and Ebel (1984), using the recent NEUSSN seismicity, calculated recurrence curves and return times for all of the northeast region and for New England respectively. The annual recurrence curves reported for all the northeast except La Malbaie, Quebec by Chiburis (1981) was

$$\log (N/yr) = 3.45 - 0.70\ M \qquad (2.8 \quad M \quad 5.2) \qquad (4.1)$$
$$\log (N/yr) = 13.36 - 2.60\ M \qquad (5.2 \quad M \quad 5.8) \qquad (4.2)$$

For the La Malbaie area, Chiburis (1981) reported

$$\log (N/yr) = 3.09 - 0.81\ M \qquad (4.3)$$

Ebel (1984) found for New England

$$\log (N/yr) = 2.87 - 0.84\ M_c \qquad (4.4)$$

The Chiburis (1981) curves were calculated from intensity data which were converted to earthquake magnitudes, while the Ebel (1984) curve was found exclusively from instrumental magnitude determinations.

Mean earthquake return times were calculated from the above equations and are given in Tables 4.1 and 4.2. Table 4.1, from Chiburis (1981), was found from equations 4.1, 4.2 and 4.3, whereas Table 4.2 was determined from Equation 4.4 above and reported by Ebel (1984). The values in Table 4.2 found by Ebel (1984) are compared also in Table 4.2 to the recurrence times for southern New England found by Chinnery (1979) and Shakal and Toksoz (1977) using historic data. The return times from the southern New England studies are greater than those for the entire New England region because they encompass a smaller area (roughly a factor of three less for the Shakal and Toksoz (1977) study area and about a factor of four smaller for that of Chinnery (1979)). The return times found for all of New England are 2 to 4 times less than those of Chinnery (1979) and are about 1.3 to 2.5 times less than those of Shakal and Toksoz (1977). Thus, the return times of earthquakes reported in the three studies are roughly comparable if all three results are normalized to the same area. Also for comparison in Table 4.2 there is a rough estimate of the site-specific seismic hazard in New England computed from a historic catalog of the entire eastern United States by McGuire (1977). In one calculation he used the assumption that seismicity of the eastern U.S. was spread uniformly over the area to determine the complete seismic hazard at any site. The seismic hazard values for different Modified Mercalli levels of ground motion found in Figure 9 of McGuire (1977) were inverted to find the return time of strong shaking at any individual site and were entered in Table 4.2. The analysis from McGuire (1977) shows that the return times of strong ground shaking at a particular site may be about an order of magnitude greater than the regional earthquake occurrence return times for earthquakes above intensity VII in New England.

TABLE 4.1

Recurrence Rates and Mean-Return-Times for La Malbaie Area, Quebec and the Rest of the Northeastern United States and Southeastern Canada

	Entire region, except La Malbaie area			La Malbaie area	
I_{MM}	M	N/yr	MRT	N/yr	MRT
	2.0	(115)	(3 days)	(30)	(12 days)
II	2.2	(83)	(4 days)	21	18 days
	2.5	(52)	(7 days)	12	31 days
III	2.8	32	11 days	7	54 days
	3.0	23	16 days	5	78 days
IV	3.4	12	30 days	2.2	163 days
	3.5	10	35 days	1.9	197 days
V	4.0	5	78 days	0.7	1.4 years
	4.5	2.1	174 days	0.3	3.5 years
VI	4.6	1.8	0.6 years	0.2	4.2 years
	5.0	0.9	1.1 years	0.1	8.8 years
VII	5.2	0.7	1.5 years		13 years
	5.5	0.1	8.8 years		22 years
VIII	5.8		53 years		39 years
	6.0		(175 years)		56 years
IX	6.4		(1923 years)		118 years
	6.5		(3500 years)		142 years
X	7.0		(70000 years)		(360 years)

Note: parenthetical values are extrapolations.

Where I_{MM} = Modified Mercalli intensity

M = Local magnitude, usually M_N or M_c

N/yr = Number of earthquakes per year

MRT = Mean return time

(From Chiburis, 1981)

Table 4.2

Return Times for New England Earthquakes

VI		VII		VIII		IX		X
4.5	5.0	5.2	5.5	5.8	6.0	6.4	6.5	7.0
This study								
10	21	31	55	98	144	312	379	994
Chinnery (1979) - Southern New England 1800 - 1959								
				229		891		3467
Shakal and Toksoz (1977) - Southern New England 1725 - 1974								
25		55		130				
McGuire (1977) - Uniform source area, entire eastern United States								
75		500		2000		7,800		39,000

The regional hazard, or probability of an earthquake in a particular time period was calculated by Ebel (1984) for New England and compiled in Table 4.3 The numbers show that the chances are roughly 1 in 3 of having a magnitude 5 earthquake in New England in a ten-year period, while they are about 1 in 2 of New England experiencing a magnitude 6 earthquake in one hundred years. This analysis, which once again is based upon the recent catalog only, indicates that the chances that a magnitude 7.0 (intensity X) earthquake could have occurred somewhere in New England during the last 255 years are about 1 in 5.

Table 4.3

New England Seismic Hazard

Values in Table represent probability of an earthquake of a particular magnitude in the specified time period.

Time (yrs)	1	7	10	50	100	200	500	1000
Magnitude								
4.6	.10	.51	.64	.99	1.00	1.00	1.00	1.00
5.0	.05	.28	.38	.91	.99	1.00	1.00	1.00
5.2	.03	.20	.28	.80	.96	1.00	1.00	1.00
5.5	.01	.12	.17	.60	.84	.97	1.00	1.00
5.8	.01	.07	.10	.40	.64	.87	.99	1.00
6.0	.01	.05	.07	.29	.50	.75	.97	1.00
6.4	.003	.02	.03	.15	.27	.47	.80	.96
6.5	.003	.02	.03	.12	.23	.41	.73	.93
7.0	.001	.01	.01	.05	.10	.18	.40	.63

5. DETAILED SEISMICITY STUDIES

5.1 Aftershock Study of the April 18, 1979 Bath, Maine Earthquake (M_c = 4.0).

An earthquake of M_c = 4.0 occurred on April 18, 1979 north of the town of Bath in coastal Maine within 12 km of the Maine Yankee nuclear power plant. This earthquake was immediately followed by a number of aftershocks that were recorded by the New England Seismic Network stations operated by Weston Observatory of Boston College and by the Massachusetts Institute of Technology. From late May until July 1979, Weston Observatory operated a portable seismic network in the epicentral area and recorded a number of additional microaftershocks.

The aftershocks recorded by the permanent stations immediately after the main shock were located relative to that event. This relative location analysis suggests that the main shock ruptured predominantly toward the southwest or west. The absolute locations of all but one of the microaftershocks recorded by the portable network were also computed. They were all located at between 3 and 7 km depth, near and possibly on the Cape Elizabeth fault (Figure 5.1). The data analysis is consistent with any of the following hypotheses: that the main shock and most of the aftershocks occurred on an ancient fault feature, the Cape Elizabeth fault; that the main shock occurred southeast of the Cape Elizabeth fault but caused micraftershocks to occur on the fault; or that the main shock coincidentally caused microaftershocks to occur near but not on the Cape Elizabeth fault. The last hypothesis notwithstanding, the microaftershock locations provide strong evidence of some modern seismicity being associated with a mapped fault in New England. However, the causative relationship between the seismicity and the fault was not clearly established by the study. This work was published by Ebel (1983).

5.2 Analysis of the Microearthquake Swarms Near Moodus, Connecticut

In 1980, 1981 and 1982, there occurred three microearthquake swarms near Moodus, Connecticut. The 1980 swarm consisted of 26 events with the largest being M_c = 0.8. The 1981 swarm lasted more than three months and was made up of over five hundred events with magnitudes greater than -2.0 (Ebel <u>et. al.</u>, 1982). The largest event of this swarmhad M_c = 2.1. The 1982 swarm also consisted of more than five hundred events of $M_c \geq -2.0$, and its largest event was M_c = 2.9. All of the swarms appear to have originated from the same location (shown on Figure 5.2) which was about eight kilometers northwest of the Connecticut Yankee nuclear power plant. One notable aspect of the hypocenters of all the

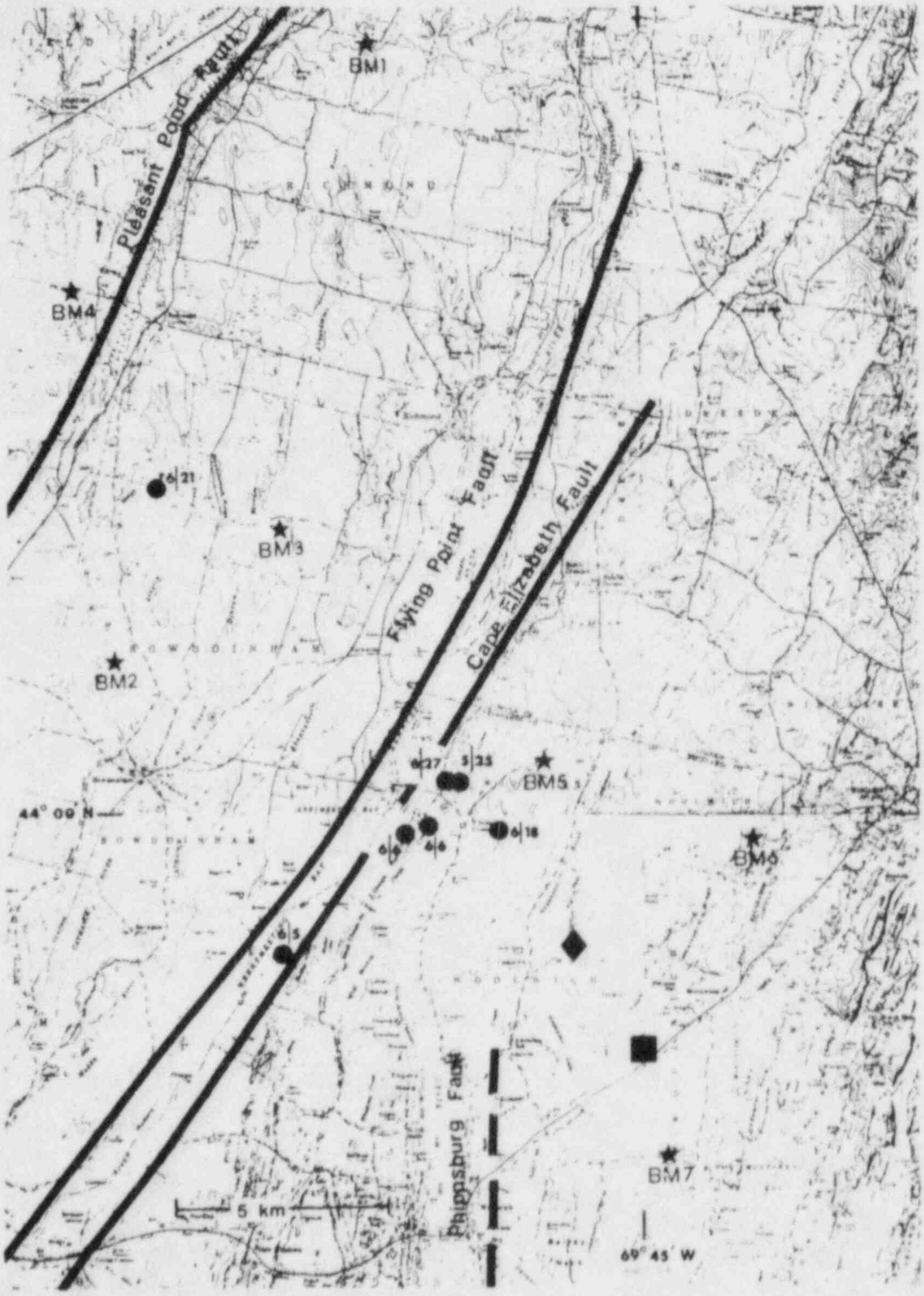

Figure 5.1 Absolute locations of the microaftershocks recorded by the portable seismic network operated in 1979 after the occurrence of the April 18, 1979 earthquake near Bath, Maine. The Cape Elizabeth and Phippsburg faults are from Hussey (1981). The location of the main shock which was used as the master event in the relative location analysis is shown as the large diamond, while the main shock epicenter from Chiburis and Ahner (1980) is denoted by the large square.

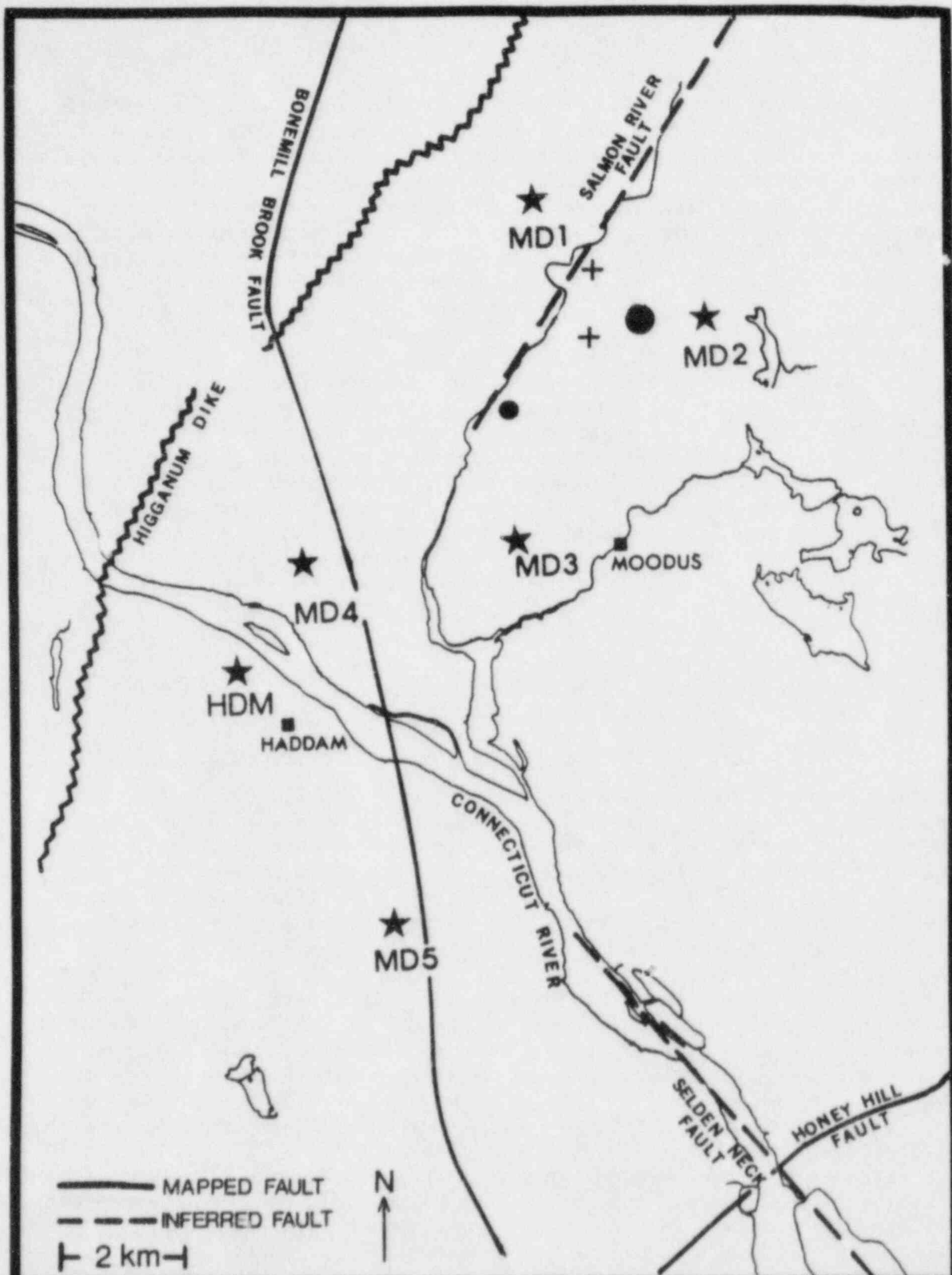

Figure 5.2 Generalized geologic map of the Moodus, Connecticut area showing the locations of the 1980, 1981 and 1982 swarms (large dot). The small dot is the location of the event resolveably to the southwest from the 1981 swarm (Ebel et. al., 1982). The stars show local seismic network stations while the crosses indicate places where portable stations were run. The geology is after Barosh (1980).

events was their very shallow depths. S-P times recorded for the events indicate that their depths are all about 1.0 km or less. The temporal histories of the numbers of events and energy releases of the two large swarms, shown in Figure 5.3 for the 1981 swarm and Figure 5.4 for the 1982 swarm, are similar. In both cases, the largest events were preceded by several tens of foreshocks and most of the events in the swarms took place within a few days of when the swarms started. The 1981 swarm differed primarily from that in 1982 in that it had several later bursts of activity, each of which was associated with the occurrence of a larger earthquake.

Recurrence curves for the foreshocks and aftershocks for the 1981 and 1982 swarms are shown in Figures 5.5 and 5.6 respectively. In both cases, the b values of the foreshocks were significantly lower than those for the aftershocks. The b values of the aftershocks, the same for the two swarms, were also lower than that for New England as a whole. Low foreshock b values have been proposed as a possible tool for predicting earthquakes (Rikitake, 1976), so the observation that the Moodus foreshocks have low b values may have important implications for identifying the imminence of future large earthquakes in the Moodus area.

5.3 Aftershock Study of the January, 1982, $M_c = 4.7$ Earthquake Near Gaza, New Hampshire

On January 19, 1982, an earthquake of $M_c = 4.7$ occurred approximately ten kilometers southwest of Laconia, New Hampshire, near Gaza, in the township of Sanbornton. The earthquake was felt throughout much of New England, eastern New York and into southern portions of the Canadian province of Quebec, inducing a maximum intensity of I = V(MM) in the epicentral region. The earthquake is significant as it represents the largest New England seismic event in nine years and the largest in this historically active part of New Hampshire since the 1940, I = VII(MM) Ossipee earthquakes. Further importance must be attached to this event as it is the first main shock in the eastern United States to trigger a significant number of strong motion instruments. Thirteen instruments at seven different locations in New Hampshire and Vermont were triggered by the event, including three instruments at Franklin Falls, which was about 7 km from the epicenter. The main shock was followed by a sequence of aftershocks which was monitored by a network of portable seismographs installed in the area from January 19 through February 17. The portable network was operated cooperatively by Weston Observatory, the Massachusetts Institute of Technology and Weston Geophysical Research, Inc. Over fifty aftershocks were recorded on the portable instruments, including two which were of a sufficient magnitude to be detected by permanent stations of the New England seismic network. These two events allowed the calculation of locations for the main shock and the two $M_c = 2.6$ aftershocks of January 19 which are relative to aftershock locations calculated using only portable network data.

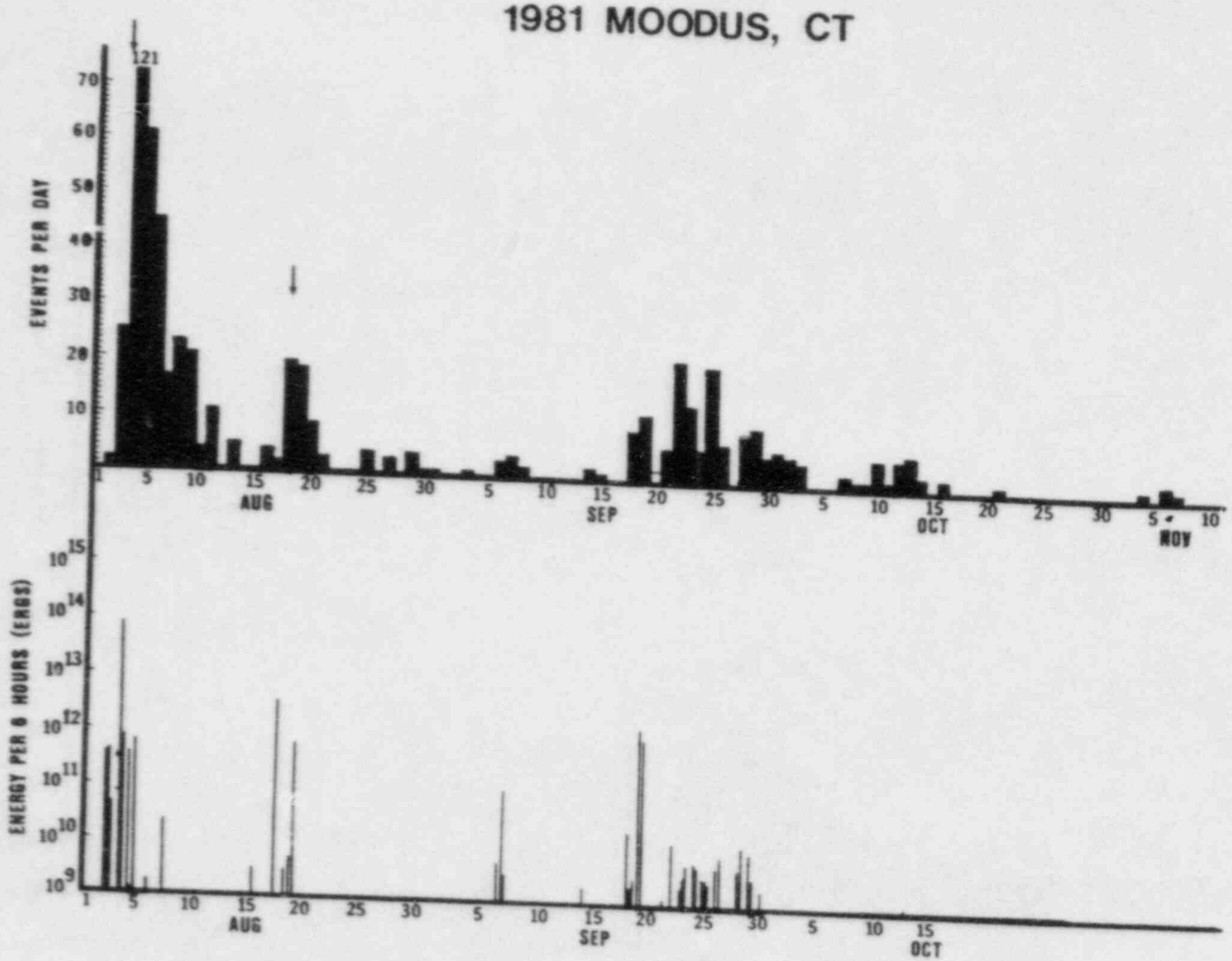

Figure 5.3 Time history of the 1981 swarm. The top histogram shows the number of events per day while the bottom histogram shows the energy release per 6 hour intervals.

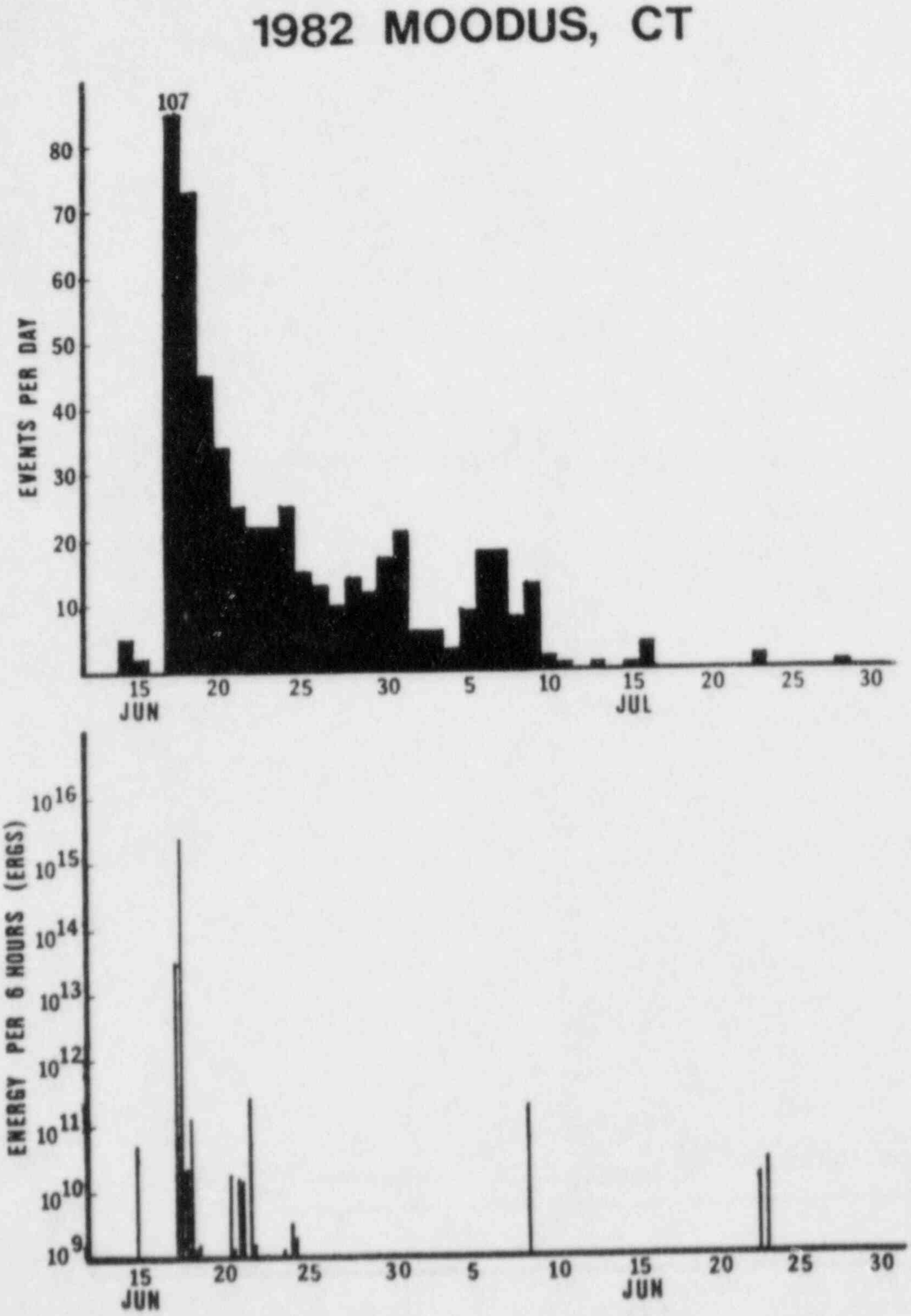

Figure 5.4 The same as Figure 5.3 but for the 1982 swarm.

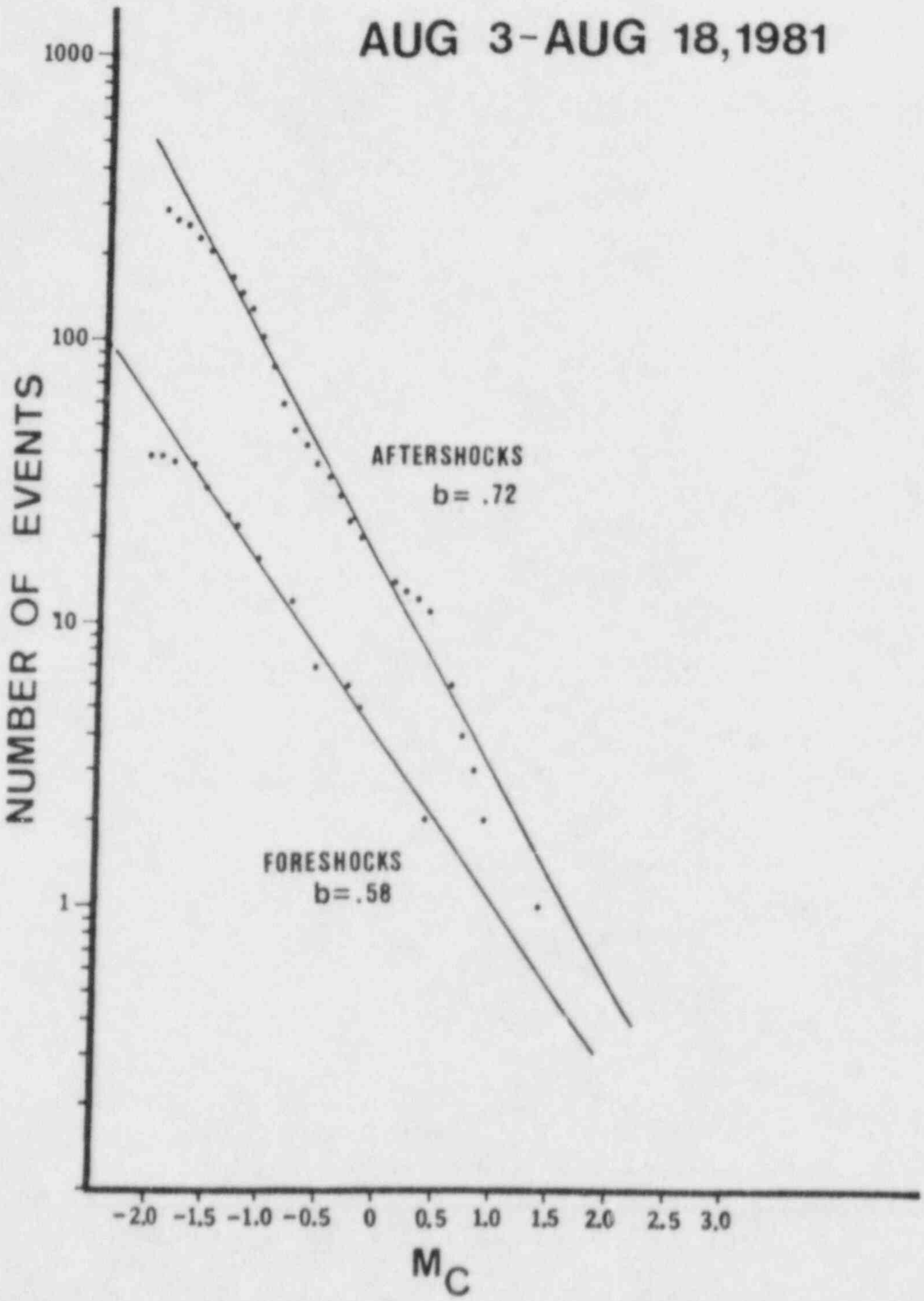

Figure 5.5 Recurrence curves for the 1981 Moodus, CT swarm.

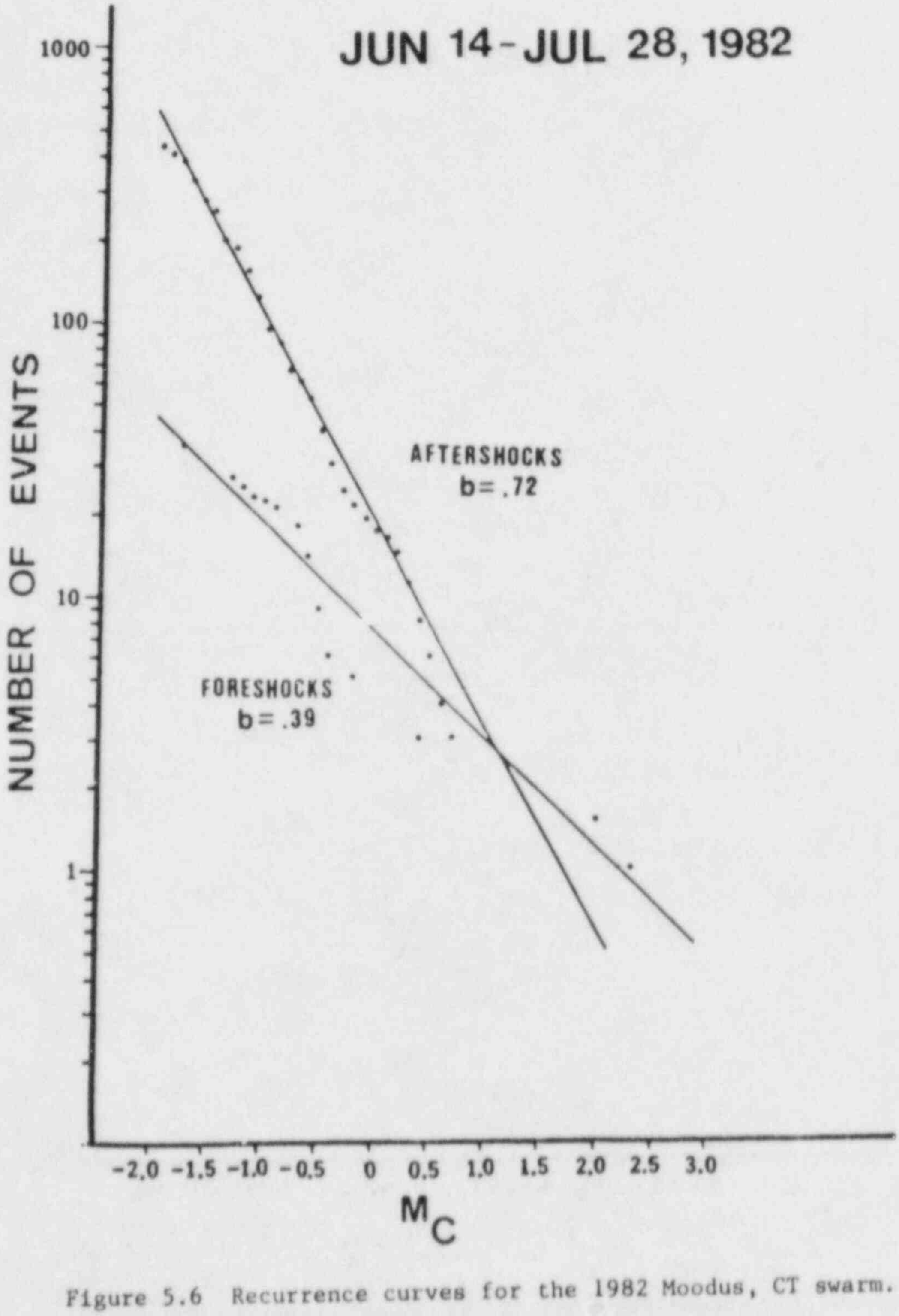

Figure 5.6 Recurrence curves for the 1982 Moodus, CT swarm.

Arrival time data for these events indicate a resolvably northward progression of epicentral location with time (Figure 5.7). In addition, a cluster of generally shallow, small magnitude events occurred in an area to the north of the January 19 activity (Figure 5.8). This northerly migration of seismicity is consistent with geological and geophysical studies conducted in the area by Englund (1976), McHone (1978) and Gresko (1980) which suggest a potential for north-south trending basement structures in the region along with the presence of shallow, multidirectional fracture patterns (Figure 5.9).

5.4 Aftershock Study of the May, 1983 M_c = 4.4 Earthquake Near Dixfield, Maine

An earthquake of M_c = 4.4 occurred southeast of Dixfield, Maine on May 29, 1983. The event was widely felt in Maine and New Hampshire and caused a maximum epicentral intensity shaking of V (MM). Five aftershocks smaller than M_c = 1.0 were recorded by portable seismographs installed after the main shock. One of the aftershocks was well-located at a depth 2.4 ± .5 km using the portable seismograph data, while the depth of the main shock was computed to be 1.8 ± 1.4 km from the regional network data. Since the epicenters of the well-located aftershock and the main shock coincided, it is likely that these events originated from the same location. First motion polarities from the main shock suggest reverse motion on north-south oriented fault planes (Figure 5.10). The earthquakes occurred about 2 km west of the northeast-striking Bald Moutain fault (Figure 5.11), but they cannot be clearly attributed to that or any other recognized brittle structure in the area. This study was published by Ebel and McCaffrey (1984).

5.5 Analysis of the Earthquakes Near Passamaquoddy Bay, Maine

The occurrence of five felt earthquakes (all $M_c \geq 3.0$) from August, 1983 through January, 1984 in eastern Maine near the New Brunswick border prompted a reexamination of the earthquake activity of that region. A search of the historic seismicity showed frequent small earthquakes throughout the 111 year historic record as well as sporadic episodes of stronger seismicity. The largest of the events from this region to date occurred on March 21, 1904. During this event intensity VIII (MM) shaking was experienced near the Calais, Maine epicenter and intensity VII (MM) motions were felt at St. John, New Brunswick 85 km away. A period of increased seismicity took place in the 1920's when six events of intensity III or greater occurred.

The two predominant tectonic features in the area are the Passamaquoddy Bay which is subsiding at a maximum rate of about 9 mm/year and the Oak Bay Fault which trends NW and is nearly perpendicular to the subsidence axis. The recent five events occurred at four separate epicenters, all of which were west of Passamaquoddy Bay and south of the Oak Bay Fault (Figure 5.12). The August 12 (M_N = 3.6) event was located

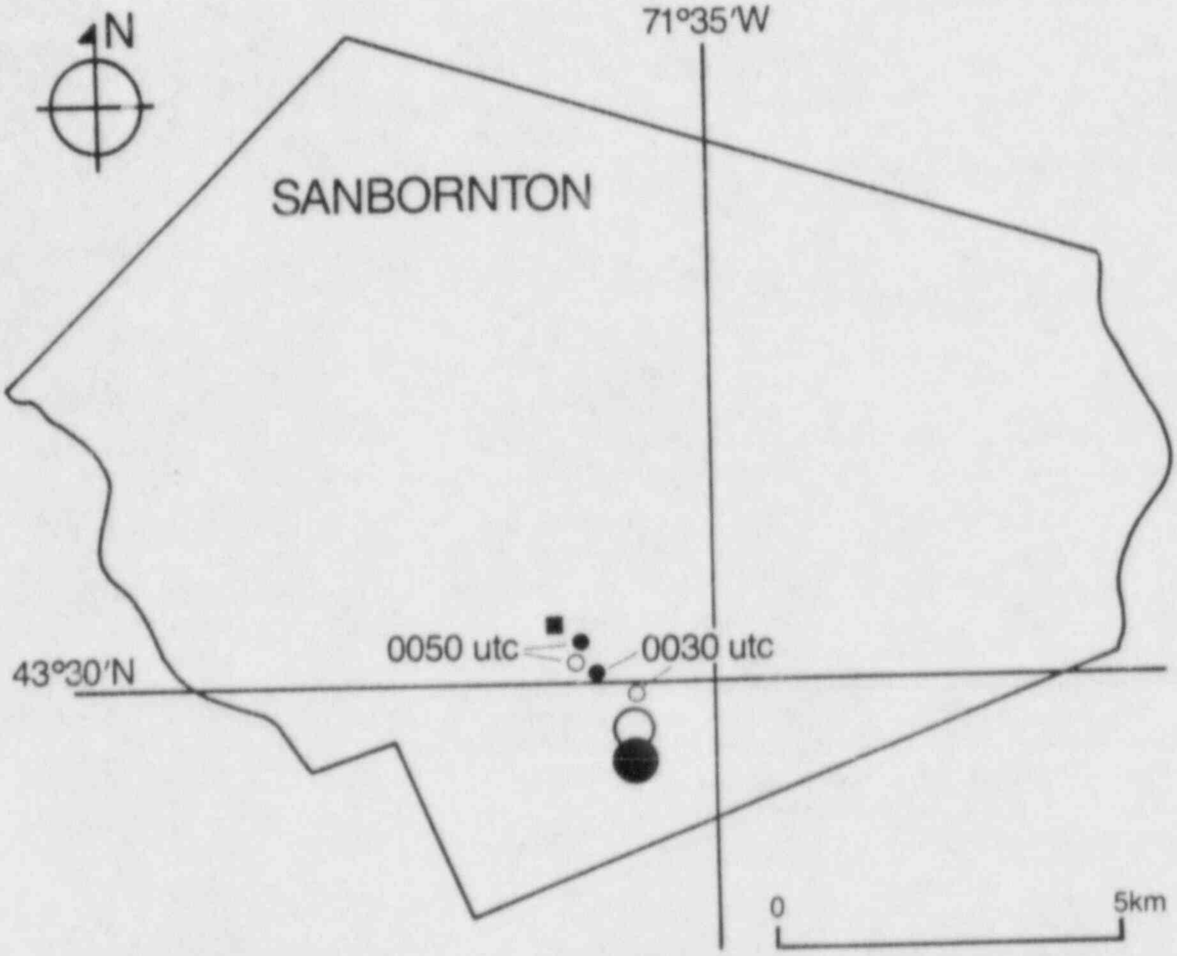

Figure 5.7 Map of Sanbornton, New Hampshire showing the relative locations determined for the main shock and the immediate aftershocks. Open circles represent relative locations determined with the use of arrival time at station BNH. Closed circles represent relative locations with data from BNH omitted. The calculated epicenter for the master event is marked with a square and the main shock epicenters are indicated by the larger circles.

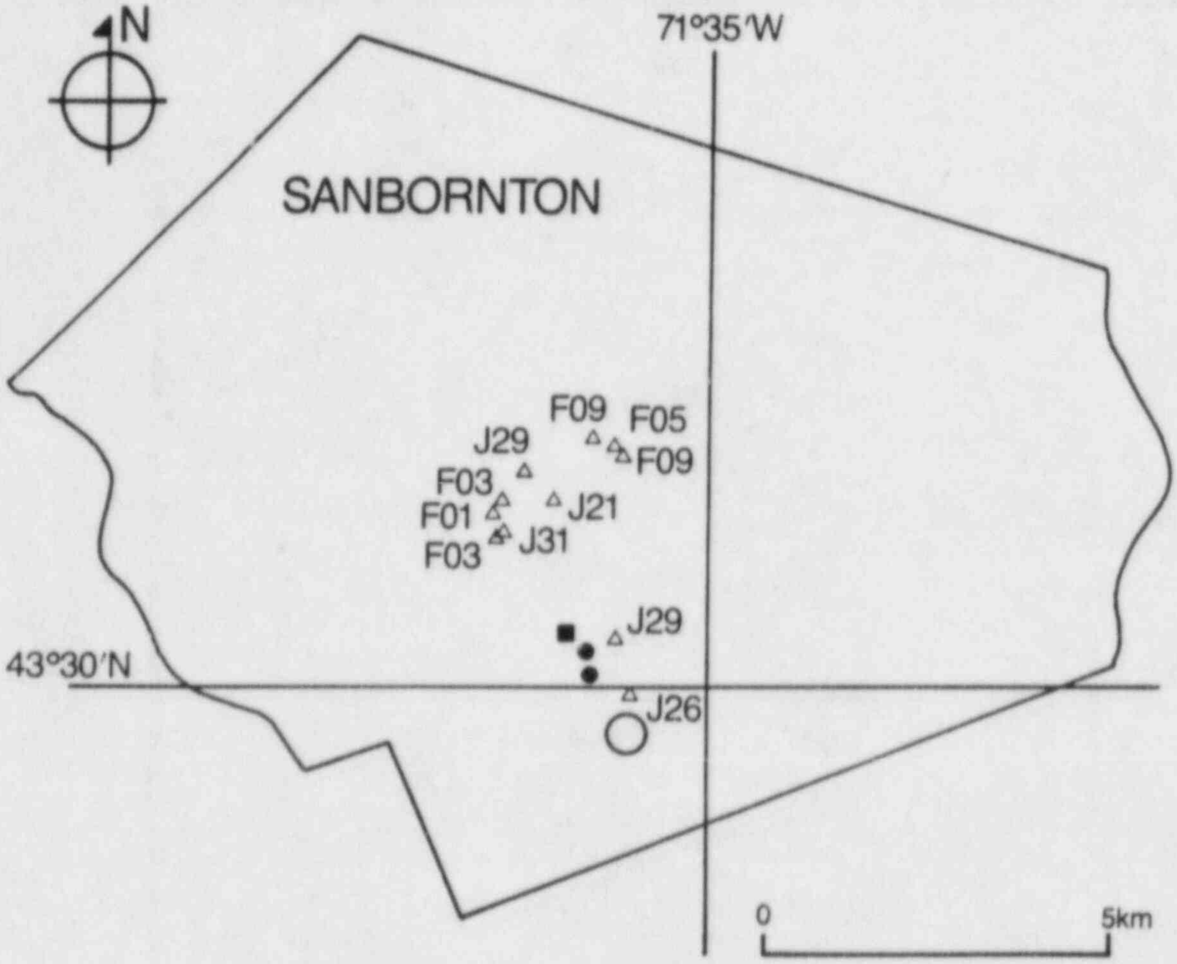

Figure 5.8 The town of Sanbornton, New Hampshire as in Figure 3. Epicenters for microaftershocks are marked with a triangle. The dates for the microaftershocks are indicated with J denoting January and F denoting February.

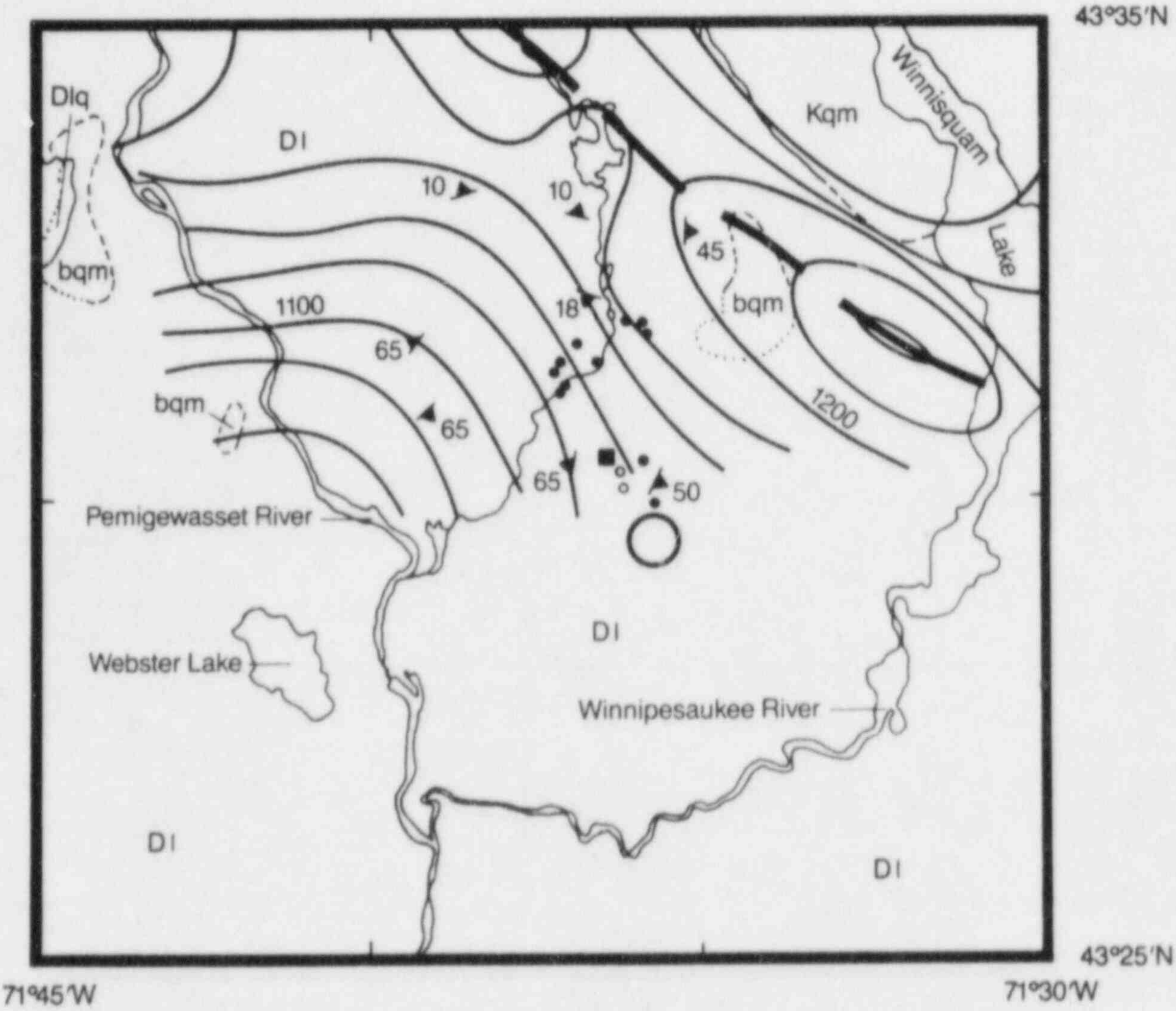

Figure 5.9 The epicenters of the 1982 earthquakes plotted on a generalized map of the geology from Englund (1976) and Gresko (1980). The large circle show the main shock epicenter, the solid square the master event location and the open and closed circles the aftershock locations. Contours of aeromagnetic anomalies in gammas are shown as is the trend of a magnetic anomaly from from Gresko (1980). The geologic symbols are: Dl and Dlg - the Littleton formation, bqm and Kqm - the New Hampshire plutonic series, solid triangles and numbers - attitudes and values of local dips.

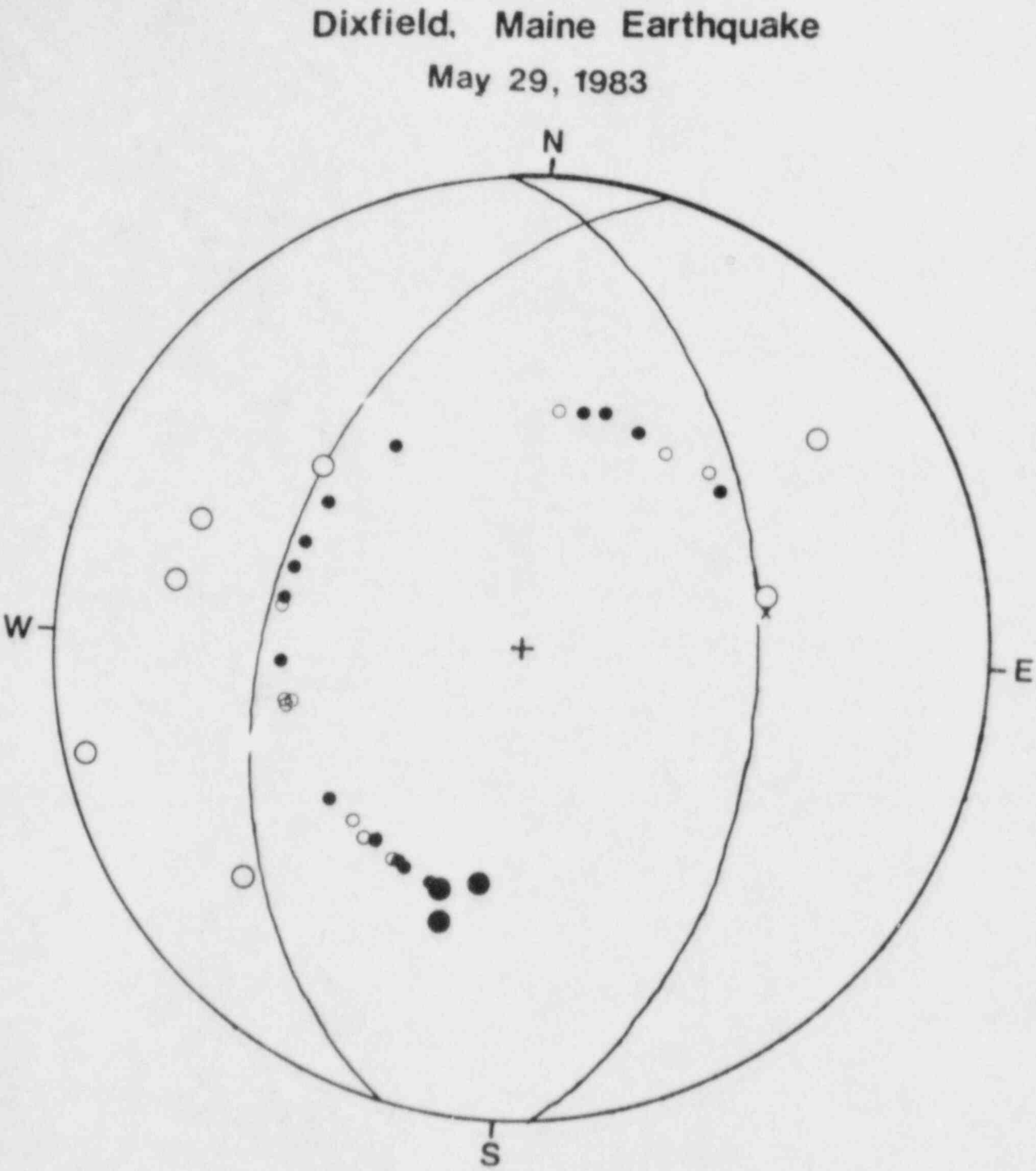

Figure 5.10 Fault plane solution for the Dixfield, Maine earthquake. Open circles show dilitations, closed circles show compressions, and the X indicates nodal arrivals at PQ0 and PQ1. The larger circles are from stations at less than 250 km from the earthquake, while the small circles are from stations at distances greater than 250 km.

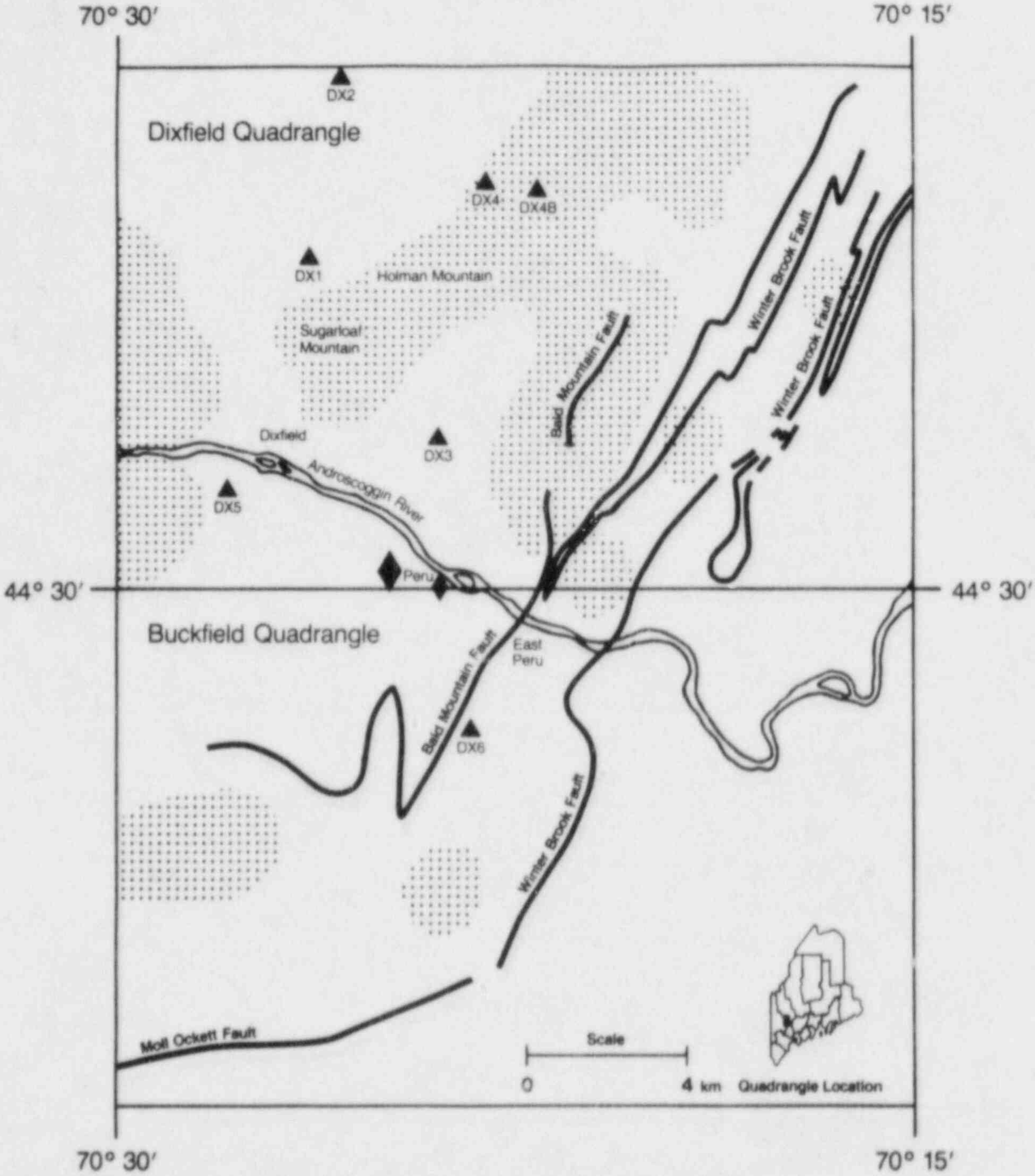

Figure 5.11 Map the the 1983 earthquake activity near Dixfield, Maine. The large diamond shows the location of the main shock, the small diamond the location of the well-located aftershock, and the triangles the sites of the portable instruments. The geology is generalized from Pankiwskyj (1981); the shaded areas are places of locally high topography.

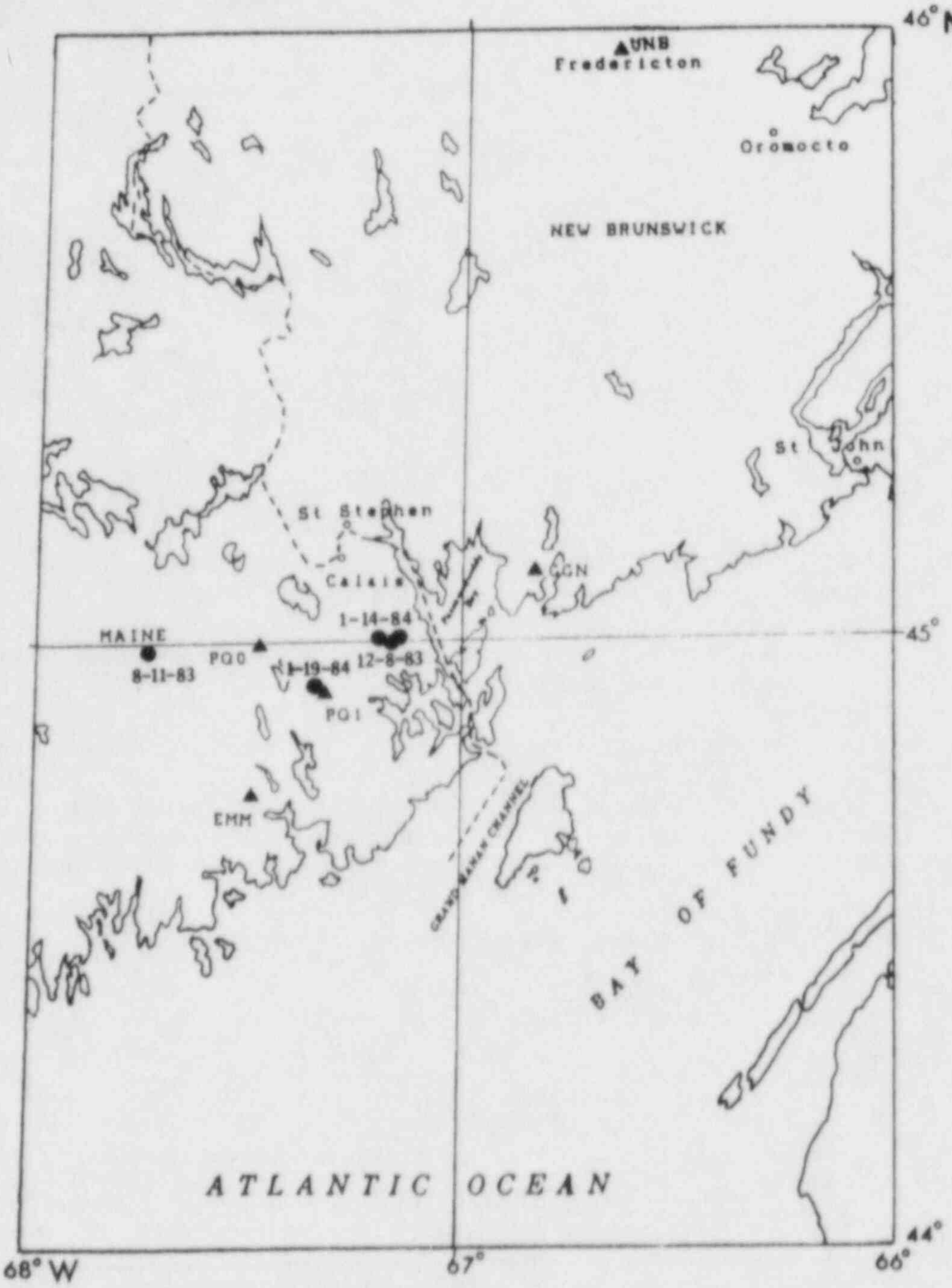

Figure 5.12 Map of the earthquakes near Passamaquoddy Bay, Maine in late 1983 and early 1984. The triangles show the locations of seismic stations and the closed circles with dates show the locations the the earthquakes.

50 km west of the bay while the December 8 (M_N = 3.2) even was located 35 km more to the east of the August 12 event. The January '4 earthquakes (M_N = 3.4 and 3.6) had origin times less than a minute apart and were located 5 km east of the December event. The latest and largest event was on January 19 (M_N = 3.8) and had an epicenter 20 km southwest of the January 14 events. Although poorly constrained, a fault plane solution of this last event indicates a thrust mechanism striking WNW (Figure 5.13). The ground acceleration of this earthquake did not trigger two accelerometers set at a 0.01 g threshold, each of which was located less that 20 km from the epicenter.

5.6 Analysis of the Earthquake Activity of the Greater New York City Area.

Kafka et al. (1985) reexamined earthquake data recorded by stations of the Lamont-Doherty seismic network in the greater New York City area to determine magnitudes and the relationship between seismicity and geologic structures. Between 1974 and 1983 the configuration of stations in this region remained relatively constant and the type of recording devices (visual drum recorders and 16 mm photographic recorders) did not change. This distribution of stations and stability recording devices allowed for a uniform measurement of seismicity and magnitudes. Magnitudes of these earthquakes were determined by comparing amplitudes and signal durations measured from high-frequency (5-10 Hz) data recorded by the local network with M_{bLg} and M_L determined from data at frequencies near 1 Hz. During the period of time studied (nearly ten years) sixty-one earthquakes were located in this region, but none of these earthquakes exceeded 3.0 on the M_{bLg} scale. The largest event (M_{bLg} = 3.0) occurred in the Coastal Plain province of northern New Jersey.

The magnitude threshold for uniform detection of events throughout this region during the period of time studied is estimated to be M_{bLg} = 1.6. With events below this threshold removed from the catalogue of network seismicity, it was found that about half of the earthquakes studied occurred within 10 km of the Ramapo fault system. This fault system lies about 30 km northwest of New York City and has been interpreted by several investigators to be the most active fault system in the greater New York City area (Aggarwal and Sykes, 1978). However, earthquakes at least as large as those recorded near the Ramapo fault were located as far as 50 km from this fault (and within 20 km of New York City) in geologic structures that surround the Newark basin (Figure 5.14). While the Ramapo fault can by no means be ruled out as a possible source zone for earthquakes in the greater New York City area, the geologic structures associated with most (if not all) earthquakes in this region remain unknown.

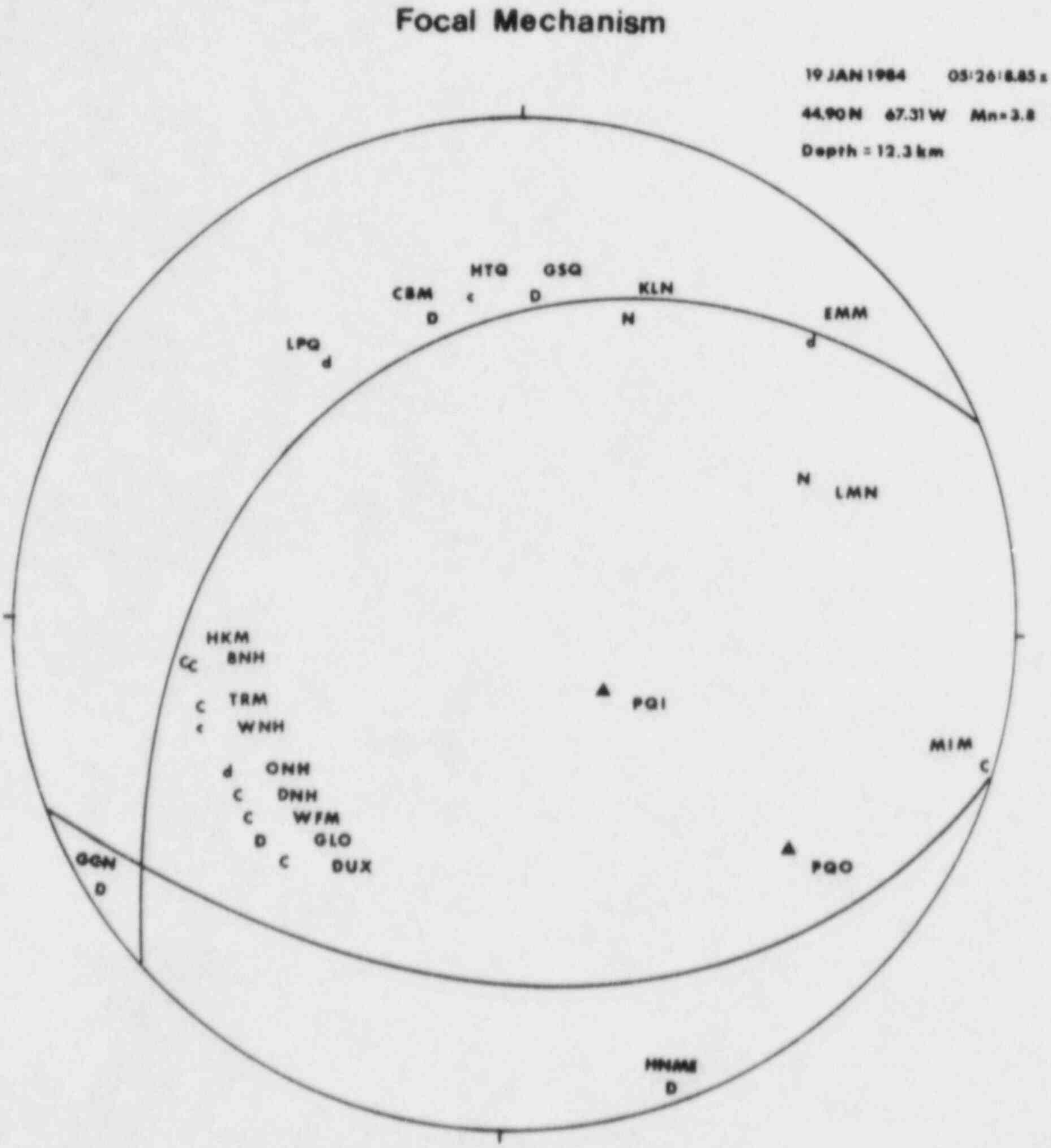

Figure 5.13 Focal mechanism (lower hemisphere) of the January 19, 1984 earthquake near Passamaquoddy Bay, Maine. C represents compressional arrivals, D represents dilitational arrivals and the lower case letters indicate that readings were of a poorer quality.

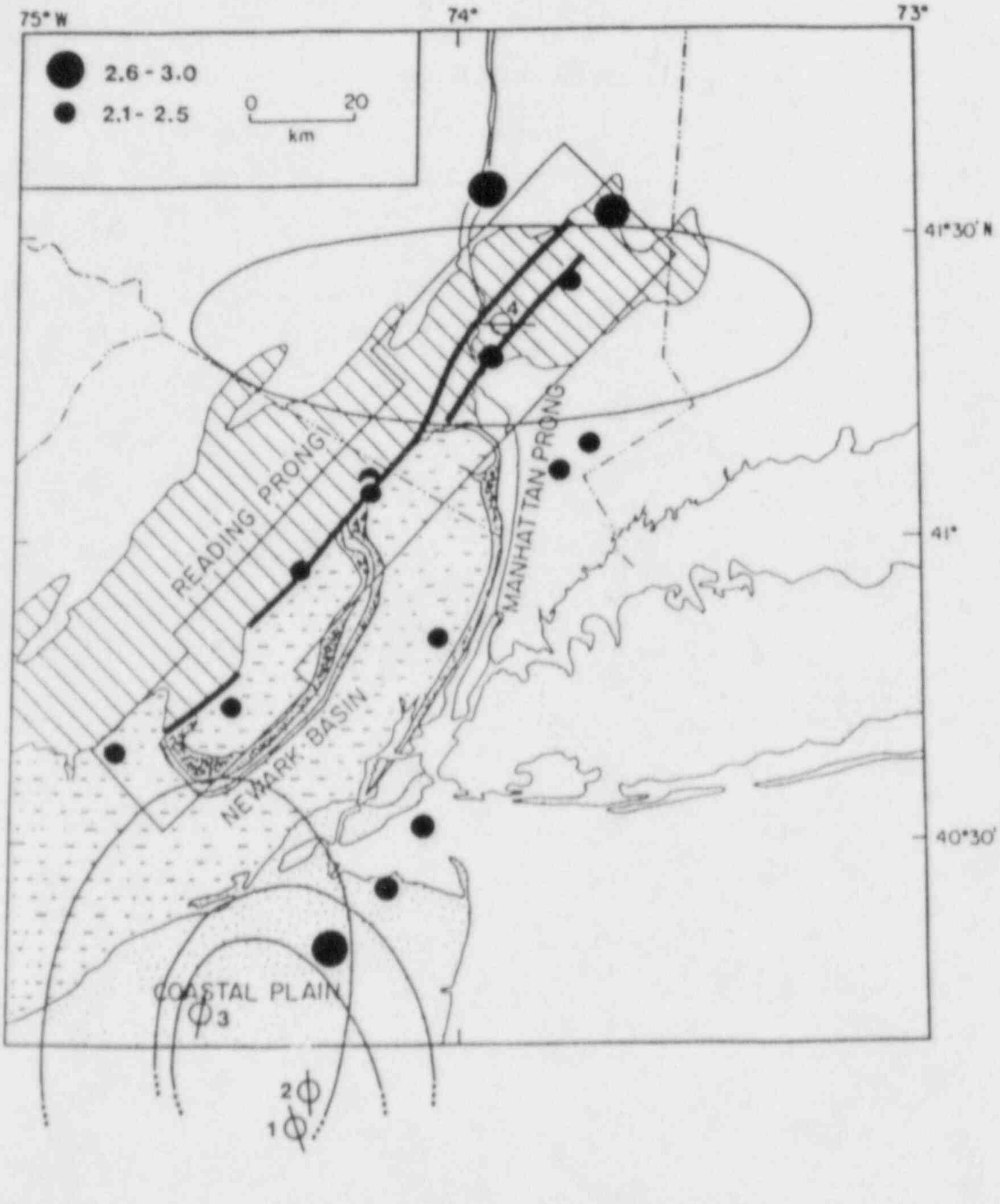

Figure 5.14 Map of the seismicity of the New York City area (from Kafka et. al., 1985). The closed circles show epicenters from January 1974 to September, 1983. The open circles labeled 1, 2 and 3 occurred on August 23, 1938 and were located by Dewey and Gordon (1984) (the error ellipses of each event are indicated). The open circle labeled 4 which took place on August 23, 1938 and its error ellipse are also shown (Dewey and Gordon, 1984).

5.7 Other Aftershock Studies

There were three occasions where portable seismographs were deployed but where no aftershocks were recorded. The three events for which this took place were the Long Island Sound earthquake of October 21, 1981 (M_c = 3.5), the Taunton, Massachusetts earthquake of January 27, 1982 (M_N = 3.0) and the Manchester, New Hampshire earthquake of March 24, 1983 (M_c = 2.9). The Long Island Sound earthquake occurred well offshore, so it was not possible to locate the portable instruments closer than about 20 km from the epicenter. The instruments were operated for about four days after that event. In the cases of the other two earthquakes, two days of recording within 5 km of the epicenters did not reveal any aftershocks. It was not surprising that no aftershocks were detected considering the small sizes of these three earthquakes.

6. WAVE PROPAGATION STUDIES

6.1 Travel Time Studies

A number of experiments to measure the crustal seismic velocity structure were undertaken during the contract period. Most of these experiments utilized travel-time measurements from natural and man-made sources to infer the average seismic structure of a part of New England. Chiburis and Graham (1978) used travel-time measurements in Connecticut to obtain a velocity model for that region (Graham and Chiburis, 1980; Chiburis and Ahner, 1980). Breunig (1980) performed a refraction experiment in Vermont and New Hampshire to find crustal thickness and velocities in that region. Eichorn (1980) used a number of travel-time determinations from paths crisscrossing Maine in calculating an average curstal velocity structure for that region. All of these models, as well as some determined by other investigators for parts of the region, are listed in Table 6.1.

Throughout the contract period Weston Observatory used the Chiburis and Ahner (1980) crustal model for Connecticut to locate all earthquakes in all of New England. It was found through trial-and-error testing that use of other models did not improve the location errors of any event, and in some cases the model appropriate for the region where the event occurred gave a larger location error than the Chiburis and Ahner (1980) model. The reason for this apparently is twofold. First, the Chiburis and Ahner (1980) model seems to be appropriate for the coastal and near-coastal regions in New England where many of the earthquakes have occurred. Second, the relatively large station spacing in the region (typically 50 km or more) means that even though an earthquake occurred in a region where one crustal model is appropriate, many of the recording stations lie atop crust which is best described by another of the velocity models. And again, most of the stations are situated in locations below which the Chiburis and Ahner (1980) model appears to describe the structure fairly well. Thus, this latter point means that all models which are a good average for the region should yield approximately the same location errors. This is the case observed for the data from New England.

6.2 Receiver Structure Studies

The existence of long-period and short-period, three component WWSSN seismographs at Weston Observatory (WES) provided a unique opportunity to study the crustal structure directly under WES. This study, reported by Foley (1984), involved the calculation of long-period synthetic seismograms for teleseisms following the technique of Langston (1977) and which were then compared to data collected at WES. The crustal structure used to compute the synthetics was perturbed and the

TABLE 6.1

New England Crustal Models Determined by Weston Observatory

V_p km/sec	Depth to Top km	Thickness km	Region	Reference
5.99	0.00	2.92	Vermont	Breunig (1980)
6.61	2.92	38.54		
8.21	41.45			
5.31	0.00	0.88	Connecticut	Chiburis and Ahner (1980)
6.06	0.88	12.21		
6.59	13.09	21.51		
8.10	34.60			
6.03	0.00	12	Maine	Eichorn (1980)
6.60	12.00	21		
6.73	34.00	5		
7.20	39.00			
6.00	0.00	8.10	Weston, MA	Foley (1984)
6.60	8.10	21.00		
8.10	29.10			
3.46	0.00	0.53	Southeast New England	Welch (1982)
3.51	0.53	14.17		
3.84	14.70	22.30		
4.80	37.00			

comparisons repeated until the crustal model which gave synthetics which best fit the data was found (Figure 6.1). It was clearly shown that the crust beneath WES is different than that described in the Chiburis and Ahner (1980) and Taylor and Toksoz (1979) models. The best fitting crustal model, which is listed in Table 6.1, has a 29.1 km thick crust with an 8.1 km thick layer upper crust (v_p = 6.0 km/sec) and a 21.0 km thick lower crust (v_p = 6.6 km/sec). The best fitting model is different from all refraction results developed for New England and may reflect the unique location of WES just east of Bloody Bluff Fault, a proposed Devonian suture of the Avalon block to the North American Plate. Since this was the first crustal structure study done entirely east of this boundary, the results may be displaying a previously unknown difference in crustal thickness across the Bloody Bluff Fault.

Another study of the teleseismic receiver structure of the Weston Observatory seismic network was performed by Monahan (1981). In this study, he measured relative arrival time residuals of teleseisms across the network (Figure 6.2). In Figure 6.2 it can be seen that the residual relative to station MIM are as great as 0.4 sec, suggesting differences in the velocity structures beneath the stations. Monahan (1981) showed that all of these residuals could be accounted for by the differences in the structure of the crust reported in the travel time studies of New England (i.e. Table 6.1).

6.3 Synthetic Seismogram Studies of Body Waves

The initiation of the acquisition of digital data provided the impetus to begin to study the waveforms of the earthquakes and blasts being recorded. The goal of this research is ultimately to be able to use recorded waveforms of earthquakes to determine details of their source properties. As a first step toward this goal, calculations of synthetic seismograms for some possible crustal models for New England were made. Figures 6.3, 6.4 and 6.5 show profiles of the synthetic seismograms of the P waves for three different models. Differences are visible the waveforms among the different models, thus indicating that the waveforms contain important information about the structure through which the waves have propagated. Preliminary comparisons of these synthetics with data have revealed some qualitative similarities (Nutting, 1984). Refinements of the crustal models of New England and calculation of the velocity structure under individual stations should be possible with further research using this technique.

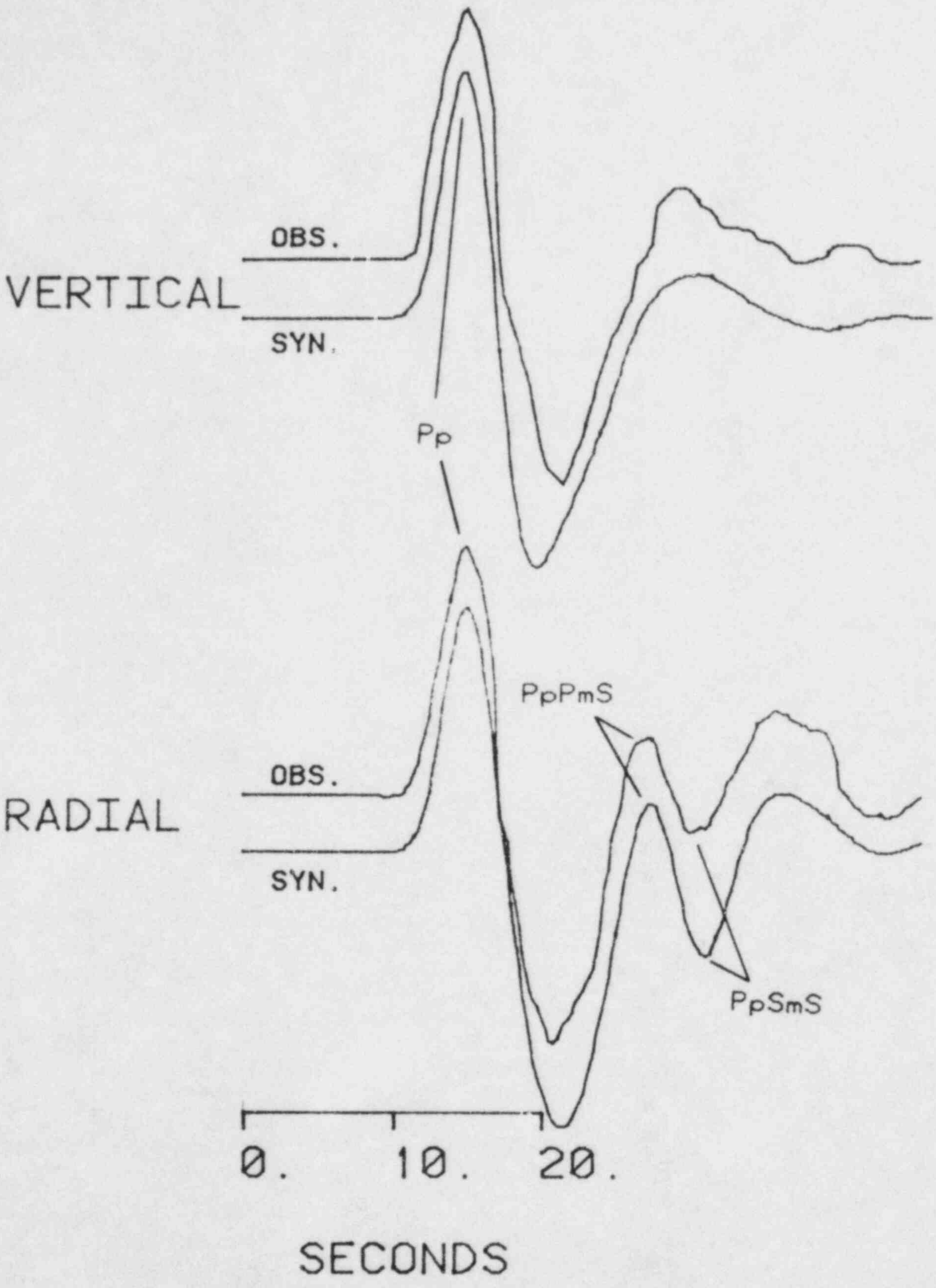

Figure 6.1 Comparison of synthetics (bottom) with observations (top) of the P wave of an earthquake from Japan recorded at Weston Observatory. The vertical and radial components are indicated as are the arrival times of some important near-receiver phases.

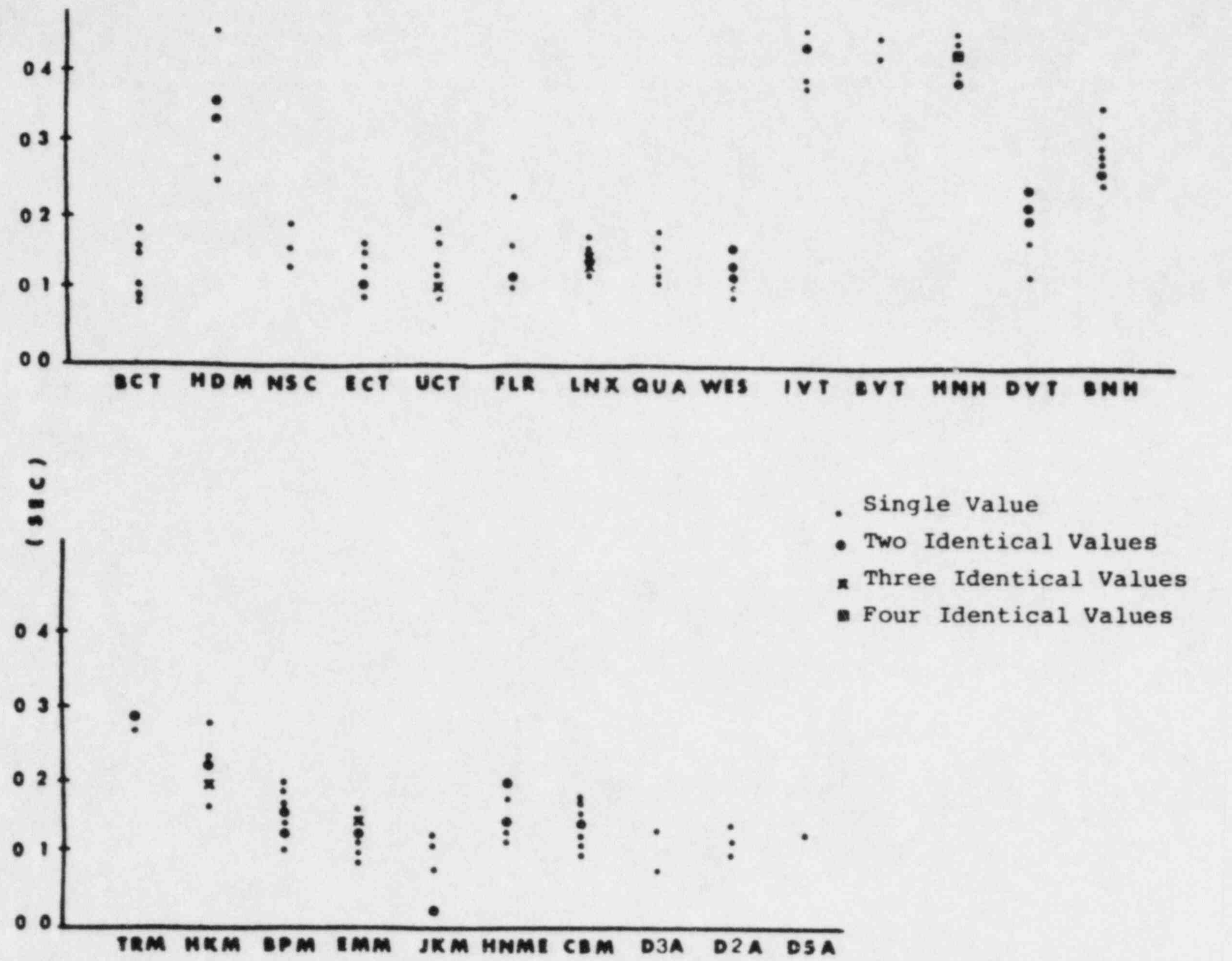

Figure 6.2 Teleseismic travel time residuals relative to station MIM for the Weston Observatory New England seismic network.

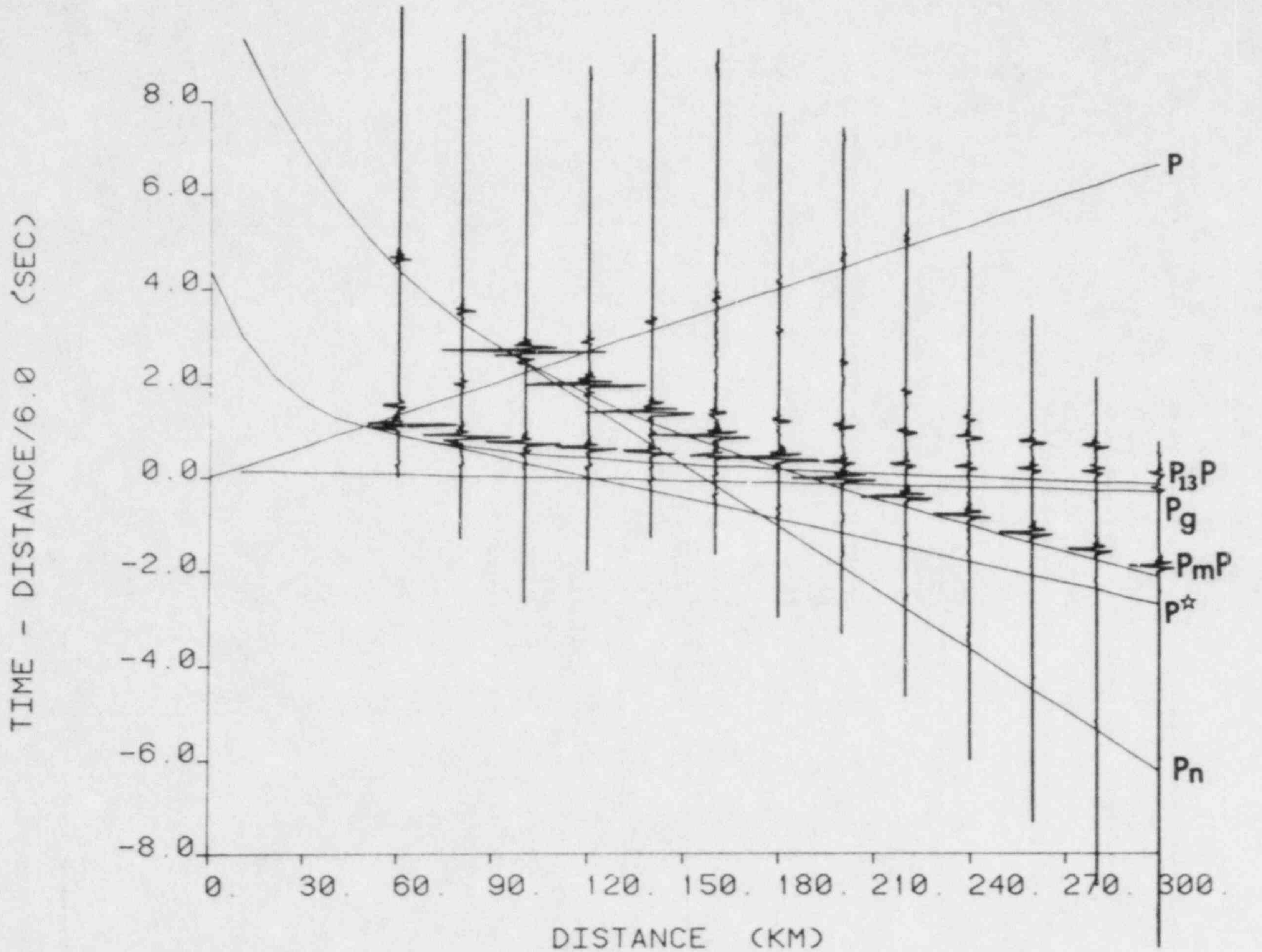

Figure 6.3 Record section of synthetic seismograms calculated from the crustal model of Chiburis and Ahner (1980). A delta function source was used as was the instrument response of the Weston Observatory New England seismic network. Prominent phases on the section are indicated.

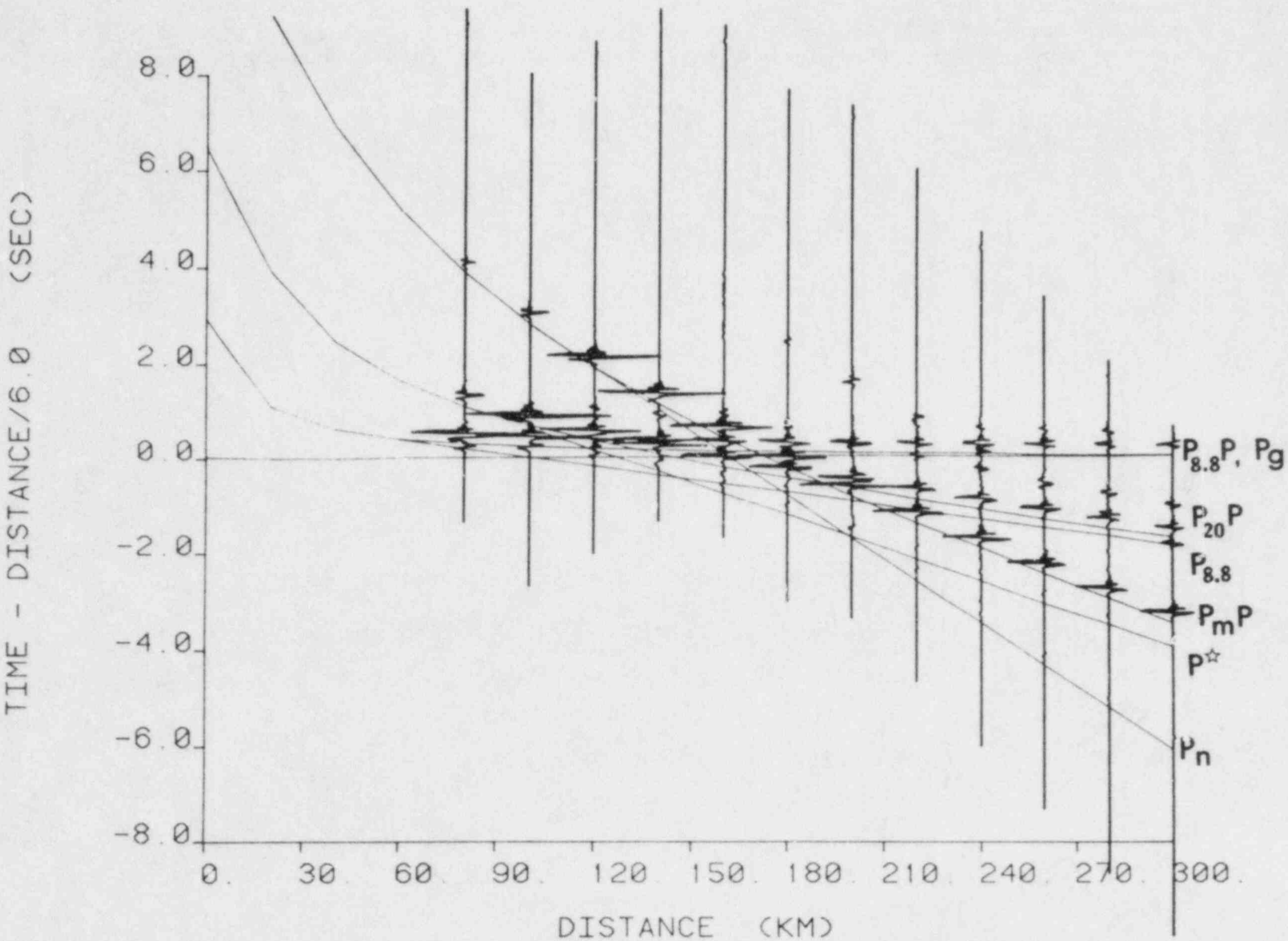

Figure 6.4 The same as Figure 6.3 but for the crustal model of Taylor and Toksoz (1979).

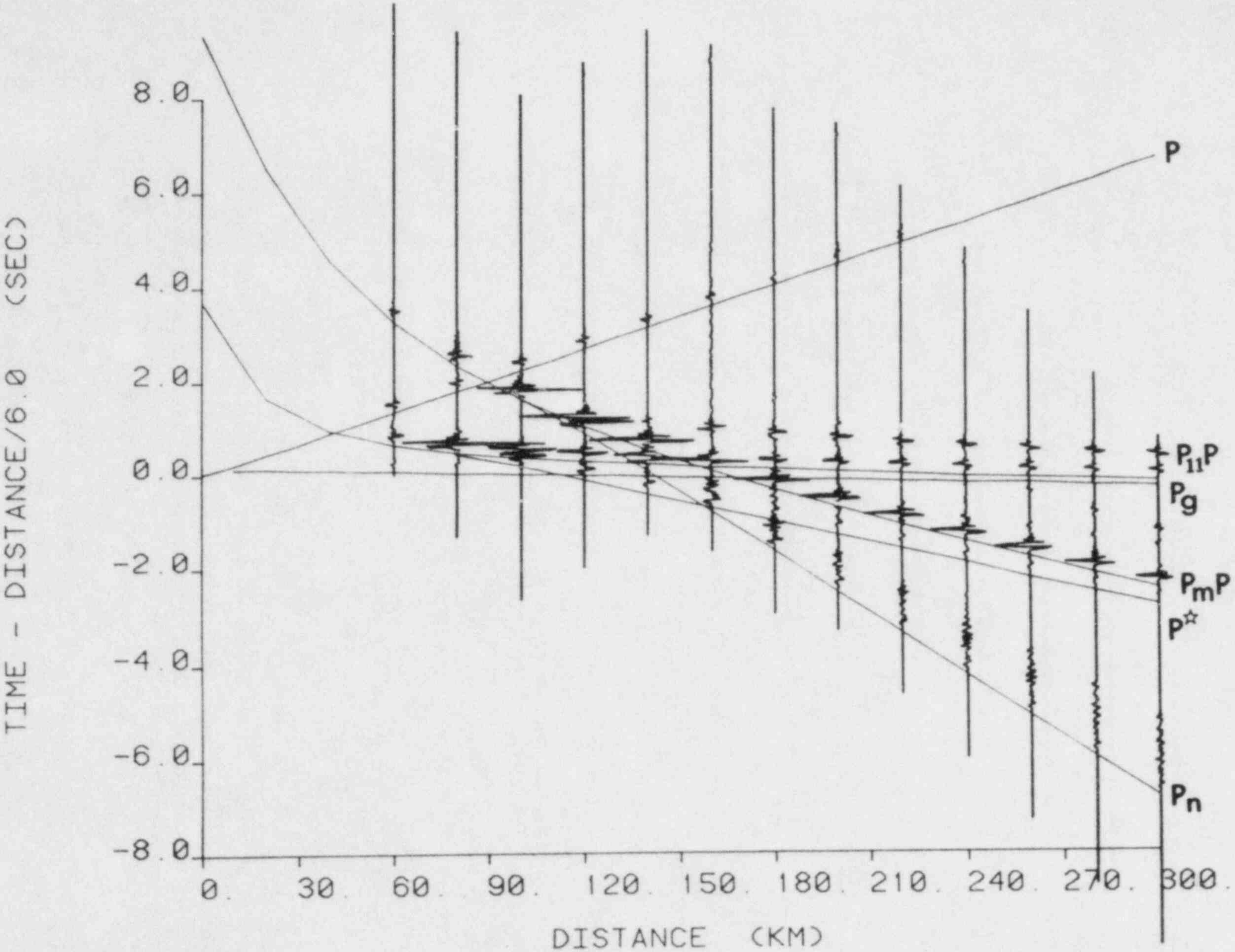

Figure 6.5 The same as Figure 6.3 but with a modification of the Chiburis and Ahner (1980) crustal model where the mid-crustal and Moho discontinuities have been replaced with velocity gradients.

6.4 R_g Wave Studies

Investigations of the source and path properties of short-period Rayleigh waves (R_g) recorded by the network have been made. These R_g waves are commonly observed on seismograms of quarry blasts throughout New England (Figure 6.6), and they have also been observed on records of shallow focus earthquakes. R_g can be modelled as a fundamental-mode surface wave propagating in a flat-layered crustal model. Thus, studies of these waves are yielding important information about upper crustal structure and depths of seismic sources in New England. The prominent signals of R_g waves that are recorded from quarry blasts throughout New England are a result of R_g being a fundamental-mode surface wave that is efficiently excited by sources at or near the surface (Figure 6.7).

The research using R_g waves has included: (1) determination of lateral variation in the upper crustal structure beneath New England from dispersion studies; (2) determination of attenuation of seismic energy in the upper crust at frequencies near 1 Hz; and (3) investigation of the use of R_g and other phases as depth discriminants for earthquakes and explosions in New England. The results found during the contract period were: (1) Group velocity dispersion has been determined for 0.5-2.0 sec R_g waves propagating across southern New England (Kafka and Dollin, 1985). The paths of these waves traverse various geologic structures and distinct dispersion regions have been identified (Figure 6.8). Normal dispersion was consistently observed in the period range studied, indicating that a superficial layer of relatively low velocity ($4.1 < V_p < 5.5$) at least 1 km thick must be present in the upper crust throughout southern New England. The existence of distinct dispersion regions has allowed the delineation of the lateral variation in the upper crust beneath the Appalachians of southern New England (Figures 6.9 and 6.10). The results from R_g dispersion studies are consistent with those from studies of local refraction data (e.g. Chiburis and Ahner, 1980) and teleseismic residuals (e.g. Taylor and Toksoz, 1979). However, the crustal models inferred from the R_g dispersion have allowed the delineation of lateral variations at much shallower depths (upper 5 to 10 km where most earthquakes in New England occur) than do the refraction studies. (2) The spatial attenuation coefficient (R_g) has been determined to be 0.0025 km^{-1} at frequencies near 1 Hz. This number compares favorably with other measurements made for New England (Cicerone, 1980, Pulli, 1984). (3) Progress has been made toward understanding how waveforms of seismic phases recorded by the network are affected by differences in source depth. The ratios of R_g amplitudes to amplitudes of Pg, Sg and Lg appear to be good indicators of source depth, particularly for events in the upper 3 to 5 km

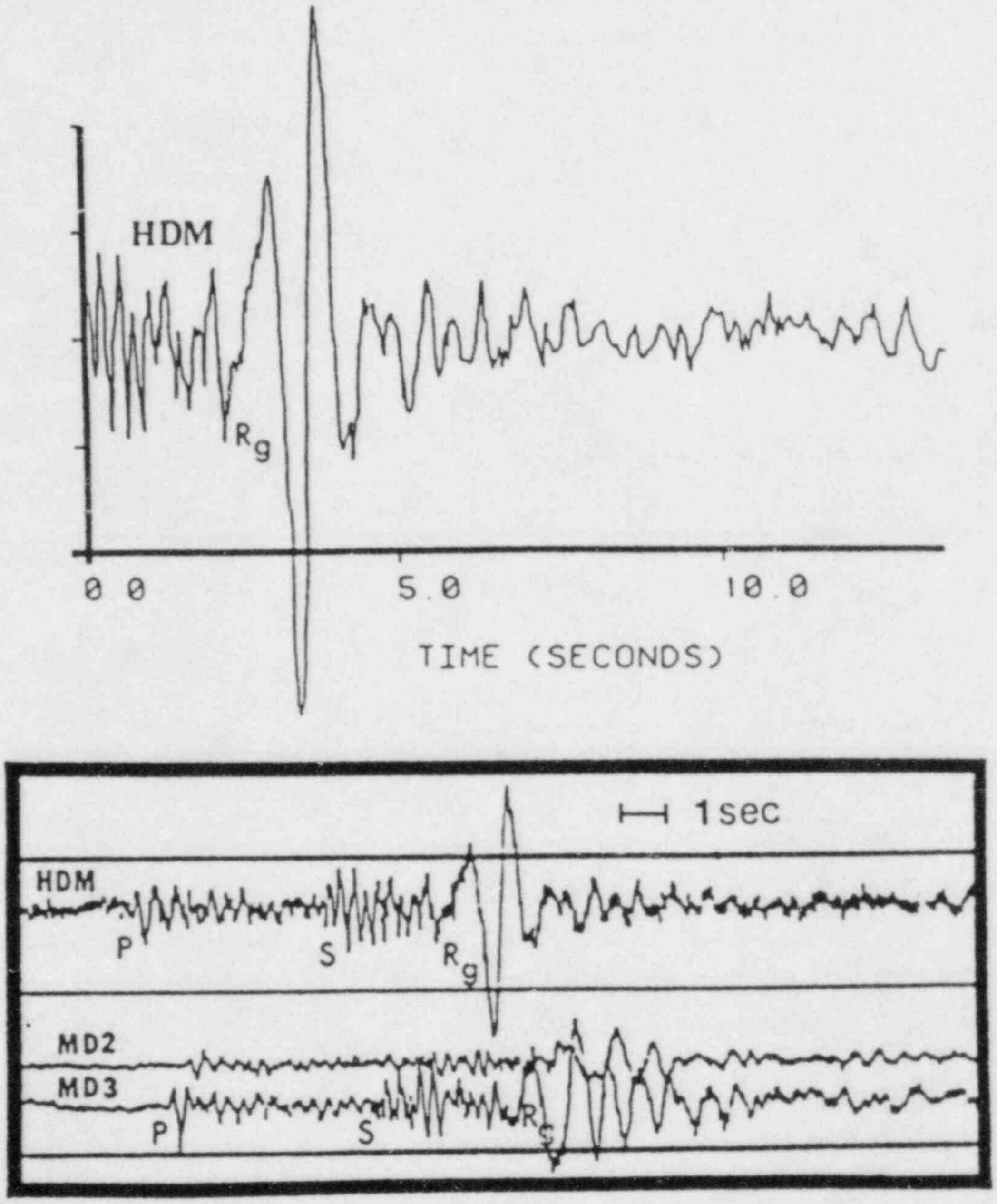

Figure 6.6 Examples of Rg waves recorded at sta'.ions HDM, MD2 and MD3 in Connecticut from a quarry blast in central Connecticut. The top part of the figure is an example of a digitized Rg wave while the bottom part of the figure shows a section of the original develocorder film at the time the blast was recorded.

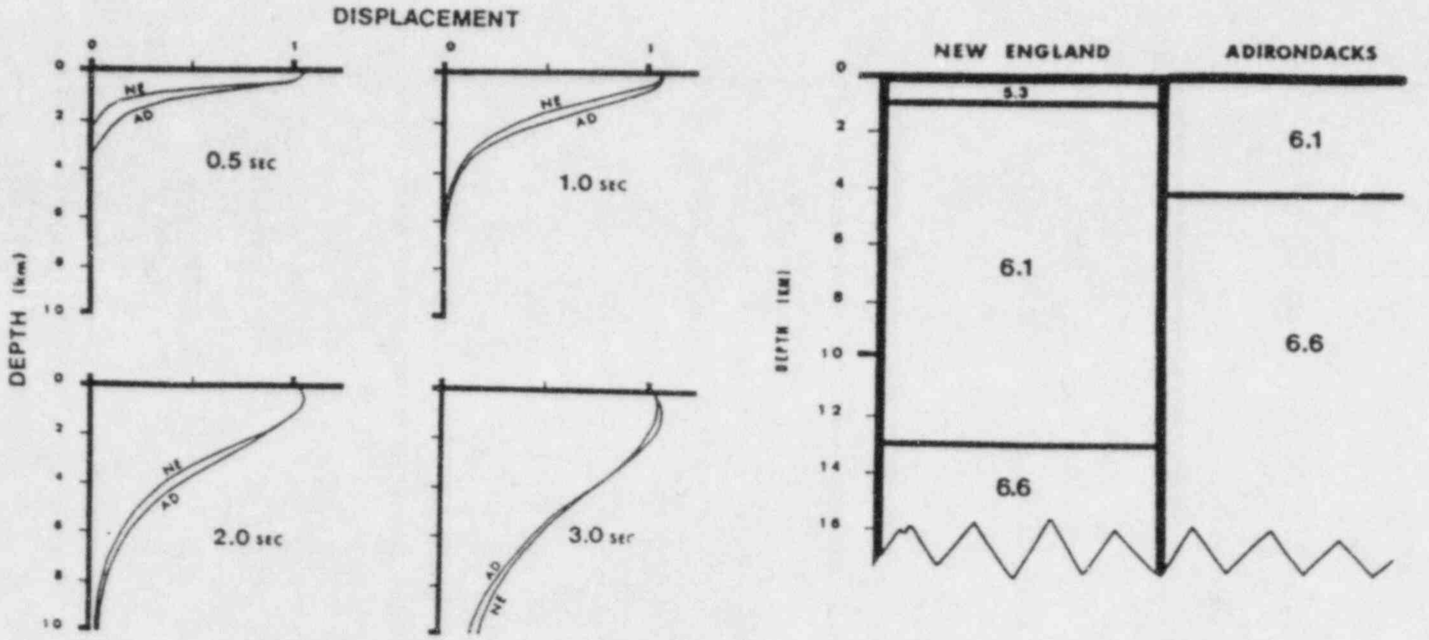

Figure 6.7 Theoretical calculations showing the effect of crustal structure on the relative excitation of Rg waves for surface focus sources. The crustal models are New England (NE) from Chiburis and Ahner (1980) and Adirondacks (AD) from Aggarwal <u>et. al.</u> (1975). For each crustal model shown on the right the P wave velocities in km/sec are indicated.

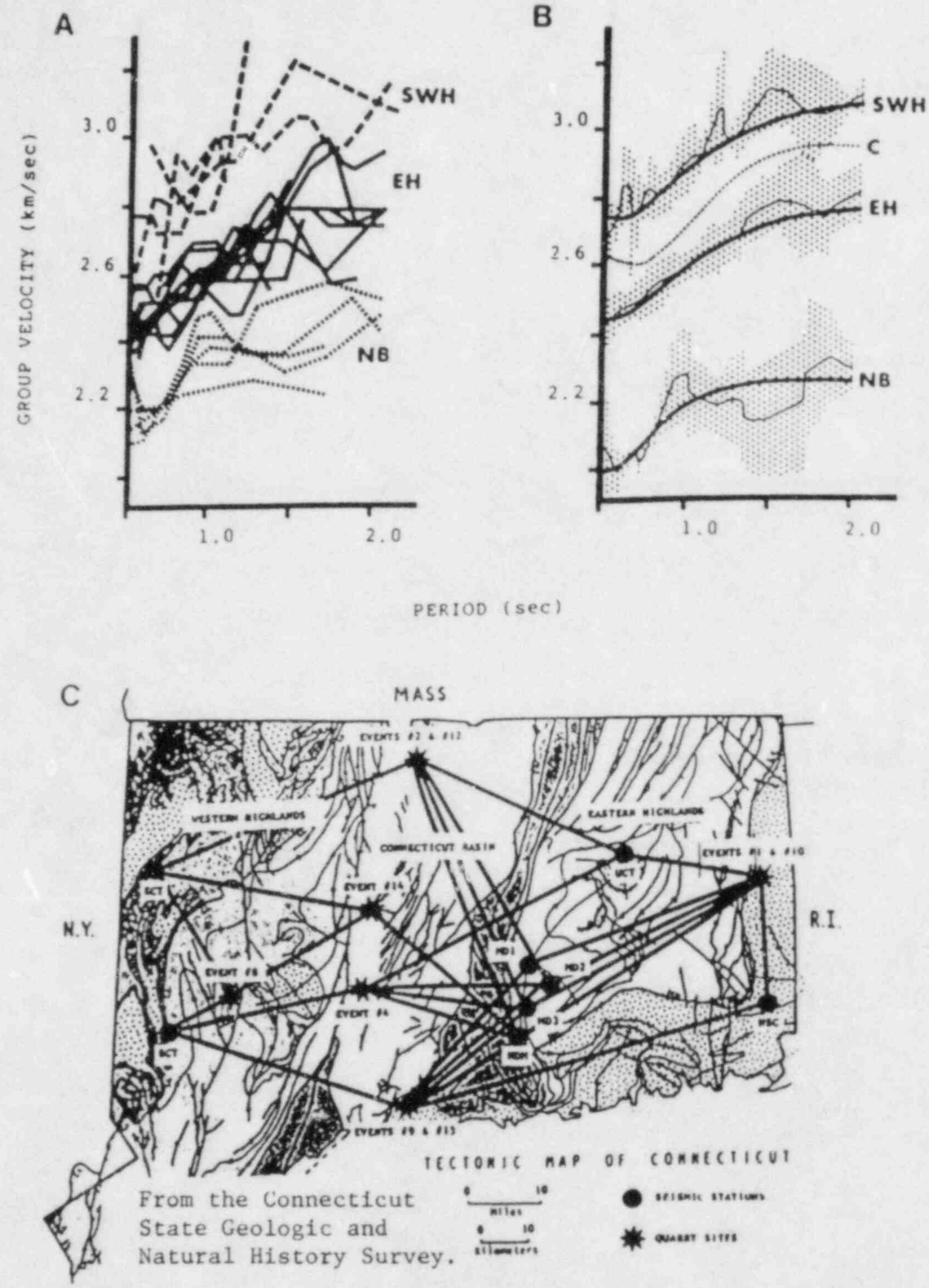

Figure 6.8 A. Observed dispersion curves for the Southwestern Highlands (SWH), Eastern Highlands (EH) and Northern Connecticut Basin (NB). B. Theoretical dispersion curves for the different models (heavy solid lines) superimposed upon the ranges of dispersion curves (stippled areas) for each of the three regions analyzed. The dotted line marked C is the theoretical dispersion for the Chiburis and Ahner (1980) crustal model. C. Map of the source-receiver paths for which Rg dispersion curves have been computed.

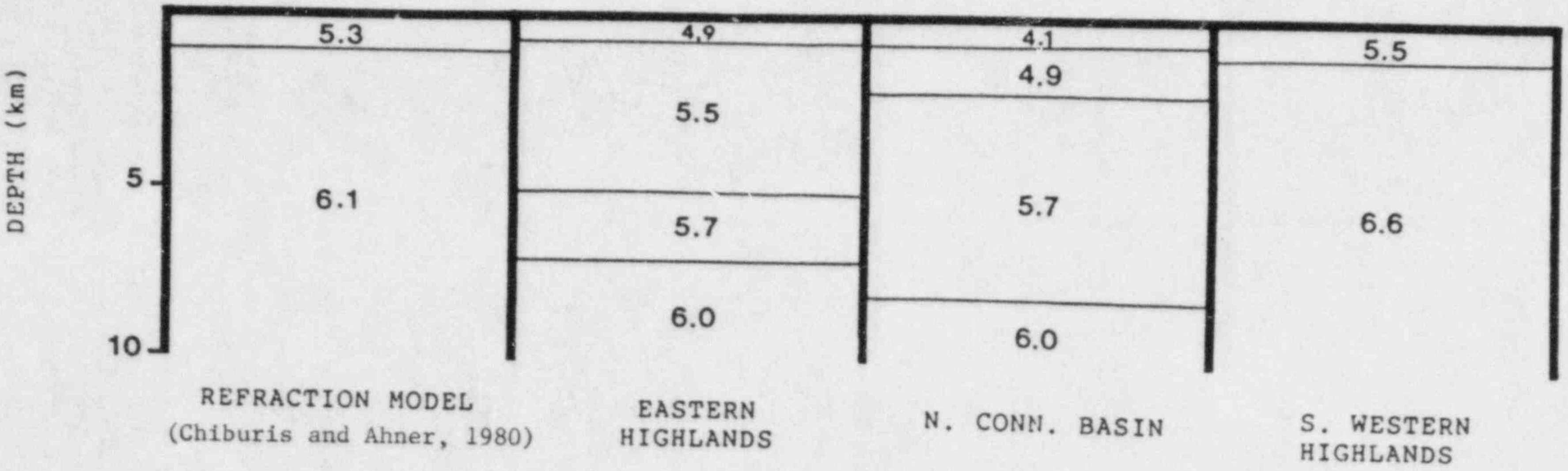

Figure 6.9 Comparison of the crustal models found for different parts of Connecticut from the Rg dispersion curves. The number indicate the P wave velocities in km/sec for each layer.

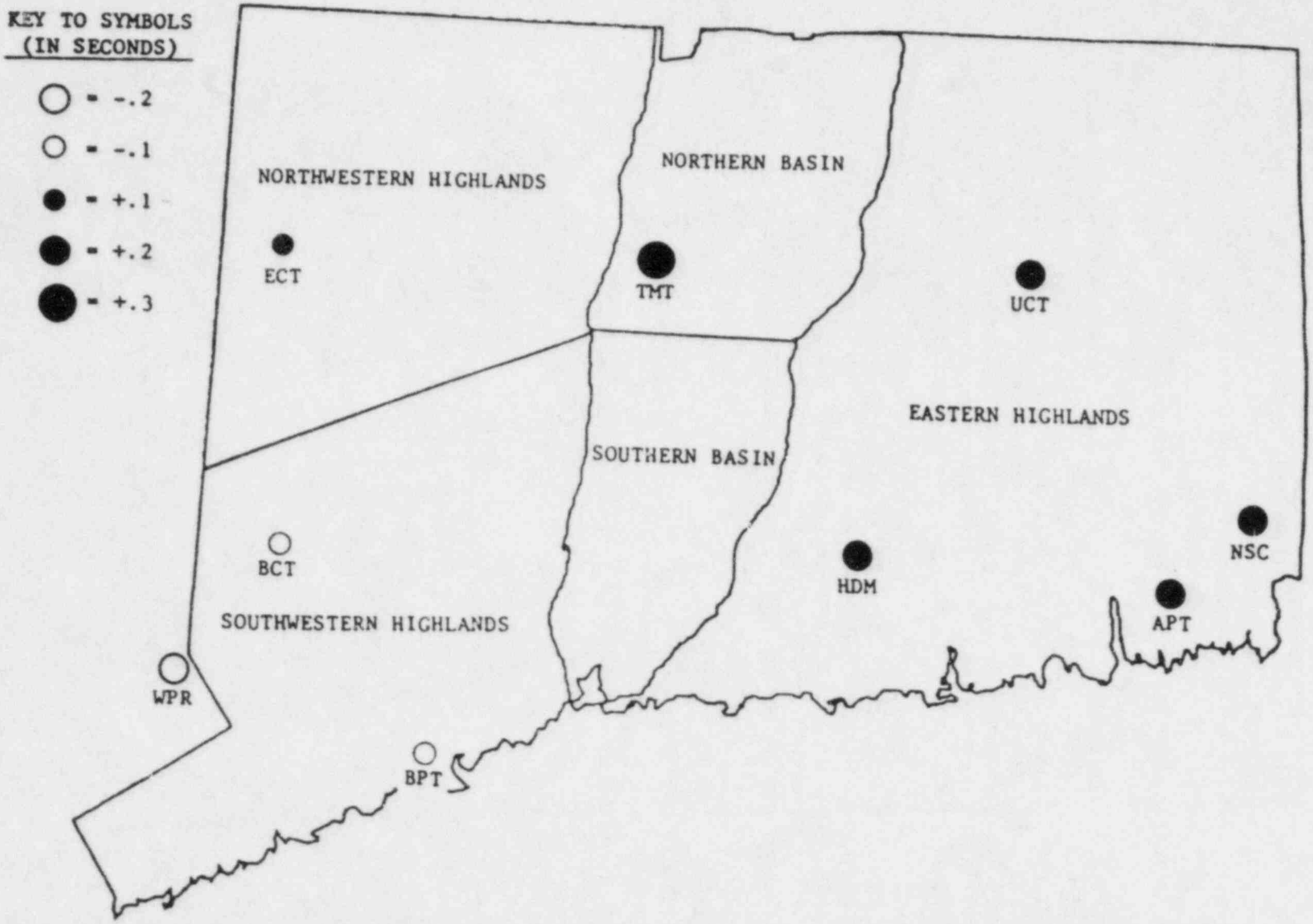

Figure 6.10 Travel time residuals for P waves at seismic stations in Connecticut. All the residuals are from Taylor and Toksoz (1979) except for station WPR which is from Peseckis and Sykes (unpublished). Negative residuals indicate higher velocity crust and positive residuals indicate lower velocity crust.

of the crust. For events with well-constrained depths greater than about 5 km, R_g has not been observed. This is consistent with the theory of fundamental-mode surface waves. (4) A preliminary investigation of Rg waves recorded by the network from explosions detonated as part of the Maine Refraction Experiment of the USGS was made. Since these blasts were recorded digitally and the origin times of the events are well known, R_g velocity differences from one path to the next were resolved with much greater accuracy than from our previous experiments in other parts of New England. Preliminary results suggest anisotropy in the upper crust beneath southeastern Maine. In this region, faster velocities were observed for paths trending NE (parallel to the grain of the regional geology) and slower velocities for paths normal to the trend of the grain. (5) Initial studies of R_g dispersion in southeastern New England suggested that velocities in the upper crust beneath the Avalonian terrain are slightly higher than velocities beneath the Eastern Highlands region of Connecticut. The presence of such a variation in crustal structure is consistent with results of geologic field studies carried out by Weston Observatory personnel which indicated that the Avalonian terrain of southeastern New England contrasts sharply with the rest of the Appalachian orogen to the west (Rast and Skehan, 1983).

7. DISCUSSION OF IMPORTANT RESULTS

The research efforts described in Sections 1 through 6 have yielded a number of important new results concerning the seismicity of the northeast region in general and New England in particular. The most important results are:

1. An improved understanding of the historic and recent rates of seismicity has been obtained. The compatibility of the earthquake return times calculated from the historic and the recent data sets have provided constraints which can be used for finding the seismic hazard in the region. It appears that the odds are good that a potentially damaging earthquake of magnitude 5.0 or greater should strike somewhere in New England once every twenty or so years. Unfortunately, the apparent randomness of the earthquake activity with time and space means that there are virtually no clues in the present seismicity as to where and when the next damaging earthquake will strike.

2. The compilations of the historic and recent earthquake locations have allowed the spatial patterns of seismicity to be mapped. While there is a general diffuse pattern to the locations of many earthquakes, there are places where the activity seems to be more concentrated. Many of these areas which were active historically also are the locations of clusters of activity in the recent seismicity. Furthermore, these areas also seem prone to experiencing larger earthquakes both historically and recently. Figure 7.1 shows the epicenters of all historic earthquakes of magnitude 4.0 or greater in New England (Ebel, 1984). This can be compared to Figure 7.2 which is a map of the same area but is of earthquakes of M_c = 3.0 or greater for the time period of October 1, 1975 to March 31, 1985. As can be seen, the areas which had the most significant earthquakes historically are the same places where the largest recent earthquakes have occurred. Figure 7.3 is a summary map of the areas of significant historic and recent earthquake activity. Those regions which have been active in both periods are the places which may have the greatest probabilities of being the sites of large earthquakes in the future.

3. The accuracy of earthquake epicentral and depth determinations have been improved. The accuracy of the historic epicenters probably averages ±20 km, while that for the instrumentally recorded recent earthquakes is better than ±5 km in most cases (Ebel 1984). The accuracy is even better (±2 km or less) in places where tight arrays of instruments have been operated (i.e. around Moodus, Connecticut and in aftershock studies). Thus the location accuracy of epicenters found using the instrumental data has improved by as much as an order of

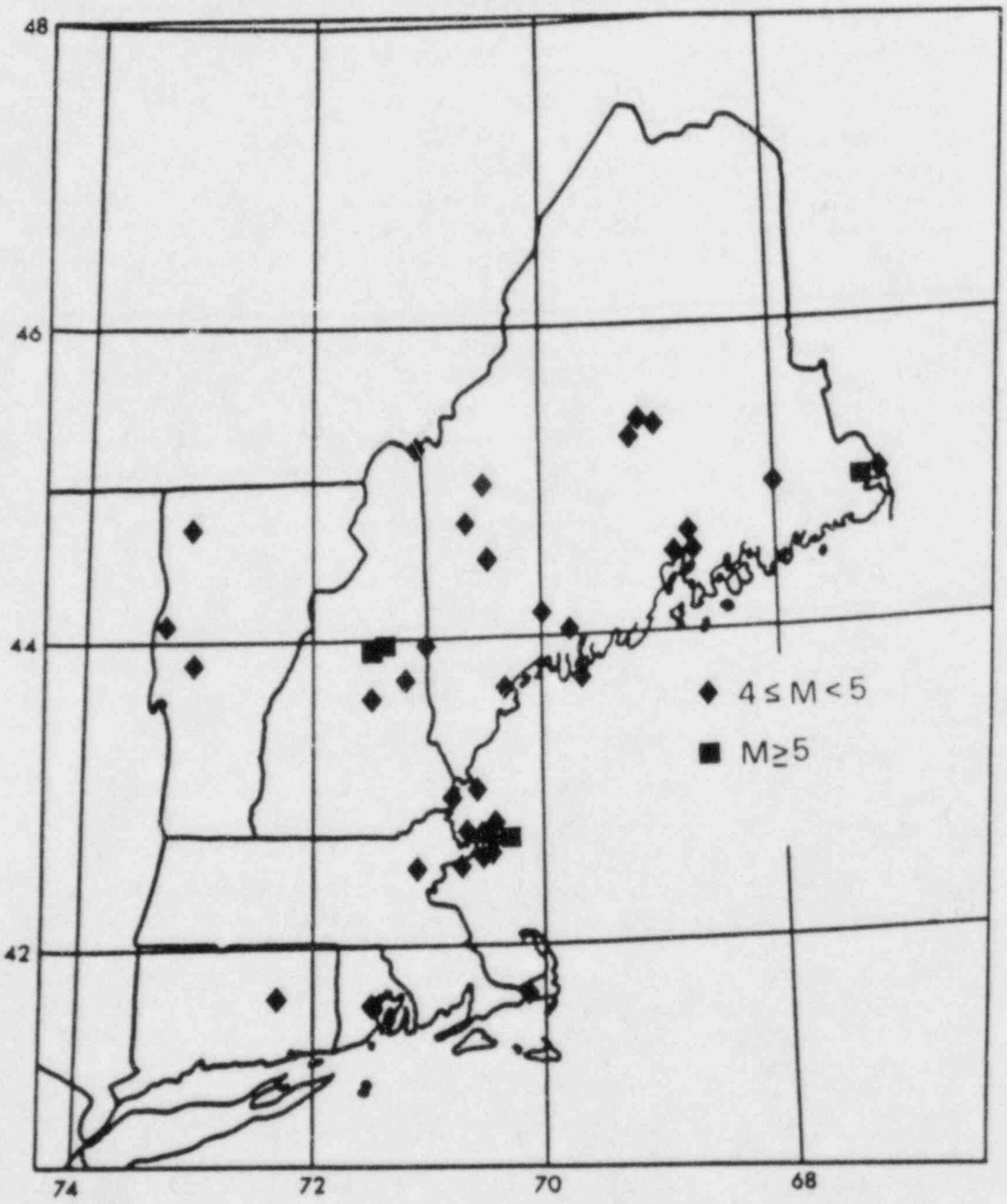

Figure 7.1 Map of the largest earthquakes in New England from 1727 to March, 1985. The magnitudes of the historic earthquakes has been estimated from the felt areas, as described by Ebel (1984).

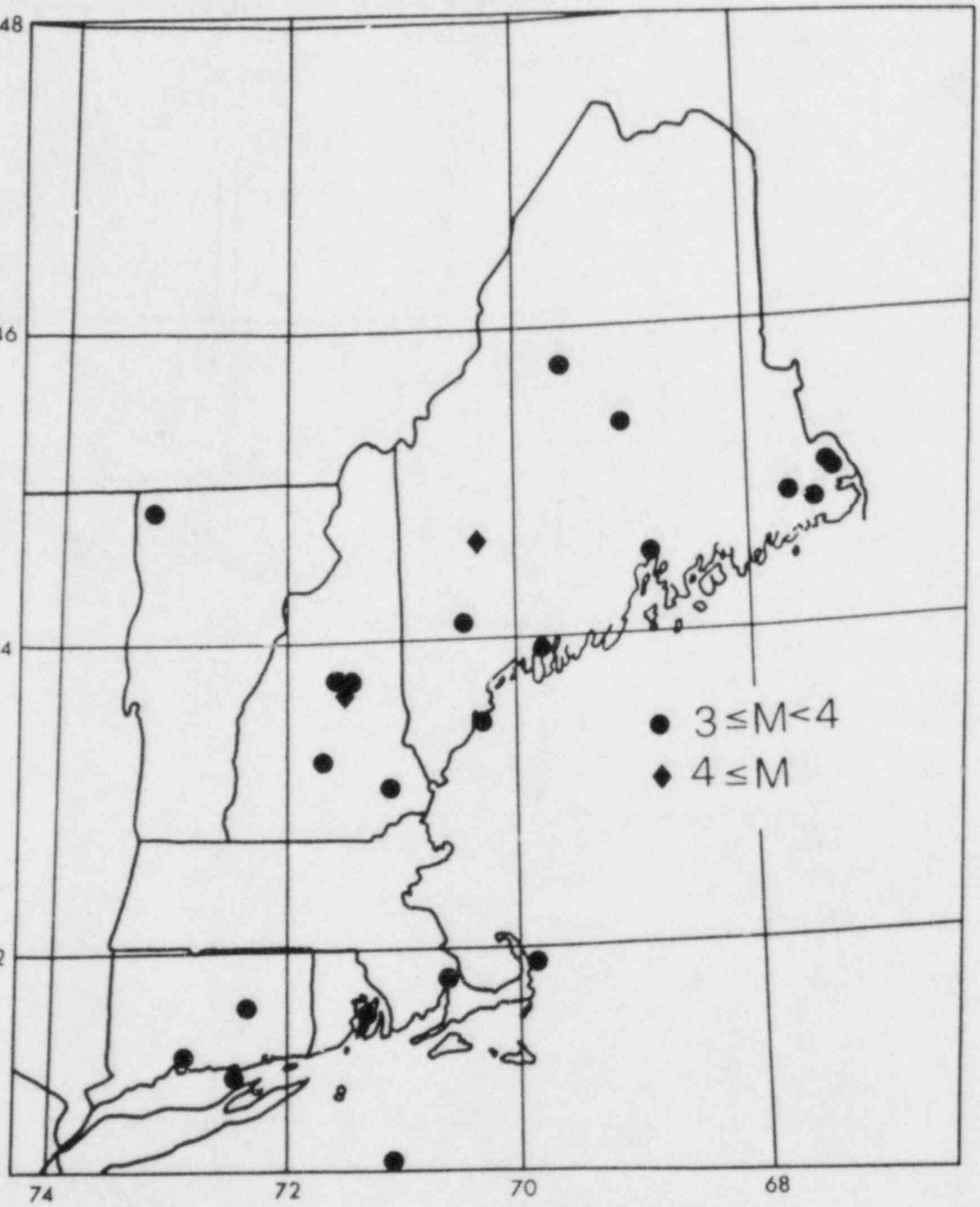

Figure 7.2 Map of the largest earthquakes in New England from October, 1975 to March, 1985.

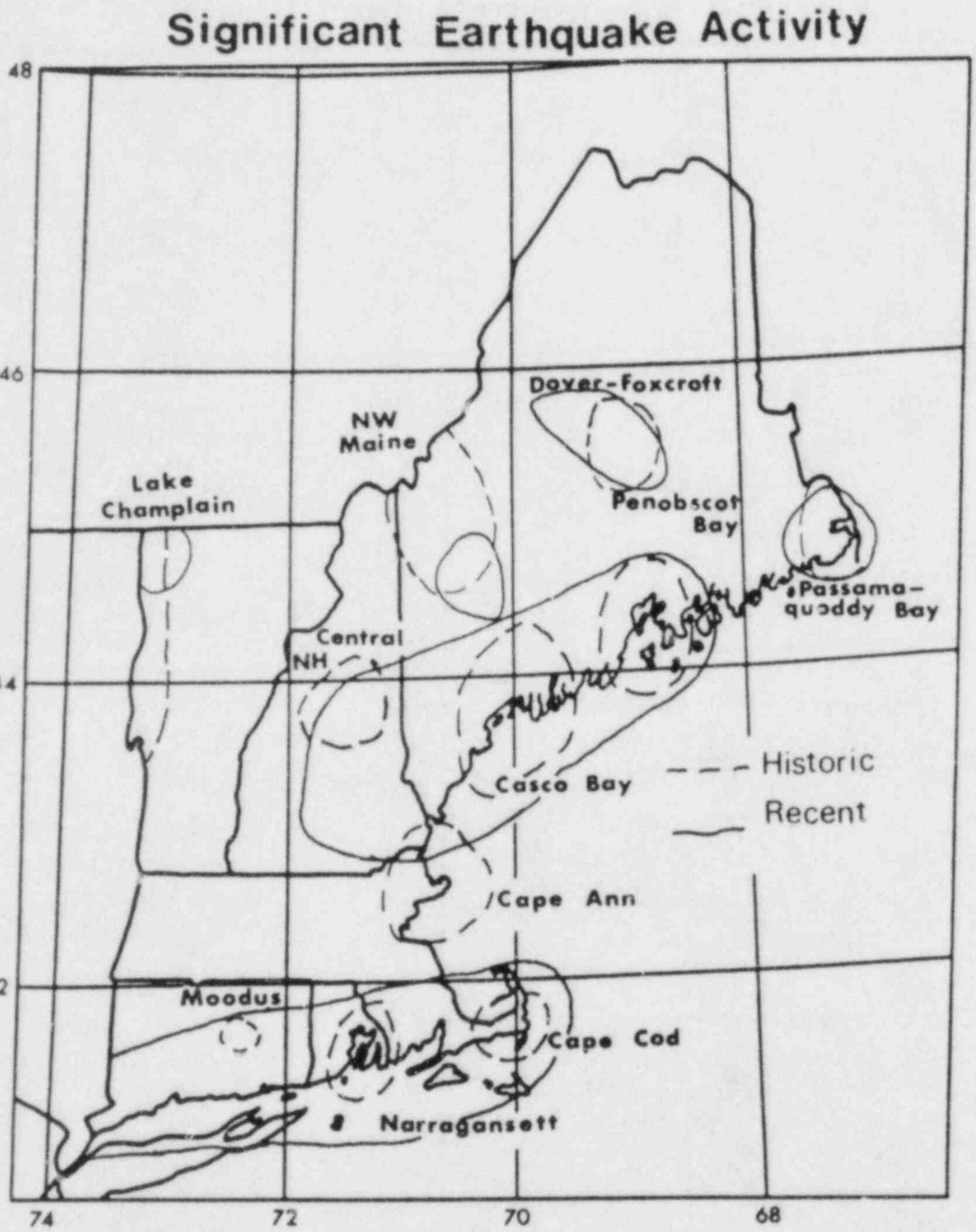

Figure 7.3 Map of the areas which have experienced larger earthquakes in the historic catalog (prior to 1975) and in the recent catalog (1975 to 1985). The boundaries of the active areas are only meant to be indicators of the more active parts of New England based upon the data in Figures 7.1 and 7.2 and in the recent and historic catalogs.

magnitude over that of the historic data. Some constraints can also be put on earthquake depths with the instrumental data, but this was somewhat limited by the relatively wide station spacing. Hypocenters were almost invariably located in the upper 10 km of the crust with many of the events locating at depths of less than 5 km. Very accurate depths computed from near-epicenter arrays of portable instrument for aftershocks confirm this range of earthquake depths in New England.

4. Details of the aftershock activity of a number of earthquakes have been documented for the northeast. The intensity of aftershock activity seems to vary widely from earthquake to earthquake, and experience indicates that the chances of detecting aftershocks to earthquakes of M_c 4.0 appear to be rather poor. The most intense aftershock occurrences seem to mitigate at the same rate as in other parts of the world, i.e. the number of aftershocks per unit time is inversely proportional to the time after the main shock. The locations of aftershock zones for earthquakes of sufficient size appear to be easily mapped if portable instrumentation is installed soon after the main shock. Such monitoring is also capable of detecting the migration patterns of aftershocks, which seem to have occurred in some instances in New England (i.e. Bath, Maine and Sanbornton, New Hampshire) as well as in other areas of the northeast (central New Brunswick). All of this experience provides important background information which can be used to make decisions about the feasibility of future aftershock studies.

5. The relationship between the occurrences of particular earthquakes and local geology cannot be clearly established in those places where the earthquakes have been located with a high accuracy. For instance, of the six areas discussed in Section 5, only the aftershocks of the Bath, Maine earthquake and some of the New York City events were found to locate on mapped fault features. The epicenters of the earthquakes from Sanbornton, New Hampshire, Dixfield, Maine and Passamaquoddy Bay, Maine did not fall on any mapped faults but all occurred near irregularities in the geology which might be inferred to be due to faults in the basement. The Moodus, Connecticut earthquakes and about half of the New York City seismicity also did not occur on mapped faults, but the events from both areas occur at the edges of Triassic basins. It should be noted that the largest earthquake in the northeast during the contract period, that in central New Brunswick in January, 1982 (m_b = 5.7), also was not associated with any confirmed geologic fault (Wetmiller et. al., 1984).

6. Some progress has been made toward determining regional and local variations in the crustal seismic velocity structure of New England. This work, comprised of the R_g wave dispersion analyses, the different seismic refraction results in the region and the studies of teleseismic travel times and waveforms, is ultimately geared toward finding different tectonic blocks within the crust of the northeast and toward improving earthquake location capabilities. Variations in crustal thickness throughout the region and locally high and low upper crustal velocities have already been found. The goal of this research is in part to improve the resolution of these studies to the point where the regional and local crustal blocks can be related to the surface geology.

Several lines of research must be pursued further if the causes of earthquakes in the northeast are to be understood and the calculation of local seismic hazard is to be most meaningful. First, site specific seismicity rates need to be established. Because of the low-seismicity at any one locality in the region, the numbers of instrumentally recorded earthquakes have been insufficient to calculate local recurrence curves. Many more years of continuous monitoring are necessary to gather a statistically significant sample for such estimates. Second, more high resolution studies of the variations in the regional variation in the seismic velocity structure and in the relationship of individual earthquakes and aftershock with geology are necessary if the mapping of seismotectonic zones is to be accurately made. No clear patterns for New England seem to have emerged from the information gathered to date. Third, improved measurments of stresses, and especially of earthquake focal mechanisms, are needed to better delineate variations in the regional stress field. Such work must be continued if the sources of the regional stress field are to be found. A better understanding of earthquake waveforms could lead to the routine determination of the focal mechanisms of recorded events, as is presently being done with teleseisms (Dziewonski et.al., 1981). Fourth, a continued effort must be made to collect well-calibrated ground acceleration data for the region. Such data is vital for the calculation of seismic hazard curves.

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The operation by Weston Observatory of a seismic network in New England from 1974 to 1985 is described, and the results of the seismic monitoring are summarized. The network coverage of Weston Observatory increased from two operating stations in 1974 to 36 stations in 1979 and was stabilized at 30 stations in the early 1980's.

The network was used to find the locations and magnitudes of all earthquake activity detected during the study period. Most earthquakes from 1974 to 1985 were found to occur in the same places as those which have been documented historically, although the activity appears to be random both in space and time. Studies of aftershocks and detailed monitoring in selected areas did not show any strong correlations between the earthquake locations and mapped geologic structures. It is concluded that the relationship among earthquakes, tetonic or structural zones and faults exposed on the surface are not well understood. The causes of the earthquake activity in the northeast are not clearly established with the seismic data which was gathered and analyzed.

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