Seismicity in the Area Surrounding Two Mesozoic Rift Basins in the Northeastern United States

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ABSTRACT

Seismologists have long sought to discern the cause(s) of earthquakes in the greater New York City (NYC) area. After decades of research on this topic, however, there are still many unanswered questions. Seismicity in the NYC area seems to be related in some way to the locations of Mesozoic rift basins (MRBs). This pattern might be a localized expression of the more global tendency of large earthquakes in stable continental interiors to occur in crust that has been stretched or extended at some time in the geologic past. This paper has three objectives: (1) It is a review of some of the well-known hypotheses that have been proposed to explain why earthquakes occur in the northeastern United States in general, and in the NYC area in particular. This review provides a background, and places this study of earthquakes in the NYC area in a more general context. (2) It is also a summary of the network seismicity in the NYC area. The network data are compared with the historical record of seismicity, and we demonstrate that there is a correlation between the network and historical seismicity. The earthquake process in the study area, therefore, appears to be stationary on the time scale of a couple of centuries. (3) Given this result, we use the network data as a "snapshot" of this earthquake process to test the hypothesis that there is a correlation between earthquake locations and the two MRBs in the study area, the Newark and Hartford basins.

INTRODUCTION

Although the level of seismic activity is only moderate in the area surrounding New York City, there are enough earthquakes to make seismologists and many local residents curious, if not concerned. Concerned, because even this moderate level of activity poses some degree of seismic risk due to the area's high population density and aging built environment. For these reasons, seismologists have long sought to discern the cause(s) of earthquakes in the greater New York City (NYC) area. After decades of research on this topic, however, there are still many unanswered questions.

One of the motivations for this study is that for the years between about 1975 and 1990, while there was a fairly dense distribution of seismic stations, we have as complete a record of earthquake activity in this area as we are likely to have for some time. Because of cutbacks in funding of network operations, the record is not as complete for the years since 1990. This situation effectively forces seismologists, at least for the time being, to base their studies of network seismicity in this area on the 1975–1990 data set. In this study, we use that limited, high-quality data set to investigate the nature of earthquake processes in the greater NYC area.

One promising result of network monitoring efforts and historical archive analyses is that there seems to be a systematic pattern of seismicity in this area—a pattern that hasn't radically changed over the past couple of centuries (Figures 1 and 2). In fact, as we will demonstrate in this paper, the earthquake process appears to be stationary in the study area on a time scale of about two centuries. This apparently stationary earthquake process is in marked contrast with the results of a study of the southeastern United States by Seeber and Armbruster (1987), who demonstrated that there was a distinct change in the pattern of seismicity before versus after the 1886 Charleston, SC earthquake (M_s~7.5). Furthermore, they showed that the seismicity for at least 80 years before the 1886 earthquake would not have delineated the location of that earthquake.

In this paper, we base our analysis of the relationship between seismicity and geological features on the presumed stationarity of the earthquake process in the study area. We caution the reader, however, that the results of Seeber and Armbruster (1987) regarding the 1886 Charleston, SC earthquake suggest that even one large earthquake in the future might render all of our conclusions moot. In fact, if—

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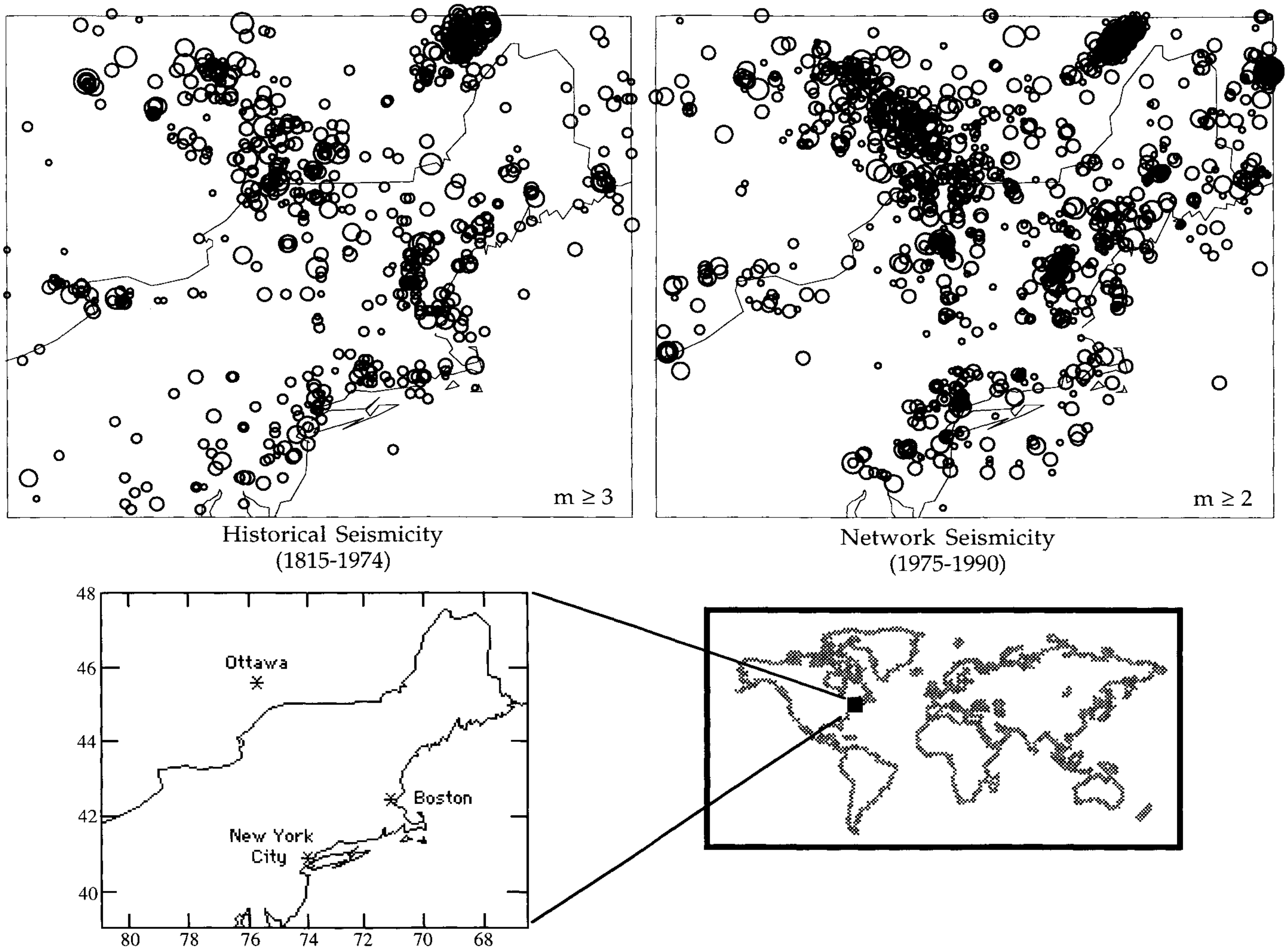

▲ **Figure 1.** Historical seismicity ($m \geq 3$) and network seismicity ($m \geq 2$) for the northeastern United States and adjacent areas. Historical data are for the time period from 1815 to 1974, and network data are for 1975–1990. Locations and magnitudes of the historical earthquakes are taken from the NCEER-91 catalog (Seeber and Armbruster, 1991), and those of the network earthquakes are taken from the Weston Observatory archives of the data recorded by the Northeastern United States Seismic Network.

as argued by Long (1988)—intraplate continental earthquakes are transient phenomena (responding to perturbations in crustal strength independent of preexisting crustal features), then the earthquake process in this area may only appear to be stationary because our earthquake catalogs represent a very short "snapshot" of the long-term (non-stationary) earthquake process.

For large earthquakes on a global scale, Johnston (1989) demonstrated that virtually all of the largest earthquakes in stable continental interiors ($M_w \geq 7.0$) occur in crust that has been stretched or extended at some time in the geologic past. Figure 2 shows that, at least in the case of the Newark basin, the seismicity seems to be related in some way to the location of a Mesozoic rift basin (MRB), an area that was stretched and extended during Mesozoic times. This pattern might be a localized expression of a more global characteristic of intraplate earthquakes.

We have three objectives in this paper. First, we review some of the well-known hypotheses that have been proposed to explain why earthquakes occur in the northeastern United States (NEUS), to provide a background and to place this study in a more general context. Second, we summarize the network seismicity and compare that seismicity to the historical record, concluding that the earthquake process appears to be stationary on the time scale of a couple of centuries. Third, given this result, we illustrate how the network data can be used as a sample of this (presumably) stationary process to investigate the relationship between geological features and seismicity. Specifically, we use the network data to test the hypothesis that there is a correlation between MRBs and earthquakes in this area. Such an approach, of course, assumes that we can infer from the observation that the seismicity appears to be stationary over the past couple of centuries that it is also stationary over much longer geological time scales.

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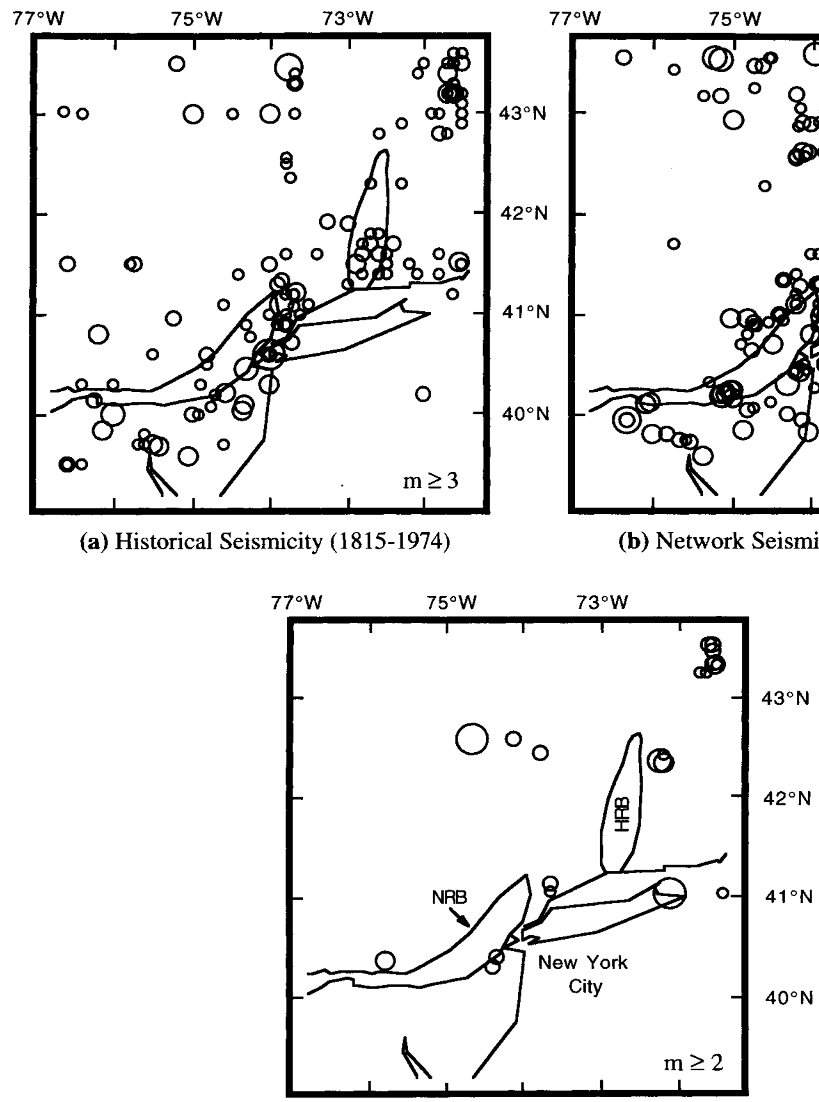

(a) Historical Seismicity (1815-1974)

(b) Network Seismicity (1975-1990)

(c) Network Seismicity (1991-1994)

▲ **Figure 2.** Historical and network seismicity in the study area. (a) 1815–1974, $3.0 \leq m \leq 5.2$; (b) 1975–1990, $2.0 \leq m \leq 4.7$; (c) 1991–1994, $2.2 \leq m \leq 4.1$. Locations and magnitudes of the historical earthquakes are taken from the NCEER-91 catalog (Seeber and Armbruster, 1991), and those of the network earthquakes are taken from the Weston Observatory archives of the data recorded by the Northeastern United States Seismic Network.

BACKGROUND: THE SEARCH FOR AN UNDERSTANDING OF EARTHQUAKE PROCESSES IN THE NORTHEASTERN UNITED STATES

The search for the cause of earthquakes in the NEUS has led seismologists down many paths. In the 1970s, with the first fruits of network monitoring beginning to provide a reliable and complete (albeit short) catalog of earthquakes for this area, there was finally a chance for seismologists to propose testable hypotheses to explain why NEUS earthquakes occur. During that time, however, there were very little data, and hypotheses sometimes appeared to be consistent with the available data, even though they turned out, in retrospect, to be inconsistent with newer data. There have been a number of summaries and reviews of this topic (*e.g.*, Sykes, 1978; Basham, 1989; Seeber and Armbruster, 1989; Ebel and Kafka, 1991), and we will not present an exhaustive summary here. Instead, we will highlight some of the well-known hypotheses, and illustrate how the network data have been helpful in testing those hypotheses. These highlights will provide the historical foundation for our study of earthquakes in the greater NYC area, and will place this study within the context of earthquakes in the NEUS in general.

Hypotheses that have been proposed to explain why earthquakes occur in the NEUS (and in other intraplate continental regions) are generally variations on one or both of the following themes: (1) pre-existing zones of weakness from ancient orogenic episodes are reactivated in the

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present-day stress field, and (2) stress concentrations due to varying material properties cause amplified stress in a localized area, eventually leading to an earthquake. Long (1988) proposed an alternative model for major intraplate continental earthquakes. In his model, intraplate continental earthquakes are caused by a short-term transient weakening of the crust, initiated by a disturbance in the hydraulic or thermal properties of the crust. An important implication of such an alternative model would be that the occurrence of a major earthquake somewhere in the NEUS could radically change the pattern of seismicity, *i.e.,* we might observe an extended aftershock sequence for many years following the main event. The seismicity resulting from this type of process would not be expected to be stationary.

Some examples of well-known hypotheses that were invoked in the 1970s to explain NEUS earthquakes were the existence of a so-called Boston-Ottawa Seismic Zone (*e.g.,* Diment *et al.*, 1972; Sbar and Sykes, 1973), the hypothesis that NEUS earthquakes are associated with plutons (*e.g.,* Simmons *et al.*, 1976), and the hypothesis that seismic activity in this area is concentrated along NE trending faults, of which the Ramapo fault appeared to be the most active (Aggarwal and Sykes, 1978). We will address each of these in turn.

Additional data recorded in the 1980s made it more difficult to argue in favor of some of the 1970s ideas, and brought a new wave of hypotheses to take their place. For example, Seeber and Dawers (1989) and Seeber and Armbruster (1989) interpreted a relatively large earthquake (m_{bLg} = 4.0) that occurred in 1985 near Ardsley, NY in terms of a hypothesis—proposed earlier by Seborowski *et al.* (1982)—that NW trending faults are important, seismically active features in the NYC area. This hypothesis was invoked by Seeber and Dawers (1989) and Seeber and Armbruster (1989) to explain the observed characteristics of the Ardsley earthquake, including the observation that the fault plane appeared to be transverse to the trend of the Ramapo fault and other presumably active NE trending faults.

Before turning to a discussion of each of these hypotheses, the overall point should be stressed that each of these hypotheses—including the one tested in this paper—might remain viable only until either the next large earthquake and/or the next advance in the level of monitoring of the smaller earthquakes. These hypotheses may, in principle, be quite reasonable on physical grounds; but whether they will, in the long term, delineate the locations of future large earthquakes remains to be seen. Again, the findings of Seeber and Armbruster (1987) on the change in seismicity before versus after the 1886 Charleston, SC earthquake should make anyone cautious in this regard.

The Boston-Ottawa Seismic Zone

The Boston-Ottawa Seismic Zone (BOSZ) was a trend, albeit discontinuous, of earthquakes extending from Boston across New England, and into Canada to the vicinity of Ottawa. An example of how the network data were helpful in testing hypotheses regarding the cause of NEUS earthquakes is the re-evaluation of the existence of a BOSZ. Before extensive network data were available, it seemed reasonable (based on the historical data) to argue that such a trend existed (*e.g.,* Diment *et al.,* 1972; Sbar and Sykes, 1973). Diment *et al.* (1972) and Sbar and Sykes (1973) also proposed that the BOSZ is located along a continental extension of the New England seamount chain, and Sykes (1978) invoked this notion as part of his general hypothesis that earthquakes in the eastern United States are located along extensions of fracture zones in the Atlantic ocean basin. Sykes (1978) recognized that there was a zone of very low activity in the middle of the BOSZ (essentially coincident with the entire state of Vermont). He argued, however, that this gap in seismicity was filled by a zone of igneous intrusions, the White Mountain Magma Series, which he conjectured was a potential source of earthquakes because it is a zone of weakness in the crust.

The additional network recording of earthquakes from the 1980s provided compelling evidence that the BOSZ was not a trend at all, *i.e.,* it now seems quite clear that there is a gap in seismicity in the part of the BOSZ that goes through Vermont and adjacent areas (Figure 1). Based on the 1975–1990 data, there is little evidence for a BOSZ or its offshore extension. In fact, Figure 1 shows that in the NEUS there are as many northeast "trends" in the seismicity (*i.e.,* transverse to the extensions of fracture zones) as there are northwest "trends" (*i.e.,* parallel to extensions of fracture zones). Thus, it now appears to have been premature to conclude that the seismicity is causally related to the extensions of fracture zones.

Stress Amplification Around Igneous Intrusions

A number of publications from the 1970s describe the hypothesis that large earthquakes in the NEUS, as well as in other stable continental areas, are caused by stress amplification in the area surrounding plutons (*e.g.,* Kane, 1977; McKeown, 1978; Simmons *et al.,* 1976). Barstow *et al.* (1981) performed a multivariate statistical analysis of the relationship between seismicity and geological or geophysical features in the central and eastern United States. One of their conclusions was that seismically active areas in the central and eastern United States have a greater likelihood of containing mafic intrusives than do non-seismic areas. Attempting to quantify the extent to which earthquakes are correlated with geological features in New England, Ebel and Spotila (1992) obtained some encouraging results. They found that 75% of the earthquakes that occurred in New England between 1975 and 1991 were spatially associated with 28 major faults and fracture zones, and they also found that about 40% of the earthquakes in that time period occurred within 10 km of a mapped pluton. Thus, there actually does appear to be evidence that eastern United States earthquakes are associated with plutons. Again though, it might take only one large earthquake in an area far from plutons or mapped faults to undermine these hypotheses. It may very well be that some earthquakes are caused by stress concentrations in the areas surrounding plu-

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tons and/or by reactivation of known faults that are mapped on the surface, but it is not at all clear that these hypotheses resolve the entire issue.

The Ramapo Fault and the Dobbs Ferry Fault

Two other hypotheses invoked to explain NEUS earthquakes focused on features in the greater NYC area: the supposed activity of the Ramapo fault and the hypothesis that the 1985 Ardsley, NY earthquake ruptured the Dobbs Ferry Fault. The Ramapo fault is a Mesozoic border fault, which forms the northwestern margin of the Newark rift basin (NRB), and which was also active during pre-Mesozoic times (*e.g.,* Ratcliffe, 1971; Ratcliffe, 1980). Aggarwal and Sykes (1978) studied locations, depths and focal mechanisms of earthquakes in the greater NYC area and concluded that seismic activity in this area is concentrated along northeast trending faults, of which the Ramapo fault appeared to be the most active. More recent studies of this area, however, produced results suggesting a more complicated relationship between earthquakes and geological features. For example, Seborowski *et al.* (1982) argued that focal mechanisms of three earthquakes that occurred on or very near the Ramapo fault have fault planes transverse to the mapped trace of the fault. Moreover, they argued that the microearthquake seismicity near the northern end of the Ramapo fault trends northwest, transverse to the trend of major geological structures mapped on the surface. In addition, based on the distribution of network seismicity, Kafka *et al.* (1985, 1989) argued that it is not clear that the Ramapo fault is any more active than some other parts of the NYC area. Although they found that about half of the network events in this area occurred within 10 km of the Ramapo fault, they also demonstrated that earthquakes at least as large as those recorded near the Ramapo fault are located as far as 50 km away in a variety of geological structures that surround the northern part of the NRB. Finally, Seeber and Armbruster (1986) presented evidence that the larger historical earthquakes in the greater NYC area were located on the southeastern side of the NRB, at a significant distance from the Ramapo fault.

The 1985 Ardsley, NY earthquake, while laying to rest (for now?) the notion that the Ramapo fault is the most active feature in the greater NYC area, generated another notion in its place. Since the preferred fault plane of the Ardsley earthquake fault plane solution was interpreted to be transverse to local, northeast-striking structures, Seeber and Dawers (1989) reasoned that there must be a northwest-striking fault. Based on focal mechanism studies, aftershock surveys, and extensive field mapping, they extrapolated from the surface mapped features to the 5 km depth of the earthquake, and proposed that the "Dobbs Ferry fault zone" (which is parallel to the inferred fault plane of the earthquake) was reactivated in the present-day stress field, causing the Ardsley earthquake. Since they did not find evidence for extensive displacement along the Dobbs Ferry fault zone, they then proposed that "low displacement faults" might be typical of intraplate earthquakes. If this hypothesis is correct, then faults with little or no displacement could be the source of significant earthquakes in intraplate areas (Seeber and Dawers, 1989).

Ancient Continental Rifting and Intraplate Earthquakes

On a global scale, research on intraplate earthquakes seems to be yielding a clearer picture of the cause of earthquakes in stable continental interiors. For large earthquakes on a global scale, Johnston (1989) has demonstrated that virtually all of the largest earthquakes in stable continental interiors occur in crust that has been stretched and extended in the continental rifting process. Thus, a reasonable place to look for correlations between geological features in the NEUS is in the vicinity of the two prominent MRBs in the NEUS, the Newark rift basin (NRB) and the Hartford rift basin (HRB). Furthermore, it has long been recognized that there is a concentration of earthquakes in the area surrounding the NRB. Although we are far from formulating a real "theory" of NEUS earthquakes, we can at least quantify the extent to which there is an *empirical* correlation between MRBs and earthquakes in this area. In the following section, we demonstrate that there is evidence that the earthquake process in this area is stationary on the time scale of about two centuries, and that the network data provides us with a statistical realization of that process. Using that network data, we test the hypothesis that earthquakes are correlated with MRBs in the study area.

CORRELATION BETWEEN NETWORK AND HISTORICAL SEISMICITY: IS THE EARTHQUAKE PROCESS STATIONARY?

The 1975–1990 catalog resulting from relatively dense network monitoring is probably the best earthquake catalog we will have of the greater NYC area for a while. Before we use that catalog to investigate the relationship between seismicity and geological features, however, we need to address the question of whether the seismicity is stationary, at least over the past couple of centuries during which we have good observations.

Seismicity Data

Figure 2 shows the historical and network seismicity in the study area. For the historical period, that figure shows 160 years of data (1815–1974, $m \geq 3$), and for the network period, 16 years of data (1975–1990, $m \geq 2$). The locations and magnitudes of the historical earthquakes are taken from the NCEER-91 catalog (Seeber and Armbruster, 1991), and those of the network earthquakes are taken from the Weston Observatory archives of the data recorded by the Northeastern United States Seismic Network (NEUSSN). There are many complications regarding how magnitudes are determined from the network seismograms (*e.g.,* Kafka *et al.,* 1985, 1989; Hermann and Kijko, 1983; Ebel, 1994). Furthermore, there are certainly many uncertainties in the estimation of magnitudes during the early

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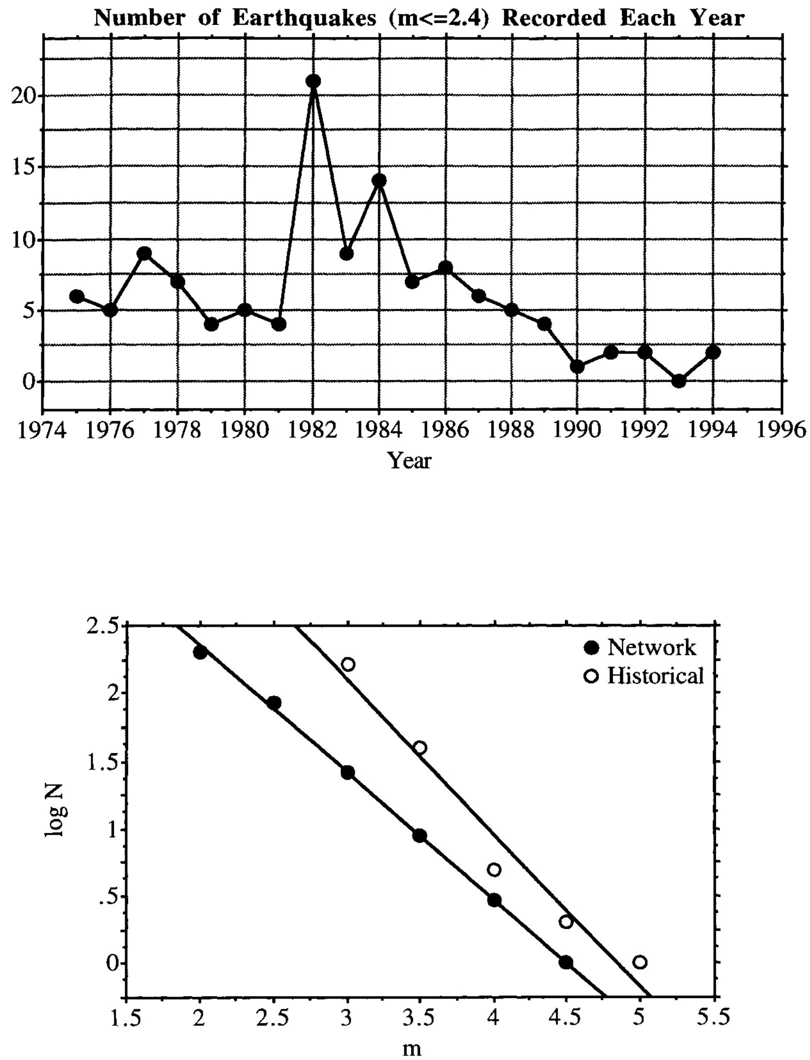

(a)

(b)

	Mean b-value	90% Confidence Interval
Network	0.94	0.90 - 0.97
Historical	1.14	0.73 - 1.55

▲ **Figure 3.** (a) Plot of the number of smaller earthquakes ($m \leq 2.4$) recorded each year in the study area for the time period 1975–1990. (b) Recurrence relations for the network and historical data for the study area.

instrumental and historical periods (*e.g.*, Seeber and Armbruster, 1991). We do not deal with those magnitude uncertainty issues here (except in the sense of investigating correlations between earthquake catalogs for various time periods), and we assume that the magnitudes in the NCEER-91 and NEUSSN catalogs are sufficiently accurate for statistical purposes. For all magnitudes in this paper, we use the notation "*m*," regardless of how the magnitudes were calculated.

The reason why we chose the 1975–1990 time period for the network data is because there appears to have been uniform recording during that period of time (Figure 3). Figure 2(c) shows the seismicity in the study area for the period 1991–1994. During that period of time the number of stations decreased in the study area; a small number of stations are currently operating in this area as part of a transitional measure in preparation for a new generation of seismic networks. To determine when the effect of the decline in funding and number of stations occurred in the study area, we plotted the number of smaller earthquakes ($2.0 \leq m \leq 2.4$) recorded each year [Figure 3(a)]. Notice in Figure 3(a) that, between 1975 and 1989, the number of smaller earthquakes recorded annually in the study area was consistently four or greater. Beginning in 1990, the number of smaller earth-

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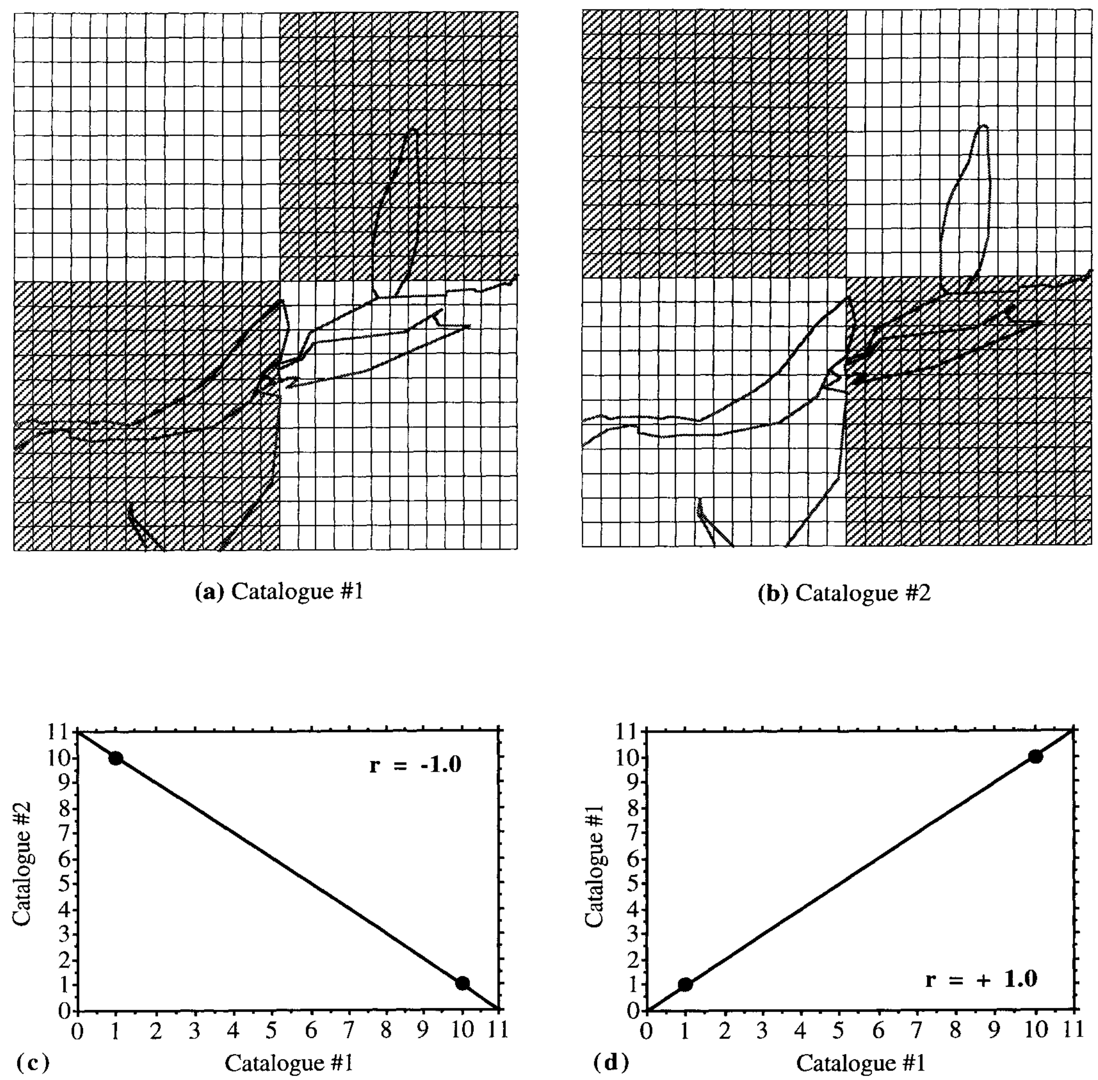

▲ **Figure 4.** Illustration of the methodology used in this study for comparison of earthquake catalogs and for comparison of MRBs with seismicity. The study area was divided into blocks, 0.2° on a side, and the number of earthquakes per unit area was calculated for each block. (a) Hypothetical earthquake Catalog #1. (b) Hypothetical earthquake Catalog #2. (c) Scattergram of number of earthquakes/unit area in a given block for Catalog #1 versus number of earthquakes/unit area in that same block for Catalog #2. (d) Scattergram of the number of earthquakes/unit area in a given block for Catalog #1 versus the number of earthquakes/unit area in that same block for an identical catalog.

quakes recorded each year was two or less. Thus, we suggest that the detection of these smaller events is incomplete after about 1990 (rather than that the seismicity has decreased), because funding of the network has steadily decreased in recent years.

Figure 3(b) shows recurrence relations for the data sets in Figure 2(a) and 2(b), as well as the *b*-values obtained from a least squares fit to those data. For the network data, the *b*-value was found to be 0.94 ± 0.04, and for the historical data, the *b*-value was found to be 1.14 ± 0.41. These results are consistent with the results of Seeber and Armbruster (1991), who obtained a *b*-value of 1.05 ± 0.05 from the NCEER–91 data for the entire eastern US. In the statistical analysis of seismicity and geological features presented below, we view the two data sets shown in Figures 2(a) and 2(b)—representing the 160-year historical and the 16-year network time periods, respectively—as two different realizations of the same earthquake process. If that process is stationary, then we would expect a statistically significant positive correlation between these two data sets.

Methodology

Figure 4 shows a schematic of the methodology we used, both for comparing catalogs with each other and also for comparing seismicity with locations of MRBs. The study area was divided into blocks, 0.2° on a side, and the number of earthquakes per unit area was calculated for each block. Suppose for example that the shaded blocks represent 10

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TABLE 1
Correlation Coefficients (r_s, τ_c)[a]

	HIST160 (1815–1974)	NET8-1 (1983–1990)	HIST80 (1895–1974)
NET16 (1975–1990)	0.26, 0.25		
NET8-1 (1975–1982)		0.18, 0.17	0.19, 0.18
NET8-2 (1983–1990)			0.18, 0.17
Correlation Coefficients (r_s, τ_c) (9-Point Moving Average)			
	HIST160 (1815–1974)	**NET8-1 (1983–1990)**	**HIST80 (1895–1974)**
NET16 (1975–1990)	0.56, 0.42		
NET8-1 (1975–1982)		0.46, 0.37	0.35, 0.28
NET8-2 (1983–1990)			0.42, 0.33

a. Note: All statistically significant at the 99% confidence level.

earthquakes/unit area, and that the unshaded blocks represent 1 earthquake/unit area. Figure 4(c) shows a scattergram of the number of earthquakes/unit area in a given block for a hypothetical Catalog #1 versus the number of earthquakes/unit area in that same block for a hypothetical Catalog #2. Since the catalogs in this example are exactly the opposite of each other, all 10s pair up with 1s and vice versa, yielding a correlation coefficient of –1.0. On the other hand, if we compare Catalog #1 with itself, we obtain a correlation coefficient of +1.0. Any real catalogs will, of course, yield a correlation coefficient between 1.0 and –1.0.

The size of the 0.2° blocks was chosen to represent a rough approximation of the average location accuracy of the events in the historical catalog. Later in this paper, when we describe the application of this methodology to evaluate the extent to which seismicity is correlated with locations of MRBs, we maintain the same 0.2° block size. This approach was taken because we are basing the correlation between MRBs and seismicity on the notion that, at that scale of spatial resolution, the results of this part of our analysis indicate that the earthquake process is stationary on the time scale of a couple of centuries.

The type of analysis described above assumes that the number of earthquakes/unit area in the blocks corresponding to each catalog is normally distributed, which of course we don't know to be true. In fact, the distribution is quite far from being normally distributed, with many zeros representing blocks with no observed seismicity. Thus, it is more appropriate to use non-parametric measures of correlation for these comparisons, and we chose the Spearman rank correlation coefficient (r_s) and the Kendall rank correlation coefficient (τ_c). These non-parametric measures of association are based on the ranks of the data rather than on the actual values of the data, and neither of them require the distributions to be normal (*e.g.*, Henley, 1981). Each of these correlation coefficients is a measure that ranges between –1.0 (for exactly negatively correlated ranks) and +1.0 (for exactly positively correlated ranks). For each test of correlation in this study, the results for r_s and τ_c are similar, and although the values of both are shown in the tables and figures, for simplicity we usually discuss only the r_s results in the text.

Results of Catalog Comparison

Figure 5 shows the results of applying this methodology to the 160-year historical catalog versus the 16-year network catalog. The resulting value of r_s is 0.26. Using a standard test of statistical significance, this result was found to be statistically significant at the 99% confidence level. In the tests of statistical significance of r_s and τ_c used in this study, the null hypothesis is that r_s (or τ_c) is zero, and the observed values are evaluated to determine at what level of confidence the null hypothesis can be rejected. Based on these tests, the r_s and τ_c values for all of the historical versus network catalog comparisons (Table 1) were found to be statistically significant at the 99% confidence level.

A word of explanation regarding the principles underlying the tests of significance of correlation coefficients used in this study: consider the situation for r_s (the tests for τ_c are similar). Let ρ_s be the Spearman correlation coefficient for a large population of 160-year historical versus 16-year network catalogs for this region. The null hypothesis is that these historical and network catalogs are not spatially correlated with each other. Equivalently, ρ_s—the Spearman correlation coefficient corresponding to the population of all possible historical and network catalogs sampled in the same way—is zero. Now imagine randomly selecting many samples of pairs of historical and network catalogs, and each time calculating the value of r_s using the methodology described above. Since by assumption $\rho_s = 0$, we would expect the resulting (observed) r_s's to be close to zero. In the case described in the previous paragraph, the observed value of r_s is 0.26. The statement that this result was found to be statistically significant at the 99% confidence level is equivalent to saying that there is only a 1% chance that the observed value would have been as high as 0.26 if the sample was selected from a population for which $\rho_s = 0$. Thus, we

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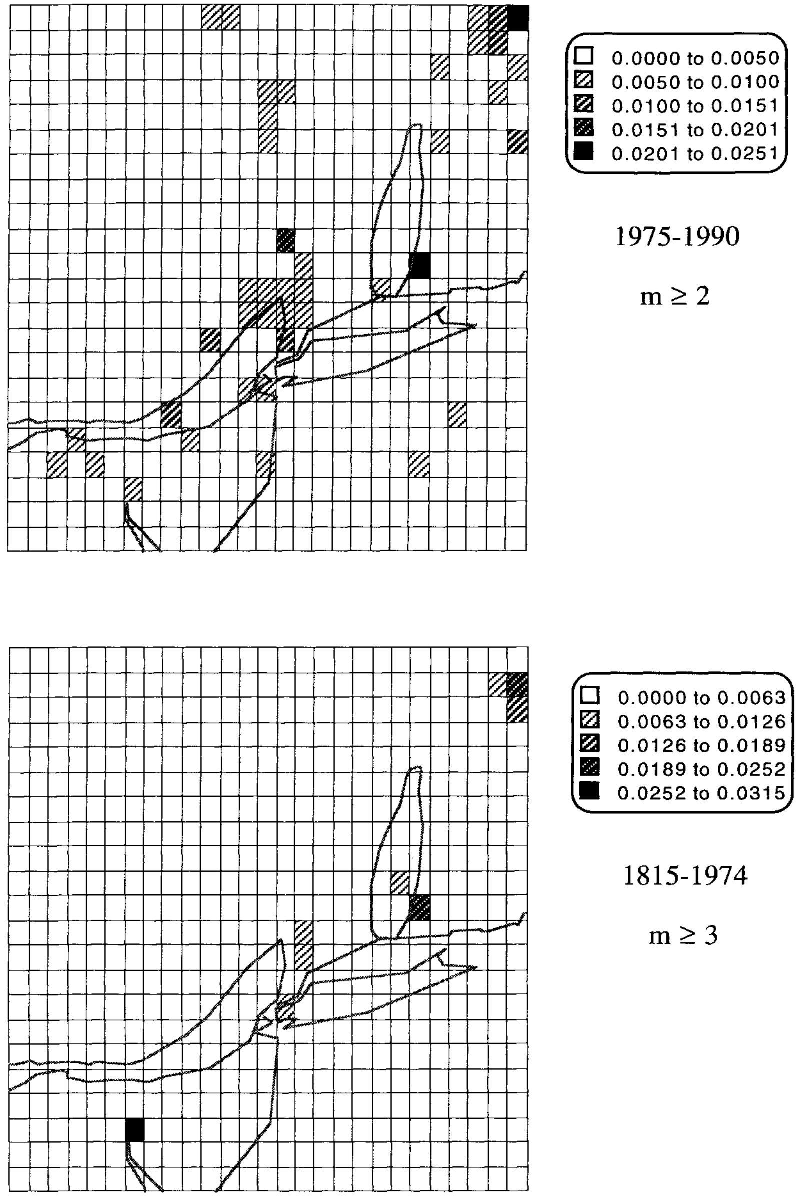

(a)

(b)

▲ **Figure 5.** Number of earthquakes per unit area for 0.2° × 0.2° blocks in the study area for (a) the network data, and (b) the historical data. The correlation coefficients between these two catalogs are 0.26 and 0.25, for r_S and τ_c, respectively. Note that the historical catalog is "peaked" at certain spots. There actually are scattered earthquakes within the unshaded areas that happen to fall within the lowest category, indicated by the unshaded blocks. This artifact no longer appears in the plot of number of earthquakes/unit area when we apply a spatial smoothing filter to that same data [see Figure 6(b)].

reject the null hypothesis and conclude that it is very unlikely that there is no correlation between the historical and network catalogs. To find the probability that the null hypothesis can be rejected, we used standard methods for testing the statistical significance of r_s and τ_c (*e.g.*, Glenberg, 1988; Downie and Heath, 1974). We make no claim to have done an exhaustive analysis of all the issues regarding how well our sampling procedure satisfies the assumptions underlying these standard methods (*e.g.*, Glenberg, 1988), nor have we addressed issues such as prior inspection and the fact that we are analyzing the one sample of pairs of catalogs that is available (*e.g.*, Glenberg, 1988; Wheeler, 1985, 1986; Steinberg and Leonard, 1986). ("Prior inspection" refers to the situation in which an hypothesis is developed after examining a data set, and then that same hypothesis is tested using the same data set, *e.g.*, Wheeler, 1985.) These issues should be addressed in future studies, and the statements in this paper regarding statistical significance should, therefore, be interpreted with an appropriate level of caution.

Figure 6 shows the results of using a 9-point moving average smoothing on the 160-year historical and the 16-year network data, and then performing the same type of comparison we described above. The moving average procedure consists of plotting the average of all values at and

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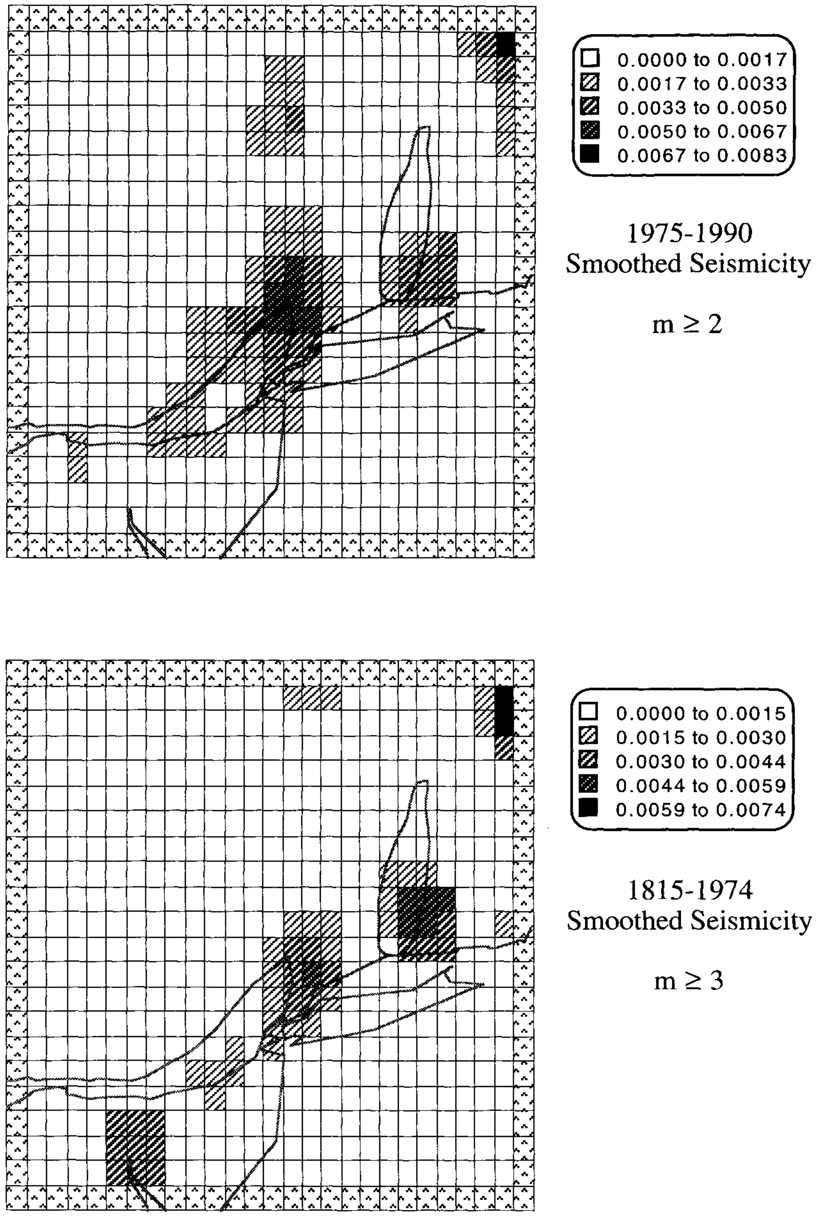
(a)

(b)



▲ **Figure 6.** Number of earthquakes per unit area for 0.2° × 0.2° blocks in the study area, after being smoothed by a 9-point moving average filter as described in the text for (a) the network data, and (b) the historical data. Blocks along the edges of the study area represent no data, which is a result of applying the moving average filter to the data shown in Figure 5. The correlation coefficients between these two catalogs are 0.56 and 0.42, for r_s and τ_c, respectively.

immediately adjacent to each block (*i.e.*, an average over a total of nine blocks). The value of r_s between these two spatially smoothed catalogs is 0.56. This value of 0.56 for r_s (and the corresponding value of 0.42 for τ_c, see Table 1) are essentially the highest correlation coefficients that we obtained for any catalog comparison in this study. Note that eliminating offshore areas (which are poorly characterized in the historical catalog) from the analysis yields a similar r_s value (0.58, and a corresponding τ_c value of 0.43) for the smoothed 160-year historical versus the smoothed 16-year network data. Although it is, in principle, a reasonable approach to remove blocks corresponding to offshore areas from the statistical analysis, the results were not significantly affected by eliminating those blocks. Thus, we do not address this issue further in this paper.

It is also instructive to apply the same reasoning to the two non-overlapping eight-year catalogs that it is now possible to construct from the network data. When we did this for 1975–1982 versus 1983–1990 (using the same moving average smoothing method) we obtained an r_s value of 0.46. Furthermore, we did this for the most recent 80 years of the historical catalog compared with each of the eight-year network catalogs, and we obtained r_s values of 0.35 and 0.42 (for 1975–1982 and 1983–1990, respectively). The results

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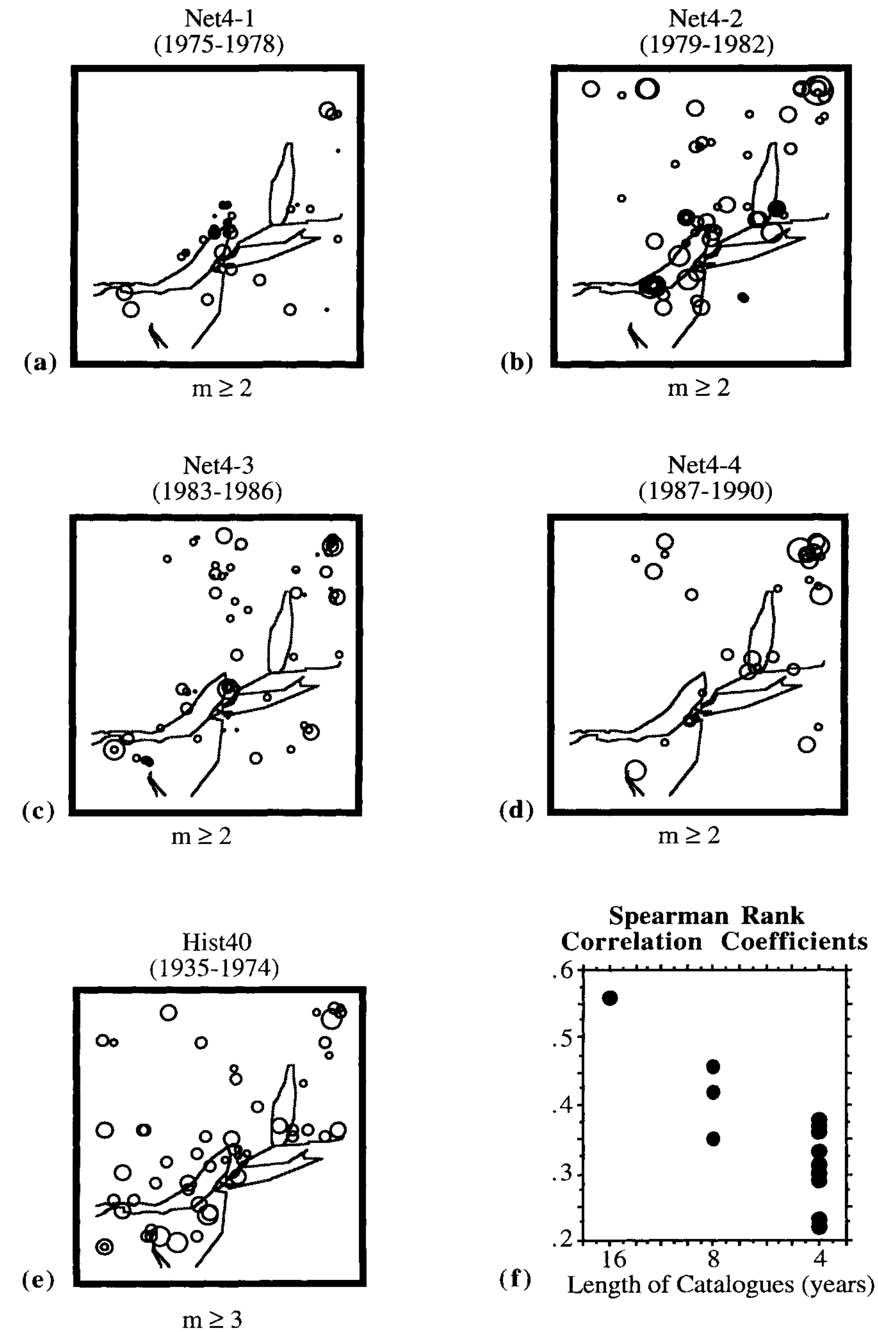

▲ **Figure 7.** Seismicity in the study area for four non-overlapping time segments of the network catalog, and for the most recent 40 years of the historical catalog. (a) 1975–1978, (b) 1979–1982, (c) 1983–1986, (d) 1987–1990, and (e) 1935–1974. (f) Values of r_s as a function of the length of network catalog used for estimating the correlation.

of all these comparisons are shown in Table 1. All of the comparisons in Table 1 yielded positive correlation coefficients and were (based on the tests described above) found to be statistically significant at the 99% confidence level.

Figure 7 shows an additional perspective on the extent to which the seismicity is stationary in the study area. We divided the 16-year network catalog into four non-overlapping four-year time segments. Also shown in Figure 7 is the most recent 40 years of the historical catalog (using $m \geq 3$ as a cutoff instead of $m \geq 2$). Notice that although the seismicity is quite different in each of these time segments, there is a general similarity in the pattern. In fact, the statement that "the seismicity seems to bear some relationship to the locations of MRBs" can be reasonably made based on just about every one of the four-year time periods. The correlation coefficients corresponding to each possible pair of these four-year (and 40-year historical) catalogs are given in Table 2. These results show that (based on the tests described above) there is

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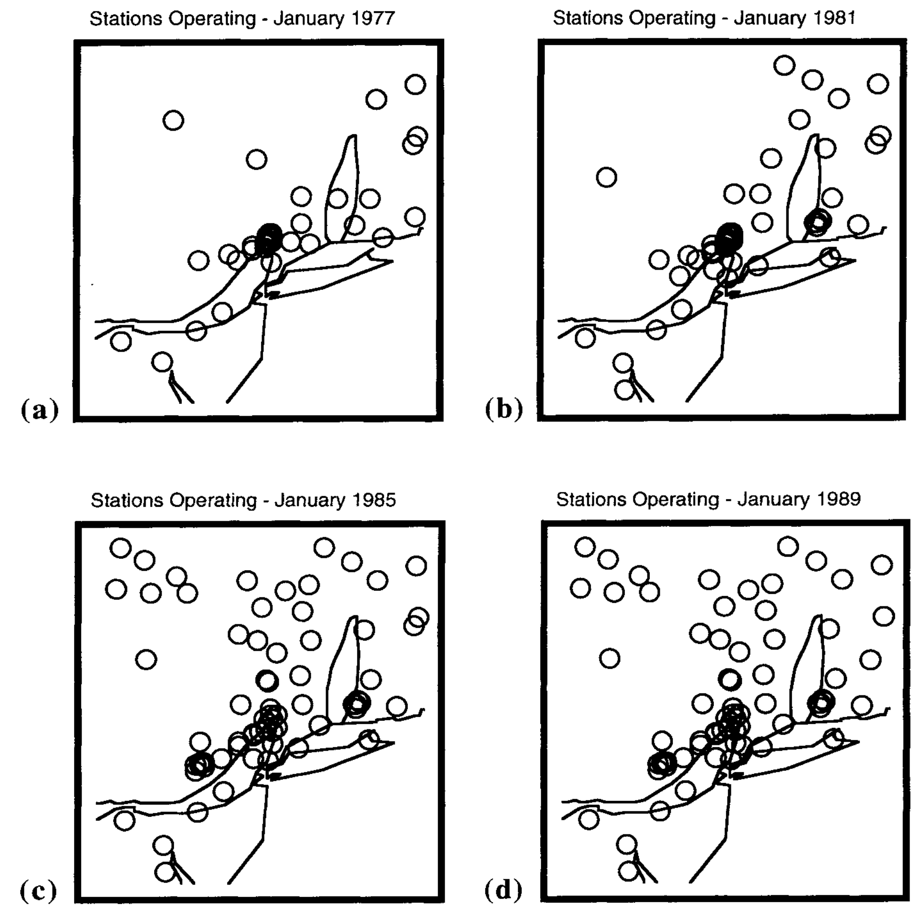
TABLE 2
Correlation Coefficients (r_s, τ_c) (9-Point Moving Average)[a]

	NET4-2 (1979–1982)	NET4-3 (1983–1986)	NET4-4 (1987–1990)	HIST40 (1935–1974)
NET4-1 (1975–1978)	0.29, 0.25	0.37, 0.32	0.22, 0.19	0.36, 0.31
NET4-2 (1979–1982)		0.30, 0.24	0.38, 0.32	0.23, 0.19
NET4-3 (1983–1986)			0.33, 0.28	0.31, 0.25
NET4-4 (1987–1990)				0.22, 0.18

a. Note: All statistically significant at the 99% confidence level.



▲ **Figure 8.** Station distribution for the middle of each of the time periods of the seismicity plots in Figues 7(a-d). These maps were constructed by plotting the stations listed in the NEUSSN bulletins of the first quarter of the third year of each of the four-year time periods represented in Figures 7(a-d).

a statistically significant correlation between each of these catalogs, with r_s correlation coefficients ranging between 0.22 and 0.38. To provide a sense of the possible effects of station distribution on the recording of earthquakes in the four network sub-catalogs shown in Figure 7, we show the station distribution for the middle of each corresponding time period in Figure 8. These maps were constructed by plotting the stations listed in the NEUSSN bulletins of the first quarter of the third year of each four-year time period.

It is interesting to note that a reasonable picture of the general distribution of seismicity (albeit not as complete a picture as we now have) could be obtained from only *four* years of monitoring earthquakes with magnitude 2.0 or greater. Similarly, the most recent 40 years of the historical catalog (which is really the early instrumental catalog) also provide a reasonable picture of the general distribution of seismicity. These observations suggest that, if we were, for example, forced to have a network that could only provide

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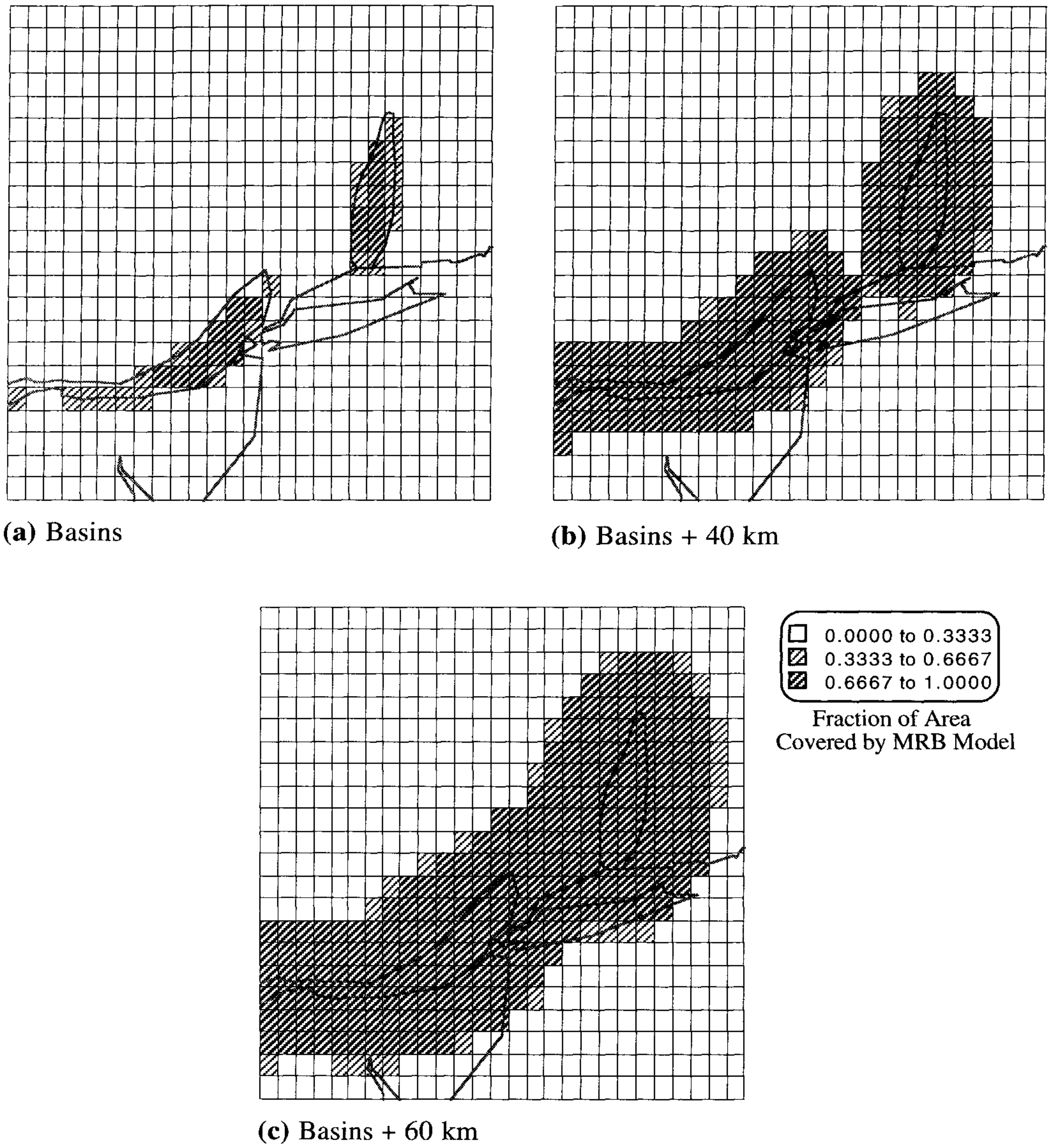

▲ **Figure 9.** Examples of models of MRBs and surrounding areas. (a) Basins, (b) Basins + 40 km, (c) Basins + 60 km.

complete monitoring at the $m \geq 2.5$ level (and if the b-value is about 1.0), we could obtain the same level of statistical basis for correlation between seismicity and geological features in about 50 years as we now have after 16 years.

In Figure 7, we also show the values of r_s as a function of the length of network catalog used for estimating the correlation. If each of the sub-catalogs are realizations of the same earthquake process, then we would expect that, on average, the longer the period of time used for the correlation analysis, the higher the correlation coefficient. Thus, we find it encouraging that the observed pattern shown in Figure 7 exhibits higher correlation coefficients for longer length sub-catalogs. This observation also suggests that, at this level of spatial resolution, the historical data yield a reliable statistical sample of the earthquake process.

Although we still have only a snapshot available of the long-term earthquake process, the fact that all of these correlation coefficients are positive—and that all of the tests of statistical significance indicate very high levels of confidence—provides some evidence that the seismicity is stationary, at least on the scale of a couple of centuries. Next, based on the (presumably) stationary nature of the earthquake process in this area, and the fact that the network was fairly stable and uniform between 1975 and 1990, we tested for correlation between seismicity and MRBs.

CORRELATION BETWEEN SEISMICITY AND MESOZOIC RIFT BASINS

To evaluate the extent to which seismicity is correlated with MRBs in this area, we created spatially varying functions that are intended to represent the spatial distribution of MRBs in the study area (Figure 9). The procedure involves calculating the percentage of area in a given block that is cov-

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ered by the surface mapped areas of an MRB. Thus, in Figure 9(a), an unshaded block represents the fact that 0–33% of the area in that block is covered by an MRB (with 0% representing no basin at all), and the darkest shading represents 67–100% (with 100% meaning that the entire block is within a basin).

Next we used the same methodology that we used for the catalog comparisons to evaluate the correlation between the MRB functions and the smoothed network catalog. We obtained an r_s value of 0.25 for the correlation between the MRB function shown in Figure 9(a) and the smoothed network seismicity shown in Figure 6(a). Using the same tests of statistical significance that we used in the previous section, this result was found to be statistically significant at the 99% confidence level. Thus, there is an *observed* positive spatial correlation between the locations of MRBs and seismicity in this area. Whether this observed correlation means that MRBs are *causally* related to earthquakes in this area is, of course, a different question.

It is clear from Figure 2 (and it has long been recognized) that *the areas surrounding the MRBs,* particularly in the case of the NRB, are more active than the basins themselves. This seismic activity in the area surrounding the MRBs might be a result of the crust being weakened in the area surrounding the basins as part of the continental rifting process (or even before continental rifting began), and this part of the crust may still be weaker than adjacent areas. To quantify the relationship between the network seismicity and the MRBs and surrounding areas, we tested a range of models that included areas surrounding the basins (in addition to the basins themselves).

Our approach was to extend the boundaries of the basins in successive steps by 10 km. For each of these extended basin models, we drew boundaries representing the extended basin models, and then calculated (for each block) the percentage of area covered by the model in the same way that we did for the basin models [see Figures 9(b) and 9(c)]. The results of this analysis are shown in Figure 10(a), where we have plotted r_s as a function of the number of kilometers that the basins were extended. Note that as we increased the area around the basins, the value of r_s increased until we reached about 40 to 50 km beyond the basin boundaries, when r_s began to decrease.

Since it is also clear that there is more activity around the NRB than around the HRB, we performed this same analysis for the individual basins. These results are shown in Figures 10(b) and 10(c). Several things should be noted in those figures. First, not surprisingly, r_s is consistently higher for the NRB than for the HRB. Second, there is a peak in the r_s value (r_s = 0.42) at 40 km for the NRB, while no such peak (or only a hint of such a peak) is observed for the HRB. Third, although the r_s values are quite low for the HRB, it is somewhat surprising that our analysis of the r_s values of 0.12, 0.11 and 0.14 (for the model of the HRB +40, 50 and 60 km, respectively) indicated that all of these results are statistically significant at the 99% confidence level (also see Table 3). Finally, we note that, in the case of the HRB, the boundaries of the extended basin models begin to encounter the higher level of activity around the northern edge of the NRB at about the "HRB +50 km" model, which may be the reason why we don't see a distinct peak in the HRB case.

TABLE 3
Correlation Coefficients (9-Point Moving Average)[a]

Basins+	r_s	τ_c
0	0.25	0.20
10	0.31	0.26
20	0.34	0.28
30	0.38	0.32
40	0.43	0.36
50	0.43	0.36
60	0.40	0.34
NRB+	r_s	τ_c
0	0.27	0.22
10	0.33	0.28
20	0.35	0.29
30	0.39	0.32
40	0.42	0.35
50	0.39	0.33
60	0.36	0.30
HRB+	r_s	τ_c
0	0.04**	0.04**
10	0.06*	0.06**
20	0.08*	0.07*
30	0.09*	0.08
40	0.12	0.10
50	0.11	0.10
60	0.14	0.11

a. Note: All values are statistically significant at the 99% confidence level, except those indicated by asterisks: *90% confidence level, **<90% confidence level.

The differences in the level of activity between the NRB and the HRB might result from different rifting processes for the two basins, and/or from different orientations relative to the present-day stress field. Alternatively, the different levels of activity could be a result of our two-century "snapshot" being a relatively active period of time for the NRB, but a relatively quiet time for the HRB.

We also tested models of areas surrounding the basins, *excluding the basins themselves* (Figure 11). At the level of resolution that we have assumed for this study (*i.e.,* blocks that are 0.2° on a side), the methodology did not discern any dif-

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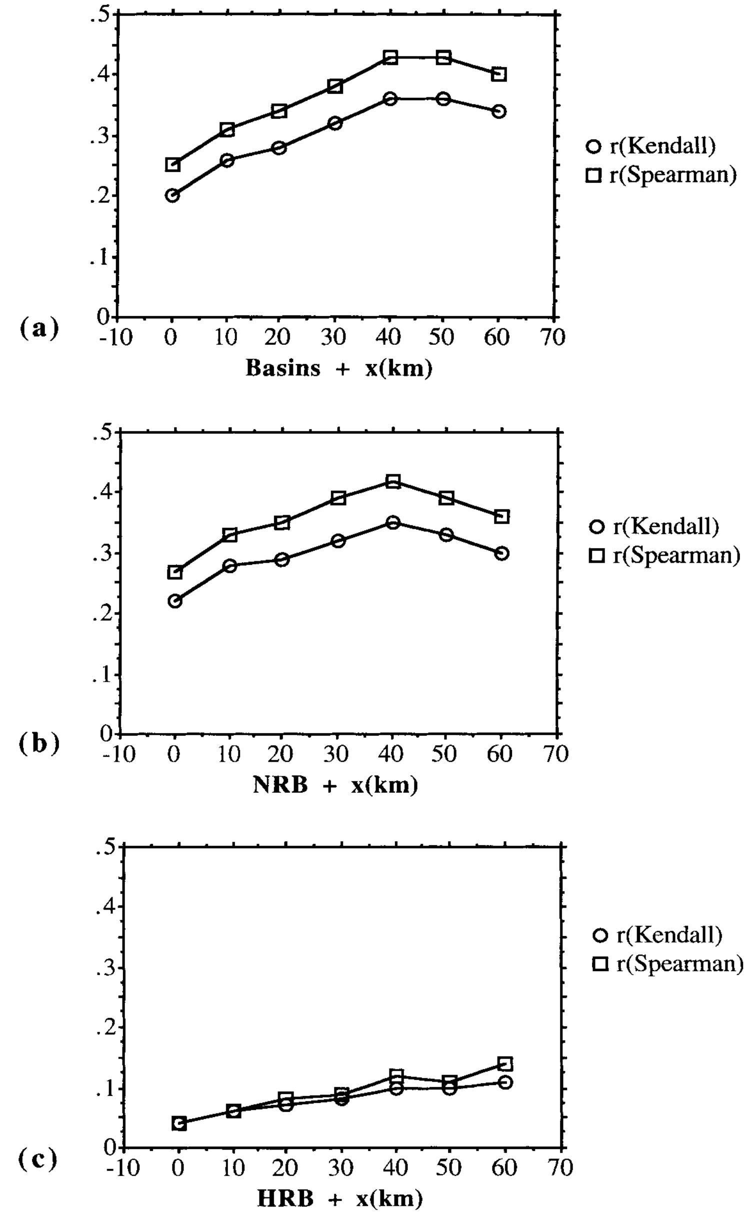

▲ **Figure 10.** Variation of correlation coefficients for different models of MRBs and surrounding areas for (a) both basins, (b) the Newark basin, and (c) the Hartford basin. The correlation coefficients (r_s and τ_c) are plotted as a function of the number of kilometers that the basins were extended.

ference when we excluded the interiors of basins. In fact, we obtained a slightly lower correlation for the model with the interiors excluded.

The highest values of r_s were generally found for the cases where the basins were extended by about 40 km (Figure 10 and Table 3). Figure 12 shows our results for the "Basins +40 km" model, along with the smoothed network seismicity data.

DISCUSSION AND CONCLUSIONS

The past few decades have seen a fascinating array of hypotheses proposed to explain why earthquakes occur in the northeastern United States in general, and in the greater New York City area in particular. Some of those hypotheses have (so far) survived the test of new data from network monitoring and analyses of historical archives, while some required major revision in light of these additional data sets. For example, the currently available evidence is sufficient to rule out (for now?) a Boston-Ottawa Seismic Zone or a concentration of earthquake activity along the Ramapo fault. Nonetheless, the more general (and commonly accepted) hypotheses of reactivation of pre-existing zones of weakness and of stress concentration in areas with contrasting material properties appear to still be viable today. These hypotheses might eventually turn out to be key elements in the search

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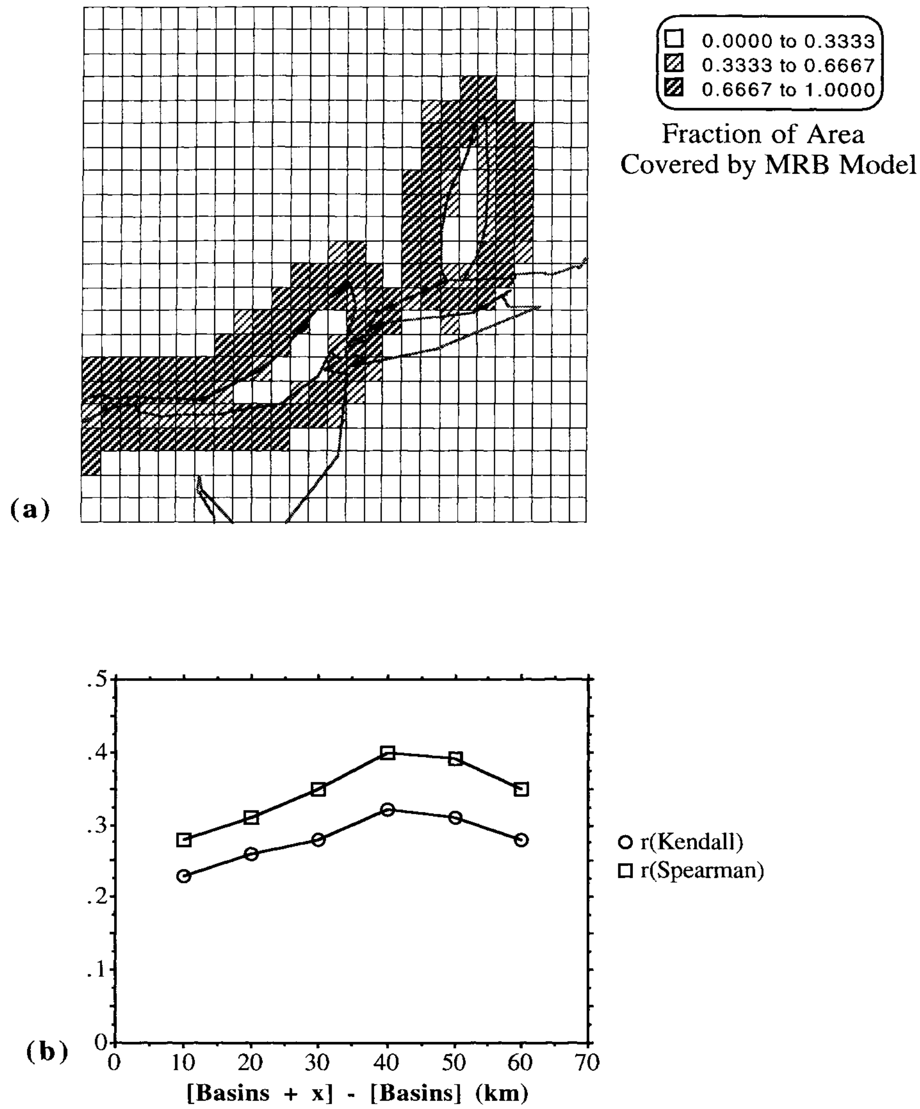

▲ **Figure 11.** (a) Example of models of areas surrounding MRBs. (b) Variation of correlation coefficients (r_S and τ_c) for different models of areas surrounding MRBs.

for the ultimate cause of earthquakes in the area surrounding the Newark and Hartford basins. Even after about two decades of network monitoring, however, there are still insufficient data to decide between these commonly accepted hypotheses and alternative hypotheses, such as the hypothesis proposed by Long (1988) that large intraplate earthquakes are transient phenomena responding to perturbations in crustal strength. In the model proposed by Long (1988), the locations of large intraplate continental earthquakes are essentially independent of preexisting faults and crustal structures. Thus, future large earthquakes in the study area might occur in places that are not currently suspected as being related to seismogenic crustal features, and the seismicity pattern following a large intraplate earthquake might be a transient, extended aftershock sequence of the main event.

On a global scale, a fairly clear picture seems to be forming in which the larger intraplate earthquakes are concentrated in areas where the crust has been stretched or extended in the continental rifting process. This observation may, in fact, hold the key to an eventual explanation of the cause of earthquakes in the greater New York City area. Additional monitoring and historical archive analyses will continue to be necessary as the primary component in research on the identification of active features in this area.

For the time being, the network catalog of events between 1975 and 1990, with $m \geq 2.0$, is probably the best catalog to use for evaluating the relationship between seismicity and geological features. Beginning in 1991, the number of smaller events recorded by the networks began to decrease, suggesting that incomplete recording of events with magnitude below about 2.5 began at about that time (presumably due to the decrease in the number of network

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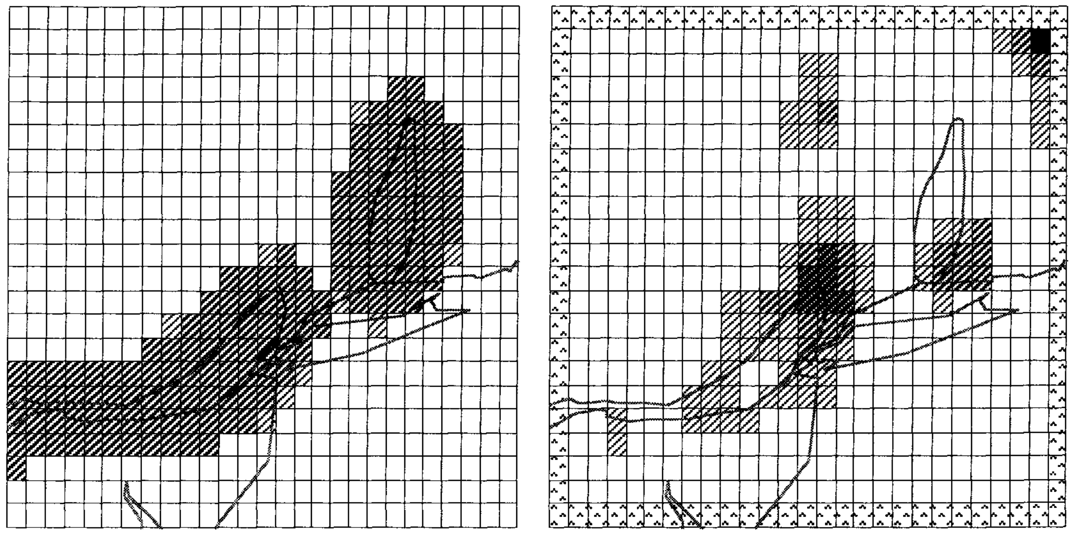
r_s = 0.43



(a) Basins + 40 km (b) Smoothed Network Seismicity

▲ **Figure 12.** Results of analysis of correlation between (a) model of MRBs plus 40 km areas surrounding MRBs, and (b) smoothed network seismicity. The correlation coefficients between these two spatially varying functions is 0.43 (r_s) and 0.36 (τ_c).

stations). Based on our analysis of that 1975–1990 data, as well as the analysis of historical data for the 160 years preceding 1975, the seismicity appears to be stationary, on the time scale of a couple of centuries.

There is an *observed* positive correlation between seismicity and the locations of Mesozoic rift basins in the greater New York City area. Based on standard statistical tests, the associated correlation coefficients were found to be statistically significant at a very high level of confidence. The correlation is stronger for the Newark basin than for the Hartford basin. Nonetheless, the data also indicate, at a high level of confidence, that there is a correlation between the seismicity and the area surrounding the Hartford basin. Whether that means that Mesozoic rift basins are *causally* related to earthquakes is, as we have said, a whole other question. If the reason why earthquakes are correlated with these basins is that the crust has been weakened in the area surrounding the basins, then it would appear that the weakened zone encompasses an area extending to about 40 to 50 km around the basins.

We were surprised that our method of analysis resulted in such high levels of confidence (99% in most cases) for the statistical significance of our findings. This may be an artifact of the specific way in which we chose to set up the analysis procedures, and may be related to such issues as prior inspection of the data and the fact that we are analyzing the one sample that is available. These issues should be addressed in future studies, as the techniques presented in this paper are further developed and refined.

With the lower number of stations currently operating in this area, the threshold for complete recording is probably about magnitude 2.5. Given that level of monitoring, we will have to wait a long time for $m \geq 2.5$ seismicity to reveal the longer-term pattern. An important issue to resolve in future studies of this area will be whether or not the pattern delineated in this study changes. Is the apparently stationary pattern of seismicity a fundamental property of the earthquake process in this area, or are we only seeing a short time segment of a process that is changing on a longer time scale? For now, perhaps all we can conclude is that, the longer this pattern persists, the more confident we can be that we are observing a complete picture of the earthquake process.

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