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Frequency-size distributions for intraplate earthquakes

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ABSTRACT

We examine the question of a possible difference in the frequency-size statistics of intraplate earthquakes, as opposed to their more numerous interplate counterparts. We use both the Harvard Centroid Moment Tensor catalogue and the data set of the National Earthquake Information Center. In the former case, we quantify earthquakes through their seismic moment and describe their population distribution through the β-value introduced by Molnar. In the latter case, we use traditional *b*-values computed from both body-wave magnitudes (m_b) and surface-wave magnitudes (M_s). We conclude that both β- and *b*-values for true intraplate earthquakes (i.e., not occurring in areas of broad tectonic deformation) are essentially equivalent to those of interplate earthquakes in similar ranges of moments or magnitudes. This is consistent with a fractal dimension of two for the intraplate seismogenic zones, suggesting that, like along plate boundaries, they consist of two-dimensional faults and not of volumes with greater dimensions. The distribution of earthquakes in deformed regions, principally the Mediterranean-Tethyan belt, follows that of worldwide interplate earthquakes but with a greater value for the critical moment expressing the saturation with depth of the width of the fault at the brittle-ductile transition, suggesting that the latter would take place at greater depths under large-scale orogens.

Keywords: *b*-values, intraplate earthquakes, seismic scaling laws.

INTRODUCTION AND BACKGROUND

The purpose of this paper is to explore the possibility that intraplate earthquakes could feature frequency-size population statistics that differ significantly from those of their counterparts at plate boundaries. We conclude that "true intraplate" earthquakes, i.e., those associated neither with deformed, diffuse plate boundaries, nor with intraplate magmatic centers (hotspots), do not exhibit recognizably different properties in this respect.

It has long been observed in all seismic provinces that there are, simply speaking, more small earthquakes than large ones. Following the introduction by Richter (1935) of the concept of magnitude, Gutenberg and Richter (1954) modeled this behavior for homogeneous populations of earthquakes (e.g., belonging to a particular seismic area) using a frequency-size relation of the form

$$\log_{10} N = a - bM, \qquad (1)$$

where N is the number of earthquakes in the group with a magnitude equal to or greater than M. The absolute value of the slope of the regression in Equation 1, universally known as the *b*-value of the population, has been found to be remarkably constant and close to unity for a large number of data sets of tectonic earthquakes. It is only when earthquake sources of a different nature are considered that the *b*-value departs significantly from 1, rising, for example, to $b \geq 2$ during volcano-

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seismic swarms (e.g., Mogi, 1963; Minakami, 1974). Such variations have indeed been used to identify volcanic swarms.

Following the generalization of the use of the seismic moment M_0 as a physical measure of the size of an earthquake, Molnar (1979) proposed the symbol β to similarly describe earthquake statistics based on seismic moments, according to

$$\log_{10} N = \alpha - \beta \log_{10} M_0, \qquad (2)$$

where N is now the number of earthquakes with a moment that equals or exceeds M_0. As discussed in detail by Okal and Romanowicz (1994), the relationship between β and b depends on the variation of any particular magnitude scale with seismic moment, which is itself controlled in part by the saturation of magnitude scales as the seismic source grows, which affects first m_b (around 6.3), then M_s (around 8.2) (Geller, 1976).

A considerable amount of literature has been published on the subject of frequency-size distributions, and the reader is referred to, for example, Båth (1981) and Frohlich and Davis (1993) for reviews. Perhaps the most seminal among such papers was Rundle's (1989), in which the author justified a β-value of 2/3 based on the following argument: In a given population of earthquakes, the process of faulting is assumed to be scale-independent (in Rundle's own words "the fault area available to produce events of all sizes is the same" p. 12,338 of Rundle [1989]). This is equivalent to the number of earthquakes of a given size, dN in the notation of Equation 2, being inversely proportional to the area of faulting, S. Earthquake scaling laws (e.g., Geller, 1976) predict that S grows like $M_0^{2/3}$ (at least in a range of source sizes not affected by the intrinsic limits of the dimensions of seismogenic zones), resulting in the theoretical values $\beta = 2/3$ and $b = 1$, as a factor 1.5 has often been introduced between various magnitude scales and $\log_{10} M_0$. Note that the values predicted in Rundle's (1989) derivation of Equations 1 or 2 based on simple physical laws are exact numbers ($\beta = 2/3$ or $b = 1$).

In this respect, it is interesting to note that in the first edition of their classic work, Gutenberg and Richter (1941) described the "rule of tenfold increase," in other words a b-value of exactly 1 as defined by Equation 1. In their more definitive work (Gutenberg and Richter, 1954), they refined this estimate to $b = 0.9$ for shallow earthquakes. Later, Turcotte (1992) related the b-value of a population to the fractal dimension D of its source, assigning $D = 1.8$ on the basis of $b = 0.9$ as reported by Gutenberg and Richter (1954). In this context, Okal and Romanowicz (1994) argued that the exact value $b = 1$ is physically more realistic, since it expresses the geometrical dimension ($D = 2$) of the seismogenic zone, constrained to a two-dimensional *fault*; any small departure from $b = 1$ then illustrates artifacts of the physical saturation of the seismogenic zone, as well as variations with earthquake size in the expected relation between magnitude and moment. Due to the generally small number of large events, there remains some controversy about the actual behavior of β at very large moments—some studies suggest a decrease in β, while others advocate an increase (Romanowicz and Rundle, 1993).

In the case of deep earthquakes, Okal and Kirby (1995) argued that the seismogenic zone may extend over a *volume* of fractal dimension $D = 3$, thus explaining larger β and b-values, as already reported by Gutenberg and Richter (1954). This trend is strongly affected, however, by the finite size of the seismogenic zone inside slabs, which explains the large diversity of observations in various subduction zones (Frohlich and Davis, 1993).

In this general context, we examine here the values of β and b for various groups of intraplate earthquakes, not directly associable with tectonic processes occurring along plate boundaries. We were motivated by the idea that intraplate earthquakes may not necessarily be constrained to a given fault system of fractal dimension $D = 2$. For example, rupture in a seismogenic *volume* of fractal dimension $D = 3$, rather than along a two-dimensional fault, would lead to $\beta = 1$ and similarly to larger values of b (Okal and Kirby, 1995). We also note that Frohlich and Davis (1993) explored the concept of possible regional variations in b-values according to prevailing tectonic regime. Similarly, Bird et al. (2002) presented a formal regionalization of tectonic boundaries and studied the variation of frequency-size relationships among them. While the earlier study by Bergman and Solomon (1980) did address the question of b-values in a few specific oceanic intraplate areas, it lacked a worldwide scope and predated the routine quantification of earthquakes through seismic moment; the other studies mentioned did not consider intraplate earthquakes in their regionalizations.

METHODOLOGY

The two principal catalogues used in the present study were the Harvard Centroid Moment Tensor (CMT) data set (Dziewonski et al., 1983, and subsequent quarterly updates) (1977–March 2003) and the database of the National Earthquake Information Center (NEIC) of the U.S. Geological Survey. The two catalogues were filtered for depth, and only shallow earthquakes ($h < 100$ km) were retained. This threshold may appear exceedingly deep, as most intraplate events are known to occur at lesser depths. However, it has no effect on the final results, as our search algorithms retained insignificant populations at the greater depths (e.g., only 18 earthquakes with $50 < h < 100$ km out of 2737 for the intraplate NEIC population; see following). Such events are generally small and poorly located, making their published depths suspect. The use of a conservative depth threshold (100 km) may indeed guard against inadvertently excluding genuine shallow intraplate events, with once again no tangible effect on our final results.

The intraplate character of an earthquake was assessed by testing the minimum distance from its epicenter to Bird's (2003) discrete set of 12,148 plate-boundary locations. We used a threshold distance of 400 km from the nearest plate boundary to define an earthquake as intraplate. This rather conservative estimate is larger than previously used, for example by Wysession et al. (1991) in the study of Pacific Basin intraplate seismicity. It ensures the elimination of events potentially associated with boundary processes, such as buckling of the plate seaward of a subduction zone. Also, since it is significantly larger than the typical sample spacing in Bird's data set (70 km), it ensures that the identification of an intraplate earthquake is not affected by the discrete nature of the latter.

THE CMT DATA SET

Focusing first on the CMT catalogue, which provides the most homogeneous data set based on the quantitative inversion of a physical model of the seismic source, the above procedure resulted in the retention of only 821 events among the 16,349 shallow CMT solutions (Fig. 1). This data set was further refined by separating those earthquakes that belong to so-called "deformed" provinces, to diffuse plate boundaries, and to identifiable intraplate magmatic centers (hotspots). Deformed provinces are mostly continental, mostly compressional, belts over which intense deformation takes place across widths significantly larger than the 400 km threshold. They include the Tethyan belt, extending from Italy to Burma, including the Tibet-Mongolian system, as well the Rocky Mountain system in North America, and additional pockets of activity in South America and Africa. The 565 events in those provinces are shown as downward-pointing triangles on Figure 1.

An additional 22 events, shown as upward-pointing triangles on Figure 1, were classified as belonging to areas of diffuse plate boundaries, as defined, for example, by Stein and Sella (2002). These earthquakes were located in the eastern Indian Ocean (between the Indian and Australian plates) and near the Macquarie triple junction, the latter of which consisted of the large 1998 earthquake and its aftershocks.

Finally, we classified separately 16 earthquakes evidently associated with activity at hotspots, principally in Hawaii and the Canary Islands; these are shown as squares on Figure 1. This left 218 "truly intraplate" events in the CMT catalogue, shown as circles on Figure 1. Note that this number represents only a very small fraction of the total shallow seismicity of the catalogue (1.3% of the number of earthquakes; 0.3% of the seismic moment released).

β-Values from the CMT Data Set

In order to provide a worldwide reference for the study of intraplate data sets, we first analyzed the frequency-moment relationship for the entire CMT catalogue of 16,349 shallow earthquakes. Individual events were sorted by moment into bins with a width of 0.2 units of $\log_{10} M_0$, with the corresponding populations shown as plus signs on Figure 2A. The larger symbols represent cumulative populations, i.e., for each bin, the number of events (N in Equation 2) with a moment equal to, or greater than, that of the bin. The β-value was obtained by a least-squares regression of N against $\log_{10} M_0$ over specific ranges of moments. The open symbols denote parts of the data set that lacked completeness and thus were ignored from the regressions. The full data set clearly exhibits a change in slope at a critical moment $M_0^c = 10^{27.5}$ dyn-cm, as previously observed by many investigators (Pacheco et al., 1992; Romanowicz and Rundle, 1993; Okal and Romanowicz, 1994) and interpreted as expressing the saturation of fault width W upon reaching the brittle-ductile transition. Our β-values (0.67 below M_0^c, 1.42 above M_0^c) are in good agreement with Okal and Romanowicz' (1994) values (0.70 and 1.35), which were obtained from a data set roughly half the size of ours. We note however that our critical moment is slightly larger than that observed ($10^{27.2}$) in the previous study. The excellent agreement between the β-value at low moments and its theoretical value (2/3) also confirms the exact fractal dimension $D = 2$ of sources unaffected by saturation.

We then applied the same algorithm to the "deformed" and "true intraplate" populations, of 565 and 218 earthquakes, respectively. (With only 22 and 16 earthquakes, respectively, the diffuse and hotspot populations were too small to provide meaningful results.) As shown on Figure 2B, the "deformed" data set features a slightly lower $\beta = 0.58$, with only the hint of a possible elbow at $M_0 = 10^{27.2}$ dyn-cm. In the case of the "intraplate" population (Fig. 2C), $\beta = 0.66$ is indistinguishable from its worldwide value and from the theoretical value of 2/3 predicted in Rundle's (1989) model. However, because of the absence of large events in the data set, no critical moment M_0^c can be defined in that case.

b-Values from the CMT Data Set

We next examined the same data set of CMT solutions, but considered its conventional magnitudes. The regression algorithm is similar, but uses bins of 0.2 units of magnitude. We computed b-values from the surface-wave magnitude M_s (measured at 20 s) and from the body-wave magnitude m_b (measured at 1 s).

About three-fourths of the 16,349 events in the global CMT database are assigned M_s values; this data set features a stepwise increase in b-value with magnitude (Fig. 3A) from $b = 0.73$ in the unsaturated range ($M_s < 6.6$) to around $b = 2$ for $M_s \geq 8$; these values are in full agreement with Okal and Romanowicz' (1994) results (their Figs. 9b and 9c). The 492 events in "deformed" provinces similarly follow the expected b-value of 2/3 at low magnitudes, for which M_s is directly proportional to $\log_{10} M_0$, but fail to involve a definitive change of slope at larger magnitudes (Fig. 3B). In contrast, the 177 "intraplate" events, which also closely follow $b = 2/3$ at low magnitudes, appear to involve a different behavior beyond $M_s = 6.5$ (Fig. 3C), but this observation is based on only a handful of events.

When regressed as a function of body-wave magnitude m_b, the full CMT data set features an almost continuous increase in b-value from 1.04 for $m_b < 6.0$ to more than 2.5 around $m_b \approx 7$ (Fig. 3D). These results are, once again, in agreement with Okal and Romanowicz' (1994) (see their Fig. 12). Results for the "deformed" and "intraplate" data sets were essentially equivalent (Figs. 3E and 3F).

In conclusion, the data set of earthquakes inverted as part of the CMT project suggests that intraplate earthquakes follow the same population distributions as their much more numerous counterparts at plate boundaries.

CMT Data set

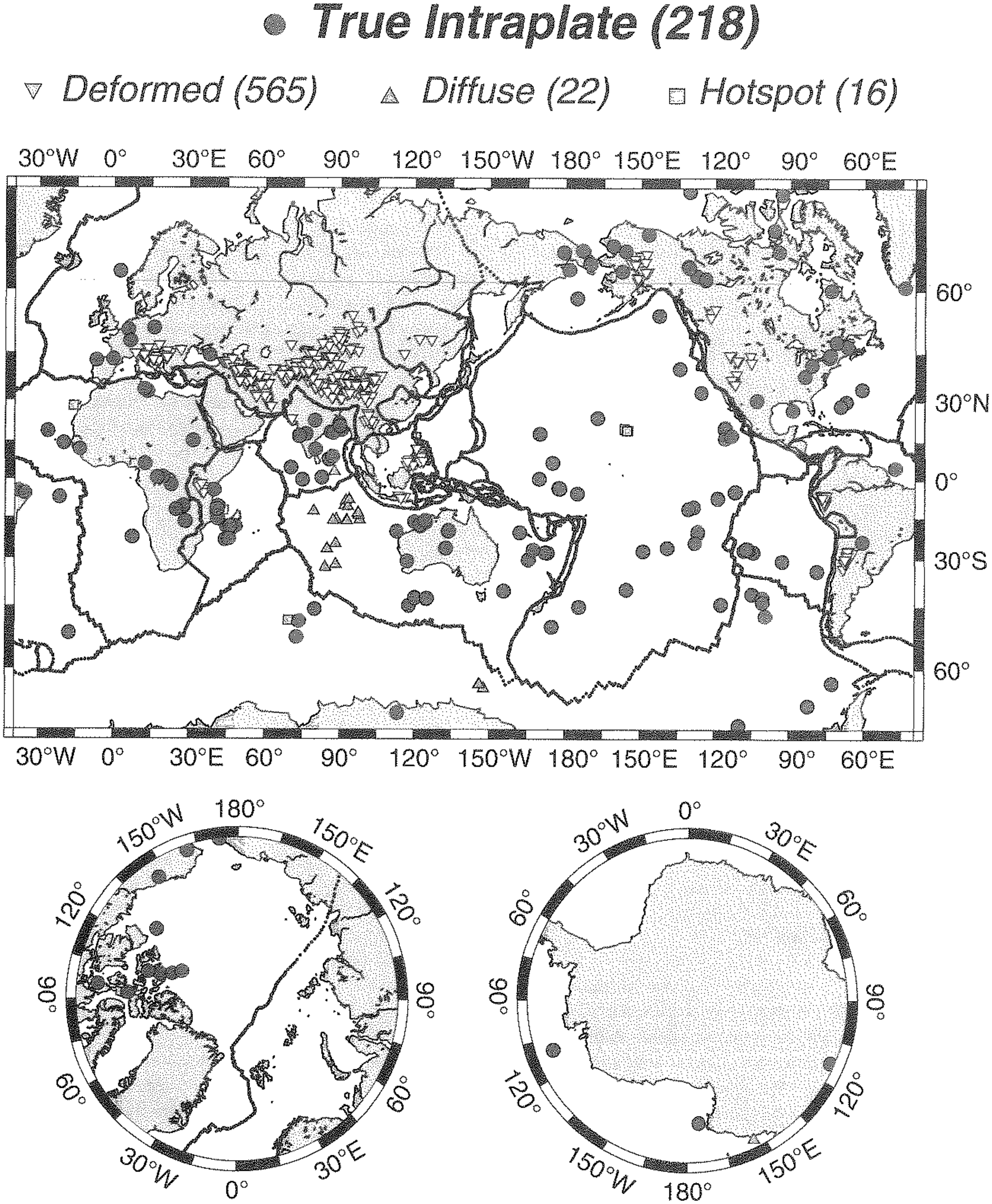

Figure 1. Maps of the Harvard Centroid Moment Tensor (CMT) data set considered in this study. The solid lines (actually composed of 12,148 individual dots) are Bird's (2003) data set of plate-boundary locations. Downward-pointing triangles show events occurring inside deformed areas, upward-pointing triangles events located at diffuse plate boundaries, and squares events associated with intraplate hotspots. The remaining truly intraplate earthquakes are shown as solid circles. Numbers in parentheses show the total populations of the various groups.

THE NEIC DATA SET

While the CMT data set is unique in providing a homogeneous catalogue of earthquake solutions inverted through a common algorithm, it suffers from its youth (no events prior to 1976 and a coarser sampling for that year) and its relatively large size threshold for completeness (in principle, no events below $m_b = 5.0$ are processed into the catalogue, resulting in completeness only for $M_0 \geq 2 \times 10^{24}$ dyn-cm). Accordingly, we consider in this section b-values computed from data sets extracted from the NEIC catalogue, keeping in mind that their interpretation may be more delicate than corresponding β-values due to the progressive saturation of magnitude scales (especially m_b) with earthquake size and its effect on apparent b-values (Okal and Romanowicz, 1994).

For this purpose, we used the NEIC database, extending back to 1963, when systematic reporting started for m_b, as defined by the Prague formula (Váněk et al., 1962). M_s, defined in the same forum, is catalogued starting in 1968. The NEIC database was used up to and including 2002. Out of the 474,203 earthquakes listed in the NEIC database, and applying the same distance threshold as for the CMT database, we identified 2737 truly intraplate shallow earthquakes with $m_b \geq 4$, after eliminating 8614 events belonging to "deformed" regions, 124 to areas of diffuse plate boundaries, and 257 earthquakes presumably correlated with hotspots. Figure 4 shows the geographical repartition of these various data sets. Note that their occasional clustering may reflect the existence of local networks (e.g., in Australia), rather than true regional variations in seismic activity.

Similarly, we identified 2019 events in the NEIC catalogue with a reported M_s, of which 1427 were in deformed areas, 33 were in diffuse regions, and 40 were associated with hotspots, leaving 519 true intraplate earthquakes.

b-Values for the m_b Populations

Figure 5A illustrates a two-segment population for truly intraplate earthquakes, with $b = 0.84$ at lower magnitudes and $b = 2.08$ for $m_b \geq 6.0$. These values are in agreement with those predicted theoretically by Okal and Romanowicz (1994): $b = 2$ in the range $5.7 \leq m_b \leq 6.6$ and b increasing from 2/3 to 1 at lower magnitudes. They are also similar to those obtained by the same authors for a much larger worldwide data set of 90,074 events.

Results are essentially similar for NEIC data sets in the "deformed" regions, which feature an increase from $b = 1.16$ at low magnitudes to $b = 2.06$ for $m_b > 6.0$ (Fig. 5B). Regarding the hotspot and diffuse data sets (Figs. 5C and 5D), the predicted steep trend at large magnitudes ($b = 2$) cannot be resolved on account of their small populations. The increase from $b \approx 2/3$ to $b \approx 1$ at lower magnitudes is well resolved for hotspot events, probably due to the existence of local networks, but not for diffuse regions, where detection capabilities are poorer at low mag-

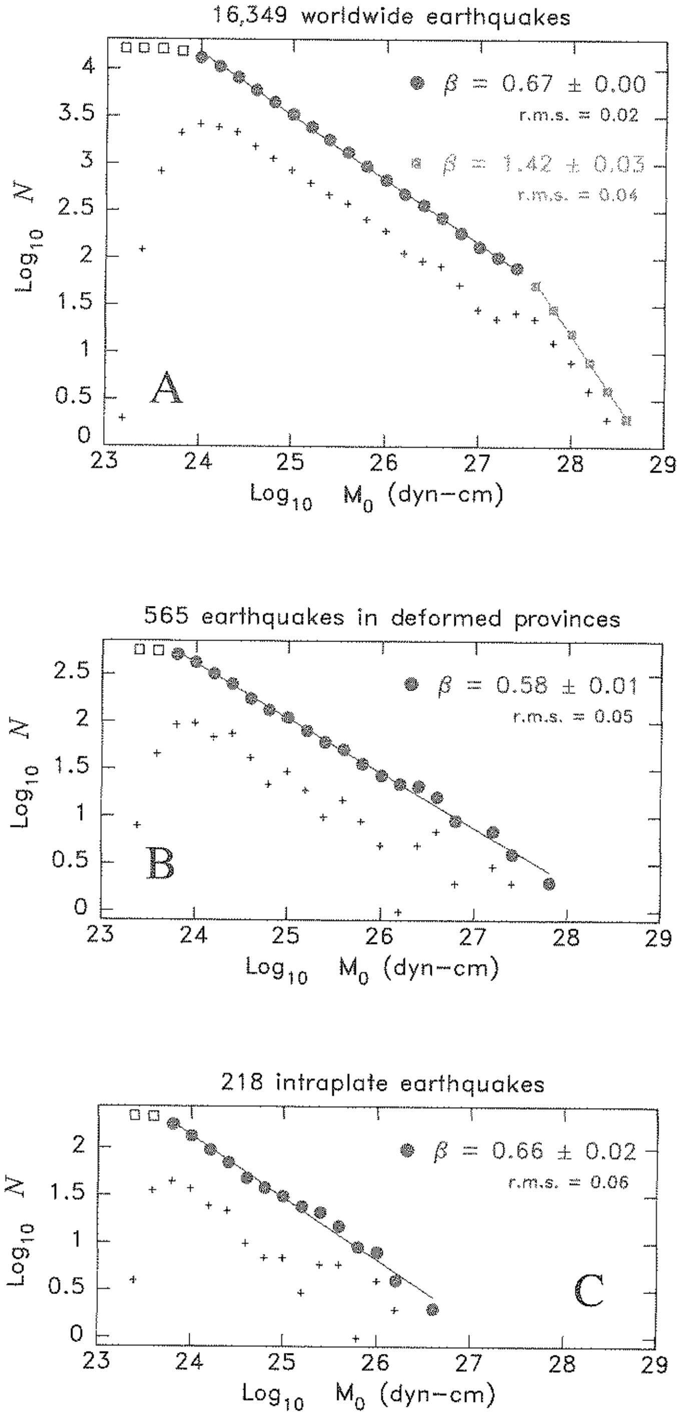

Figure 2. Determination of β-values for the Harvard Centroid Moment Tensor (CMT) data sets (r.m.s. is root mean square). We used bin widths of 0.2 logarithmic units of seismic moment M_0. On each frame, the + signs represent the populations of earthquakes in each bin, and the larger symbols denote the cumulative populations. The open squares at low moments represent bins affected by undersampling due to loss of completeness of the data set. The large solid symbols (circles and squares) show ranges of satisfactory linear regression, with corresponding β-values listed at right. (A) Full CMT data set of shallow earthquakes. (B) CMT solutions in deformed provinces. (C) True intraplate earthquakes. See text for discussion.

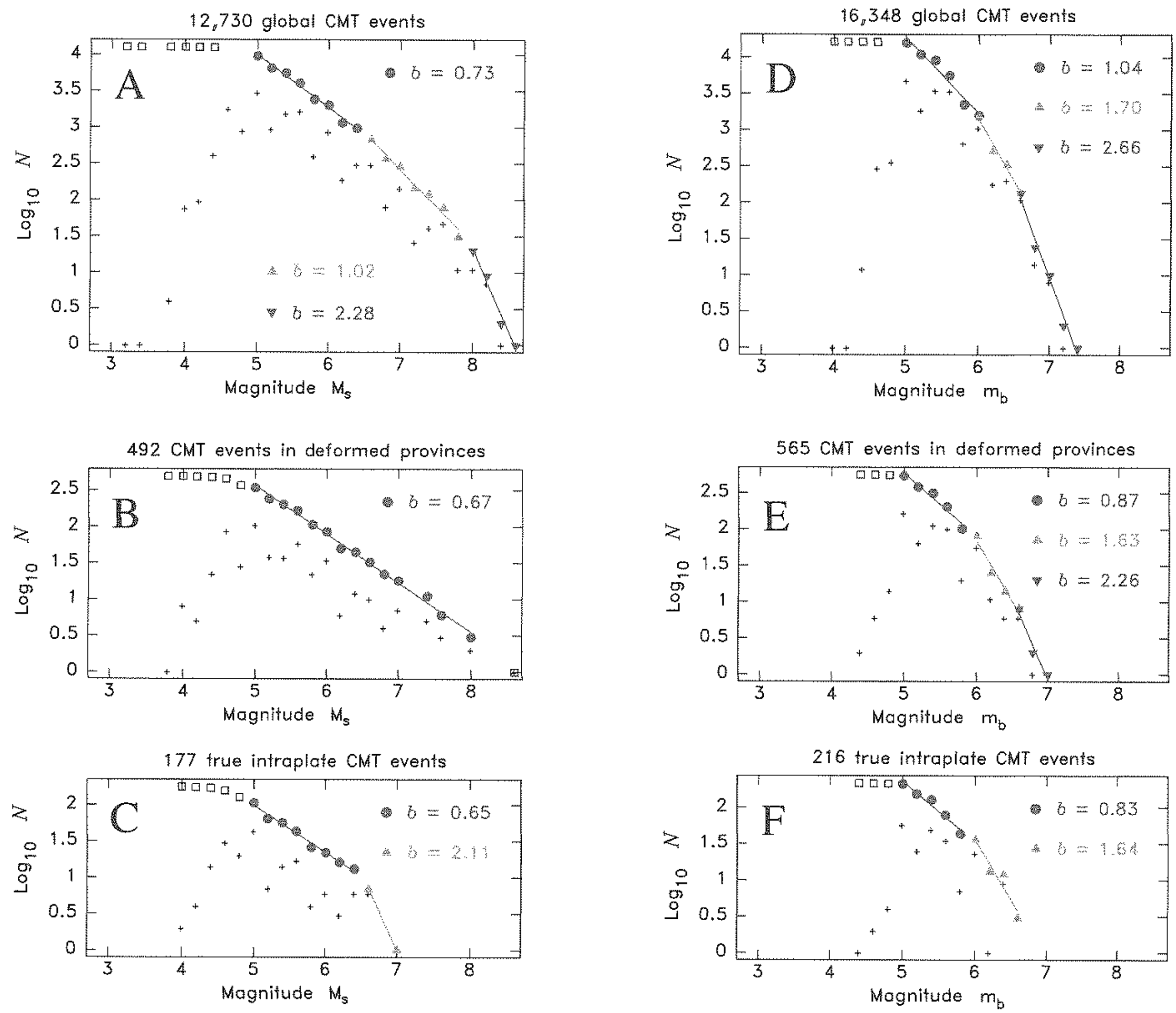

Figure 3. Determination of b-values for the Harvard Centroid Moment Tensor (CMT) data sets. The layout of the individual frames is similar to that in Figure 2, except that events are now binned according to their catalogued magnitudes M_s (left) or m_b (right). See text for discussion.

nitudes. In summary, none of the data sets examined exhibits a behavior of the m_b populations recognizably different from that of the global data set, as investigated and justified theoretically by Okal and Romanowicz (1994).

b-Values for the M_s Populations

The largest data set, relative to the deformed regions, features $b = 0.63$ at lower M_s, which is in reasonable agreement with the theoretical value of 2/3 (Okal and Romanowicz, 1994), although the expected transition to higher values of b is not as prominent as for global data sets (Fig. 6B). By contrast, the 519 "true" intraplate earthquakes have $b = 0.78$ for $M_s \leq 6.6$, which is in acceptable agreement with the predicted $b = 2/3$, and to $b = 2.39$ at higher magnitudes (Fig. 6A). The position of the elbow ($M_s = 6.7$) is remarkably consistent with Okal and Romanowicz' (1994) predictions and observations. The hotspot and diffuse data sets were too small for meaningful regressions.

TARGETED REGIONAL DATA SETS FROM THE NEIC CATALOGUE

In this section, we complement the previous approach with a detailed look at a number of specifically targeted areas, namely the Pacific, North American, and African plates, which regroup most of the intraplate CMT data set, as illustrated on Figure 1.

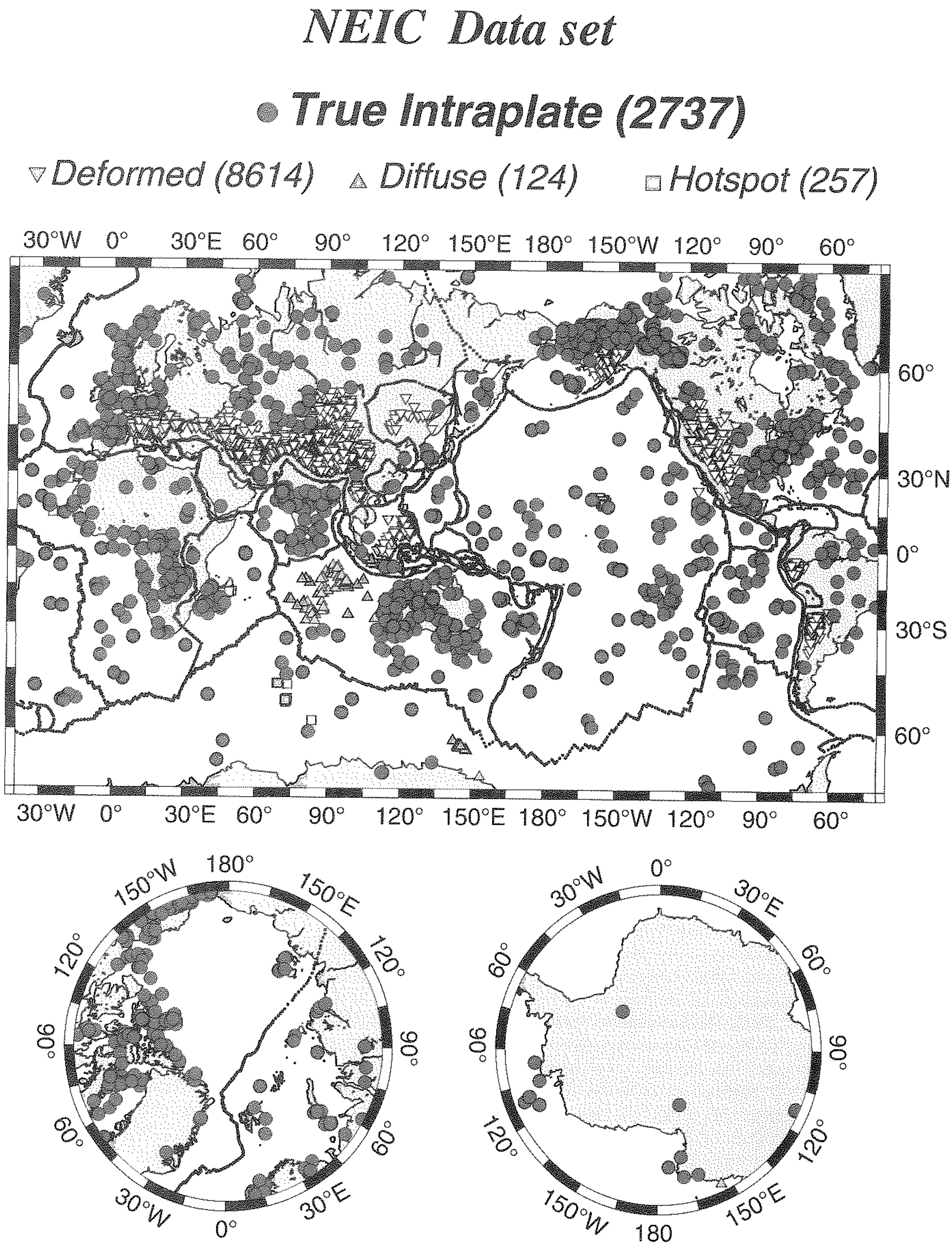

Figure 4. Maps of the National Earthquake Information Center (NEIC) data set. Symbols as in Figure 1.

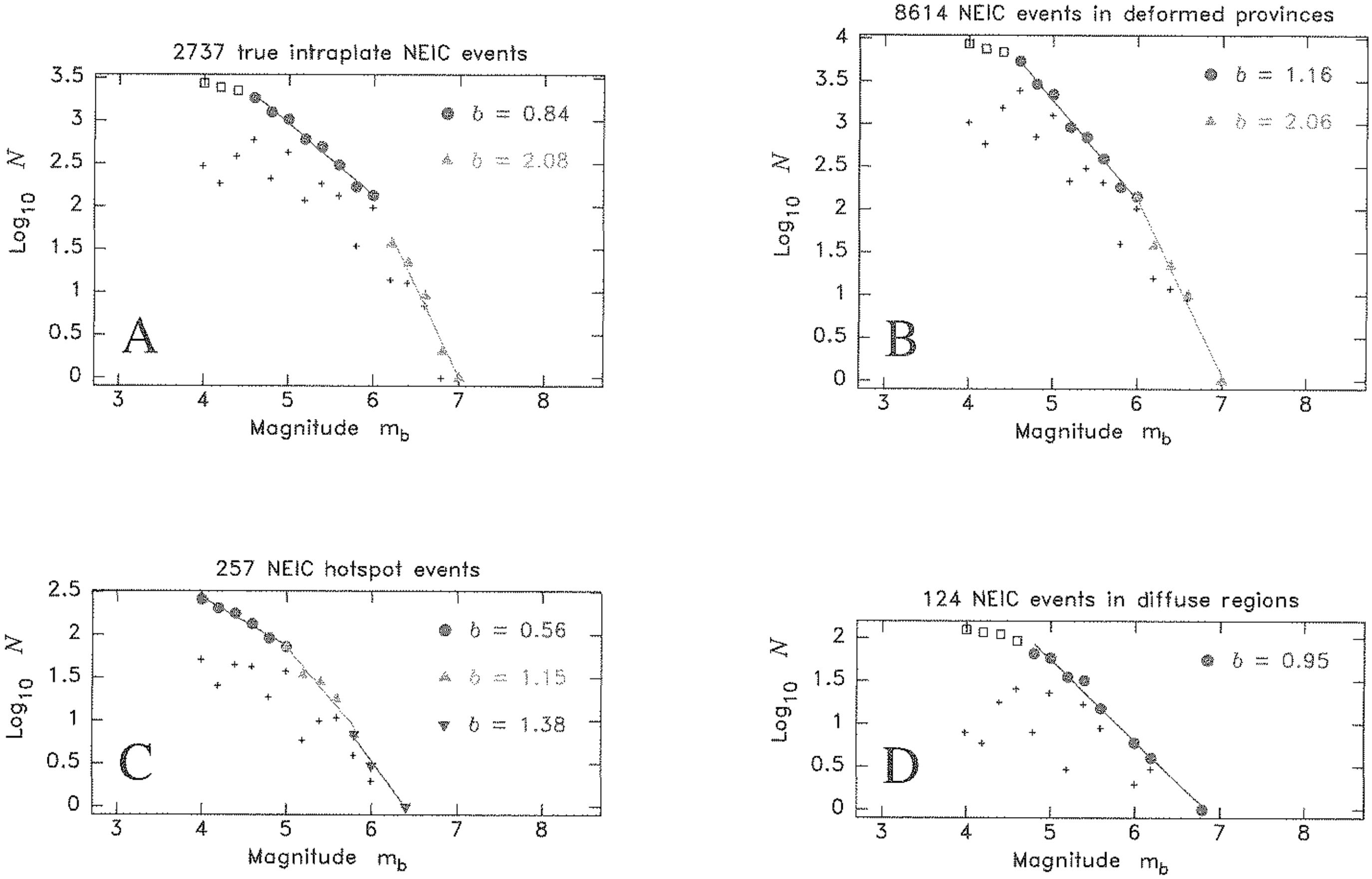

Figure 5. Determination of b-values for the National Earthquake Information Center (NEIC) data sets, using the body-wave magnitude m_b and the same procedure as in Figure 3.

The Pacific Data Set

Figure 7A shows the NEIC data set for the Pacific plate, which includes 435 sources. As mentioned by Wysession et al. (1991), a significant portion of the intraplate activity of the basin is expressed through swarms, i.e., episodes of seismicity concentrated in space and time but lacking a clear main shock. The most intense among them was the Gilbert Island swarm of 1981–1984, which included 225 teleseismically recorded events (Lay and Okal, 1983; Okal et al., 1986). The International Seismological Center (ISC) catalogue has reported two additional events in the area in 1997 and 2001, and there have been reports during the 1990s, of tremors felt by residents of the nearby islands. The b-value of the Gilbert swarm has previously been reported as 1.35 (Lay and Okal, 1983) and 1.40 (Wysession et al., 1991). Our study (Fig. 7B) confirms these relatively high values and further suggests that a gradual increase in b-value (from 1.02 to more than 2) is resolvable around $m_b = 5.4$.

Wiens and Okal (1987) also analyzed a swarm of earthquakes that occurred in the Pacific plate, ~700 km west of the Rivera fracture zone, for which they reported b = 1.05. We reprocessed this data set after including 11 additional earthquakes (for a total of 80), listed by the ISC, but not by the NEIC, and found an equivalent b = 1.08 (Fig. 7C). In this respect, the East Pacific swarm does not feature the increase in b with size found in the Gilbert swarm. We note that this difference in behavior may also be reflected in the fact that the East Pacific swarm clearly features a main shock (on 02 December 1984), but it remains rather unusual because most of the activity occurred prior to the main shock, with comparatively fewer aftershocks.

After removal of the Gilbert and East Pacific swarms, the Pacific plate NEIC data set is composed of 141 earthquakes, for which b-values are presented on Figure 7D. The body-wave coefficient, b = 1.29, is larger than expected theoretically (b = 1) in the corresponding range of magnitudes, but, on the other hand, it is comparable to the observed worldwide average (b = 1.35; Okal and Romanowicz, 1994; their Table 4). Any possible increase in b-value with m_b is impossible to resolve because of the narrow band of magnitudes available above the relatively high threshold of completeness in this region ($m_b \geq 5.0$). Similarly, the M_s-derived value (b = 0.74) is equivalent to the observed

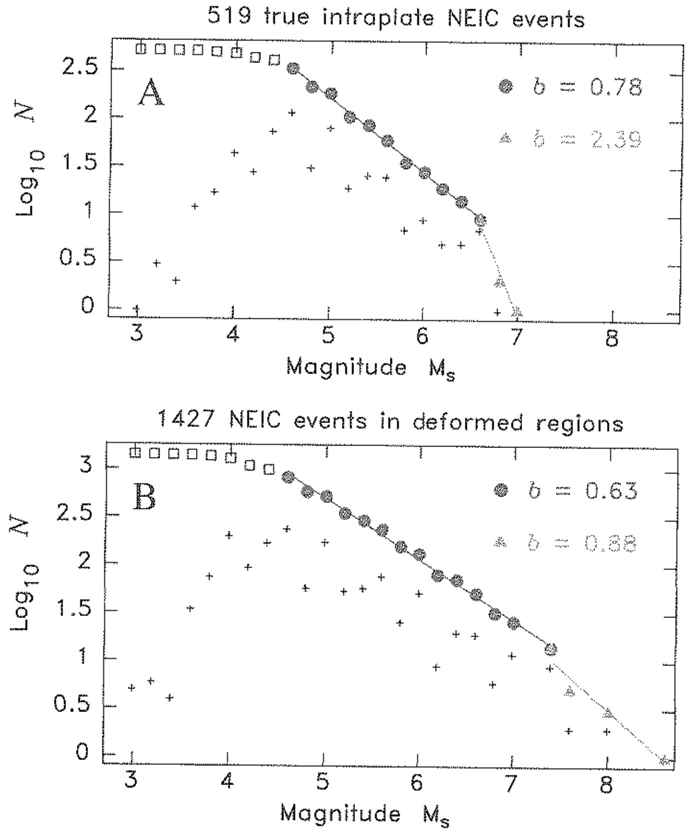

Figure 6. Same as Figure 5, but using the surface-wave magnitude M_s. The "deformed" and "diffuse" data sets are too small for regression.

worldwide average (b = 0.75; Okal and Romanowicz, 1994; their Table 3), and it is comparable to the predicted value of 2/3 (Fig. 7E). We note however that the population with reported M_s (43 events) becomes undersampled.

This Pacific data set does include a significant number of earthquakes clustered at locations initially described as regions B and C by Okal et al. (1980), for which these authors proposed very low b-values based on local magnitudes. When further extracted, this data set of only 40 east-central Pacific earthquakes features b = 1.02, and the remaining nonclustered data set has b = 1.31, both values using m_b. This would generally support the observation by Bergman and Solomon (1980) of a slightly higher b-value inside the Pacific plate but probably not their interpretation that such b-values express a more important contribution of volcanic earthquakes.

The African Plate

We originally extracted from the NEIC database a data set of 285 earthquakes located inside the African plate. Among them, 63 were located in the vicinity of Lake Kariba, on the Zambezi River at the border between Zambia and Zimbabwe (Fig. 8). Gough and Gough (1970a, 1970b) modeled the occurrence and evolution of this seismicity as induced by the loading stresses incurred during and after the filling of the reservoir in 1959–1963. They calculated b-values of 1.03 for the full data set spanning 1959–1968, and b = 1.14 for the filling stage, culminating with the largest shocks in 1963. They argued that, since these values were computed using local magnitudes M_L, the corresponding m_b-based value should be 1.4 times greater, based on the empirical relationship between m_b and M_L proposed by Gutenberg and Richter (1956). In this context, it seemed warranted to conduct an independent frequency-size study at Kariba based on the NEIC catalogue updated to 2003. Interestingly, we found that activity detected teleseismically continues at Kariba, with a total of 63 events with a reported m_b since 1963, at a rate (~1.5 earthquake per year) essentially equivalent to that during and immediately after filling of the reservoir. Figure 8A shows that this group features an m_b-based b-value of 0.96, which is not significantly different from the M_L-based values reported by Gough and Gough (1970b). In particular, these authors proposed, on their Figure 6, two slightly different regressions for the whole swarm (their open symbols) and the generally smaller events during the filling phase (1959–1963). These expressions lead to b = 1.02 for the complementary data set of earthquakes subsequent to filling (1964–1968). The agreement between this value and ours (m_b-based and pertaining to 1963–2003) would suggest first that the seismogenic process subsequent to filling is still going on 35 yr later, and second, that the use of Gutenberg and Richter's (1956) relation to convert M_L into m_b may not be warranted in the context of the Kariba earthquakes. We note in particular that Gutenberg and Richter's work pertained to aftershocks of the 1952 Kern County earthquake, in the magnitude range 5–7, much larger than the Kariba events, and that other relationships were later proposed for M_L versus m_b (Båth, 1981) for various magnitude ranges in various parts of the world, with slopes varying from as low as 0.8 (Chhabra et al., 1975) to as high as 2.04 (Båth, 1978).

After removal of the Kariba seismicity, the remaining 222 NEIC intraplate earthquakes feature a regular b = 0.92 between m_b = 4.6 and 6, with a lone event (m_b = 6.4 in Guinea) suggestive of an increase in b (Fig. 8B). These results are in general agreement with the theoretical models by Okal and Romanowicz (1994), which predict an increase from b = 2/3 to b = 1 in the range of source sizes covered by the African data set. The data set of events with a reported M_s was too small (39 events) for meaningful processing.

Eastern North America

As presented on Figure 9, we extracted 432 events belonging to the passive (eastern) part of the North American continent. This data set features b = 1.23 beyond its limit of completeness (m_b = 4.6), with the possible hint of an increase in b for $m_b > 6$. After removal of 57 earthquakes clustered in the New Madrid seismic zone, the b-value is only marginally altered, to b = 1.19. The New Madrid events feature a poorly constrained $b \approx 1.25$.

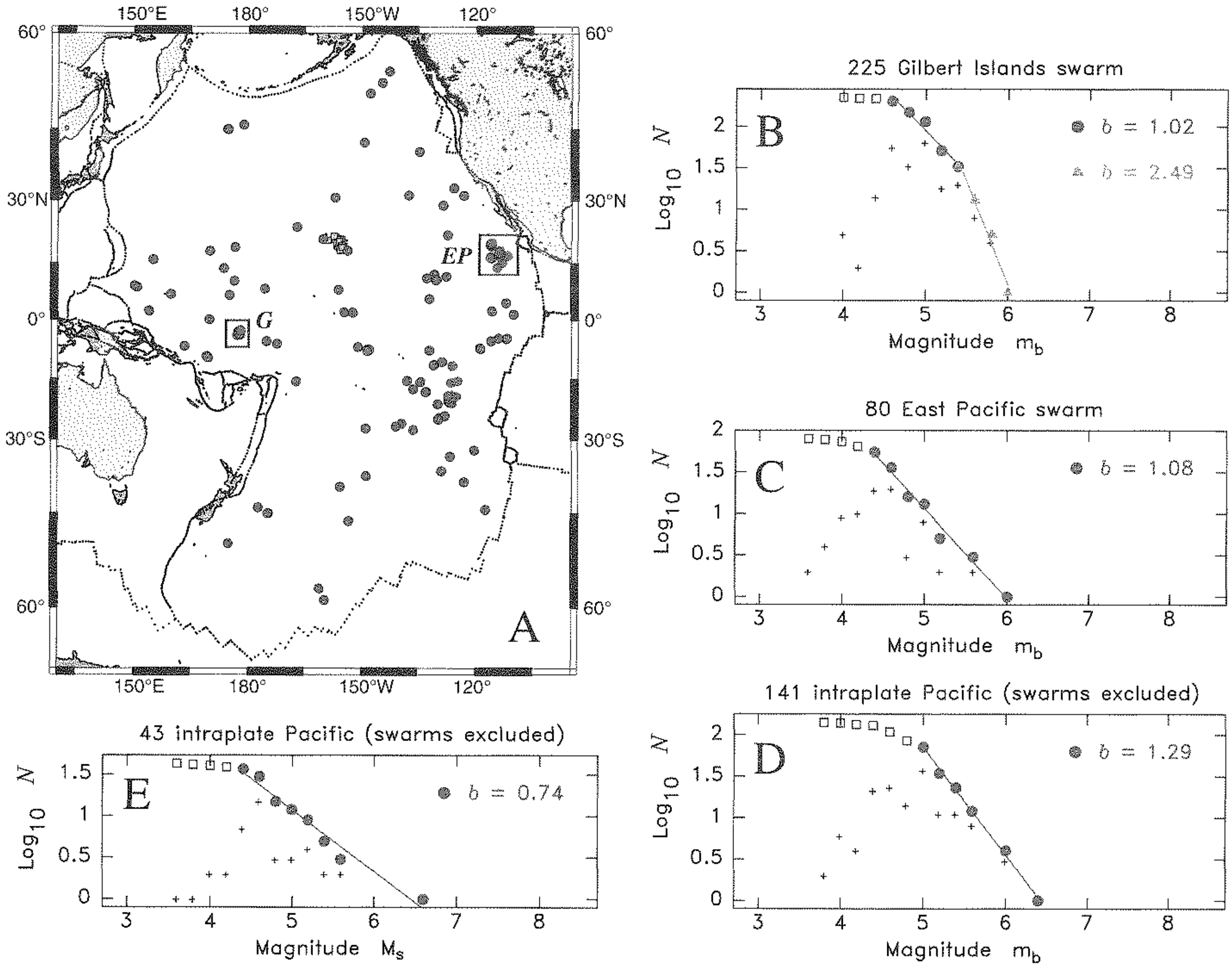

Figure 7. Detailed study of the Pacific plate. (A) The map is a close-up of Figure 4, with the locations of the Gilbert Islands (G) and East Pacific (EP) swarms outlined. The frames at right show the populations of body-wave magnitudes m_b for the two swarms (B, C) and for the remaining 141 true intraplate events (D). (E) The small data set of surface-wave magnitudes M_s.

DISCUSSION AND CONCLUSION

Our examination of global data sets of intraplate earthquakes, extracted both from the CMT and NEIC catalogues, fails to reveal a compelling difference in frequency-size distribution of their populations as compared to their much more numerous interplate counterparts. This suggests a fundamental similarity of the relevant scaling laws governing the populations of intraplate and interplate events. Most importantly, the β-value of 0.66 for the intraplate population (derived from seismic moments, and thus not subject to the saturation affecting the standard magnitude scales) is identical both to its worldwide counterpart (0.67), and to the theoretical value (2/3) predicted at low moments by Romanowicz and Rundle (1993). In the framework of Rundle's (1989) theory, it argues for a fractal dimension $D = 2$ for the source of intraplate earthquakes, which suggests that they take place on two-dimensional systems of *faults*, rather than inside a seismogenic volume, as do certain deep earthquakes (Okal and Kirby, 1995).

Expectedly, the interpretation of results derived from standard magnitude scales becomes less clear-cut as the characteristic period of the magnitude decreases and the effect of saturation becomes more prevalent for any given size of event. For true intraplate earthquakes, the b-value (0.78) computed at low magnitudes from the 20 s surface-wave magnitude M_s is somewhat higher than predicted theoretically (2/3), but similar to that observed by Okal and Romanowicz (1994) on a global data set (0.75). When derived from the 1 s body-wave magnitude m_b, b-values (0.83 for the CMT data set; 0.84 for the NEIC data set) are lower than reported by Okal and Romanowicz (1994)

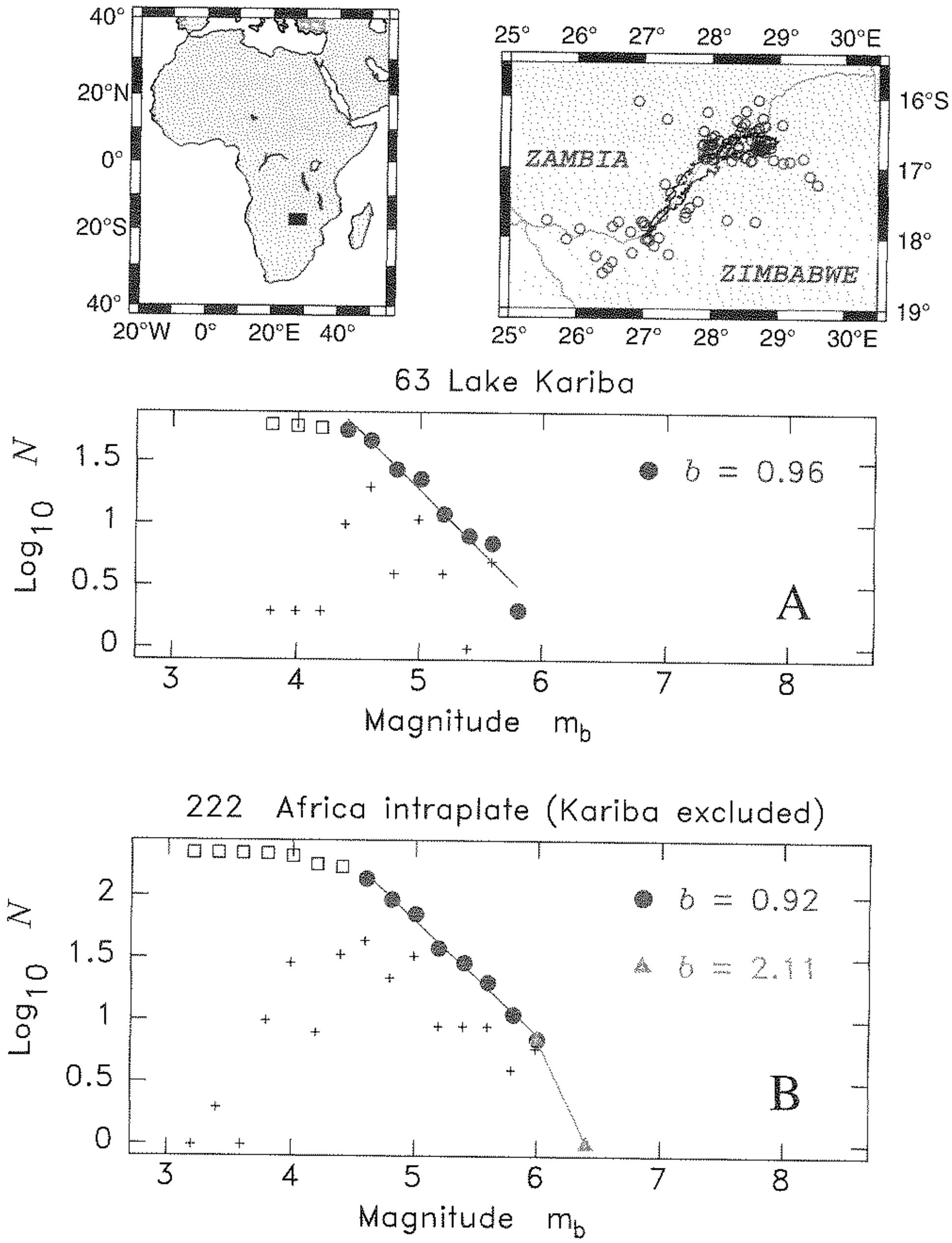

Figure 8. Populations of body-wave magnitudes m_b in the African plate. (A) Sub–data set of earthquakes associated with the filling of Lake Kariba. (B) Remaining true intraplate earthquakes. The maps at top show the distribution of seismicity around Lake Kariba (right) and (left) the location of the study area in Africa (dark rectangle).

for CMT events (1.17) or the full NEIC population (1.35) but still fall in the range (2/3–1) predicted theoretically. We confirm the trend toward larger b-values for intraplate events in the Pacific, but only for m_b (1.29), while the M_s-derived b-value is the same as in other plates.

On the other hand, events in the so-called "deformed" provinces, consisting mainly of the Mediterranean-Tethyan belt, exhibit slightly reduced values of β and b in the low-magnitude range, but perhaps more interestingly, greater values of the critical earthquake size controlling the elbows in the frequency-size distributions: in the case of the well-sampled M_s distribution, Figure 6B could suggest $M_s^c \approx 7.4$, significantly greater than predicted theoretically (6.7) or observed in true intraplate areas (6.6); in the case of the CMT data set, the absence of any detectable elbow could be explained by a critical moment larger than the global value of $10^{27.5}$ dyn-cm, which would thus fall into a

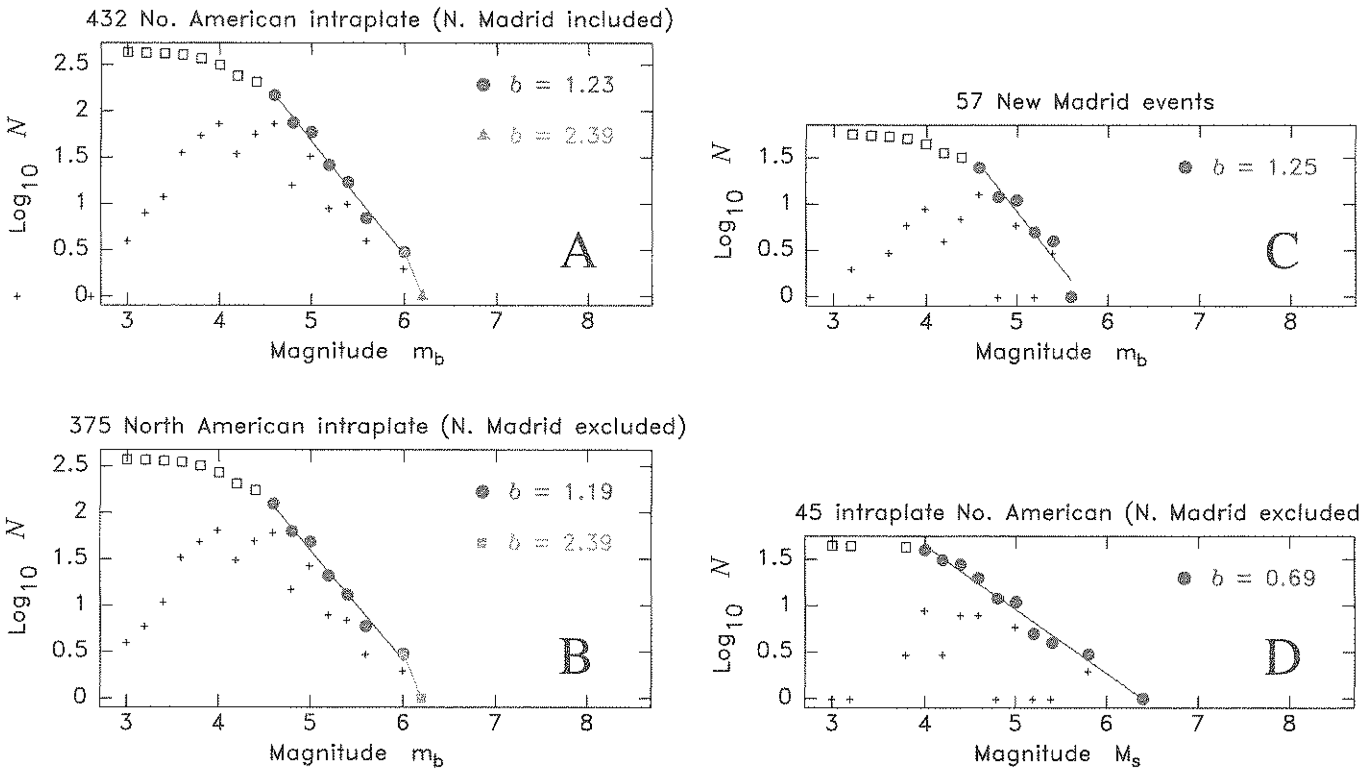

Figure 9. b-value determinations for North American intraplate earthquakes. Body-wave b-values are shown for the full population (A) and after removal of New Madrid earthquakes (B). The latter are examined in detail (C). (D) The small data set of surface-wave magnitudes M_s.

poorly sampled part of the moment population. Since the size of the critical earthquake reflects the saturation of fault width when it reaches the maximum depth of the seismogenic brittle zone, a larger critical earthquake would argue for a thicker seismogenic zone, which in turn has been both documented and explained in the context of the ongoing deformation and orogeny in the deformed provinces.

Finally, it is interesting to compare our results for the North American data set, which is overwhelmingly continental (Fig. 9A), with those for the Pacific plate, which is exclusively oceanic (Fig. 7D). The similarity between b-values (1.23 versus 1.29) suggests a commonality of scaling processes in both provinces. The smaller nature of the Pacific data set precludes the determination of a corner magnitude and, hence, its comparison with the value obtained in North America, which at any rate is only tentative.

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