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Combined satellite and surface-based estimation of the intracloud / cloud-to-ground lightning ratio over the continental United States

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ABSTRACT

Four years of observations from the NASA Optical Transient Detector (OTD) and Global Atmospherics National Lightning Detection Network (NLDN) are combined to determine the geographic distribution of the climatological intracloud (IC) / cloud to ground (CG) lightning ratio, termed *Z*, over the continental United States. *Z* over this region is 2.64-2.94, with a standard deviation of 1.1-1.3 and anomalies as low as 1.0 or less over the Rocky and Appalachian mountains and as high as 8-9 in the central-upper Great Plains. There is some indication that *Z* covaries with ground elevation, although the relationship is nonunique. Little evidence is found to support a latitudinal covariance. The dynamic range of local variability is comparable to the range of values cited by previous studies for latitudinal variation from the deep tropics to midlatitudes. Local high *Z* anomalies in the Great Plains are coincident with anomalies in the climatological percentage of positive CG occurrence, as well as in the occurrence of large positive CGs characteristic of organized or severe storms. This suggests that storm type, morphology and level of organization may dominate over environmental cofactors in the local determination of this ratio.

1 Introduction

In recent years, significant advances have been made in determining regional or global estimates of variability in cloud-to-ground (CG) and total (intracloud and cloud-to-ground, IC+CG) lightning flash rates (Orville and Henderson 1986; Goodman and Christian 1993; Cummins et al. 1998; Huffines and Orville 1999; Christian et al. 1999; Goodman et al. 2000). However, the relative proportions of intracloud and cloud-to-ground lightning remain one of the greatest unknowns in regional and global lightning estimates. This partitioning (commonly expressed as the IC:CG ratio, and denoted by Z) is important from several standpoints. First, much of our knowledge of lightning energetics (currents, total charge transfer) derives from observations of cloud-to-ground flashes, which are comparatively easier to detect over larger distances than intracloud flashes. Extrapolation from such measurements when estimating regional or global $\mathrm{NO_x}$ production, or the contribution of lightning to the global electric circuit, is contingent upon accurate knowledge of Z and its regional variability. Second, attempts to relate observed total lightning flash rates to storm kinematic and microphysical processes also require a knowledge of this partitioning if they are to proceed beyond empirical correlations. Specifically, intracloud and cloud-to-ground lightning represent two different mechanisms by which lightning helps balance storm generator currents, which are driven by local microphysical and large scale gravitational and kinematic charge separation. Recent empirical studies (to be discussed below) suggest that the intracloud component is more closely coupled to storm evolution aloft, and assessment of a given sensor's (network's) ability to monitor lightning-convection relationships may thus indirectly depend on Z (through its relative IC and CG detection efficiencies). Third, accurate cross-sensor validation of the performance characteristics of next-generation orbital 'total' lightning sensors such as NASA's Optical Transient Detector (OTD) and Lightning Imaging Sensor (LIS) (Christian et al. 1996; Christian et al. 1999) requires knowledge of Z. Finally, more accurate knowledge of climatological Z values may be relevant to the

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use of IC discharge detection as an early warning indicator of ground strike occurrence (the utility of which has been empirically demonstrated by Weber et al. 1998; Murphy and Cummins 2000), at least in cases where the incremental advanced warning utility derives from IC flashes in developed, advecting storms.

As noted above, prior large scale or climatological estimates of Z have been few and far between, mostly a result of the difficulty in *robustly* measuring the intracloud lightning component using traditional ground-based RF detection techniques. The literature is dominated by the results of two prior studies, those of Pierce 1970; Prentice and Mackerras 1977 (revisited by Mackerras et al. (1998)). These efforts are notable in that they examined Z variability (using effectively 'point' measurements) over many different (land) remote regions, including both midlatitude and tropics. Furthermore, they employed long periods of continuous monitoring. These authors found some indication that Z varied with latitude, with high values observed in the deep tropics and comparatively smaller values at midlatitudes. The actual point estimates, along with the fitted curves of Pierce 1970; Prentice and Mackerras 1977, are summarized in Fig. 1. It is clear that among these estimates, the variance is high and the fitted latitudinal dependence is at best tentative. Over the years, heuristic arguments have been offered to explain the apparent latitudinal variability (Rutledge et al. 1992), including the idea that deeper cloud tops (higher mean tropopauses) in the tropics favor intracloud flashes as a dissipation mechanism for generator currents. Clearly, statistical tests of this hypothesis are constrained by data availability.

Additional knowledge of the magnitude and variability of Z derives from field-campaign based case studies of individual storms or small storm ensembles. Rutledge et al. (1992) found Z in moderate flash rate tropical continental storms (Darwin, Australia) of comparable magnitude to midlatitude estimates (2-5) but noted a covariance between Z and storm total lightning flash rate f, scaling as approximatly $f^{0.5}$. They further noted that at high flash rates, Z can reach very high values. This was consistent with earlier observations of high Z in an Alabama storm with severe surface weather (with much of the time-evolving signal in Z driven by variability in the intracloud flash rate, and correlated to storm kinematic (Kingsmill and Wakimoto 1991) and microphysical (Goodman et al. 1988) properties). This was also consistent with findings by MacGorman et al. (1989), who observed high Z values during the tornadic phase of a severe midwest thunderstorm, similarly driven by increases in the IC and decreases in the CG flash rates. MacGorman et al. (1989) further summarized sferics-based observations of severe storms by prior investigators and drew similar inferences; i.e. that the intracloud flash rate grows disproportionately from (and sometimes inversely with) the CG flash rate as storms become more severe and perhaps have stronger updrafts. Recent merged satellite-NLDN observations of tornadic storms in Oklahoma by the OTD (April, 1995; (Buechler et al. 2000)) and LIS (May, 1999; K. Driscoll (pers. comm. 1999)) help confirm the occurrence of extraordinarily high Z in such storms (values of 16-25 for the 1995 storm [without and with detection efficiency corrections], and 29-35 for the 1999 storm, comprised of both unusually high intracloud flash rates and unusually low ground flash rates. Similarly, in a study of severe storms in central Florida with the Kennedy Space Center LDAR network, high Z ratios (driven by high intracloud flash rates) were found to precede surface mesocyclone occurrence by at least 10 minutes (Williams et al. 1999). Significant temporal variability of Z during the evolution of a severe (hail-producing) front-range storm during STERAO-A was also documented by Lang et al. (2000).

Clearly the high variance in observed Z both regionally and within the time-evolution of individual storms should caution against large scale extrapolation from point or case-study measurements. We further note that our knowledge of Z over most of the world's open oceans is essentially unknown. While observational limitations still preclude an accurate assessment of this ratio over global scales, the necessary remote sensing systems are now in place to assess its values over a large region, the continental United States. The U.S. National Lightning Detection Network (NLDN) continuously monitors ground strike occurrence with very high detection efficiency (Cummins et al. 1998) and little spatial bias. The OTD satellite-based sensor detects total lightning (IC and CG) asynoptically from low-earth orbit over most of the globe, and has been in operation for over four years (long enough to begin building climatologies with moderately high spatial resolution). While some ambiguity remains as to the absolute performance characteristics (detection efficiencies) of these two systems, the data are of high enough quality to generate spatial maps of Z, and the performance unknowns are constrained enough to allow basic sensitivity studies and assign error ranges to these maps. This spatial analysis is the main focus of this study. In section 2, we describe the basic data quality control and processing methodology. Section 3 presents the geographically varying results, and compares these against other relevant parameters which might help explain the observed geographic patterns. These patterns are not explainable by uncertainty in the performance characteristics of either the OTD or NLDN (demonstrated through a sensitivity analysis in

Appendix A). Section 4 summarizes the key results.

2 Methodology

Four years of satellite- and ground-based lightning observations are utilized in this study. The analysis period ran from 1 May 1995 to 30 April 1999. The analysis domain is the continental United States, land-only regions. During this time, the network configuration (number and deployment of receivers, waveform acceptance criteria) of the NLDN remained essentially constant (exceptions included sensors added in the far south of Texas and in southern Louisiana, changes which should not affect estimates over most of the domain). CG flash occurrences were composited into a 0.5×0.5 degree grid; reported flashes with peak current amplitudes between 0 and +10 kA were discarded from the dataset due to a strong suspicion that they may be contaminated by intracloud lightning (Cummins et al. 1998; Wacker and Orville 1999a; Wacker and Orville 1999b) (this contamination seems to represent at most 5% of the total dataset). With the exception of remote coastal regions, the network CG detection efficiency is expected to be geographically uniform (Cummins et al. 1998) and is initially assumed to have a constant value of 90% (i.e., gridded flash rates are scaled by (1.0/0.9)). The implications of this assumption are examined in a sensitivity analysis in Appendix A.

OTD flash data are similarly composited into a 0.5×0.5 degree grid. The OTD sensor views surface locations at midlatitudes approximately three times every two days, for a duration of anywhere from 1-240 seconds, depending on the orientation of the sensor field-of-view. Data have been aggregated in 55-day blocks to minimize bias from precessing sampling of the local diurnal lightning cycle (Boccippio et al. 2000). Sensor spatial accuracy is typically comparable to this grid resolution (Boccippio et al. 2000) although it can occasionally degrade to about 2-3 degree accuracy due to navigational drift. OTD version 1.1 ”flash” data products are used in this study; under this algorithm, collections of ”groups” (time-concurrent and adjacent CCD pixel illuminations) are assembled into a nominal flash if they are separated by no more than 333 ms and 25 km. The OTD CG flash detection efficiency has been estimated to be between 49-65% (Boccippio et al. 2000), and is a function of the sensor threshold (gain) setting. For the purposes of this study, we assume that the OTD IC and CG detection efficiencies are equal and that they are geographically invariant. We examine the implications of these assumptions in Appendix A. The prescribed detection efficiencies are [64%, 49%, 56% and 62%] for the periods [05/01/95-6/12/95, 6/13/95-7/19/95, 7/20/95-10/22/96 and 10/23/96-04/30/99], respectively. Each OTD orbit undergoes manual Quality Assurance (QA) inspection, and any OTD orbits which raised manual QA alerts were discarded from the final dataset. Each OTD flash is assigned an automatic quality metric termed the ’Thunderstorm Area Count’ (TAC), which roughly indicates the confidence that the reported flash is indeed lightning and not ambient noise, based upon the flash information content (number and clustering of optical pixels) and occurrence of other flashes in its geographic location at other times during the satellite overpass. Flashes with TAC less than 140 were discarded from the gridded composites; this value has been found by manual inspection to provide reasonable discrimination between true lightning and radiation noise (K. Driscoll, pers. comm. 1997), and is the same threshold applied in the detection efficiency estimates of Boccippio et al. (2000). In addition to the DE-scaled flash totals, the actual duration of viewing was recorded for each grid location, allowing conversion of flash counts to bulk flash rate density (i.e., $\frac{fl}{km^2 sec}$).

The intracloud to cloud-to-ground ratio Z is thus calculated from:

$$Z = \frac{\sum_{\tau=1...4}\left(\frac{f_{OTD}(\tau)}{DE_{OTD}(\tau)}\right) - \frac{f_{NLDN}}{DE_{NLDN}}}{\frac{f_{NLDN}}{DE_{NLDN}}} \tag{1}$$

where τ denotes the OTD subperiods of different threshold values as listed above, $f_{OTD}(\tau)$ and f_{NLDN} are observed (climatological) flash rate densities, DE_{NLDN} = 0.9, and $DE_{OTD}(\tau)$ is as given above.

It should be noted that the severe subsampling associated with the low-earth orbiting OTD is the limiting factor in this computation (0.15% of the (x, y, t) sampling afforded by the continuously-observing NLDN). With four years of observation, large scale climatological total flash rate estimates are possible, although estimates at the full grid resolution (0.5 deg) are not. As such, we smooth both OTD and NLDN grids using a moving filter; i.e., a new 0.5×0.5 deg grid is constructed in which each grid location contains the average of n adjoining grid cells in all directions. All grid cells are equally weighted in this and all other smoothing/averaging results presented herein. We inspected smoothed grids with $n = 1, 2, ...9$, and found that 3.5 degree spatial smoothing $(n = 3)$ was the minimum smoothing which yielded regionally continuous Z fields. All results presented below are computed with this smoothing factor, applied to both OTD and NLDN data.

3 Results and Discussion

In this section we present the basic geographic distribution of Z across the continental U.S., and analyze it in the context of prior hypothesized relationships to latitude, elevation, storm total flash rate and storm morphology. The most significant features of this distribution are insensitive to uncertainty in the performance characteristics of either the OTD or NLDN, hence the results may be slightly biased but still contain robust geographic variability (this is demonstrated through a detailed sensitivity analysis in Appendix A).

a. Geographic Distributions

The four-year mean $\overline{Z}$ for the continental U.S. (average of all 0.5×0.5 deg grid values *after* spatial smoothing) is found to be 2.94, with a standard deviation of 1.28. This is clearly within the bounds of prior midlatitude climatological estimates (Pierce 1970; Prentice and Mackerras 1977; Mackerras et al. 1998) (Fig. 1), lending some confidence to the joint data technique and assumed sensor detection efficiencies. These local mean values are, however, shown to have significant regional variability (Fig. 2). The most pronounced features are:

1. A significant high anomaly extending from southwest to northeast along the Rocky Mountain front range / upper Great Plains region. Local mean Z values reach as high as 6-9 in this region, with two distinct local extrema evident.

2. Low anomalies over the Rocky Mountains and Appalachian Mountains, with values as low as 1.0 in the west and even lower than 1.0 in the east. The geographic alignment of these anomalies roughly follows, but is offset west of, the local topography. A secondary low anomaly is evident near the Sierra Nevada range.

3. A broad region of high anomalies in the Pacific Northwest.

4. Local high anomalies along the California coast and northeast New England.

We shall primarily focus on anomalies (1) and (2) in the discussion below. Anomaly (4) has a significant likelihood of being attributable to lack of geographic uniformity in NLDN network detection efficiency, and we will not attempt to explain it here. Anomaly (3) is harder to attribute to network limitations. It is possible that this arises from very low flash rate storms which mostly or only produce IC flashes during their lifecycles, or from a local winter bias in the population of lightning producing storms (which for unspecified reasons might exhibit higher Z). We do note that the climatological mean flash rate (both as observed by OTD and NLDN) in the regions of anomalies (3) and (4) is extremely low (less than 1.0 and 0.2 $\frac{fl}{km^2 yr}$, respectively), and Z values computed for these regions are thus high variance estimates. Given this, we are not yet confident that anomalies (3) and (4) are real).

b. Latitudinal dependence

Clearly there is no monotonic dependence on latitude evident in Fig. 2, and the very notion of taking zonal (latitude) means seems suspect when faced with such regional variability. Fig. 3 shows the zonally averaged Z values plotted against latitude. While there does appear to be a weak latitudinal dependence, it is also evident that this structure seems attributable to the zonally averaged elevation, and hence may be merely an artifact of the covariance of this zonally averaged profile. Fig. 4 shows the (smoothed) 0.5×0.5 deg grid point Z estimates overlaid on the prior results of Pierce 1970; Prentice and Mackerras 1977; Mackerras et al. 1998. We note that the variability in local estimates is *comparable* to the total equator-to-midlatitude drop in Z suggested by these prior estimates (this, of course, was also a feature of the earlier point data).

Prior hypotheses to explain apparent equator-to-midlatitude drops in local Z measurements have either invoked differences in the height of the local freezing level, arguing that higher main negative charge centers disfavor cloud-to-ground flashes, or differences in the height of the local tropopause, arguing that deeper mixed phase and upper positive charge regions in the tropics favors intracloud over cloud-to-ground flashes. Price and Rind (1993) have suggested that the differences in freezing level altitude in summer midlatitude and annual tropical tropospheric profiles are generally small, and unlikely to cause broad variability in Z. Climatological maps of the tropopause pressure over the continental U.S. (Hoinka 1998) are instructive in assessing the second explanation. Considering the northern hemisphere summer months (season of maximum lightning occurrence), there is a local low tropopause pressure anomaly over the northern Great Plains, presumably a feature of climatological stationary wave structure over the continent. However, this anomaly (10 mb or so) is small compared to the overall increase (about 80 mb) in tropopause pressures from the southernmost to northernmost latitudes considered here (about 25N and 50N). Further, the mean tropopause pressure at these southermost latitudes is fairly close to that typical of the deep tropics (within 30 mb). While this still corresponds to a significant altitude difference, it is clear that a wide enough dynamic range of tropopause pressures is explored in the present analysis to determine whether a dependency (manifest as a function of latitude) exists as a ”zeroth-order” controlling variate in establishing Z (it does not).

We of course do not explore tropical regions in the current analysis. Thus, while the present results do not refute earlier suggestions of a broad tropics-to-midlatitude latitudinal dependence in Z (which was largely driven by tropical observations), they offer no evidence to support one, despite covering a significant dynamic range of one indirectly hypothesized physical cofactor (tropopause pressure). Alternatively, the majority of the midlatitude grid locations shown in Fig. 4 show reasonable agreement with earlier midlatitude point measurements from both hemispheres, which taken by themselves offered little indication of latitudinal dependence.

c. Elevation dependence

Anomaly (2) (low Z near significant mountain ranges) is a striking feature of the climatological results. Elevation dependencies in Z are physically plausible possibilities; the proximity of ground to the main negative charge region in electrified storms could certainly be suspected to help control the direction of travel of initiated discharges. Fig. 5 demonstrates an (apparent) relationship between Z and mean ground elevation. However, this relationship should be treated with caution. First, it is clear from Fig. 2 that a physical dependency on elevation, if it exists, is clearly nonunique. Fig. 6 shows the mean ground elevation for the continental U.S., derived from a high resolution (30 second) topographic database and smoothed using the same 3.5 deg operator as the Z field. We note that the lowest anomalies in Z occur over the Appalachian Mountains, which of course have significantly lower altitude than the Rocky Mountain range. Closer inspection also reveals that the local minimum in Z, which extends from western Wyoming through eastern Utah and Arizona, is offset distinctly to the west of the elevation maxima of the Rocky Mountain range. Together these suggest that the influence of elevation may be broader than simple geometric effects via the relative electrical position of charge regions and ground. Specifically, the effects of topography on the convective initiation, evolution, organization and life cycle of storms are likely cofactors of comparable importance.

Fig. 5a further shows the mean IC and CG flash densities in each elevation bin. While these curves should again be interpreted with caution (each bin may encompass many different local convective "regimes"), it appears that the Z-elevation dependency is driven by variability in the IC rate, at least in the large scale composites. Cummins (pers. comm, 1999) has recently demonstrated clear positive correlations between CG flash density and elevation at very small scales (near Tucson, AZ); however, we cannot determine the local-scale variability of Z with the present (heavily undersampled) low earth orbit satellite data.

Fig. 5b examines whether an elevation dependency was possibly found in prior point estimates of Z, specifically those of Prentice and Mackerras (1977). Here, their data are aggregated into 250 m elevation bins and averaged. Three aggregation techniques are attempted: (1) use of the nominal elevations reported by Prentice and Mackerras (1977) for some of their locations, (2) use of 0.5 deg mean elevations from a high resolution topopgraphic grid, using nominal (poorly resolved) latitudes and longitudes reported by Prentice and Mackerras (1977), and (3) use of elevations from a 3.5 degree smoothed elevation grid and the same nominal latitudes and longitudes. Standard deviation bars are overlaid; in bins with only one measurement, these bars instead span the entire plot. There is perhaps some indication that Prentice and Mackerras (1977)'s data contained an elevation signal similar to that found over the continental United States, although given the small sample size and in absence of a formally articulated elevation dependency mechanism, this suggestion should be viewed with caution.

We thus conclude by noting that while some univariate elevation dependency appears to exist, it is at best *nonunique*, and its structure (minima for very low and very high elevations) does not lend itself to simple geometrically-based explanations.

d. Total flash rate dependence

As discussed in section 1, Rutledge et al. (1992) found a clear dependency in Darwin, Australia between Z and the storm total flash rate f, with Z scaling as roughly $f^{0.5}$. This is qualitatively consistent with prior observations of significantly elevated intracloud flash rates in severe storms without commensurate increases (and sometimes decreases) in the CG flash rate (Goodman et al. 1988; MacGorman et al. 1989). The geographic distributions of Fig. 2 *might* call into question the large-scale relevance of this observation, given that the maxima occur in different locations than those found in conventional thunderday maps or annualized CG regional flash rate f_r climatologies (e.g., Huffines and Orville (1999)); these instead find extrema over Florida and the Gulf coast. Fig. 7a illustrates this via the OTD-derived regional total flash rate f_r. However, it has recently been noted (from OTD and LIS observations) that the annualized regional flash rate is largely controlled by differences in flashing cell frequency of occurrence rather than differences in per-cell flash rate f_c (Williams et al. 2000; Boccippio et al. 2000). We thus must examine f_c rather than f_r to determine whether Z varies with cell flash rate over this domain.

We test this by examining the distribution of per-cell flash rates observed by the OTD. Contiguous optical flashes observed during an overpass are clustered into data units termed "areas", which correspond roughly to individual electrified "cells". OTD version 1.1 areas are nominally comprised of flashes whose centroids are within 25 km of each other, although in the dataset, this algorithm appears to fail approximately 20% of the time, resulting in occasional area fragmentation (this occurs in scenes with high data rates, primarily scenes with many concurrent cells).[1] These algorithm errors and occasional navigational errors found in the OTD dataset place some limits on the usefulness of this data product, although it is noted in Boccippio et al. (2000) that the broad properties of lightning-derived cells (particularly the separation into number of cells and per-cell flash rate) behave similarly in the OTD and LIS datasets (data from the latter sensor do not suffer from algorithmic failure or navigational ambiguity, and utilize a slightly different area clustering algorithm). Williams et al. (2000) further corroborated the frequency of occurrence / per-cell flash rate distinction using a completely different storm identification technique. We finally note that the asynoptic OTD sampling means that instantaneous cell flash rate observations represent randomly sampled snapshots taken throughout the life cycle of the complete spatio-temporal spectrum of electrified storms.

Fig. 7b and 7c show the number of observed cells during the four years of study and the mean per-cell total flash rate. Clearly, here as in the tropical results reported by Boccippio et al. (2000), spatial variability f_r is dominated by the frequency of occurrence of flashing cells. (Williams et al. (2000) further observed that this effect dominated in determination of the local diurnal cycle of total flash rate). Weak local maxima are found in $\overline{f}_c$ in roughly the same locations as the maxima in Z, although additional extrema occur in $\overline{f}_c$ which are not matched in Z. Low values of both are observed over the mountain ranges. There is thus weak qualitative support in the climatological averages for the broad, continental U.S. applicability of some $Z(f_c)$ dependence as reported by Rutledge et al. (1992), although it appears such a dependency is nonunique or mitigated by other factors (direct examination of $\overline{f}_c$ vs. Z grid cell by grid cell scatterplots (not shown) reveals little statistical correlation between the two). As with altitude, there is the suggestion that other variates are more directly involved in determination of Z.

[1]The median radius of an OTD area over the continental United States is 16 km, although this value may be high-biased due to occasional navigational drift during individual overpasses.

e. *Storm morphology dependence*

The prior three subsections fail to show simple univariate relationships between Z and geometric or geographic (environmental) parameters, suggesting that the structure, convective vigor and degree of organization (i.e., the morphology of flashing storms) may be important in determination of Z, an idea certainly qualitatively consistent with prior findings of high intracloud flash rates occurring in association with severe storms. The regional variability in Z (especially the northern Great Plains maxima) is clearly reminiscent of some spatial patterns previously identified using climatological NLDN data, specifically the percentage of positive CG occurrence (Orville and Silver 1997; Orville and Huffines 1999), frequency of large (greater than 75 kA peak current) positive CGs (Lyons et al. 1998), and mean positive CG peak current (Orville and Huffines 1999). Furthermore, the distribution of local severe surface weather from 1989-1998 (hail and tornado occurrence) has been found to have similar spatial structure, with greatest similarity found when this distribution is limited to severe storms which exhibit predominantly positive CGs (PPCGs) (L. Carey, pers. comm. 1999, in a statistical confirmation of more limited earlier observations by Branick and Doswell (1992)). We might expect even closer corrlation of Z with PPCG hail-producing storms, given heightened correlation between positive CG occurrence and hail (as compared against other severe weather, i.e. damaging winds and tornadoes) (Reap and MacGorman 1989).

We present the positive CG distributions here in the hopes of stimulating further research and discussion. Fig. 8a shows the percentage of positive CG (%PCG) occurrence in the total NLDN dataset for the same period as this study (May 1995-April 1999), filtering small positive CGs as described in section 2. As found in Orville and Silver 1997; Orville and Huffines 1999, this field exhibits a strong maximum running southwest-northeast through the upper Great Plains, with an extremum coincident with the secondary maximum in Z and local maxima in $\overline{f}_c$. Analysis by L. Carey (pers. comm, 1999) suggests that this extremum is not driven by preferential occurrence of positive CGs in trailing stratiform regions of mesoscale convective systems, but rather by (high peak current) +CG occurrence in the convective regions of high flash rate storms. Fig. 8b shows the occurrence of large positive CGs (LPCG) during the study period, using criteria selected by Lyons et al. (1998). As with the distribution of %PCG, the LPCG rate peaks in the same location as the northernmost maximum in Z (Fig.

2).

Fig. 9 shows the 0.5×0.5 deg grid values (after spatial smoothing) of Z and %PCG plotted pairwise (central U.S. only), with the LPCG flash rate color coded in a relative 'cool-warm' scale. As suggested by the spatial maps, anomalously high mean Z values only occur in regions of high %PCG, although the converse is not true (a wide range of Z is found in the highest %PCG locations, although the impacts of necessary smoothing are not known). There appear to be much weaker and certainly nonunique statistical relationships between Z and the LPCG flash rate, although again, the effects of the 3.5 degree smoothing operator may be manifest here.

The apparent correlation between Z and positive CG occurrence is tantalizing but difficult to directly explain. Recent findings by Stolzenburg et al. 1998; Stolzenburg et al. 1998a; Stolzenburg et al. 1998b (using electric field soundings) may have some bearing; these authors identify not only an elevated main negative charge region in the updraft regions of strong MCS' but an apparent deepening and elevation of the lower positive charge center (Fig. 10) over traditional 'tripole' models (Simpson and Scrase 1937; Williams 1989). The upward displacement of vertical charge distributions appears correlated with updraft intensity, to the extent that in the most intense storms studied - supercells - the lower positive charge center elevation and enhancement could equivalently be viewed as an 'inverted dipole', although the lack of direct causal links to changes in underlying charging mechanisms should caution against this terminology. Regardless of terminology, it is clear that in the strongest updraft cases studied by Stolzenburg et al. (1998b), the lower positive charge region occupies the same position in altitude as the main negative charge region in 'garden variety' thunderstorms (e.g., mountain New Mexico storms as studied by these authors). [2] Together, the sounding results offer corroboration of the 'elevated dipole' hypothesis put forward by MacGorman et al. 1989; MacGorman and Nielsen 1991 to explain unusually high Z ratios observed in severe (tornadic) storms. Specifically, the observations imply that in strong-updraft storms: 1) intracloud flashes may be favored mechanisms for adjustment between the (now elevated) upper positive and main negative charge regions, 2) intracloud flashes may also be favored mechanisms for adjustment between of the (now more vertically separated) main negative and lower positive charge regions, and 3) positive CGs may be favored ground discharges from the (now more developed and vertically isolated) lower positive charge region (and further favored by the increased distance of the main negative charge region from ground).

The above speculations are indirect and heuristic, and should be approached with caution. First, our synopsis of the results of Stolzenburg et al. 1998; Stolzenburg et al. 1998a; Stolzenburg et al. 1998b ignores the fact that significantly more complex vertical charge structures are found by these authors outside updraft cores. Second, we have only indirect linkage between the observed lightning properties and the geographic distribution of realized updraft spectra across the continental U.S., and issues such as the relative positioning of the Z extrema are not directly handled by these hypotheses. For example, the Z and %PCG extrema appear shifted north and west of regions of climatological peak tornado occurrence (Concannon et al. 2000); similarly, Branick and Doswell (1992) presented evidence that predominantly positive CG supercells appeared shifted north of Oklahoma and may have been restricted to the low precipitation (LP) end of spectrum of supercell storms. Use of climatological Z inferences may be obscuring underlying physics (these inferences are an average of instantaneous measurements of both 'garden variety' and severe storms, whose probability of occurrence may vary regionally; e.g. Fig. 7b), or the elevated dipole hypothesis alone may be insufficient to explain the spatial variability.

Additional caution is recommended in extrapolating these results to the behavior of deep tropical convection, as it is unclear how much of the influences of storm morphology on Z are dependent on uniquely midlatitude factors which modulate storm organization, such as strong vertical wind shear or midlevel dry intrusions. We note that the majority of electrified cells observed over the continental U.S. (Fig 7b) occur at more southern latitudes in which strong $Z(\overline{f}_c)$, $Z(\%PCG)$, $Z(LPCG)$ relationships do not appear to dominate; this 'regime' of thunderstorm occurrence *might* be more typical of deep tropical electrified cells than the extreme storms of the Great Plains.

4 Conclusions

The intracloud to cloud to ground lightning ratio, Z, has been computed over the continental United States using four years of combined OTD and NLDN data. This represents the first regional (non-point) determination of climatological values of this ratio ever constructed. There is no evidence to support prior infer-

[2] Additional evidence that the lower negative charge center is not significantly elevated in "garden variety" thunderstorms might be found in examination of the altitudes of the lower channels of intracloud flashes in Florida thunderstorms, as mapped by VHF/Time-of-Arrival systems (Krehbiel et al. 1984) and Heckman and Ushio (pers. comm. 2000) using a much larger sample. In these storms, lower channel altitude remains relatively constant with time, while upper channel altitude appears to rise with storm cell updrafts.

ences of broad latitudinal dependency in this parameter, although the current study cannot rule out the possibility of variability between the deep tropics and the southernmost latitude considered here (about 25N). Low Z values are found over mountain regions, although a general Z-elevation dependency does not appear to be unique (lower Z anomalies occur over the Appalachian mountains than the Rocky Mountains). It is impossible to determine from the present analysis whether this 'mountain' signal arises from geometric effects (surface proximity to the main negative charge region) or meteorological effects (limited storm organization in mountain regions).

Local high Z anomalies are found running southwest-northeast through the upper Great Plains, coincident with high anomalies in climatological NLDN percent positive CGs (%PCG) and large positive CGs (LPCG). The anomaly region is also consistent with the climatologically favored region for MCS occurrence, and with local high anomalies in reported severe storm occurrence, particularly severe storms with predominantly positive CGs (PPCG). We tentatively ascribe this anomaly to a local high-bias of the complete spectrum of thunderstorm updraft strength, and invoke the 'elevated dipole' hypothesis of MacGorman et al. 1989; MacGorman and Nielsen 1991 and corroborating electric charge structure evidence of Stolzenburg et al. 1998; Stolzenburg et al. 1998a; Stolzenburg et al. 1998b to explain the Z, $LPCG$ and $\%PCG$ anomalies. Under this hypothesis, stronger updrafts loft both the main negative and upper positive charge centers to higher altitudes, disfavoring CG discharges, and yield an enhanced lower positive charge center, favoring positive CGs over negative CGs. The northward offset of Z maxima from local maxima in severe storm occurrence might be attributed to a broader spectrum of non-severe ('garden variety') thunderstorms occurring at lower latitudes (i.e., in the upper Great Plains, severe storms represent a larger proportion of all flashing storms, and hence have greater impact on mean Z). The overall suggestion is that storm intensity, morphology and/or level of organization have far more significant impacts on realized Z values than 'environmental' variates such as the freezing level height, troposphere depth or surface elevation.

The observed mountain and severe storm anomalies cannot be explained solely by either constant or geographically variable bias in actual OTD or NLDN detection efficiency, given plausible uncertainties in both. Constant bias in assumed detection efficiency at best amplifies or dampens preexisting geographic variability in Z by a modest amount. Geographically variable sensor detection efficiency would both have to be unrealisitically large and operate contrary to the expected physics to explain the observed signal; hence these anomalies appear to be real. Sensor bias cannot yet be ruled out as a factor in high Z anomalies inferred in coastal regions and in the Pacific Northwest. It is speculated that if real, these Northwest anomalies might correspond to very low flash rate storms which produce mostly or only intracloud flashes during their lifetime, or might correspond to a winter bias in the occurrence of lightning in which mechanisms other than those discussed above might determine Z. Further analysis on a seasonal basis (using the complete OTD dataset, through its end-of-mission in 2000) could yield additional insight into the geographic distribution of Z, as well as its causes.

5 Acknowledgements

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6 Appendix: Sensitivity Analysis

Robust extraction of geographic variability from even one lightning detection system is at best difficult, given limitations in the performance characteristics of most remote sensing systems (and in our ability to diagnose these limitations over large domains). Extraction of such variability from multi-sensor datasets is even more risky, and skepticism of merged sensor results is both wise and warranted. We address the three most likely sources of potential bias in our climatological Z estimates: uniform bias in our estimates of OTD or NLDN detection efficiency, possible differences between OTD intracloud and cloud-to-ground detection efficiency, and possible geographic variability in OTD and NLDN detection efficiency. For the first two bias sources, a direct (forward) calculation is performed to demonstrate that the inferred regional variability cannot be accounted for by such uniform detection efficiency estimate bias, only magnified (enhanced) to a small degree. For the third bias source, we perform reverse calculations from the inferred Z fields to demonstrate that the magnitude of detection efficiency bias required to solely *explain* the inferred spatial pattern is beyond that considered plausible given prior validation studies of the OTD and NLDN.

a. Uniform bias in OTD or NLDN detection efficiency estimates

The first possible bias source is uniform error in our estimate of the actual OTD or NLDN detection efficiencies (DE); for OTD, this is a total lightning DE, for NLDN, a CG DE. Here we examine only the possibility that these estimates contain uniform bias (i.e., no geographic variability). We can directly represent the resulting bias in Z estimates using the following framework:

Let $\hat{f}$ denote a regional flash rate density estimate (total for OTD, CG for NLDN) using the estimated detection efficiency $\widehat{DE}$, i.e.:

$$\hat{f} = \frac{n_f}{(A)(\delta t)(\widehat{DE})} \tag{2}$$

where n_f is the number of flashes observed by the sensor during the study period, δt is the total duration of observation, and A is the area of a 0.5×0.5 degree grid cell in the composited data. Hence, as in Eq. 1,

$$\hat{Z} = \frac{\hat{f}_{OTD} - \hat{f}_{NLDN}}{\hat{f}_{NLDN}} \tag{3}$$

Further define a general bias factor β to describe actual biases in our estimated detection efficiency $\widehat{DE}$ and IC:CG ratio $\hat{Z}$:

$$\beta_{DE} = \frac{\widehat{DE}}{DE_{true}}, \beta_Z = \frac{\hat{Z}}{Z_{true}} \tag{4}$$

After some algebraic rearrangement, the resulting (fractional) bias in $\hat{Z}$ can then be represented by:

$$\beta_Z = \frac{\hat{f}_{OTD} - \hat{f}_{NLDN}}{\hat{f}_{OTD}\frac{\beta_{DE,OTD}}{\beta_{DE,NLDN}} - \hat{f}_{NLDN}} \tag{5}$$

We use this expression to examine the spatial variability in β_Z for two possible scenarios: 1) $\widehat{DE}_{OTD}$ is 15% too high, (2) $\widehat{DE}_{OTD}$ is 15% too low. Note that the structure of Eq. 5 indicates that the second case is also mathematically equivalent to a situation with $\widehat{DE}_{NLDN}$ being 15% too high. The resulting β_Z is shown in Fig. 11a,b. The biases have expected effects on $\hat{Z}$: overestimates of DE_{OTD} (underestimates of DE_{NLDN}) lead to an exaggeration of regional variability, while underestimates of DE_{OTD} (overestimates of DE_{NLDN}) tend to dampen regional signals. We note that the magnitude of the β_Z fields for these reasonable values of detection efficiency uncertainty *cannot* by itself explain the observed spatial patterns.

b. OTD IC, CG detection efficiencies not equal

As discussed in Section 2, we have operationally assumed that the OTD intracloud detection efficiency is the same as its cloud-to-ground detection efficiency. Since the limited navigational accuracy of the OTD precludes accurate validation of $DE_{OTD,IC}$ by short range surface total lightning sensors, this is an essentially unconfirmed assumption. Its implications for the present analysis warrant discussion here.

To do so, we recognize that the OTD total lightning flash rate density is the sum of the intracloud and cloud-to-ground contributions. If we assume that the NLDN provides the 'true' climatological CG flash rate, and denote 'true' quantities as those without hats, we have:

$$\hat{Z} = \frac{f_{CG}\frac{DE_{OTD,CG}}{\widehat{DE}_{OTD}} + f_{IC}\frac{DE_{OTD,IC}}{\widehat{DE}_{OTD}} - f_{CG}}{f_{CG}} \tag{6}$$

Since $\widehat{DE}_{OTD}$ is taken here as the empirically estimated CG detection efficiency $\widehat{DE}_{OTD,CG}$ from Boccippio et al. (2000), this reduces to the simple expression:

$$\hat{Z} = (\frac{DE_{OTD,IC}}{\widehat{DE}_{OTD,CG}})Z + (\frac{1}{\beta_{DE_{OTD,CG}}} - 1) \tag{7}$$

The second term on the right is additive and likely small (i.e., we have reasonable confidence in $\widehat{DE}_{OTD,CG}$ and $\beta_{DE_{OTD,CG}}$ is near unity). Hence, the estimated IC:CG ratio is directly and linearly biased by the actual differences in OTD intracloud and cloud-to-ground detection efficiencies.

Two *extreme* limiting cases can be examined. First, if $\widehat{DE}_{OTD,CG}$ were completely and erroneously computed from intracloud flashes (implausible, given clear time correlation between OTD flashes and NLDN CGs reported in Boccippio et al. (2000), and direct measurements of optical pulses from CGs in optically thick storms by the U2 aircraft Optical Pulse Sensor (Goodman et al. 1988)), then $\widehat{DE}_{OTD,CG} \sim DE_{OTD,IC}$ and $DE_{OTD,CG} \sim 0$, hence from Eq. 6 or 7, $Z \sim \hat{Z} + 1$. Alternatively, if $\beta_{DE_{OTD,CG}} \sim 1$ but $DE_{OTD,IC} = 1$ (worst possible bias), $Z \sim \hat{Z}/2$. These represent extreme bounds on two likely cases ($DE_{OTD,CG}$ lower than estimated, $DE_{OTD,IC}$ higher than estimated), and hence are upper and lower bounds on true Z. Actual OTD-based bias (*including* possible regional variability in these detection efficiencies) must lie somewhere between these bounds.

Physical inference suggests that, if anything, $DE_{OTD,IC}$ is likely equal to or higher than $DE_{OTD,CG}$

(less optical scattering occurs between the lightning source and cloud top). This has possibly been demonstrated by a limited case study of the LIS (OTD follow-on sensor) by Thomas et al. (2000). Furthermore, *total* lightning DE estimates for on-orbit operational OTD sensitivity (55%-70%), based on U2 aircraft Optical Pulse Sensor data (Koshak et al. 2000), are within the range of uncertainty for $\widehat{DE}_{OTD,CG}$ as reported in Boccippio et al. (2000). This places a reasonable upper bound on $DE_{OTD,IC}$, and indicates that the magnitude of the bias should be small. We note that even the extreme limits discussed above cannot explain the high Great Plains anomaly in $\hat{Z}$, hence the more plausible DE uncertainty certainly cannot.

Finally, it is observed that if differential $DE_{OTD,CG}$ and $DE_{OTD,IC}$ exists and is regionally invariant, Eq. 7 dictates that such differences cannot alone account for regional variability in the estimated $\hat{Z}$; they can only amplify existing regional variability.

c. Geographic variability in OTD or NLDN detection efficiency

As discussed above, another likely source of bias in our inferred $\hat{Z}$ distribution is the possibility of geographic nonuniformity in the detection efficiencies of either the NLDN or OTD. This might occur with the ground network in coastal regions with limited network coverage, in mountain regions with possible signal blockage, etc.[3] It might occur with the satellite sensor if storm optical depth covaried with geographic location (i.e., greater optical depth leads to greater lightning optical pulse attenuation and reduced detection efficiency). Our approach here will be to determine what level of regional DE variability would be required to explain *all* of the observed regional Z variability, and assess whether this corresponds to plausible values given prior validation studies of both sensors.

We can thus perform a direct reverse calculation from Eq. 5, solving for the OTD or NLDN DE biases 'necessary' to yield the observed local departures from the continental mean $\overline{Z}$:

$$\frac{\beta_{DE,OTD}}{\beta_{DE,NLDN}} = \frac{\frac{\hat{f}_{OTD}-\hat{f}_{NLDN}}{\beta_Z} + \hat{f}_{NLDN}}{\hat{f}_{OTD}} \qquad (8)$$

where β_Z is taken as the ratio between the apparent local $\hat{Z}$ and the domain-averaged $\overline{\hat{Z}}$ (assumed to be a 'true', geographically invariant [i.e., bias-free] value). As discussed above, this technique equivalently yields the geographic bias necessary in *either* sensor to yield the observed patterns. Fig. 12 shows the computed $\frac{\beta_{DE,OTD}}{\beta_{DE,NLDN}}$ fields.

For anomaly (1) (the Great Plains Z maxima), our estimated $\widehat{DE}_{OTD}$ would have to be 50% of the actual sensor efficiency to account for the anomaly (i.e., we would need to be overcorrecting observed OTD flash counts by a factor of 2). This would not only suggest that the true DE_{OTD} is close to unity in this region (highly unlikely), but it would demand higher local detection efficiency in a region characterized by severe storms whose high liquid water contents, if anything, should lead to a lower optical detection efficiency. This observation extends to differential $DE_{OTD,IC}$ and $DE_{OTD,CG}$ discussed in the previous section; the extreme bias limits ($Z \sim \hat{Z}/2$, $Z \sim \hat{Z}+1$) cannot account for the high anomaly. Similarly, a local NLDN DE low anomaly of about 45% (a twofold overestimate on our part) could account for the signal, but the network sensor configuration in this region shows no spacing extremum necessary to explain such an extreme departure (Cummins et al. 1998; Orville and Huffines 1999).

Alternatively, for anomaly (2) (mountain region Z minima), the reverse situation holds: we would need to be locally overestimating $\widehat{DE}_{OTD}$ by a factor of 1.6-2.0 or so in a region where high cloud bases and concomitantly lower expected adiabatic liquid water contents should lead to relatively high DE_{OTD} anomalies, if anything. Similarly, a twofold underestimate of $\widehat{DE}_{NLDN}$ on our part could yield the mountain anomaly, but this would require DE_{NLDN} to be locally truly greater than unity. Thus, in order to be explainable by geographic variability in the sensors' detection efficiency alone, the two main U.S. Z anomalies would require not only excessive (factor of two) variability over previously estimated values, but would also force this variability to occur in ways which are inconsistent with its most likely physical causes (optical attenuation for OTD, sensor spacing for the NLDN).

For the coastal (Maine and California coast) anomalies, the 'necessary' bias ratio is also found to be extreme (0.4-0.8) but within the realm of possibility given spatial maps of detection efficiency estimates presented in Cummins et al. (1998). As such, we recompute the Z field using a digitized version of the geographic distribution of DE_{NLDN} presented by these authors. The resulting Z is shown in Fig. 13; under these assumptions, the continental mean is found to be 2.64 with a standard deviation of 1.1. Use of DE_{NLDN} as estimated by Cummins et al. (1998) will represent a slight *overcorrection* at northern latitudes, due to improvements in detec-

[3] Errors in the data processing approach, e.g. our exclusion of <10 kA nominal +CG flashes, could also conceivably yield an effective regional DE bias.

tion efficiency arising from inclusion of solutions from Canadian Lightning Detection Network stations during the last few months of the study period (Fournier and Pyle 1998; Cummins et al. 1999). Nonetheless, it is evident from Fig. 2 and 13 that inclusion of best estimates of spatial variability in DE_{NLDN} does not fully eliminate anomalies (3) or (4), especially in Oregon, Idaho and western Montana. Similarly, high anomalies along the California coast and in northeast New England are lessened but not eliminated. Given that the climatological flash rates are low in these regions (i.e., the $\hat{Z}$ is high variance) and the DE_{NLDN} estimate imperfect, we cannot yet confidently claim that these latter anomalies are real.

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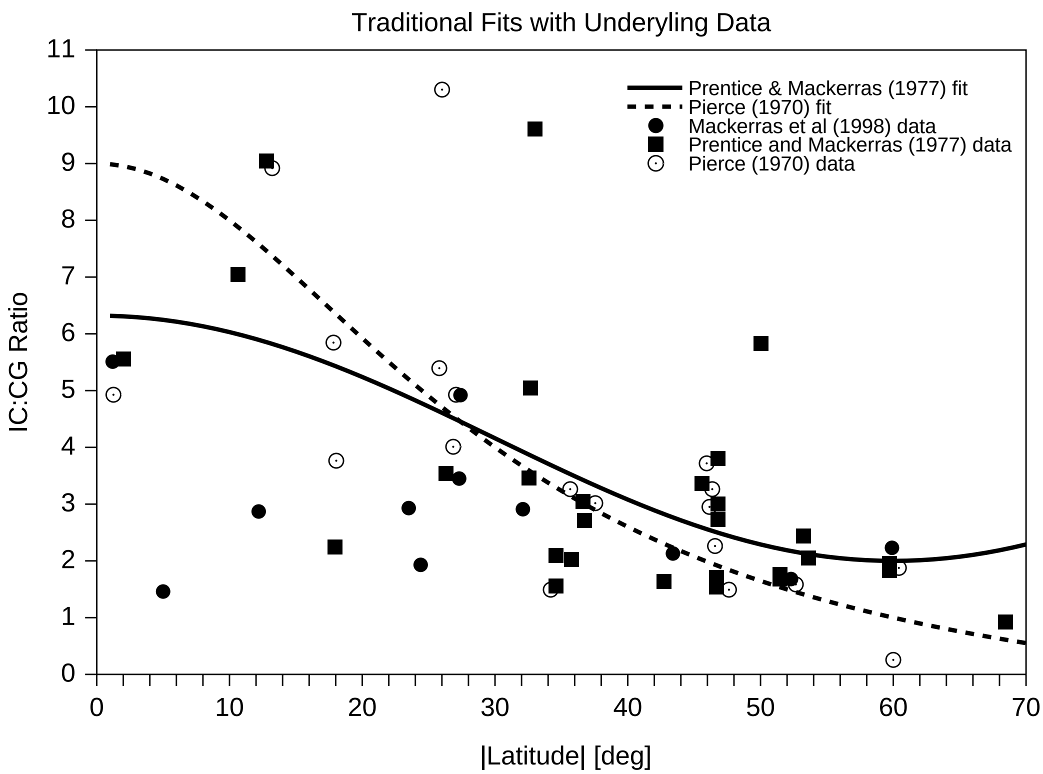


FIGURE 1: Prior ensemble observations of regional IC:CG ratio Z and its inferred dependence on latitude.

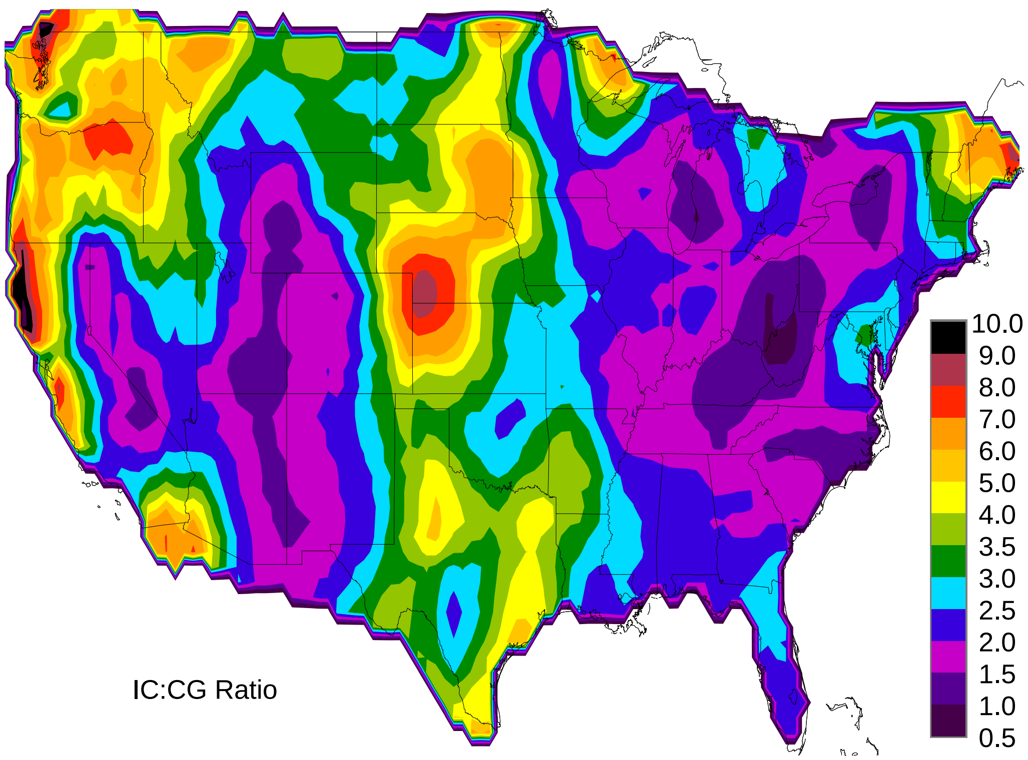


FIGURE 2: Climatological mean IC:CG ratio Z estimated from four years of OTD and NLDN observations. Estimates are computed from 0.5 degree composite grids smoothed with a 3.5 degree operator.

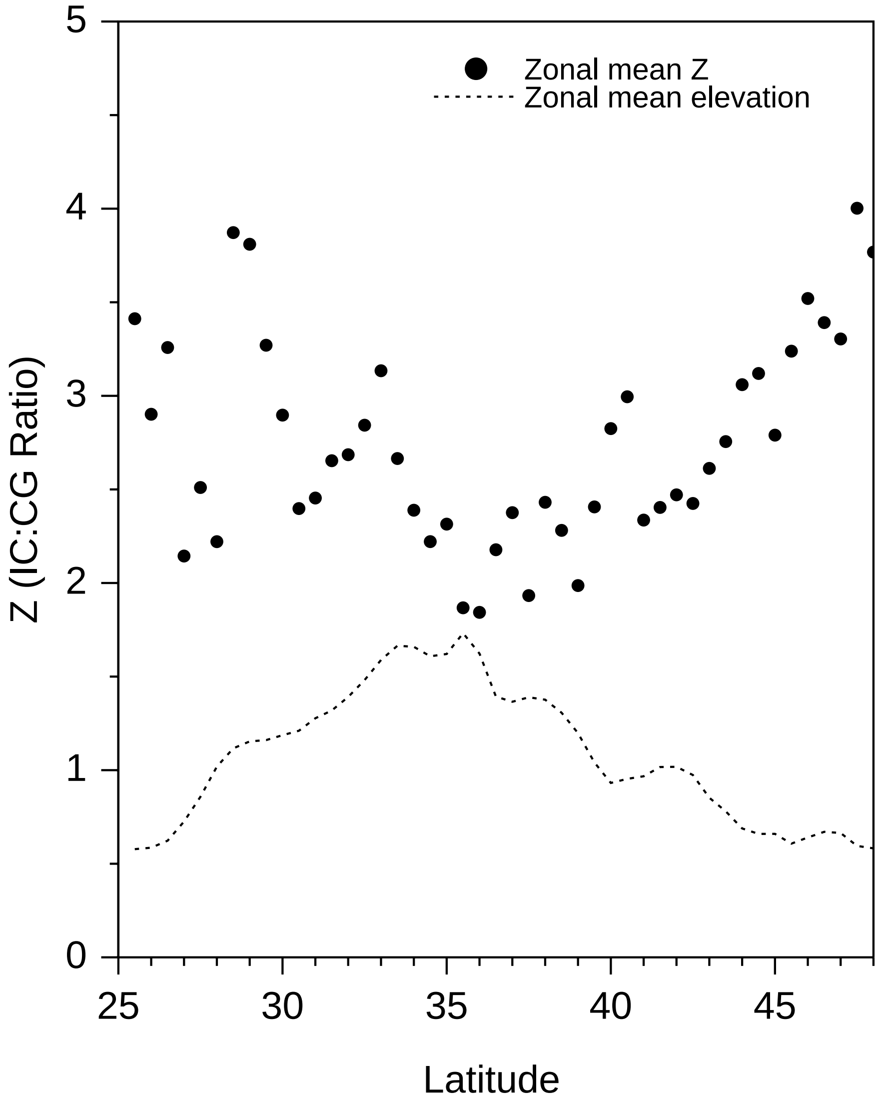


FIGURE 3: Zonally averaged (0.5 degree bins) estimates of Z as a function of latitude; zonal average elevation (in km) is overlaid. While some latitudinal variability is evident, it is likely due to coincidental aliasing of elevation effects.

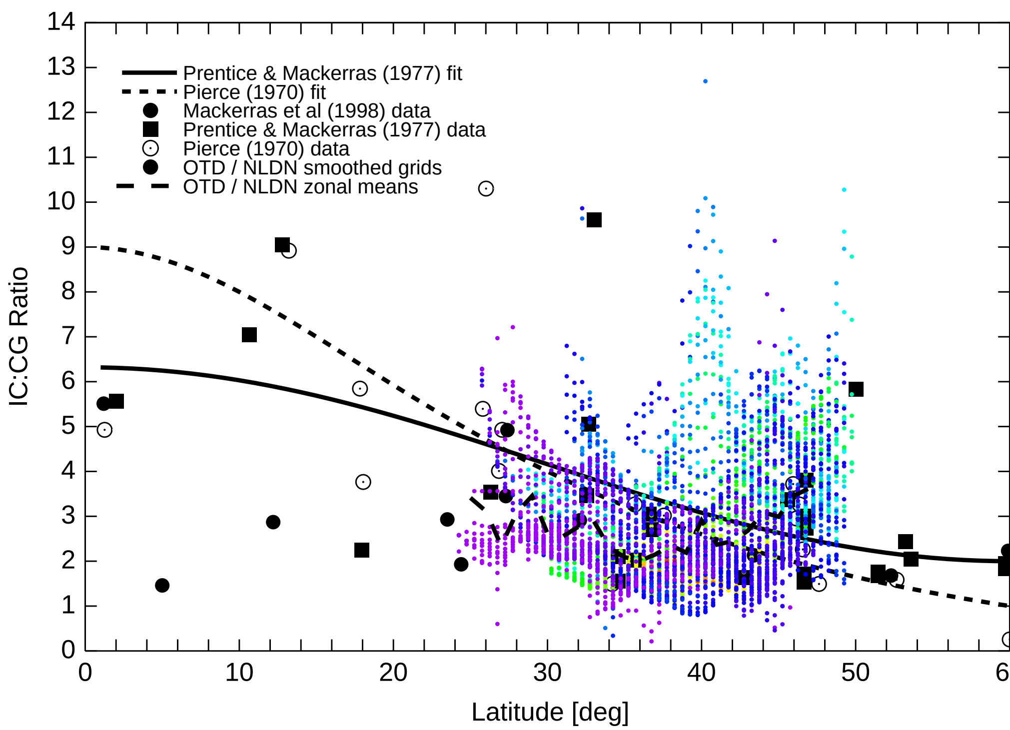


FIGURE 4: Individual (smoothed) 0.5 deg grid estimates of Z overlaid on the prior observations of Pierce (1970), Prentice and Mackerras (1977), Mackerras et al (1998). While most of the values are consistent with prior estimates, the variability in local mean Z is comparable to the total tropics-to-midlatitude drop suggested by earlier investigators.

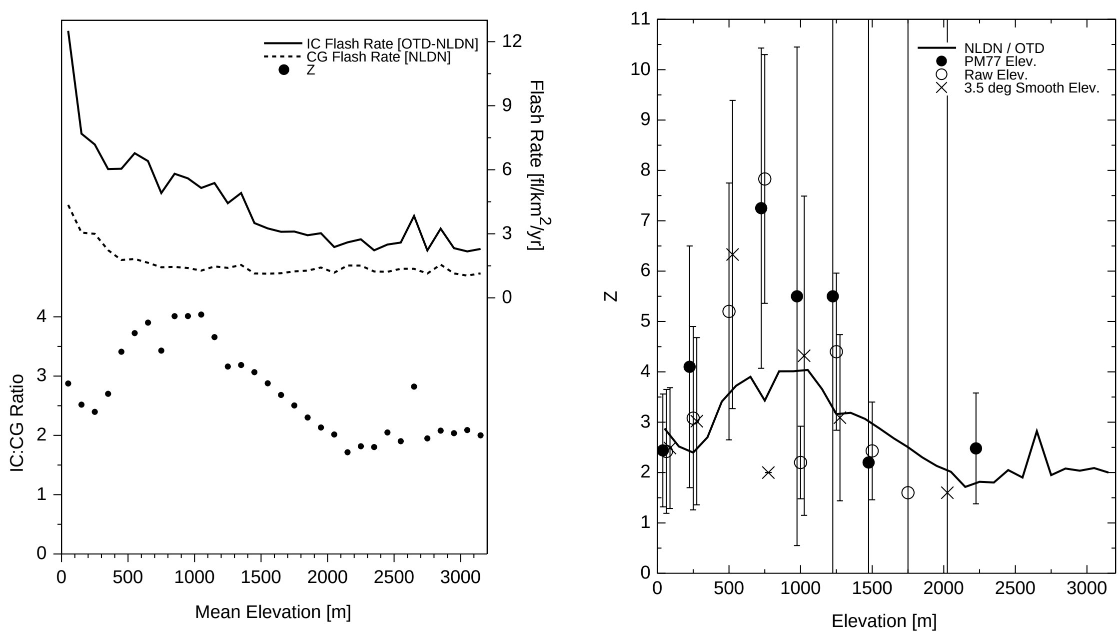


FIGURE 5: *(a)* Mean Z value for all 0.5 deg grid locations within individual 100m elevation bands. *(b)* Point data (PM77) of Prentice and Mackerras (1977) aggregated into elevation bins using three different elevation assignment techniques (see text for details). Standard deviations within each bin are overlaid.

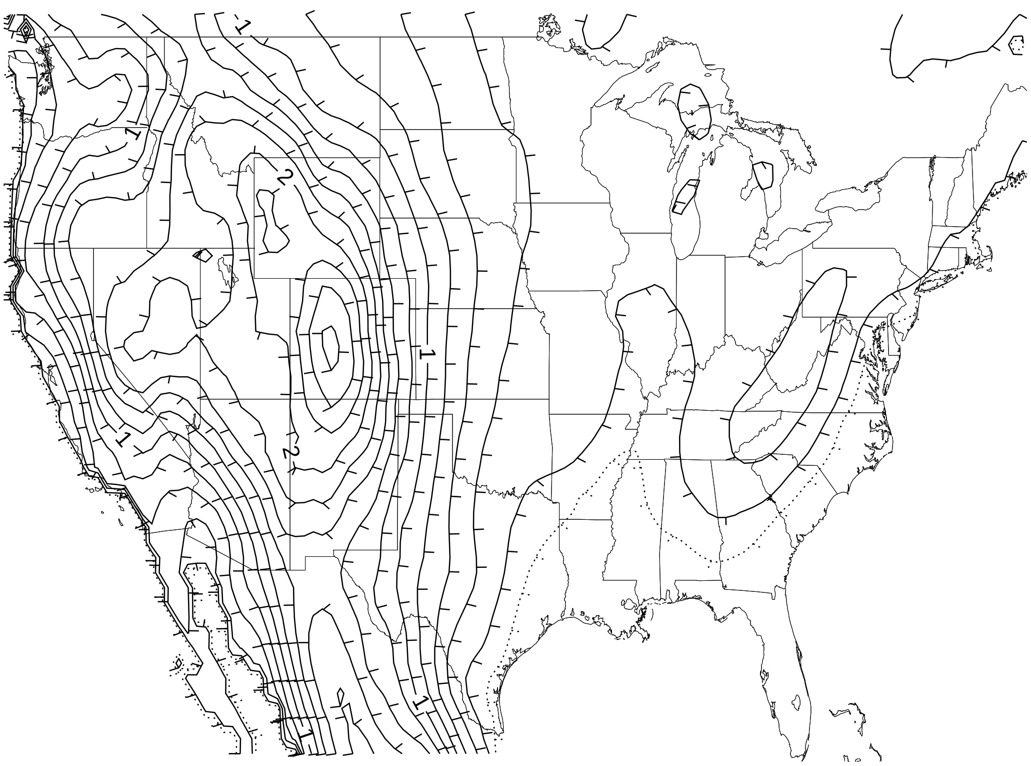


FIGURE 6: Ground elevation smoothed using the same as operator as in Fig. 2. Contours are at 100 m (dotted) and every 200 m above that; 1 km and 2 km contours are labelled. Ticks indicate ”downslope” direction.

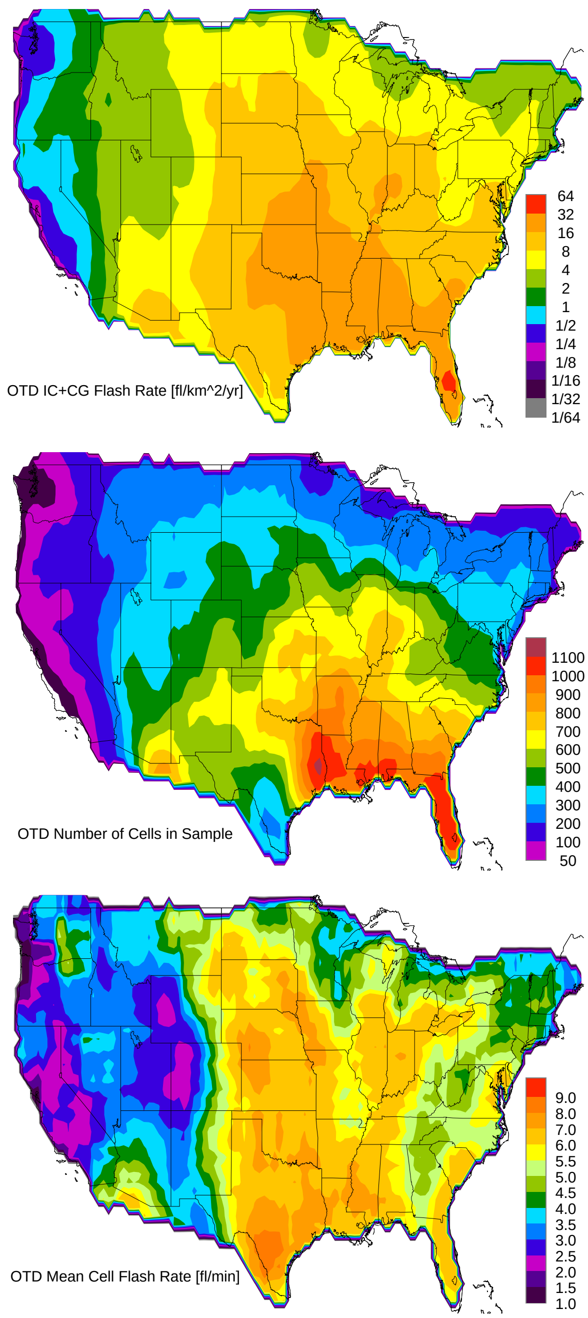
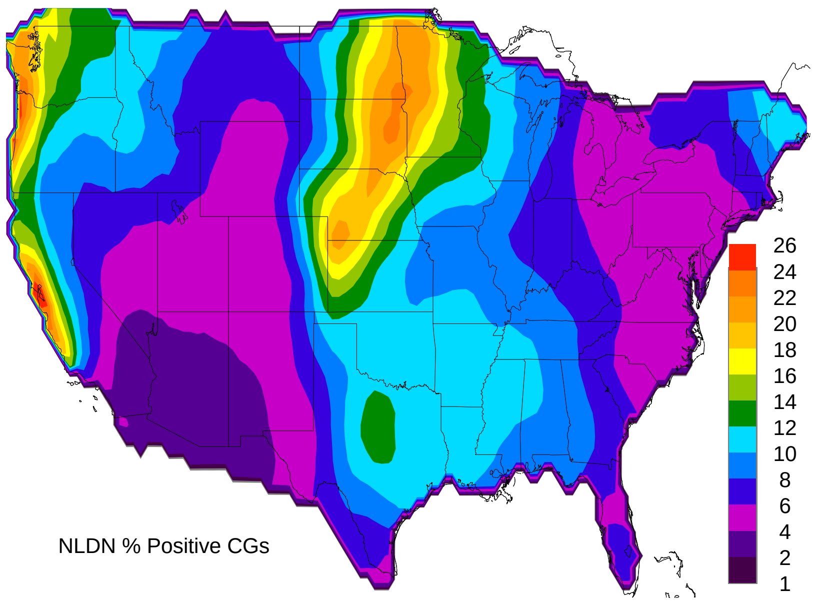


FIGURE 7: *(a)* Climatological regional flash rate density f_r observed by the OTD (detection efficiency corrections are applied as discussed in the text). *(b)* Number of cells ("area" products) n_c observed by OTD during the study period (May 1995 - April 1999). The map is not normalized for geographic differences in total satellite viewing time. Over the U.S., the regional flash rate density signal f_r is dominated by the frequency of occurrence of flashing cells. *(c)* Mean per-cell flash rate f_c during the study period.



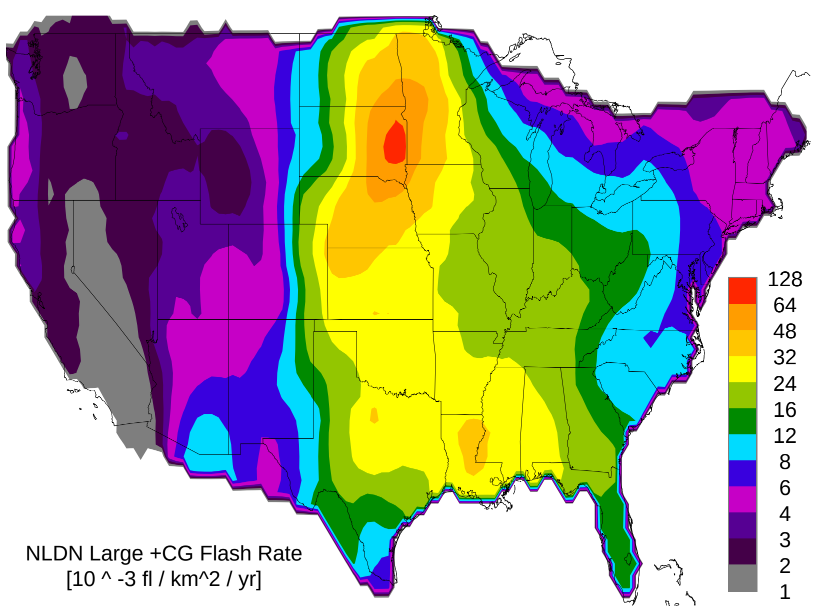


FIGURE 8: *(a)* Percentage of positive CGs (%PCG) observed by NLDN during the study period (May 1995 - April 1999). *(b)* Flash rate density of large positive CGs (LPCG; peak current greater than 75 kA) during the study period.

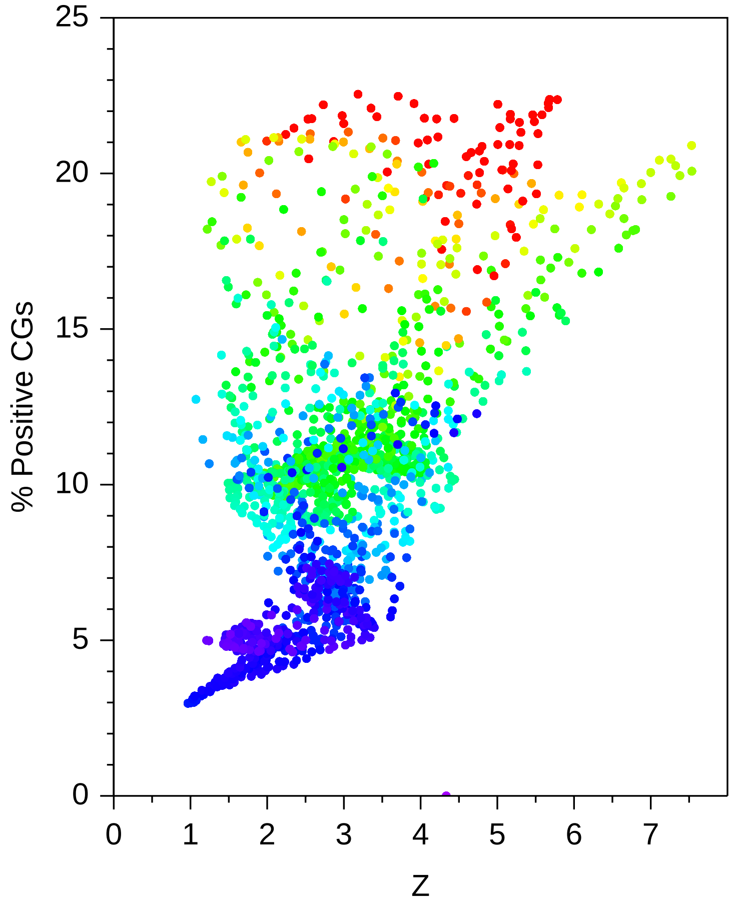


FIGURE 9: Pairwise values of Z and %PCG for each 0.5×0.5 degree grid location, after 3.5 degree spatial smoothing. Values are plotted for the central U.S. (89W to 109W, all latitudes) only. Color shading denotes the LPCG flash rate density using a 'cool-warm' color table, with maxima corresponding to the maximum values in Fig. 8b.

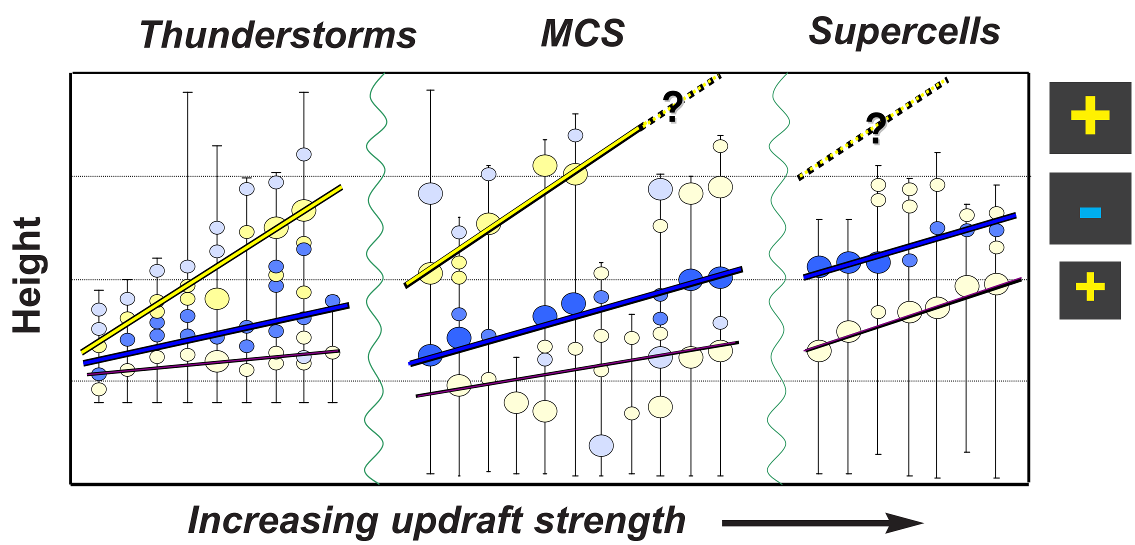


FIGURE 10: Reinterpretation of charge center locations inferred by Stolzenburg et al (1998b) from electric field balloon soundings in various storm types, arranged in order of increasing balloon ascent rate (loosely, updraft velocity) within each storm type. Charge ellipses are centered at the center height of each analyzed charge layer, with larger ellipses for layers deeper than 1 km. Soundings extend vertically until balloon burst or loss-of-signal. Uppermost negative charge regions in the "thunderstorm" category are hypothesized to be cloud-edge screening layers, and hence not central to the trends shown here.

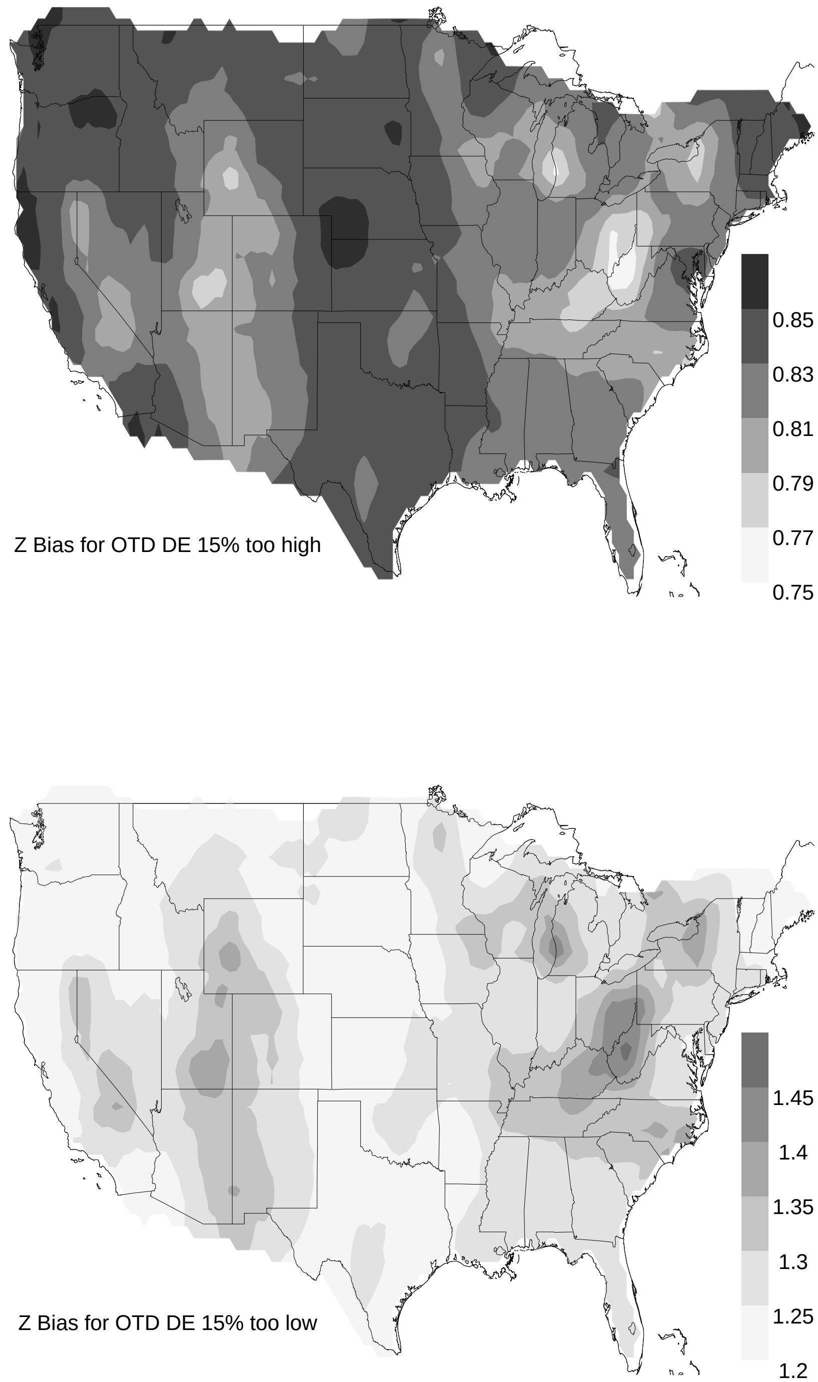


FIGURE 11: Estimated Z bias β_Z resulting from *(a)* overestimating (underestimating) OTD (NLDN) detection efficiency by 15%, and *(b)* underestimating (overestimating) OTD (NLDN) detection efficiency by 15%. In the first case, an overall bias is introduced and regional differences are slightly exaggerated; in the second case, an overall bias is introduced and regional differences are damped.

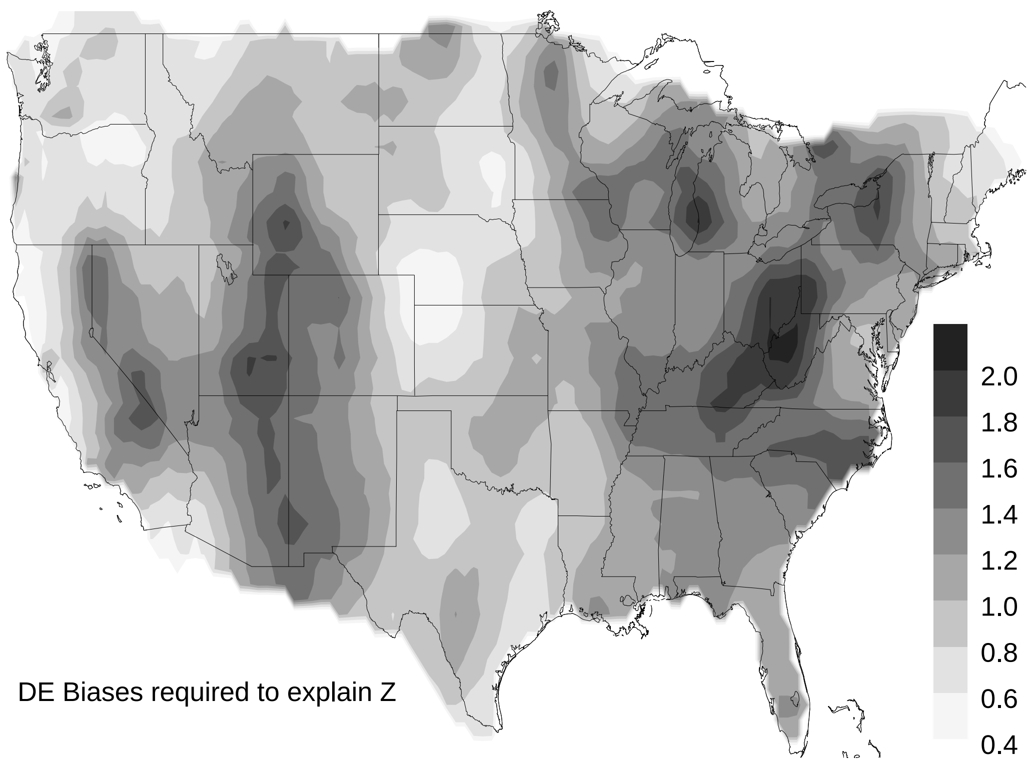


FIGURE 12: Ratio of OTD and NLDN detection efficiency biases necessary to solely account for regional variability in the inferred Z. Low values require that OTD detection efficiency is locally underestimated or NLDN detection efficiency is locally overestimated in order to account for the local Z anomaly; high values require the converse.

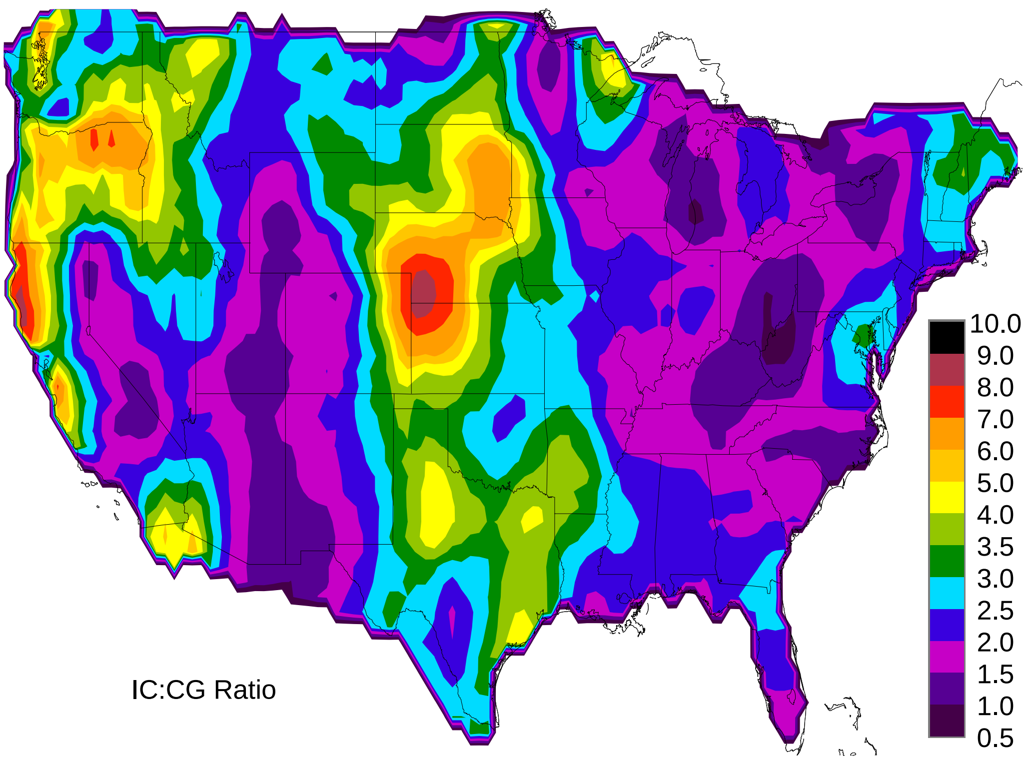


FIGURE 13: Intracloud to cloud-to-ground ratio Z computed using NLDN spatially-variant detection efficiency as computed by Cummins et al (1998).