Down to Earth With Nuclear Electromagnetic Pulse: Realistic Surface Impedance Affects Mapping of the E3 Geoelectric Hazard

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Abstract An analysis is made of Earth-surface geoelectric fields and voltages on electricity transmission power-grids induced by a late-phase E3 nuclear electromagnetic pulse (EMP). A hypothetical scenario is considered of an explosion of several hundred kilotons set several hundred kilometers above the eastern-midcontinental United States. Ground-level E3 geoelectric fields are estimated by convolving a standard parameterization of E3 geomagnetic field variation with magnetotelluric Earth-surface impedance tensors derived from wideband measurements acquired across the study region during a recent survey. These impedance tensors are a function of subsurface three-dimensional electrical conductivity structure. Results, presented as a movie-map, demonstrate that localized differences in surface impedance strongly distort the amplitude, polarization, and variational phase of induced E3 geoelectric fields. Locations with a high degree of E3 geoelectric polarization tend to have high geoelectric amplitude. Uniform half-space models and one-dimensional, depth-dependent models of Earth-surface impedance, such as those widely used in government and industry reports informing power-grid vulnerability assessment projects, do not provide accurate estimates of the E3 geoelectric hazard in complex geological settings. In particular, for the Eastern-Midcontinent, half-space models can lead to (order-one) overestimates/underestimates of EMP-induced geovoltages on parts of the power grid by as much as ±1,000 volts (a range of 2,000 volts)—comparable to the amplitudes of the geovoltages themselves.

Plain Language Summary A nuclear explosion in the near-Earth space environment can produce an electromagnetic pulse (EMP) at the Earth’s surface. A low-frequency part of the EMP signal, known as E3 and covering periods from about a tenth of a second to a few hundred seconds, can induce geoelectric fields in the conducting solid Earth, interfering with the operation of electricity power grids. To investigate this, accurate estimates are required of the Earth’s surface impedance—that is, the relationship between geomagnetic and geoelectric field variation. Surface impedance is a function of the electrical conductivity of subsurface rock structures. Using impedance tensors obtained from survey measurements, time-dependent scenario maps are constructed of the E3 geoelectric fields and power-grid geovoltages that would be generated by a hypothetical nuclear explosion above the United States. Over the course of the scenario, geoelectric amplitude, polarization, and variational phase are shown to differ significantly from one location to another, mostly as the result of geographic granularity in impedance. It is concluded that extremely simple impedance models, such as those widely used in government and industry reports concerned with power-grid vulnerability assessment, do not generally provide accurate estimates of the E3 geoelectric hazard in complex geological settings.

1. Introduction

Tests of nuclear weapons during the 1960s showed that explosions set hundreds of kilometers above the Earth’s surface can adversely impact ground-based electric and electronic systems across a wide geographic area (e.g., Dupont, 2004). The United States (U.S.) Starfish Prime test of July 9, 1962, 09:00 UT (July 8, 1962, 11:00 p.m. local Hawaii time) was relatively isolated: A 1.4 megaton explosion was set 400 km above Johnston Island in the Pacific Ocean (Berkhouse et al., 1962) (16.5°N, 190.4°E). But 1,350 km away from the ground-zero (epicenter) point, nighttime Hawaii was illuminated with a bright flash of light, radios
felled in Honolulu and then became noisy, burglar alarms were tripped (Glasstone & Dolan, 1977), and local street-lights in Honolulu went black (Vittitoe, 1989). The Starfish Prime test also damaged satellites and produced unusual aurorae (e.g., Osmundsen, 1962). Perhaps even more dramatic was Soviet test 184 (also designated K3) on October 22, 1962 (UT): A 300 kiloton explosion was set 300 km above Jezkazgan, Kazakhstan (47.75°N, 64.0°E) (e.g., Zetser et al., 2004). It caused fires in power supplies and blew fuses on a 500 km long communication-cable system located hundreds of kilometers from the epicenter (Greetsai et al., 1998; Seguin, 1995).

These impacts were the results of electromagnetic pulses (EMPs) interacting with the atmosphere, the geomagnetic field, and the solid Earth. A nuclear explosion of several hundred kilotons set at altitudes of several hundred kilometers generates an EMP having three semi-distinct phases (Gombosi et al., 2017; Legro et al., 1985; Rivera et al., 2016). The first, called E1, lasts for less than a microsecond. X-rays and gamma-rays directed toward the Earth ionize a layer of the atmosphere between altitudes of 30 and 100 km; electric currents induced in this layer broadcast a secondary pulse of electromagnetic radiation that can reach the Earth’s surface and interfere with electrical and electronic systems, including possibly damaging protective systems on electric-power grid infrastructure, rendering them vulnerable to later EMP phases (e.g., Graham et al., 2008). The lingering effects of E1 are encompassed by an intermediate phase, or E2, that lasts from a microsecond to about a hundredth of a second. This period band corresponds, roughly, to that of lightning, and since many terrestrial technological systems are lightning hardened, it is often assumed that E2 would not be especially impactful. As a consequence, it is not very well studied (Gombosi et al., 2017, their Chapter 2.2; Rivera et al., 2016, their Section 3.2).

Our focus, here, is on the E3 phase, also known as magnetohydrodynamic (MHD) EMP or late-time EMP. This phase covers a period band from a tenth of a second to a few hundred seconds. It consists of a two-part jerking of the Earth’s magnetic field (Gombosi et al., 2017; Legro et al., 1985; Rivera et al., 2016). During the first part, E3A, the blast creates a diamagnetic plasma bubble that pushes outwards against ambient geomagnetic field lines. During the earliest parts of this phase, the ionized atmosphere below the explosion tends to pin geomagnetic field lines in their pre-blast locations, and the amplitude of abrupt changes in the geomagnetic field are damped before reaching the Earth—the ionization of the atmosphere acts like a shield of early E3A field variation. After about 10 s, however, the ionization begins to subside, and diffusion permits the amplitude of E3A geomagnetic field variation to increase at the Earth’s surface. During the second part, E3B, the ionized and heated atmosphere below the explosion buoyantly rises, a heaving that brings the field lines beneath the explosion hypocenter back together (e.g., Gombosi et al., 2017; Legro et al., 1985; Rivera et al., 2016). E3 geomagnetic disturbance induces geoelectric fields in the electrically conducting solid Earth, and these, in turn, drive uncontrolled currents in grounded electric-power grids that can interfere with transmission operations and damage high-voltage transformers (e.g., Electric Power Research Institute, 2019; Tesche et al., 1991).

The proliferation of nuclear weapons has led to concerns about the potential impact that nuclear EMPS would have for national and international security (e.g., Baker, 2019; Broad, 1981; Graham et al., 2008; Popik et al., 2017). Some scenario simulations anticipate, for example, that a high-altitude detonation of even a relatively primitive nuclear device could cause widespread and long-lasting damage to a nation’s electric-power grid (e.g., Popik et al., 2017). For this reason, EMP is frequently covered by the popular press (e.g., Bump, 2016, January 15; Burnham, 1983, June 28; Conca, 2020, June 25; Hambling, 2017, September 28; Nikolewski, 2017, March 16). Recognizing the threats posed by EMP weapons, the United States Congress has commissioned informational reports on EMP (Graham et al., 2008; Wilson, 2006), and the Department of Energy (DOE) and the commercially supported Electric Power Research Institute (EPRI) have developed strategies for improving understanding of the EMP threat to the U.S. national electric-power grid (Department of Energy, 1983, 2017; Department of Energy and Electric Power Research Institute, 2016). A March 2019 executive order (13,865) issued by the President of the United States directs Federal agencies to improve national resilience to EMPS. In particular, the Department of the Interior and the U.S. Geological Survey (USGS) are directed to support research that enhances the understanding of variations of Earth’s magnetic field associated with EMPS.

Time-dependent maps of E3 EMP geoelectric fields can, in principle, be obtained by convolving a model of bomb-generated geomagnetic disturbance with estimates of the Earth-surface impedance tensor field. In
developing such maps, numerical methods pioneered in the 1980s by the Oak Ridge National Laboratory (ORNL) have been widely influential (Barnes, Rizy, et al., 1993; Legro et al., 1985, 1986). ORNL's methods have been used in numerous E3 scenario mapping projects across the conterminous United States (CONUS) (e.g., Electric Power Research Institute, 2017; Electromagnetic Pulse Commission, 2017; Gilbert et al., 2010; International Electrotechnical Commission, 1996; Lee et al., 2019; Rackliffe et al., 1988; Tesche et al., 1991), and these maps have been used in projects for assessing the vulnerability of power grids to the E3 hazard (e.g., Barnes, Tesche, et al., 1993; Electric Power Research Institute, 2019; Electromagnetic Pulse Commission, 2017; Tesche et al., 1991). The models of E3 geomagnetic disturbance resemble magnetometer measurements made during the high-altitude nuclear tests of the 1960s, and the maps of E3 geoelectric fields are visually compelling, but the surface impedances used to develop those maps commonly assume uniform half-space models of Earth conductivity.

Recognizing that the Earth's conductivity structure is complicated and, generally, three-dimensional across a wide range of spatial scales, and recognizing that this structure distorts the amplitude, polarization, and variational phase of induced geoelectric fields (e.g., Bedrosian & Love, 2015; McKay & Whaler, 2006), we investigate the effects of geologically realistic Earth-surface impedance on the time-dependent mapping of E3 geoelectric fields. Our methods are broadly similar to those used for analyzing geoelectric fields induced in the solid Earth during magnetic storms (e.g., Blake et al., 2016; Kelbert et al., 2017; Love et al., 2019; Lucas et al., 2020; Marshalko et al., 2020; Marshall et al., 2020; Simpson & Bahr, 2020; Torta et al., 2017; Wang et al., 2020). Indeed, comparisons are sometimes drawn between magnetic-storm variation and E3 electromagnetic variation. But important quantitative differences affect hazard and vulnerability assessments and possible mitigation measures (e.g., Meliopoulos et al., 1994; Neal et al., 2011; Rivera et al., 2016). In particular, E3 electromagnetic variation has a different geographic expression than that of magnetic-storm disturbance, and electromagnetic amplitudes are concentrated at higher frequencies than are characteristic of magnetic storms. In developing results that are specific for E3, we use a parameterization of a scenario EMP geomagnetic disturbance similar to that used by ORNL (and other investigators), but instead of using idealized models of surface impedance, we use impedance tensors derived from wideband magnetotelluric measurements acquired during a recent survey of the eastern-midcontinental United States. To our knowledge, our E3 analysis is the first in which realistic impedances are used. We convolve the scenario E3 geomagnetic disturbance with the magnetotelluric impedance tensors to construct scenario E3 geoelectric waveforms across the survey region. These are then interpolated onto power-grid lines and integrated to estimate geovoltages as a function of time. Our results can be used to quantify the errors associated with estimates of the E3 geoelectric hazard obtained using idealized half-space impedances.

2. Idealized Half-Space Impedance

It is already understood, from analyses of magnetic storms, that induced geoelectric fields can be significantly distorted by heterogeneous solid-Earth conductivity structure. But since many E3 scenario analyses assume half-space models of Earth conductivity, we find it useful to briefly review induction in a simple half-space model. More complicated idealized models can also be considered, such as quarter-space models (e.g., Simpson & Bahr, 2005, their Chapter 2.6; Berdichevsky & Dmitriev, 2008, their Chapter 6), though such models are not specifically needed for our analysis that follows, and, as far as we know, they have not been considered in EMP simulations. For a right-hand geographic coordinate system (north \( \hat{x} \), east \( \hat{y} \), down \( \hat{z} \)), we consider a boundary-value problem with the atmosphere (\( z < 0 \)) treated as a vacuum and the Earth's interior (\( z > 0 \)) treated as a uniform electrical conductor \( \sigma \) (resistivity \( \rho = 1 \sigma \)); the interface (\( z = 0 \)) represents the Earth's surface. In this setting, electric and magnetic \( \{\mathbf{E}, \mathbf{B}\}\) fields are assumed to vary in time \( \tau \); they can also, with Fourier transformation, be represented in the frequency domain \( \{\mathbf{E}, \mathbf{B}\}(f) \). The quasi-static approximation applies, \( 2\pi\epsilon_0 f \ll \sigma \), where \( \epsilon_0 \) is the permittivity of free space.

An incident electromagnetic plane wave \( \{\mathbf{E}', \mathbf{B}'\} \) is assumed to propagate straight down from above onto the half-space Earth model. At the surface and in the conducting half-space, the relationship between the transmitted electric \( \mathbf{E}' \) and magnetic \( \mathbf{B}' \) fields is given, in the frequency domain, by

\[
\mathbf{E}'(f, z) = \frac{1}{\mu_0} \mathbf{Z}_{HS} \cdot \mathbf{B}'(f, z),
\]

(1)
where the impedance tensor is skew-symmetric,
\[
Z_{HS}(f|\sigma) = \begin{bmatrix}
0 & Z_{HS} \\
-Z_{HS} & 0
\end{bmatrix}, \quad \text{where } Z_{HS}(f|\sigma) = (1 + j)\sqrt{\frac{\mu_0 \sigma}{\epsilon}}
\] (2)
(Landau et al., 1984, Chapters 59 and 87; Panofsky & Phillips, 1962, their Chapter 11–6), where \(\mu_0\) is the permeability of free space, \(\epsilon\) denotes multiplication, and where \(j = \sqrt{-1}\). Impedance \(Z\) has units of \(\Omega\) (ohms); the transfer quantity \(Z / \mu_0\) has units of mV/km/nT. From this, we note that in the half-space, the transmitted electric and magnetic fields are, for a given variational frequency, orthogonal, \(\mathbf{E}'(f) \cdot \mathbf{B}'(f) = 0\). With inverse Fourier transformation of Equation 1, induction can be described in the time domain as
\[
\mathbf{E}'(t) = \frac{1}{\mu_0} \left( Z_{HS} \ast \mathbf{B}' \right)(t),
\] (3)
where, in this equation, \(Z_{HS}(t|\sigma)\) is the half-space impulse response function, and \(\ast\) denotes convolution. For a particular frequency, time variation of the electric field leads that of the magnetic field by the phase \(\phi = \tan^{-1}\left(\mathbf{E}' / \mathbf{B}'\right) = 45^\circ\).

Let us, now, consider two very different, but representative, Earth conductivities for the half-space model: for a geological setting of sedimentary rock wet with briny water, \(\sigma = 1\) S/m, and for a geological setting of dry igneous or dry metamorphic rock, \(\sigma = 10^{-4}\) S/m (e.g., Palacky, 1988; Telford et al., 2009). For these conductivities and for variation at 1 s (approximately representative of E3), \(Z_{HS}\) ranges from \(3 \times 10^{-3}\) to \(3 \times 10^{-1}\) \(\Omega\). In comparison to the impedance of free space, \(Z_0 = \epsilon \mu_0 = 377\) \(\Omega\), where \(\epsilon\) is the speed of light, the solid Earth is a low-impedance medium, \(|Z_{HS}| \ll Z_0\). At the surface of the conducting half-space, the amplitude of the transmitted electric field, relative to that of the incident wave, is given by
\[
E' = T_E(f|\sigma) \cdot E_i, \quad \text{where } T_E(f|\sigma) = \frac{2Z_{HS}}{Z_0 + Z_{HS}}
\] (4)
(e.g., Stratton, 1941, their Chapter 9.10, Equation 10). For rock conductivity ranging from 1 to \(10^{-4}\) S/m, and for variation at 1 s, the transmission coefficient \(T_E\) ranges from \(1.6 \times 10^{-3}\) to \(1.6 \times 10^{-2}\). These tiny transmission coefficients mean that only a small fraction of an E3 wave is transmitted into the Earth—most of it is reflected. That the transmission coefficients have a wide range (two orders of magnitude) means that, for a given incident E3 wave, the amplitude of the transmitted electric field is very much dependent on local conductivity (and, therefore, dependent on local geology). On the other hand, the amplitude of the transmitted magnetic field is
\[
B' = T_B(f|\sigma) \cdot B_i, \quad \text{where } T_B(f|\sigma) = \frac{2Z_0}{Z_0 + Z_{HS}}.
\] (5)
Noting, again, that \(|Z_{HS}| \ll Z_0\), we obtain Lenz's law,
\[
B' = 2 \cdot B^i,
\] (6)
which, we emphasize, for the variational frequency range of interest here, is essentially independent of local conductivity (and, therefore, independent of local geology).

For an electromagnetic wave directed onto the half-space, with an incident angle \(\theta^i\) of propagation measured relative to vertical, so that \(\hat{z} \cdot \mathbf{E}' = E' \cos \theta^i\) and \(\hat{z} \cdot \mathbf{B}' = B' \cos \theta^i\), the refracted angle \(\theta'\) is given by Snell’s law,
\[
\sin \theta' = \frac{1}{n} \sin \theta^i, \quad \text{where } n = \frac{Z_0}{Z_{HS}}
\] (7)
(Stratton, 1941, their Chapter 9.8; Yuferev & Ida, 2010), where \(n\) is the index of refraction of the conducting medium, and where we have assumed that air has a unit index of refraction. Since \(|Z_{HS}| \ll Z_0, \sin \theta' \approx 0\), this means that, independent of incident angle, \(\theta^i < 90^\circ\), the direction of transmission is very close to downward, and the transmitted electromagnetic wave is close to horizontally polarized, \(\{\mathbf{E}', \mathbf{B}'\} \approx \{\mathbf{E}_h, \mathbf{B}_h\}\). Therefore, to a good approximation, Equation 1 applies for electromagnetic waves that are obliquely incident onto the half-space.
3. Magnetotelluric Impedance

Magnetotelluric impedance tensors \( \mathbf{Z}(f, x, y) \) are derived (e.g., Egbert, 2007b) from time-series measurements made simultaneously, at a given survey site \((x, y)\), of natural geomagnetic variation \( \mathbf{B}_g(t, x, y) \) and geoelectric field variation \( \mathbf{E}_g(t, x, y) \) (e.g., Ferguson, 2012; Simpson & Bahr, 2005), such that

\[
\mathbf{E}_g(f, x, y) = \frac{1}{\mu_0} \mathbf{Z}(f, x, y) \cdot \mathbf{B}_g(f, x, y).
\] (8)

In contrast to Equation 2, the elements of a magnetotelluric impedance tensor are fully populated,

\[
\mathbf{Z}(f, x, y|\sigma(r)) = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{bmatrix} (f, x, y|\sigma(r)).
\] (9)

Each magnetotelluric impedance tensor is a nonlinear function of the electrical conductivity \( \sigma(r) \) beneath and surrounding the survey site, where \( r \) is the position vector. These tensors are usually used in inversions for solid-Earth conductivity structure (e.g., Egbert, 2007a; Rodi & Mackie, 2012), but we use them, here, to estimate E3 geoelectric fields. Because subsurface conductivity can be complicated in structure, induced geoelectric fields do not, in general, have a straightforward relationship with geomagnetic field variation; in contrast to the idealized half-space, generally, \( \mathbf{E}_g(f) \cdot \mathbf{B}_g(f) \neq 0 \) and \( \phi \neq 45^\circ \).

4. Geographic Scales

It is important to consider geographic scales in an analysis of E3 geoelectric fields. Fundamentally, for an idealized half-space Earth model, electromagnetic amplitudes are attenuated across a diffusive length scale, a skin-depth, given by

\[
\delta(f|\sigma) = \frac{1}{\sqrt{\pi \mu_0 \sigma f}}.
\] (10)

For electromagnetic field variation realized across the Earth, at a particular site over conductivity structure that is (possibly) heterogeneous, \( \sigma \) in Equation 10 is a bulk-average of the conductivity within a roughly hemispherical volume beneath and surrounding the site with radius \( \delta \). For either half-space or bulk-average conductivities ranging from \( 1 \) S/m for sedimentary rock to \( 10^{-4} \) S/m for metamorphic or igneous rock, and for E3 variation at \( 1 \) s (or \( 1,000 \) s), \( \delta \) ranges from \( \sim 0.5 \) to 50 km (or from \( \sim 16 \) to 1,600 km).

Consider, now, the empirical parameterization given by the magnetotelluric Equation 8. This is often a reasonable description of the relationship between \( \mathbf{B}_g(f) \) and \( \mathbf{E}_g(f) \)—provided geomagnetic field variation across the Earth’s surface, for a given variational frequency, has a characteristic lateral length scale \( L(f) \) that is longer than the corresponding diffusive scale,

\[
\delta(f) < L(f).
\] (11)

This is called the plane-wave condition (Wait, 1982, their Chapter VI). Depending on the diffusive length scale, the plane-wave condition can be violated if the source currents generating geomagnetic field variation are both at relatively low altitude and spatially complicated, such as can sometimes be realized with disturbance current systems in the ionosphere (e.g., Jones & Spratt, 2002; Pirjola, 1992). In such circumstances, the patterns of induced currents in the solid Earth can be qualitatively different, the parameters in Equation 9 can be quantitatively different, each from those for a plane wave.

In terms of the lateral length scale for an E3 EMP waveform, maximum E3A disturbance is realized at the Earth’s surface for explosions set at altitudes of \( \sim 500 \) km, and maximum E3B disturbance is realized at the Earth’s surface for explosions set at altitudes of \( \sim 150 \) km (e.g., Rivera et al., 2016, their Figure 23). Therefore, we can plausibly assume that E3 disturbance is generated by electric currents at heights greater than 150 km, and even quite a bit higher. By geometric attenuation, \( L \) characterizing the ground-level expression of geomagnetic variation will exceed 150 km and possibly, even, exceed the altitude of the explosion. As discussed in Section 6, we choose an E3 parameterization for which \( L \approx 800 \) km. With this, for variation at \( 1 \) s and with \( \delta \) ranging from \( \sim 0.5 \) to 50 km, the plane wave condition (11) is well satisfied. On the other hand, for variation at \( 1,000 \) s and \( \delta \) ranging from \( \sim 16 \) to 1,600 km, the plane wave condition (11) is not necessarily...
well satisfied. This means that the empirical parameterization given by Equation 8 likely works well for the high-frequency parts of E3, but it is less accurate for the lowest-frequency parts of E3 in resistive geological settings.

Next, it is important to remain mindful of the effects that geological structure (e.g., Pollard & Fletcher, 2005; Tarbuck et al., 2017) can have on the length scale of induced E3 geoelectric fields. In some geological settings, such as in basins filled by a thick accumulation of sediments, structure can be relatively one-dimensional, depth-dependent. In other settings, tectonic forces have shaped terranes by faulting, deformation, and intrusion, resulting in structure that is fundamentally two- or three-dimensional. In such settings, subsurface conductivity structure and surface impedance can differ significantly across a wide range of geographic scales (e.g., Bahr, 2005; Lovejoy & Schertzer, 2007). Although the shortest of these scales is limited by diffusion, Equation 10, site-to-site differences in impedance strongly affect E3 local geoelectric field, Equation 4. On the other hand, the geomagnetic field is only weakly affected by surface impedance, Equation 6. Therefore, given the possibly shortish length scale of surface impedance, a reasonable method for calculating an E3 geoelectric field is to convolve a broad regional model of E3 geomagnetic field variation with local magnetotelluric estimates of surface impedance as per Equation 8.

5. Eastern-Midcontinental Tensors

We use 127 wideband magnetotelluric tensors derived from measurements acquired by the U.S. Geological Survey from 2016 to 2019 during a survey of a part of the eastern-midcontinental United States (35.0° – 38.5°N × 88.0° – 93.0°W). The study area encompasses the cities of St. Louis, Missouri (MO) and Memphis, Tennessee (TN); it includes parts of Illinois (IL), Kentucky (KY), and Arkansas (AR) (Bedrosian et al., 2020). In Figure 1, we show a map of the study region and magnetotelluric survey sites. The Eastern-Midcontinent is underlain by Precambrian basement rock that is, relatively electrically resistive; this is overlain by differing depths of younger and relatively electrically conductive sedimentary rock, the thickness of which is the difference between surface elevation (Danielson & Gesch, 2011) and the Great Unconformity elevation (Marshak et al., 2017). Notable, for our purposes, is the Ozark Dome (e.g., Anderson et al., 1979), where the sedimentary cover is relatively thin and basement granitic and dolomite rocks are exposed in some places. In contrast, the Reelfoot Rift (e.g., Dart & Swolfs, 1998) and the Illinois Basin (e.g., Kolata et al., 2005) are both deeply filled with sedimentary rock. Survey site spacing ranges from about 6 to 70 km. Each tensor $Z_{mn}(f, x, y)$ for each site $(x, y)$ is parameterized for a discrete set of frequencies, $f_1, f_2, f_3, \ldots$ within the band from $10^0$ to $10^{-5}$ Hz (10^{-3} to 10^3 s). Localized geologic complexities (e.g., DeLucia et al., 2019; Van Arsdale & Cox, 2007; Van Schmus et al., 1993) are manifest in the properties of the impedance tensors.

It is useful to plot the frequency dependence of “apparent” resistivity and phase, given, respectively, by

$$\rho_{mn}(f) = \frac{|Z_{mn}(f)|^2}{2\pi\sigma_f}$$

and

$$\phi_{mn}(f) = \tan^{-1}\left[\frac{\text{Im}(Z_{mn}(f))}{\text{Re}(Z_{mn}(f))}\right]$$
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Figure 2. Apparent resistivities and phases, \( \rho_{xx}(T), \phi_{xx}(T) \) (yellow), \( \rho_{xy}(T), \phi_{xy}(T) \) (blue), \( \rho_{yx}(T), \phi_{yx}(T) \) (red), \( \rho_{yy}(T), \phi_{yy}(T) \) (green), given by Equations 12 and 13, each as a function of variational period \( T \), for survey sites (a), (b) SFM06 and (c), (d) RF111. Note that the vertical ranges in (a) and (c) are different.

(e.g., Berdichevsky & Dmitriev, 2008, their Equation 1.17), where the subscript \( mn \) denotes the various Cartesian tensor components, \( xx, xy, yx, yy \). In Figure 2, we plot these quantities for two different survey sites: SFM06 near Nashville, southern Illinois (38.29°N, 89.41°W), and RF111 near Success, northeast Arkansas (36.44°N, 90.74°W). Site SFM06 is located on and surrounded by sedimentary rocks that are part of the Illinois Basin. From Figures 2a and 2b, we note that the apparent resistivities and the phases of the primary elements are approximately equal, \( \rho_{xx} \approx \rho_{yy} \) and \( \phi_{xx} \approx \phi_{yy} \pm 180^\circ \), across a wide range of variational periods (frequencies). In other words, impedance at this site is approximately skew-symmetric, but the off-diagonal apparent resistivities are not constant across variational periods. These observations are consistent with conductivity structure beneath and surrounding the survey site that is close to one-dimensional, depth-dependent. At 1 s, apparent resistivity is \( \sim 5.5 \, \Omega \cdot m \), and, from Equation 10, we estimate that induction occurs across a diffusive depth of \( \sim 1.2 \) km. In contrast, RF111 is located atop the Ozark Dome and adjacent to the thick sedimentary fill of the Reelfoot Rift. From Figures 2c and 2d, we note that, for variational periods shorter than 10 s, \( \rho_{xy} \ll \rho_{xx}, \phi_{xy} \approx \phi_{xx} \pm 180^\circ \), but between 100 and 1,000 s, \( \rho_{yy} \gg \rho_{xx} \) and \( \phi_{yy} \neq \phi_{xx} \pm 180^\circ \). Furthermore, the non-primary element resistivity \( \rho_{yy} \) is comparable to the primary resistivities \( \rho_{xx} \) and \( \rho_{xy} \) between 0.1 and 1,000 s. Impedance at this site is not skew-symmetric. These observations are consistent with conductivity structure beneath and surrounding the survey site that is far from simply one-dimensional, depth-dependent. At 1 s, apparent resistivity is \( \sim 1.5 \times 10^4 \, \Omega \cdot m \); from this, we estimate that induction occurs across a diffusive scale of \( \sim 50 \) km.

Another way to examine impedance tensors is to plot the transfer-function amplitude,

\[
\frac{1}{\mu_0} Z(f) = \frac{|E_h(f)|}{|B_h(f)|} = \frac{E_h(f)}{B_h(f)}.
\] (14)
as a function of the azimuthal polarization \( D^E \) of the induced horizontal-component geoelectric vector, \( Z(D^E(f)) / \mu_0 \) (e.g., Berdichevsky & Dmitriev, 2008, their Equation 1.91), and as a function of the azimuthal polarization \( D^B \) of the horizontal-component geomagnetic vector, \( Z(D^B(f)) / \mu_0 \) (e.g., Berdichevsky & Dmitriev, 2008, their Equation 1.89). When plotted as a function of geoelectric polarization, transfer-function amplitude is elliptical on a polar plot. For a given tensor and at a specific variational period, the length of the major (minor) axis of the ellipse corresponds to the maximum (minimum) principal amplitude of the geoelectric field per unit-amplitude geomagnetic field. In contrast, when plotted as a function of geomagnetic polarization, transfer-function amplitude is roughly peanut-shaped on a polar plot. Furthermore, an impedance phase tensor is elliptical when plotted in polar coordinates as a function of geographic direction, \( \phi(D(f)) \) (e.g., Booker, 2014; Caldwell et al., 2004). Here, the major (minor) axis of a phase ellipse corresponds to the maximum (minimum) principal phase.

As examples, in Figures 3a and 3b, for site SFM06, we plot the transfer-function amplitudes, \( Z(D^E) / \mu_0 \) and \( Z(D^B) / \mu_0 \) for five different variational periods (0.1, 1.0, 10.0, 100.0, and 1,000.0 s). Amplitudes for this site are relatively circular across azimuth and for a range of variational periods. In Figure 3c, we plot phase ellipses \( \phi(D) \) for site SFM06; for variation at 0.1 and 1,000.0 s, the phase ellipses are also close to circular. As we noted from Figures 2a and 2b, these observations are consistent with conductivity structure beneath and surrounding the survey site that is close to one-dimensional, depth-dependent. On the other hand, in Figure 3d, for site RF111, \( Z(D^E) / \mu_0 \) is highly elliptical, with 1-s variation having a maximum (minimum) amplitude of 387 mV/km/nT (61 mV/km/nT) and polarization along the 119° ± 180° azimuth, or approximately east-southeast (west-northwest). In Figure 3e, \( Z(D^B) / \mu_0 \) is peanut-shaped, with a narrow waist and, with 1-s variation, polarization along the 37° ± 180° azimuth. For this tensor, for a given variational period, the maximum polarization \( Z(D^B) / \mu_0 \) is nearly orthogonal to the maximum polarization of \( Z(D^B) / \mu_0 \). In Figure 3f, for 1 s variation, the phase \( \phi(D) \) has maximum (minimum) of 58° (35°), with maximum along the 126° ± 180° azimuth. As we noted from Figures 2c and 2d, these observations are consistent with conductivity structure beneath and surrounding the survey site that is far from one-dimensional, depth-dependent.

In Figures 4a and 4b, we show a map of transfer-function amplitude ellipses/peanuts for the various magnetotelluric survey sites for 1 s variation. These maps demonstrate the relationship between surface impedance and prominent geologic and tectonic structures. Over and surrounding the resistive Ozark Dome in southeast Missouri, transfer functions generally have high amplitude, and some are polarized. We note, in particular, that the transfer function for RF111 has one of the highest maximum amplitudes and is one of the most polarized of all the sites in the study region. In contrast, over the conductive Illinois Basin and Reelfoot Rift, transfer functions, such as that for SFM06, have relatively low amplitude and are generally less polarized. In Figure 4c, we show a map of transfer-function phase ellipses for the various survey sites for 1 s variation; here, we see significant differences between the Ozark Dome and the Illinois Basin and Reelfoot Rift. Judging from Figure 4, we can anticipate that the geography of surface impedance will impart significant spatiotemporal distortion to E3-induced geoelectric field variation.

6. Spatiotemporal Model of Disturbance

Recognizing that nuclear EMP is an extremely complicated phenomenon, and accepting the fact that many numerical EMP models are classified (e.g., Department of Energy, 2017), we appreciate that most publicly published studies of E3 EMP and its effects rely on simplified parameterizations of EMP electromagnetic variation at the Earth’s surface. While these parameterizations are physically motivated, they do not represent all of the intricacies and possibilities of an EMP event. They assume that the explosion height (several hundred kilometers) and yield (several hundred kilotons to a few megatons) are such that the E3 signal is of nearly maximum amplitude (e.g., Rivera et al., 2016, their Figure 23), but the assumed explosion altitude and yield are not usually specified. The parameterizations are generic, but they serve as useful benchmarks for comparison. In this regard, ORNL reports provide a parameterization of E3 electromagnetic variation that has had widespread influence. Here, the spatiotemporal form of horizontal-component, ground-level geoelectric field variation is represented by the multiplication of separate geographic and temporal functions for each of the blast A and heave B phases.
The time-invariant vector functions $\mathbf{e}^{A, B}(x, y)$ describe the geographic dependence of the amplitude and polarization of the induced E3 geoelectric field, and the geographically invariant functions $E^{A, B}(t)$ describe the time dependence of E3 geoelectric field variation. Some investigators (e.g., Barnes, Rizy, et al., 1993; Electric Power Research Institute, 2017) choose a normalization in which the $\mathbf{e}(x, y)$ functions have absolute amplitude (with units), and the $E(t)$ functions serve as basis...
functions describing relative variation in time. Other investigators (e.g., Gilbert et al., 2010) choose a normalization in which the \( E(t) \) functions have absolute amplitude, and the \( e(x, y) \) functions serve as basis functions describing relative amplitude in geography.

As discussed following Equations 4 and 6 and in Section 4, geoelectric field variation is a strong function of Earth conductivity, while geomagnetic field variation is not. Therefore, to estimate E3 geoelectric induction across heterogeneous Earth structure, we use a model of geomagnetic field variation and magnetotelluric impedance tensors to estimate E3 geoelectric field variation. Instead of Equation 15, we use the following (very similar) parameterization of horizontal-component, ground-level geomagnetic field variation

\[
B_h(t, x, y) = b^A(x, y) \cdot B^A(t) + b^B(x, y) \cdot B^B(t),
\]

which can be motivated on physical grounds similar to those for Equation 15. We use a normalization in which the \( B(t) \) functions have absolute amplitude, and the \( b(x, y) \) functions serve as basic functions describing the geographic dependence of relative amplitude.

In considering the idealized constitutive functions in Equation 16, let us begin with the function \( b^A(x, y) \) describing the geography of the E3 disturbance geomagnetic field generated by the blast-phase plasma bubble. It expands very rapidly, and in a matter of seconds, it attains a lateral dimension of thousands of kilometers. Electric currents \( J \) on the underside surface of this bubble, but in the atmosphere, are assumed to be horizontal, uniform, and eastward directed (Barnes et al., 1993, their Figure 3b; Gilbert et al., 2010, their Figures 2–4). By Ampère's law, \( \mu_0 J = \nabla \times B \), these currents generate a horizontal, uniform, and northward-directed disturbance geomagnetic field (e.g., Gilbert et al., 2010, their Figures 2–3), so that

\[
b^A(x, y) = b^A \hat{x}.
\]

Given our chosen normalization, we take \( b^A = 1 \); this means that \( B^A(t) \) is a positive function. While this geomagnetic field component increases in intensity, a westward geoelectric field is induced in a half-space Earth model, Equation 15, and, so, \( E^A(t) \) is a negative function. Technically, \( b^A \) should be parallel to the local horizontal component of the Earth’s main geomagnetic field, which has a non-zero declination almost everywhere over the Earth’s surface. However, over most of CONUS, geomagnetic declination is relatively small, and the zero declination line passes through the middle of the eastern-midcontinental study region. We note, furthermore, that other investigations of E3 (e.g., Barnes, Rizy, et al., 1993; Gilbert et al., 2010) do not make a correction for local geomagnetic declination.

We also need to prescribe the function \( b^B(x, y) \) describing the geography of the E3 disturbance geomagnetic field generated by the heave phase. For a buoyant plasma bubble rising upward with velocity \( u \) through the ambient geomagnetic field, a westward dynamo electric field \( \nabla \times u \times B \) is generated. This drives a westward current that, outside of the dynamo, closes through two oppositely circulating current gyres (Barnes et al., 1993, their Figure 3c; Gilbert et al., 2010, their Figures 2–7), clockwise to the north and counterclockwise to the south of the rising plasma bubble. It is conventional, here, to work in traditional spherical coordinates (radial up

Figure 4. Transfer-function amplitude \( Z / \mu_0 \) for variation at 1.0 s as a function of (a) The induced geoelectric polarization \( D^E \), and as a function of (b) Geomagnetic polarization \( D^B \), and (c) Phase \( \phi \) for variation at 1.0 s as a function of geographic direction \( D \).
\( \hat{r} \), colatitude \( \hat{\delta} \), longitude \( \hat{\lambda} \), specified by the Earth’s rotational axis. The basis in the heaving region at \((\hat{\vartheta}, \hat{\lambda})\) and at height \(r_e\) is given by

\[
\mathbf{J}^{(\hat{\vartheta}, \hat{\lambda}|_{r_e}, \hat{\vartheta}, \hat{\lambda}, \Delta)} = \mathbf{J}(\hat{\vartheta} + \Delta / 2, \hat{\lambda}) - \mathbf{J}(\hat{\vartheta} - \Delta / 2, \hat{\lambda}),
\]

where the gyres are spherical elementary currents,

\[
\mathbf{J}(\hat{\vartheta}, \hat{\lambda}|_{r_e}, \hat{\vartheta}, \hat{\lambda}) = \frac{I_0}{4\pi r_e} \cos \left( \frac{1}{2} \mathbf{J}(\hat{\vartheta}, \hat{\lambda}, \hat{\vartheta}, \hat{\lambda}) \right) \hat{\lambda},
\]

(Amm & Viljanen, 1999; Rigler et al., 2019; Vanhamäki & Juusola, 2020), defined in spherical coordinates \((\hat{\vartheta}, \hat{\lambda})\) specified by the focal point of each gyre, and where \(I_0\) is a scaling factor. The halfway point between the northern and southern focal points of the two gyres is \((\hat{\vartheta}, \hat{\lambda})\); this is, itself, the geographic epicenter of the heaving plasma. The colatitudinal separation of the gyres is \(\Delta\), and

\[
\cos \vartheta = \cos \hat{\vartheta} \cos \hat{\lambda} + \sin \hat{\vartheta} \sin \hat{\lambda} \cos (\hat{\vartheta} - \hat{\lambda}).
\]

The ground-level horizontal-component geomagnetic field generated by each current gyre is poloidal and can be derived using the Biot-Savart law,

\[
B_\vartheta(r, \hat{\vartheta}, \hat{\lambda}|_{r_e}, \hat{\vartheta}, \hat{\lambda}) = -\frac{\mu_0 I_0}{4\pi} \frac{1}{r \sin \hat{\vartheta}} \left( \frac{r - r_e \cos \hat{\vartheta}}{\sqrt{r^2 - 2rr_e \cos \hat{\vartheta} + r_e^2}} - 1 \right),
\]

(Amm & Viljanen, 1999, their Equation A.8), and, therefore,

\[
\mathbf{b}_\vartheta(x, y) = \mathbf{b}(r, \hat{\vartheta}, \hat{\lambda}|_{r_e}, \hat{\vartheta}, \hat{\lambda}, \Delta) = B_\vartheta(r_e, \hat{\vartheta} + \Delta / 2, \hat{\lambda}) \hat{\vartheta} - B_\vartheta(r_e, \hat{\vartheta} - \Delta / 2, \hat{\lambda}) \hat{\vartheta}.
\]

Given our chosen normalization, we take \(\mathbf{b}(r, \hat{\vartheta}, \hat{\lambda}) = 1\). Since the geomagnetic field beneath this current system and between the two gyre focal points is southward directed, \(\mathbf{b}(r, \hat{\vartheta}, \hat{\lambda})\) is a negative function. While this geomagnetic field component increases in intensity between the focal points, an eastward geoelectric field is induced in a half-space Earth model, Equation 18, and, so, \(E(t)\) is a positive function.

In Figures 5a and 5b, we show the two ground-level E3 geographic basis functions for the blast phase \(\mathbf{b}_\vartheta(x, y)\) and for the heave phase \(\mathbf{b}_\vartheta(x, y)\) centered on our chosen explosion epicenter \((36.75^\circ N, 90.50^\circ W\), which is also shown in the center of Figure 1\). We choose \(\Delta = 800\) km so that the pattern of E3B disturbance resembles Barnes, Rizy, et al. (1993, their Figure 3b) and Gilbert et al. (2010, their Figures 2–10). With this, and in consideration of the discussion in Section 4, we recognize that the lateral scale of these geographic functions satisfy the plane-wave condition given by Equation 11 for all but the longest spatial scales at the lowest E3 frequencies (longest E3 periods). In Figure 5, we also show as a box the study region where we map E3 geoelectric fields. As idealized functions, the forms of \(\mathbf{b}_\vartheta(x, y)\) and \(\mathbf{b}_\vartheta(x, y)\) far outside this box are immaterial to our analysis—over spatial scales broader than those considered here, realistic treatment of E3 fields should include an attenuation term so that E3 amplitudes properly diminish with increasing distance from the explosion epicenter.

Next, we consider the waveform functions in Equations 15 and 16. The International Electrotechnical Commission (1996, IEC) has published reference waveforms of the horizontal-component EMP geoelectric field, including for E3. The IEC E3 standard geoelectric waveform is a scalar time series describing ground-level variation for a uniform half-space conductivity of \(\sigma = 10^{-4}\) S/m. The applicable latitudes are given in the IEC document as between 30\(^\circ\) and 60\(^\circ\); no specification is given of explosion parameters. The exact physical foundation of the geoelectric waveforms is not described in the IEC document, but mention is made of reports from high-altitude nuclear tests conducted over the South Pacific during the early 1960s.
In more detail, the IEC E3 geoelectric waveform is

\[
E(t) = E^A(t) + E^B(t),
\]

(International Electrotechnical Commission, 1996), which, presumably, includes the Starfish Prime test summarized in our introduction.

The eight constants \(\{E_0^{A,B}, k^{A,B}, a^{A,B}, b^{A,B}\}\) are given in the report of the International Electrotechnical Commission (1996, their Chapter 5.3.3). But, in light of the discussion following Equation 4 and in Section 4, we recognize that a geoelectric waveform depends on surface impedance. Since the IEC geoelectric waveform \(E(t)\) pertains to a uniform half-space impedance for a particular conductivity, it cannot, in general, be applied in regions of different conductivities, let alone any region of complicated geology. We also understand that direct measurements of geoelectric fields made during EMP tests in particular geological settings (above particular subsurface conductivity structures) (e.g., Bomke et al., 1964; Burch & Green, 1963; Gill, 1962; Poletti & Gadsden, 1963) cannot, in general, be treated as EMP geoelectric fields that might be realized in other geological settings.
On the other hand, and in light of the discussion following Equation 6 and in Section 4, we understand that a geomagnetic waveform \( B(t) \) is insensitive to geological setting. Although an E3 geomagnetic waveform is not given in the IEC document, for the assumed uniform half-space conductivity, we can recover it using Equation 1 and taking an inverse Fourier transform to obtain the convolution

\[
B(t) = -\frac{\mu_0 \sigma}{\pi} \int_0^1 \frac{1}{\sqrt{t - \xi}} E(\xi) d\xi,
\]

(25)

(see also, Gilbert et al., 2010); the minus sign in this equation assumes a westward (eastward) oriented electric field, which is induced by a northward (southward) oriented geomagnetic field of increasing amplitude. Performing the integration, we obtain

\[
B(t) = B^A(t) + B^B(t),
\]

(26)

where

\[
B^{A,B}(t) = 2 \sqrt{\frac{\mu_0 \sigma}{\pi}} \frac{L_0}{k} \left( F_{A,B} \left( \sqrt{\frac{A}{B}} \right) - F_{B} \left( \sqrt{\frac{A}{B}} \right) \right) \quad \text{for} \; t \geq 0,
\]

(27)

and where \( F(x) \) is Dawson’s integral (e.g., Oldham et al., 2009, their Chapter 42; Press et al., 1992, their Chapter 6.10).

**Figure 7.** Comparison of International Electrotechnical Commission ground-level E3 (a) Geomagnetic \( B(t) \) waveform (black) and (b) Geoelectric \( E(t) \) waveform (black) together with the corresponding wideband-limited waveforms (purple), \( 10^{-3} - 10^{-5} \) Hz (\( 10^3 - 10^5 \) s) and long-period band-limited waveforms (brown), \( 10^{-1} - 10^{-3} \) Hz (\( 10^1 - 10^3 \) s). Note that, for clarity, time is shown on a logarithmic axis. The moment of the electromagnetic pulse explosion is at \( t = 1.0 \) s.
In Figures 6a and 6b, we show the IEC E3 geomagnetic $B(t)$ waveform, Equation 26, and the geoelectric $E(t)$ waveform, Equation 23, together with their individual blast E3A-phase and heave E3B-phase components. For clarity, time is given on a logarithmic axis, with the moment of the explosion at $t = 1.0$ s. Immediately after the explosion, the amplitudes of the geomagnetic and geoelectric E3 waveforms increase rapidly. This is followed by a more gradual decline and a small overshoot of the opposite sign. In detail, the E3 geomagnetic waveform has a rise time of 20.9 s from its start to its maximum of 1,458 nT and a full-width at half-maximum of 65.8 s. The induced geoelectric waveform has a rise time of 2.1 s from its start to its maximum and a full-width at half-maximum of 19.9 s; for the assumed uniform half-space conductivity of $\sigma = 10^{-4}$ S/m, the maximum value attained is 38.7 V/km. By $t = 1000.0$ s, the E3 signals have mostly faded. Qualitatively, the IEC geomagnetic waveform resembles direct geomagnetic recordings made during tests in the 1960s (Chavin et al., 1979; Dyal, 2006, their Figure 12; Electromagnetic Pulse Commission, 2017, their Chapter 3; Legro et al., 1985, their Figure 1); the IEC geoelectric waveform resembles those in other influential reports (Barnes et al., 1993, their Figure 3a; Department of Energy, 2021; Electromagnetic Pulse Commission, 2017, their Figures 15–20).
Before proceeding, recall that the magnetotelluric impedance tensors we use are wideband limited from $10^{-3}$ to $10^{-3}$ Hz ($10^{-3}$ to $10^{3}$ s). Recalling that the Fourier ingredients of a (non-periodic) impulse cover a broad range of frequencies (e.g., Bracewell, 2000), we need to establish that the magnetotelluric wideband is sufficient to resolve the E3 geoelectric impulse. In Figure 7a, we compare both the complete geomagnetic $B(t)$ IEC waveform and the wideband-limited counterpart, which generally follows the complete waveform. The offset in the wideband-limited waveform is due to lack of resolution at the low-frequency (long period) end of the signal spectrum, $< 10^{-3}$ Hz ($> 10^{3}$ s). This offset does not significantly affect the IEC (half-space) geoelectric field $E(t)$; as we see in Figure 7b, the wideband-limited waveform almost perfectly matches the complete waveform. From this, we understand that wideband magnetotelluric tensors, such as used here, are sufficient for mapping E3 geoelectric fields. For comparison, and in support of discussion in Section 10 concerning the possibility of using magnetotelluric impedance tensors for E3 mapping in other parts of CONUS, in Figure 7, we also show band-limited geomagnetic and geoelectric waveforms, $10^{-1}$–$10^{-4}$ Hz ($10^{4}$–$10^{5}$ s). In this case, the band-limited $E(t)$ IEC waveform does not match the complete waveform—significant Gibbs ringing is seen for the early blast phase from $t = 1.0$ to $10.0$ s; we note that the long-period band-limited $E(t)$ is acausal, commencing before the explosion, $t < 1.0$ s. From this, we understand that long-period magnetotelluric tensors are inadequate for detailed mapping of E3 geoelectric fields.

Figure 9. Snapshot in time of scenario ground-level E3 effects at $t = 1.5$ s (for electromagnetic pulse explosion at $t = 1.0$ s): (a) IEC geomagnetic waveforms; (b) Geomagnetic disturbance; (c) Geoelectric field for the half-space model with $\sigma_{HS} = 10^{-3}$ S/m; (d) Geoelectric field for magnetotelluric impedance tensors; (e) Geovoltage on power grid using the half-space model; (f) Geovoltage $V$ on power grid derived using magnetotelluric impedance tensors.
8. E3 EMP Scenario

We calculate E3 geoelectric fields for an EMP explosion with an epicenter (36.75°N, 90.50°W) centered on the eastern-midcontinental study region. Using the magnetotelluric Equation 8, magnetotelluric tensors $\mathbf{Z}(f, x, y)$ derived from the survey measurements, Section 5, and Fourier transformation of the model E3 geomagnetic field variation, $\mathbf{B}_h(f, x, y)$, given by Equation 16, we obtain the frequency domain expression of the geoelectric field $\mathbf{E}_h(f, x, y)$ at each survey site; inverse Fourier transformation gives $\mathbf{E}_h(t, x, y)$. These calculations are similar to those that we (e.g., Bedrosian & Love, 2015; Kelbert et al., 2017; Love et al., 2019) and others (e.g., Dimmock et al., 2019; Marshalko et al., 2020; Marshall et al., 2020; Wang et al., 2020) use (at lower frequencies) for analysis of magnetic-storm induction of geoelectric fields in the three-dimensional solid Earth. For comparison with previously published EMP results (e.g., Barnes, Rizy, et al., 1993; Electric Power Research Institute, 2017; Electromagnetic Pulse Commission, 2017; Gilbert et al., 2010; Legro et al., 1985; Siebert & Witt, 2019), we also calculate the E3 geoelectric fields that would be induced in a hypothetical Earth with a half-space impedance $Z_{HS}(f, \sigma)$, using Equation 1 and $\sigma_{HS} = 10^{-3}$ S/m (the referenced authors use a conductivity that is a factor of 10 higher than that used for the IEC geoelectric benchmark).

In the dynamic-content that the time-dependent movie-map given as Supporting Information S1 a scenario simulation is depicted of ground-level E3 electromagnetic field variation (and geovoltages on power-grid lines, discussed in Section 9) across the study region, covering 60 s of time after the moment of the explosion at $t = 1.0$ s, derived for both the half-space Earth model and for the magnetotelluric tensors. In Figure 8, we show a snapshot of this movie-map at the moment of the explosion, but before any E3 effects have been realized. In Figure 9, we show a snapshot from the movie-map at time $t = 1.5$ s, just half a second after the explosion. At this instance in time, the geomagnetic field, Figures 9a and 9b, is primarily $\mathbf{B}_h(t, x, y)$, close to
uniform northward, and its amplitude is \( \sim 170 \) nT. The geoelectric field for the half-space model, Figure 9c, is close to uniform westward, with an amplitude of 8.2 V/km. In contrast, the geoelectric field for the magnetotelluric tensors, Figure 9d, is far from uniform. Geoelectric amplitudes across the study region range from 0.8 V/km at one location to 29.0 V/km at another location; the median amplitude is 4.6 V/km, and the 68% interval is [1.2, 10.4] V/km.

In Figure 10, we show a snapshot at time \( t = 3.0 \) s, two seconds after the explosion. At this instance, the geomagnetic field, Figures 10a and 10b, is still primarily \( B^A(t, x, y) \), still close to uniform northward, but its amplitude has grown to \( \sim 609 \) nT. The geoelectric field for the half-space model, Figure 10c is still close to uniform westward, but its amplitude has grown to 12.2 V/km. The geoelectric field for the magnetotelluric tensors, Figure 10d, while still generally westward, has grown in amplitude, now ranging from 1.7 V/km at one location to 33.2 V/km at another location; the median amplitude is 6.2 V/km, and the 68% interval is [2.7, 12.3] V/km. In Figure 11, we show a snapshot at time \( t = 20.0 \) s. At this instance, the geomagnetic field amplitude is \( \sim 1,455 \) nT; deviation from northward polarization is due to an increased contribution from \( B^B(t, x, y) \); related deviation from westward polarization in the half-space geoelectric field is seen in Figure 11c. The decrease in the rate of change of the geomagnetic field corresponds to a decrease in geoelectric amplitudes for both the half-space model and for the magnetotelluric tensors, Figure 11d. The geoelectric field for the magnetotelluric tensors has lost much of its westward polarization.

Recall from Section 5 that the magnetotelluric impedance for site SFM06 is nearly skew-symmetric, something seen in geological settings that are close to one-dimensional, depth-dependent. As a result of the site's simple impedance, and given that the E3 geomagnetic field waveform is approximately northward, Figure 12a, the geoelectric waveform is approximately westward, Figure 12b. The geoelectric field attains its
maximum value of 2.2 V/km at \( t = 6.7 \) s. The geoelectric field for the half-space model is, not surprisingly, also westward, but its amplitude is much higher than the magnetotelluric amplitude since SFM06 impedance is significantly lower than that of the half-space. At this site, the half-space geoelectric field attains its maximum value of 12.2 V/km at \( t = 3.1 \) s, or 3.6 s before that for the SFM06 magnetotelluric tensor. The difference in amplitude could have been anticipated from Figure 2a and Equations 2 and 12, which indicate that \( Z_{HS} / Z > 1 \).

Very different from induction at site SFM06 is that at RF111. Recall from Section 5 that, as a result of complicated Earth structure, the magnetotelluric impedance for RF111 is far from skew-symmetric. From Figure 3d, we note that the geomagnetic polarization inducing the largest amplitude E3 geomagnetic field is along the \( 37^\circ \pm 180^\circ \) azimuth. But from Figure 13a, we see that the E3 geomagnetic field is approximately northward. Despite this, the E3 geomagnetic field is sufficient to induce high-amplitude E3 geoelectric fields. From Figure 13b, the geoelectric field attains its maximum amplitude of 37.3 V/km at \( t = 2.1 \) s. This exceeds the 15.0 V/km maximum value used by the U.S. Defense Threat Reduction Agency (Siebert & Witt, 2019). It exceeds the 24.0 V/km maximum value from an ORNL report (Barnes, Rizy, et al., 1993) that EPRI has used in an analysis of EMP effects on the U.S. power grid (e.g., Electric Power Research Institute, 2017). It is less than the 84.6 V/km maximum value given by the Electromagnetic Pulse Commission (2017). The geoelectric vector at RF111 is directed along the \( 110^\circ \pm 180^\circ \) azimuth, very close to the dominant polarization (at 1.0 Hz; 1.0 s) of the transfer function, along the \( 119^\circ \pm 180^\circ \) azimuth, seen in Figure 3c. In contrast, in Figure 13b, we see that the half-space model does not support west-northwest geoelectric polarization. At this site, the half-space geoelectric field attains its maximum value of 12.2 V/km at \( t = 3.1 \) s, or 1.0 s after that for the RF111 magnetotelluric tensor. The difference in amplitude could have been anticipated from Figure 2c and Equations 2 and 12, which indicate that \( Z_{HS} / Z < 1 \).
In comparing Figures 1 and 4 with the movie-map and with the snapshots, such as Figure 10d for \( t = 3.0 \) s, we note correlation between high (low) E3 geoelectric amplitudes and magnetotelluric sites with the high (low) impedance of the Ozark Dome (Illinois Basin). Furthermore, the orientation of the most polarized impedance tensors, seen as narrow ellipses in Figure 4a, roughly correspond to the polarizations of induced geoelectric vectors at those sites. These results are a clear demonstration that, should an EMP event actually occur over the eastern-midcontinental United States, localized differences in E3 geoelectric amplitude, polarization, and variational phase would result from geographically complicated surface impedance. Recalling from Section 3 that magnetotelluric impedance is, itself, a function of subsurface three-dimensional conductivity structure, we understand that simple half-space or one-dimensional, depth-dependent models of surface impedance will not, in general, provide accurate estimates of the E3 geoelectric hazard in complex geological settings.

9. Voltages on Power-Grid Lines

The integrated projection of a geoelectric field onto a power-grid line gives the geovoltage on that line,

\[
V = -\int_G E_A \cdot d l,
\]

where \( G \) is the grid path between grounding points (e.g., Molinski, 2002; Pirjola, 2007). Using transmission power-grid topological information from the Department of Homeland Security (DHS), our estimated E3 geoelectric fields, and numerical interpolation methods given by Lucas et al. (2018, 2020), we estimate time-dependent geovoltages on power-grid lines across the eastern-midcontinental study region. We show a map of the grid geovoltages in the movie-map of S1; we show snapshots of the grid geovoltages in Figures 8–11. In particular, in Figure 10e, we show half-space line geovoltages \( V_{hs} \) for \( t = 3.0 \) s—they are

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**Figure 13.** Scenario ground-level E3 (a) Geomagnetic waveform, \( B_x(t) \) (red) and \( B_y(t) \) (blue), and (b) Geoelectric waveform, \( E_x(t) \) (red) and \( E_y(t) \) (blue), for site RF111, Arkansas.
highest (lowest) for lines oriented east-west (north-south) but, otherwise, relatively uniform across the region. In contrast, in Figure 10f, we see that geovoltages derived using magnetotelluric tensors \( V_{MT} \) show significant geographically localized differences—notably, geovoltages are high (low) across the Ozark Dome (Illinois Basin). For a given inducing E3 geomagnetic field, geovoltages can, in principle, be close to zero due to grid-line orientation and due to the local impedance.

In Figure 14, we show a map of the geovoltage difference \( V_{MT} - V_{HS} \) at \( t = 3.0 \) s. Importantly, here, we see that geovoltages for the half-space model are generally lower (higher) than for the magnetotelluric tensors by as much as \(-1000\) V across the Ozark Dome (Illinois Basin)—the peak-to-peak range in error is \(-2000\) V. As we noted for the E3 geoelectric field, the geographic granularity in grid geovoltage is due primarily to the three-dimensional surface impedance of the solid Earth. Other candidate half-space conductivities, such as \( \sigma_{HS} = 10^{-2} \) S/m or \( \sigma_{HS} = 10^{-2} \) S/m, either of which might be reasonably used for hypothetical type calculations, would result in systematic errors in grid voltages (compared to magnetotelluric tensors) that would be even larger than shown in Figure 14.

In more detail, in Figure 15a, we show geovoltage waveforms for line 42,687 (the identification given in the DHS files), which runs 93 km approximately east-west across southern Missouri and across the Ozark Dome. The geovoltage obtained using magnetotelluric tensors attains a maximum value of 1,960 V at \( t = 2.1 \) s. The waveform for the half-space model generally gives a lower geovoltage, and it is notably different in shape from the waveform for the magnetotelluric tensors. Geovoltage for the half-space model attains a maximum value of 1,042 V at \( t = 3.1 \) s, after the maximum obtained using the magnetotelluric tensors. In contrast, in Figure 15b, we show geovoltage waveforms for line 34,886, which runs 64 km approximately east-west across the border between southern Illinois and Missouri. The geovoltage obtained using magnetotelluric tensors reaches a maximum value of 298 V at \( t = 2.5 \) s. Here, the geovoltage waveform for the half-space model is generally higher, and it is also different in shape from the waveform for the magnetotelluric tensors. Geovoltage for the half-space model attains a maximum value of 731 V at \( t = 3.2 \) s, also after the maximum obtained using the magnetotelluric tensors. Under Ohm’s law, and assuming a typical line resistance of (say) one or a few ohms, E3 induced quasi-direct currents of hundreds of amps could plausibly be realized on power grid lines in the study region. Such currents would be sufficient to cause grid system failures and blackouts, though, because of the short duration of the E3 pulse, it is not certain that high-voltage transformers would be damaged (e.g., Electric Power Research Institute, 2019).

With Kirchhoff’s circuit laws, our estimates of E3 geovoltages could, conceivably, be used to estimate the induced currents on the eastern-midcontinental power grid. To do that, we would need several types of grid parameters: the connectivity of the grid, line resistances, and network grounding resistances (e.g., Boteler & Pirjola, 2017; North American Electric Reliability Corporation, 2013). Unfortunately, these parameters are not generally publicly available, and, at least in the United States, grounding resistances are often either unknown (e.g., Bui et al., 2017) or difficult to obtain (e.g., Overbye et al., 2013, their Section VI). Indeed, in an analysis of magnetic-storm induced currents on the grid of the eastern United States, Overbye et al. (2012) used grounding resistivities that they describe as being “ballpark” estimates. Our review of the literature found published reports using hypothetical grounding resistivities for the United States grid ranging from 0.1 \( \Omega \) (e.g., Horton et al., 2012; Pulkkinen et al., 2012) to 2 \( \Omega \) (e.g., Overbye et al., 2012)—a range factor of 20 that would certainly affect quantitative estimates of E3 induced currents. Until accurate grid-system parameters become available, we choose to focus our analysis on calculations we can make with confidence, such as the E3 geovoltages depicted in the movie-map of S1 and the snapshots of Figures 8–11.
Conclusions and Discussion

From this study, we learn that E3 EMP geoelectric fields, generated by a nuclear explosion in the near-Earth space environment above our heads, can be strongly distorted by the geography of the Earth's surface impedance, a tensor that is a function of three-dimensional geological structures beneath our feet. This qualitative point might have been anticipated from previous analyses in which synthetic magnetic signals (e.g., Bedrosian & Love, 2015; McKay & Whaler, 2006) and measured magnetic-storm variation (e.g., Cuttler et al., 2018; Kelbert et al., 2017; Lucas et al., 2018) are convolved with long-period ($<10^{-1}$ Hz; $>10$ s) magnetotelluric impedance tensors. Local surface impedance can have a significant effect on the amplitude, polarization, and variational phase of local induced geoelectric fields. In our analysis of E3, based on the convolution of a standard parameterization of E3 geomagnetic field variation with wideband magnetotelluric impedance tensors, we draw a similar set of general conclusions.

But our conclusions are not merely qualitative, we also arrive at important quantitative conclusions that are specific for the E3 hazard. In particular, just two seconds after a hypothetical nuclear explosion above the eastern-midcontinental United States, Section 8, geoelectric amplitude ranges from 1.7 V/km at one location to 33.2 V/km at another location, with a median amplitude of 6.2 V/km and a 68% interval of [2.7, 12.3] V/km. Generally, sites with a high degree of geoelectric polarization tend to have high geoelectric amplitude. We furthermore find, Section 9, that E3 geovoltages on power-grid lines can, for a hypothetical nuclear explosion over our chosen study region, be as high as 1,960 V; for other grid lines, due to localized small-amplitude geoelectric fields, and due to the orientation of a grid line relative to the orientation of the geoelectric field, geovoltages can, in principle, be close to zero. But using the half-space model leads to overestimation of geovoltages on some lines and underestimation on other lines—errors have a range.

![Figure 15](image-url). Geovoltage $V_{MT}(t)$ obtained using magnetotelluric tensors (black) and half-space $V_{HS}(t)$ (green) for line (a) 42,687 and (b) 34,886.
of ≈2,000 V. Overestimating an E3 hazard might motivate implementation of overly protective (and possibly expensive) mitigation strategies. Underestimating an E3 hazard might end up leaving a power grid vulnerable.

Given the results of our E3 EMP scenario analysis for the Eastern Midcontinent, it is reasonable to envision performing similar scenario analyses for other places. A long-period magnetotelluric survey, with sparse 70 km station spacing, has been performed for most of CONUS (Schultz et al., 2006–2018; Schultz, 2010) as part of the National Science Foundation’s EarthScope project (Williams et al., 2010). Magnetic-storm geoelectric hazards have been mapped using these long-period tensors and data time series from magnetic observatories—the eastern United States and the northern Midwest exhibit both high long-period surface impedance and high magnetic-storm geoelectric hazard (e.g., Love et al., 2016; Lucas et al., 2020). But as demonstrated in Section 7, long-period tensors do not adequately resolve E3 geoelectric signals—wideband impedance tensors are needed. Still, the long-period survey serves as useful reconnaissance—we might reasonably expect the eastern United States and the northern Midwest to each exhibit high E3 geoelectric hazards. Targeted wideband magnetotelluric surveying and E3 scenario analyses in the eastern United States would support quantitative E3 vulnerability estimates for power-grid systems serving many of the nation’s largest cities.

The methods used here might be further developed for more detailed studies in the future. To date, realistic estimates of magnetic-storm induced geoelectric fields have been obtained by convolving geomagnetic field variation, such as recorded at magnetic observatories, either directly with the magnetotelluric survey tensors (an approach similar to that used for this report) or with surface impedance derived from models of Earth electrical conductivity structure that are, themselves, derived through inversion of magnetotelluric survey tensors (models like those constructed for traditional solid-Earth investigations) (e.g., Kelbert, 2020; Marshalko et al., 2020; Simpson & Bahr, 2020). In extending the latter modeling approach to a study of the E3 geoelectric hazard, joint inversion might be made for Earth models using both geographically sparse long-period survey tensors and a denser distribution of wideband tensors. Collaboration is needed between different research groups to enable physics-based modeling of the EMP source and, just as importantly, proper treatment of three-dimensional Earth-surface impedance. Such a modeling project would provide improved understanding of the geographic-temporal nature of the EMP hazard for different scenarios and in different geological settings. Then, realistic assessments could be made of the vulnerability of power-grid systems to a realistic E3 geoelectric hazard.

Data Availability Statement

The impedance tensors are available from the Data Management Center of the Incorporated Research Institutions for Seismology (ds.iris.edu/ds/products/emtf). The power-grid line data are obtained from the Department of Homeland Security open data website (https://www.geoplatform.gov).

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