A Robust Estimation of the 3-D Intraplate Deformation of the North American Plate From GPS

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Abstract Glacial isostatic adjustment (GIA) is the main cause of deformation in intraplate North America. Here we use up to 3,271 Global Positioning System station velocities to image this 3-D deformation across the entire plate. We apply a new robust strain rate estimation algorithm (median estimation of local deformation), which does not require the assumption that part of the plate is unaffected by GIA, an assumption we show to be false. Our results show extension in the area underneath the Laurentide ice sheet, contrasted with a semiannular belt of horizontal contraction of up to \(-4 \times 10^{-9} \text{ yr}^{-1}\) around the former ice sheet. This contractional belt is kinematically linked to an \(-1-2 \text{ mm yr}^{-1}\) far-field horizontal motion directed toward the ice sheet. Our results, together with a new robustly imaged vertical velocity field, are consistent with GIA as the main cause of deformation, although the contractional strain rates around the former ice sheet and the far-field horizontal velocities are significantly higher than those predicted by the ICE6G_C(VM5a) model. This finding suggests that our results, including the location of reversal in sign of horizontal motion relative to the ice sheet, will be useful to reevaluate the mantle viscosity structure used by GIA models. Besides the GIA-attributed deformation, we find almost no other region with significant strain accumulation. A plate-scale spatial correlation between strain rate and seismicity is absent, suggesting that GIA is a limited driver of contemporary seismogenesis and that intraplate seismicity must be attributable to factors other than secular strain accumulation.

Plain Language Summary The theory of plate tectonics says that tectonic plates move as rigid blocks along the Earth’s surface and that the Earth’s crust should only deform at the boundary between plates. However, the recent explosion in the number of high-precision Global Positioning System stations allowed us to capture some subtle deformation patterns inside the North American plate that only became apparent by a very careful analysis of the relative motions between thousands of stations. We found that most of the plate is moving at 1–2 mm/year towards central Canada. Consequently, around most of Canada there is a zone where the crust is contracting. Within Canada, the crust is extending outward and is moving upward rapidly. These patterns can be explained by the process of the crust and mantle still rebounding from a time when it was covered by a thick ice sheet about 16,000 years ago. The fact that this causes the land to move towards the former ice sheet is an unexpected result that will be useful in understanding the relaxation properties of the underlying mantle. Moreover, we found that earthquakes inside the North American plate do not occur where we see the crust deform, which leaves these events still enigmatic.

1. Introduction

Plate rigidity is the leading axiom in the kinematic framework of plate tectonic theory. For the North American plate, early studies using space-geodetic data confirmed plate rigidity within the uncertainty of the observations (Argus & Gordon, 1996; Dixon et al., 1996; Kogan et al., 2000). As the number of Global Positioning System (GPS) stations grew over time, their position time series lengthened, and their velocity uncertainties decreased, two types of studies have emerged: One uses the GPS velocities to estimate intraplate strain rates (Calais et al., 2006, 2016; Gan & Prescott, 2001; Ward, 1998), while the other identifies and delineates a subpart of the plate that exhibits rigidity within the measurement precision (Altamimi et al., 2012, 2017; Argus et al., 2010; Blewitt et al., 2013). The conclusion from both types of studies is that a large part of intraplate North America fails to behave as one rigid entity and that ongoing glacial isostatic adjustment (GIA) is likely the main cause. This conclusion is corroborated by the geodetically observed pattern of vertical motions (Alinia et al., 2017; Argus & Peltier, 2010; Calais et al., 2006; George et al., 2012; Goudarzi et al., 2016; Mazzotti et al., 2011; Park et al., 2002; Peltier et al., 2015; Sella et al., 2007; Snay et al., 2016; Tiampo et al., 2012).
The aforementioned geodetic strain rate models have low spatial resolutions by design, because (1) the sought-after signal is near the noise level in the data, and (2) the influence of even minor outlier data, particularly across short distances, is large. Despite these limitations, GPS data have been used for regional strain rate estimation for the New Madrid Seismic Zone (NMSZ; Calais et al., 2005; Craig & Calais, 2014; Frankel et al., 2012; L. Liu et al., 1992; Newman et al., 1999; Smalley et al., 2005; Weber et al., 1998), the Wasatch Valley Seismic Zone (WVSZ) north of the NMSZ (Craig & Calais, 2014; Galgana & Hamburger, 2010; Hamburger et al., 2002), and the Saint Lawrence Valley Seismic Zone in Québec (Lamtohe et al., 2010; Mazzotti et al., 2005). The difficulties with these attempts are exemplified by the observations that (1) based on the same data, no consensus has been reached on whether strain is accumulating across the NMSZ, and (2) the strain rate seemingly increases for areas with decreasing sizes around the WVSZ (Galgana & Hamburger, 2010). The disparate results for the NMSZ have led to some public debates on what they imply for the seismic hazard there (Frankel, 2003; Schweig et al., 1999; Stein et al., 2003) and for the origin of intraplate earthquakes there and elsewhere (Calais et al., 2016; Craig et al., 2016; M. Liu et al., 2011; M. Liu & Stein, 2016; Stein et al., 2009).

The large-scale strain rate results presented by Calais et al. (2006, 2016) are only for the central and eastern United States. Until our work, studies that took a plate-wide focus on the North American deformation field only used the horizontal velocity field (Argus & Peltier, 2010; Sella et al., 2007; Snay et al., 2016). Such studies faced the difficulty of defining a reference frame in which to present and interpret the geodetic velocities. The definition of such a frame is not obvious because GIA is often predicted to affect most, if not all, of the plate (Argus & Peltier, 2010; Klemann et al., 2008; Latychev et al., 2005; Peltier & Drummond, 2008). Indeed, when GPS data across all of the North American plate are used to fit a plate-wide rigid-body rotation, the misfit is much larger compared to plates not expected to have internal GIA-related velocity gradients (Kreemer et al., 2014). Most recent plate motion studies therefore simply exclude stations in areas expected to be affected by GIA. However, if GIA causes significant far-field horizontal motions, such spatial restriction may lead to a biased plate motion estimate that (1) may not be representative of the entire plate, (2) cannot be compared to long-term plate motion estimates, and (3) cannot be used to constrain GIA models, which are particularly sensitive to the far-field horizontal motions (James & Morgan, 1990; Mitrovica et al., 1993, 1994; Sella et al., 2007).

Some of the effects of GIA on plate motion estimation were already discussed by Klemann et al. (2008), who showed this problem to be the largest for North America. Other studies have tried to estimate plate motion by accounting for GIA (Argus & Peltier, 2010; Snay et al., 2016) or by using GIA predictions to assess which stations are not affected (Argus et al., 2010). However, displacement rates from different GIA models can vary considerably (see, e.g., Sella et al., 2007), leaving much uncertainty in such an approach. To make GIA models using horizontal velocities less sensitive to assumptions about the reference frame, strain rates that are reference frame independent should be considered instead. We are aware of only one study that used strain rates to constrain GIA, and that was for Fennoscandia, based on a sparse data set, and was limited in area (Nocquet et al., 2005).

Here we present a new plate-wide high-resolution strain rate model for entire intraplate North America. Our results are based on a new algorithm that is developed specifically for intraplate areas; it is very stable and robust against outlier data and has a resolution that is commensurate with the density of GPS stations in a given area (but here with a minimal $\sim$100-km spatial resolution). Our strain rate estimation method does not require the data to be in any specific reference frame, yet our approach produces a horizontal intraplate velocity field that is consistent with the found deformation field. Our results are presented in conjunction with a new vertical displacement map based on the GPS Imaging technique presented by Hammond et al. (2016).

There are several novelties in our approach and findings. First, our strain rate model is derived from the most complete GPS data set available to date, taking advantage of extensive geographic coverage over nearly all of North America, from the mid-Atlantic Ridge (MAR) to the western Cordillera, Yucatan Peninsula to Greenland and the Arctic Ocean (Figure 1). Second, we describe a new method for making robust estimates of crustal strain rate from GPS data and apply it to provide constraints on the GIA process. This approach overcomes the limitation of prior studies that rely on velocities, and hence a choice of reference frame, to constrain GIA. Moreover, our model includes realistic uncertainties in the local strain rate estimate. Third, we show the presence of a belt of significant contractional strain rate around the location of the former ice sheet, and this belt is more compact and of higher intensity than predicted by the ICE6G_C(VM5a) GIA model.
This contractional belt is mostly the consequence of a far-field inward motion that is greater than predicted and larger than the outward motion from the location of the former ice sheet. Fourth, we perform a quantitative examination of the relationship between GIA, intraplate strain rate, and seismicity and resolve no significant strain accumulation at most of the notable areas of seismicity in North America. This result extends earlier regional comparisons that reached similar conclusions to the entire North American plate.

2. Method

For a given evaluation point, we estimate the median strain rate tensor ($\hat{\varepsilon}$) and median rotation rate vector ($\hat{\omega}$) from a set of $N$ local estimates of strain rate tensors ($\varepsilon$) and rotation vectors ($\omega$). Each $\varepsilon$ and $\omega$ is determined from three horizontal velocity vectors at the vertices of a local spherical triangle defined below, each specified by $(v_e, v_n)$ in the local east and north direction, respectively. Our method is called MELD (median estimate of local deformation).
We remedy this by averaging the locations and velocities for all points located within 5 km of each other. We observations points on a sphere (Renka, 1997). In practice, this means that no collocated points can be used.

...point equal to the evaluation point, even when that point lies slightly outside the considered triangle. (in our case spherical triangles) and are very weakly dependent on the location of the reference point. For to include three station velocities). Those estimates are assumed to be constant within the area considered...tions at the location of the velocity observation (no subscript) or at a reference point (subscript 0):

\[ \mathbf{R} \begin{pmatrix} v_x \\ v_y \end{pmatrix} = \begin{pmatrix} (\mathbf{r}_e(\mathbf{e}_0)(\mathbf{e}_0) + (\mathbf{r}_e(\mathbf{e}_0)(\mathbf{e}_0) + (\mathbf{r}_e(\mathbf{e}_0)(\mathbf{e}_0) + (\mathbf{r}_e(\mathbf{e}_0)(\mathbf{e}_0) + (\mathbf{r}_e(\mathbf{e}_0)(\mathbf{e}_0) - \mathbf{e}_x - \mathbf{e}_y - \mathbf{e}_z) \end{pmatrix} \begin{pmatrix} \hat{e}_x \\ \hat{e}_y \\ \hat{e}_z \end{pmatrix} \]

(1)

where \( R \) is the radius of the Earth and \( \hat{e}, \hat{n}, \) and \( \mathbf{r} \) are the unit vectors in the local east, north, and radial directions at the location of the velocity observation (no subscript) or at a reference point (subscript 0):

\[ \hat{e}^T = (- \sin \phi \ \cos \phi \ 0) \]

(2)

\[ \hat{n}^T = (- \sin \theta \cos \phi \ - \sin \theta \sin \phi \ \cos \phi \)

(3)

\[ \hat{r}^T = ( \cos \theta \cos \phi \ \cos \theta \sin \phi \ \sin \phi \)

(4)

where \( \phi \) and \( \theta \) are the longitude and latitude, respectively. Using horizontal velocities on an ellipsoid in this spherical approximation leads to model errors in strain rate that are negligible. If one has velocity observations at at least three distinct and noncollinear locations, then \( \hat{e} \) and \( \hat{\omega} \) can be solved for (by expanding (1) to include three station velocities). Those estimates are assumed to be constant within the area considered (in our case spherical triangles) and are very weakly dependent on the location of the reference point. For consistency, for each triangle that contributes to the median at the evaluation point, we set the reference point equal to the evaluation point, even when that point lies slightly outside the considered triangle.

The MELD algorithm consists of the following steps. First, a Delaunay triangulation is constructed from a set of observations points on a sphere (Renka, 1997). In practice, this means that no collocated points can be used. We remedy this by averaging the locations and velocities for all points located within 5 km of each other. We then identify the unique triangle in which a given evaluation point resides. In theory that could give an estimate of \( \hat{e} \) and \( \hat{\omega} \) from the velocities at that triangle’s three vertices (we call this “level 1”). However, these estimates have no robustness because there is no redundancy in the data (i.e., the number of velocity data equals the number of model parameters) so any noise in the velocity data gets directly mapped onto the model estimates. This makes the model parameters extremely sensitive to outlier data. This problem is more noticeable when the actual strain rate is near zero and is particularly acute when the distances between the vertices are small (because strain rate is a spatial derivative). These problems can be overcome by considering more velocity data, possibly weighted by distance to the evaluation point, and by performing a least squares regression. However, such regressions come at the expense of the spatial resolution of the estimate, while the problem of outlier data always persists in any least squares estimation. The next few steps in MELD are an attempt to overcome such problems.

The second step in MELD is to consider the next level (“level 2”) of data points. These points are those that in the Delaunay triangulation are the nearest neighbors of the original three vertices. This larger set of points, which includes the original three points, is then used to construct all possible spherical triangles. To reduce uncertainty in the median value, we only consider triangles that are not too small, too skinny, or too skewed, properties that make the strain rate estimate for that triangle very uncertain. These geometric qualities are reflected in the model covariance matrix derived from the least squares solution to equation (1). To select triangles with good geometric properties, we only consider those for which each of the matrix’s components has a standard deviation smaller or equal to an imposed maximum value \( \sigma_{\text{max}} \). The choice of this value is discussed below.

Next, we only consider triangles that make a useful contribution to the model parameters at the evaluation point because we want the model estimate to be as local as possible. However, we do not want to limit this to the triangles that enclose the evaluation point, and we therefore also consider those triangles for which the evaluation point is within the triangle’s local circle of influence.” That is, we consider all triangles for which the distance between a triangle’s centroid and the evaluation point \( D \) is less or equal to the maximum distance between the centroid and any of its three vertices \( R_c \). If the number of triangles for which \( D \leq R_c \) does not
yield an imposed minimum number of triangles \( (N_{\text{min}}) \), then we consider the next level ("level 3"). That is, we additionally consider the data points that, in the Delaunay triangulation, are the outward nearest neighbors of the points considered in level 2, and we then create a new set of all possible local triangles, and so on.

If at a given level \( N \geq N_{\text{min}} \), we find the median values of \( \hat{c} \) and \( \hat{\omega} \) from all \( N \) triangles. To avoid sensitivity to the coordinate system, we do not estimate medians on each parameter individually but find the multivariate median (or "centerpoint") as defined by the point of maximum multivariate "Tukey" depth (Rousseeuw & Struyf, 1998; Struyf & Rousseeuw, 2000). For each parameter \( (y) \) we then estimate the standard deviation in the median values from the \( S_N \) estimator (Rousseeuw & Croux, 1993):

\[
\hat{\sigma} = 1.1926S_N = 1.1926\text{median}_{j} \left\{ \text{median} \left| y_j - y \right| \right\}
\]  

(5)

The definition of \( S_N \) reads as follows: For each estimate of \( y \), determine the median absolute deviation with the other estimates of \( y \), and then take the median of those \( N \) median values. The median can be efficiently found with the "quicksort" algorithm (Hoare, 1961). \( S_N \) is independent of coordinate frame transformation, and can capture the asymmetric spread in a non-normal distribution. The 1.1926 factor ensures that \( \hat{\sigma} \) equals the typical standard deviation if the estimates follow a normal distribution (Rousseeuw & Croux, 1993).

Following Blewitt et al. (2016), final values of the median for each parameter are then computed using the multivariate median estimate, but only after the tails of the distribution are trimmed beyond 2\( \hat{\sigma} \) in order to remove outliers and to make the final median even more robust. We identify these outliers for each component separately, discard each triangle that has at least one outlying parameter, and reestimate the multivariate median estimate, but only after the tails of the distribution are trimmed beyond 2\( \hat{\sigma} \).

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\[
\hat{\sigma}_{\text{final}} = \sqrt{\frac{1}{2} \hat{\sigma} \sqrt{\frac{N}{N'}}} 
\]  

(6)

where \( \hat{\sigma} \) is the standard deviation after trimming and \( N' \) is the effective number of independent triangles. It is difficult to assess how many triangles are truly independent, given that each velocity is being used in multiple estimations (i.e., most stations are on the vertex of many local triangles). As an approximation, we set \( N' = M/3 \), with \( M \) being the number of stations used after trimming. For three stations, this gives the trivial estimation of \( N' = 1 \). For six stations, one can form two independent triangles, and so on.

To evaluate whether the uncertainties are realistic we perform a test under the assumption of Gaussian noise in the data. That is, we simulate a large number of data sets in which the velocities always reflect a constant \( 1 \times 10^{-9} \text{ yr}^{-1} \) north-south strain rate across the entire network, but we perturbed the velocities with Gaussian noise with a standard deviation equal to the 68th percentile of the observed uncertainties in our data described below (i.e., 0.37 mm yr\(^{-1}\)). We then use MELD, and the model input parameters presented below, to find that in an aggregated 68\% of instances the true strain rate falls within \( \hat{\sigma}_{\text{final}} \) from the estimated strain. We therefore find \( \hat{\sigma}_{\text{final}} \) to be very realistic; the standard deviations in the model very precisely reflect the noise in the data.

We set \( N_{\text{min}} \) to 120. This value of \( N_{\text{min}} \) can in theory be reached with a minimum of 10 unique points (if no triplet of points are collinear), because the number of possible triangles is \( M(M - 1)(M - 2)/6 \). We find that the exact choice of \( N_{\text{min}} \) is not critical; either the strain rates may diffuse spatially with increasing \( N_{\text{min}} \), where data points are far apart, or estimates become less robust with decreasing \( N_{\text{min}} \), because there are not enough estimates to indicate a robust median.

In contrast, the choice of \( \sigma_{\text{max}} \) critically affects the robustness of the strain rate field. We set \( \sigma_{\text{max}} = 5 \times 10^{-9} \text{ yr}^{-1} \), which corresponds to the standard deviation in strain rate that is inherent for a triangle with a size large enough such that 0.28 mm yr\(^{-1}\) of motion across it corresponds to a strain rate of \( 1 \times 10^{-9} \text{ yr}^{-1} \). The 0.28 mm yr\(^{-1}\) is very close to the 0.3 mm yr\(^{-1}\) that is the median standard deviation in the velocity data. This choice of \( \sigma_{\text{max}} \) thus ensures that the typical noise in the strain rates in each triangle, resulting from the noise in the data, stays at the level \( 1 \times 10^{-9} \text{ yr}^{-1} \). The chosen level of \( \sigma_{\text{max}} \) implies a minimum surface area for an equilateral triangle of \( \approx 34.5 \times 10^{3} \text{ km}^2 \), which is equivalent to a circle with a radius of...
~105 km. This is therefore the best spatial resolution our model can obtain, even at places with dense coverage. Otherwise, the spatial resolution of our model automatically mirrors the resolution of the network.

To test how robust MELD is compared to least squares estimates, we create a synthetic model that is similar to the one introduced above to validate the uncertainties in our model: a uniform north-south strain rate of $1 \times 10^{-9} \text{ yr}^{-1}$ across all of North America. We create 80 different realizations of the velocity field at the stations used below using the same Gaussian noise as used above. For each realization we estimate the strain rate at 679 evaluation points using MELD or doing a least squares regression on the same velocities MELD identified to contribute to each evaluation point’s strain rate estimate. We find that in 81% of the cases MELD’s estimate is closer to the input strain rate than the least squares estimate. When we do the same comparison, but apply a median spatial filter (MSF; see section 3.1) to each synthetic velocity field, MELD still performs better than least squares for 74% of the aggregate evaluation points. The robustness of MELD is also shown below; we obtain very similar results for a model based on all observations compared to models based on alternative data sets that are smoothed or have outliers removed.

3. GPS Data

We consider all data between 1 January 1996 and 3 November 2017 within the spatial footprint of the stable North American plate as defined by Kreemer et al. (2014) but modified by a cut-off north of Alaska and Greenland. The data come from many different sources (see Acknowledgments). A large number of the stations we analyze are not from traditional continuous GPS networks and data archives (i.e., UNAVCO or National Geodetic Survey [NGS]) but are drawn from commercial and state networks and networks that were installed to study the ionosphere (Jayachandran et al., 2009), the troposphere (Ware et al., 2000), and surface subsidence (G. Wang, et al., 2015). For some of these latter networks, metadata may be lacking and the stability of the stations and the quality of their data may not be optimal. We consider them here because our method, thriving on the power of numbers, employs robust analysis steps to overcome occasional problems with the data, such as undocumented equipment changes, and monument instability. All considered stations are continuously operating, except for those belonging to the Canadian Base Network, which have infrequent measurements on stable pillars (Henton et al., 2006). We process all the RINEX data available from these stations in 24-hr batches to obtain daily three component positions. The solutions are processed using the precise point position method with the GIPSY OASIS II software provided by the Jet Propulsion Laboratory (JPL), using JPL’s fiducial-free GPS satellite orbit and clock products (Zumberge et al., 1997). Our specific observable modeling and parameter estimation procedures are outlined elsewhere (Blewitt et al., 2013; Kreemer et al., 2014) The daily station coordinate solutions are then transformed using a 7-parameter transformation from the fiducial-free orbit frame into the IGS08 reference frame using JPL products.

To estimate velocities from the position time series we use the MIDAS (Median Interannual Difference Adjusted for Skewness) median-trend algorithm (Blewitt et al., 2016). The MIDAS trend estimator is robust against outliers, offsets (even undocumented ones), and seasonality. MIDAS will not consider slopes across known offset epochs (from the metadata) in its estimation of the median trend. If the offset is not documented, MIDAS still finds a robust slope if the time series is sufficiently long (see Blewitt et al., 2016 for details). We only analyze time series that are at least 1.5 years long. This is a shorter timespan than recommended for least squares regression (Blewitt and Lavallée, 2002), but MIDAS rates for this time length are still robust against 17% outliers in the time series (Blewitt et al., 2016). Moreover, because of its robustness, MELD thrives on a maximization of the number of data.

MIDAS also gives realistic trend uncertainties that reflect the scatter in the data used in the median estimator; the median standard deviation for the velocities we consider is 0.3 mm yr$^{-1}$. While our strain rate estimation method does not include the data uncertainties, they are used to evaluate the goodness of fit of our model as well as in the median spatial filtering (section 3.2).

To enhance the spatial data coverage for the strain rate estimation, we also include horizontal velocities from two published studies: 24 velocities for Québec, Canada (Goudarzi et al., 2016) and 18 velocities for the Westfjords peninsula in northwest Iceland (Árnadóttir et al., 2009). These studies were rotated into our solution by minimizing the velocity difference at stations common between those studies and our solution (including stations outside the North American footprint). Because the Goudarzi et al. (2016) study presents rates for stations in Québec that are derived from longer time series than are available to us, we discard our
rate estimates for those stations before the transformation. We tried to include the published velocities of Galgana and Hamburger (2010) for the WVSZ but found systematic strain rates around this network that suggest a possible problem with their reference frame and consequently with our subsequent transformation, so they were not used. After transformation, we retain only our own velocity estimates at common stations. For the imaging of the vertical motion, we replace our vertical rates with those published by Goudarzi et al. (2016) for Québec stations, which are based on longer time series.

A few stations with velocity estimates within the North American footprint were manually removed beforehand. These include six stations in the middle of the Greenland ice sheet that are collocated with communication sites. For the horizontal (but not vertical) data we remove stations in central Iceland, south-central Mexico, and near the Cayman Transform fault, which may all be affected by interseismic strain accumulation on the nearby plate boundary faults. Considering these exclusions, we have 3,245 and 3,271 potentially useful horizontal and vertical velocities, respectively; their locations are shown in Figure 1 and the velocities are tabulated in the Supplemental Material.

We finally combine all velocities for stations within 5 km from each other. This combination is done in an iterative hierarchical algorithm that sorts the stations according to the number of other stations that are within 5 km distance. After the combination for the station with most collocated stations, a new sort is done, and so on, until no stations are within 5 km from each other. For the combination, we average the positions of the collocated stations and assign average horizontal velocities weighted by the individual data variances and corresponding standard deviations.

3.1. Data Filtering

We show below that MELD is robust against outlier horizontal velocities. To make the vertical velocity field robust, we filter the vertical velocities using the GPS Imaging technique of Hammond et al. (2016). First, GPS Imaging applies an MSF to replace observed velocities with weighted median values. The median velocities are determined using neighboring rates with a weighting scheme that uses the distance to neighboring stations, the velocity uncertainties, and a spatial structure function (SSF) that is derived from the vertical rates. We use the same SSF as was derived for vertical velocities for California and Nevada by Hammond et al. (2016). In a second step we use GPS Imaging to interpolate the vertical velocities onto a 0.5° grid raster using the same weighted median process. The original vertical rates and their MSF counterparts are shown in Figures S2a and S2b in the supporting information, respectively.

To assess the robustness of our strain rate model to outlier data, we create two alternative horizontal velocity fields. The first has 331 of the most egregious outlier velocities removed using a new outlier detection algorithm (Appendix A). Another input velocity field is created by applying an MSF to the horizontal velocities (including outliers). For this, median filtered horizontal velocities are derived using the same SSF as used for the vertical velocities, but for an MSF approach that is tailored to horizontal velocities. That is, because spatial gradients in horizontal velocities can be due to a rotation, we estimate a local rotation for each set of velocities used to determine a median value, subtract that rotation before doing the MSF, estimate the median, and then rotate the median velocities back into the original IGS08 frame using the same rotation vector. The resulting MSF velocity field, and corresponding model, is relatively free of speckle noise compared to the original velocity field and its associated model.

4. Results

4.1. Strain Rate Field

For the MELD model estimation we use a 0.5° grid for the evaluation points and only considered points within the stable North American footprint. To assess the robustness of our strain rate model to outlier GPS velocity data, we compare the style and magnitude of three different models: one based on the original velocity data, one based on the same data minus 331 outliers, and another based on a version of the original data that has the MSF applied to it. As a confirmation of the robustness of MELD, we find no significant differences in the tensor style (Figure S3) nor in the second invariant of the tensor (Figures S4 and S5). The style is defined as 
\[ \epsilon_{ij} = \max(|\epsilon_1|, |\epsilon_2|) \] (Kreemer et al., 2014), where \( \epsilon_1 \) and \( \epsilon_2 \) are the largest and smallest principal strain rates, respectively, and the second invariant of the tensor is defined as 
\[ q = \sqrt{\epsilon_{ee}^2 + \epsilon_{en}^2 + 2\epsilon_{en}^2} \] . The model based on data that included outliers naturally has somewhat larger
uncertainties and thus fewer places with significantly resolved strain rates; therefore, for the remainder of this paper, we present the model that had the outliers removed.

Figure 2 shows a map of the style of the strain rate tensor, but the style is only shown for places where the second invariant is larger than 2 times the standard deviation ($\sigma$) therein. This result shows that for most places with significantly resolved strain rate, it is either contractional or extensional. For this reason, we focus below on the dilatational component of the strain rate tensor; that is, $\dot{\varepsilon}_{1} + \dot{\varepsilon}_{2}$, which equals $\dot{\varepsilon}_{ee} + \dot{\varepsilon}_{nn}$. Figure 3a shows the formal error in the dilatation rate, and Figure 3b shows the spatial resolution of our model result (expressed for each evaluation point as the radius of a circle with the same area as the median area of all triangles used in that point’s estimate). Figure 4a shows the dilatation rate where it exceeds $2\sigma$, and this is superimposed with spatially averaged normalized principal strain rate axes. Figure S6 shows the dilatational strain rates for all areas. Tabulated values can be found in the Supplemental Material.

The main feature in our model is a nearly contiguous semiannular belt of (mostly) contractional strain rates up to $\sim 4 \times 10^{-9} \text{yr}^{-1}$ that stretches from the Canadian Prairies (west of Lake Winnipeg), to the Great Lakes, the Canadian Maritime Provinces, Davis Strait, and Baffin Bay. The southern part of the belt also appears in the model of Calais et al. (2016). There appears to be a gap in the belt in terms of contractional strain rates between the Great Lakes to the Maritime Provinces, but this is mostly because the style of strain rate is shear there (Figure 2). Indeed, the second invariant of strain rate (Figure S4b) shows as a more continuous belt than the dilatational strain rates. The contractional belt does not wrap around through northwestern Canada, that is, between Alaska and central Canada. This could be because it crosses the northern Cordillera of Canada, which we place outside the intraplate area. It is also possible that this absence is an artifact of poor data coverage. The motion of northern Alaska toward Hudson Bay is of the same magnitude as for Greenland,

**Figure 2.** Style of the strain rate tensor, defined by Kreemer et al. (2014) as $(\dot{\varepsilon}_{1} + \dot{\varepsilon}_{2})/\max(\dot{\varepsilon}_{1}, \dot{\varepsilon}_{2})$ (where $\dot{\varepsilon}_{1}$ and $\dot{\varepsilon}_{2}$ are the largest and smallest principal strain rates, respectively), with the style only shown for places where the second invariant of the strain rate (i.e., $\sqrt{\dot{\varepsilon}_{ee}^2 + \dot{\varepsilon}_{nn}^2 + 2\dot{\varepsilon}_{en}^2}$) is larger than 2 times the standard deviation therein. The scale is saturated at $-1$ (compression) when both principal axes are compressional, and at $+1$ (extension) when both principal axes are extensional. Results are shown for evaluation points on a 0.5° grid.

**Figure 3.** (a) Standard deviation of the dilatation rate (i.e., $\dot{\varepsilon}_{ee}$ or $\dot{\varepsilon}_{nn}$ or $\dot{\varepsilon}_{1} + \dot{\varepsilon}_{2}$) and (b) spatial resolution expressed as the radius (in km) of a circle which area equals the median area of the triangles used to determine the model parameters.
but, because of the very sparse data coverage in the Northwest Territories, the corresponding contractional strain rate signal is smeared out compared with what we see in Baffin Bay.

In the interior of the contractional belt (i.e., an area stretching from Northwest Territories in the northwest and Québec in the southeast), strain rates are systematically extensional. Specifically, we see two distinct centers of extension: one east of Hudson Bay and another one, mostly insignificant, west of the Hudson Bay. Strain rates in the eastern center are up to $4 \times 10^{-9}$ yr$^{-1}$, and there, the extension is mostly biaxial. Our model's extensional strain rates are significantly lower than those reported by Calais et al. (2016).

Standard deviation in the strain rates is generally below $1 \times 10^{-9}$ yr$^{-1}$ (Figure 3a). As a result, the low strain rates in most areas besides the described semiannular belt and large parts of the Canadian extensional zone are smaller than the 2σ significance level (Figure 4). It is still worth pointing out that in most far-field areas (i.e., the areas outward from the contractional belt) strain rates are mostly contractional with the shortening axis typically orthogonal to the contractional belt. That is, the far-field deformation is generally similar in style and orientation, but at a lower amplitude, than the strain rate within the contractional belt. We find positive dilation for some far-field areas in the conterminous United States, including the NMSZ, but these are not, or barely, significant in our current model. The only area south of the semiannular contractional belt with significant (compressional) strain rate is in north Texas and southwestern-most Oklahoma.

The root-mean-square (RMS) misfit to the data is 0.5 mm yr$^{-1}$, and the RMS of the misfit normalized by the data uncertainties (NRMS) is 2.0. These values were determined by considering for each evaluation point all unique GPS velocities that were used in the estimation and comparing them with the predicted velocities based on the evaluation point's median model parameters. Because each observed velocity contributes to the MELD estimates of multiple evaluation points, each observed velocity is used multiple times in the RMS calculation. For each GPS velocity, the median RMS scatter in the multiple velocity predictions is $\approx 0.07$ mm yr$^{-1}$. That this value is much smaller than the uncertainties in the data attests to the spatial self-consistency of the strain rate model.

To test how our model compares with some of the best observed GPS velocities, we compare the observed velocities of the 10 long-running stations used by Kreemer et al. (2014) to estimate the angular velocity for North America. For those 10 stations we find an RMS of 0.3 mm yr$^{-1}$ and an NRMS of 1.7 when comparing the observed velocities with the average predictions at each station. This confirms the above-mentioned fit when all stations are considered and is in fact better, likely because of the increased scatter in the velocity data that is inferred from short(er) time spans. When no strain rate in the plate is considered, and the plate rotates as predicted by our net rotation (see section 4.2), the RMS misfit for the velocities at those 10 stations increases to 0.7 mm yr$^{-1}$ and the NRMS to 26.0 (using the average predictions at each station).

### 4.2. Horizontal Intraplate Velocity Field

For each evaluation point, we determine the weighted median Euler vector ($\hat{\omega}$). When all those vectors (i.e., the separate components of $\hat{\omega}$) are averaged, using as weight the local area of the evaluation point, we obtain the "net rotation" of the total area considered relative to the IGS08 reference frame. We do not use the variances in $\hat{\omega}$ as an additional weight to derive the mean rotation because they vary systematically as a function of the location vector of the evaluation point relative to $\hat{\omega}$. This would cause a bias in the net rotation estimate. The net rotation vector we find corresponds to a Euler pole located at 2.3°S and 86.0°W and a rotation rate of 0.2010° Myr$^{-1}$. We discuss the net rotation vector and its comparison with published estimates in section 5.3 below.

We remove the net rotation from the mean predicted IGS08-fixed velocities at the data locations to obtain the intraplate velocity field (Figure 4b), where the mean predicted velocities are the averages based on the model predictions for all evaluation points to which the data point contributed. The velocity field thus shows

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**Figure 4.** (a) Dilatation rate at a 0.5° evaluation grid with blue colors being contractional strain rate and red colors being extensional strain rate. Results are only shown for dilatation rates that are larger than twice their standard deviation. Superimposed are the orientations and relative size of the principal axes: The white vectors are positive (extensional), and the black vectors are negative (contractional). All principal axes are normalized to unity, and they reflect averages for a set of nonoverlapping areas of equal size. (b) Same dilatation rate as in Figure 4a but now superimposed with selected predicted velocities (i.e., number of continental stations between 26°N and 48°N is downsampled by a factor of 5 for clarity). The predicted velocities are the averages based on the model predictions at all evaluation points the data point contributed to. The reference frame is one in which the area-weighted mean rotation for all evaluation points (i.e., the plate's net rotation) is subtracted from the velocities in the original IGS08 reference frame.
the component of the IGS08 velocity field related to intraplate deformation and not to the net rotation of the plate. This intraplate velocity field and found strain rate field should therefore be consistent with one another. Figure 5 shows the intraplate velocity field (both as vectors at selected evaluation points and as a color map of the velocity) that is obtained by calculating the velocity at each evaluation point based on the local $\omega$ and then removing the net rotation. Evaluation points are part of a 0.5° grid.

The intraplate velocities systematically increase in magnitude with distance away from the extensional center(s) with velocities $>2$ mm yr$^{-1}$ at the farthest distances, that is, northern Alaska, eastern Greenland, Flores Island (Azores), and Yucatan Peninsula (Mexico). Moreover, the direction of the far-field velocities is consistently toward the extensional center(s) and generally aligned with the contractional strain rate orientation. Also, in this reference frame, a distinct line of zero-velocity coincides with where the spatial gradient in the horizontal velocity is near its maximum and runs closely parallel to the contractional strain rate belt, at least within continental North America. These features of the intraplate velocity field affirm that our inferred net rotation reference frame is sensible in that it naturally links the intraplate velocity field with the strain rate field (even though the strain rate field is independent of the plate rotation). In any case, the inward velocity of the far-field is larger than the outward velocity of places previously covered by ice. The contractional strain rate belt is therefore more the result of the inward motion of the far-field than of near-field outward motion. In other words, we would not see the contractional belt if the far-field horizontal motions were not inward and of the magnitude they are.
Strain rates in the crust beneath the west Atlantic are controlled by only a few stations, and it is therefore worth noting that the observed velocities at two distinct stations on Bermuda (BRMU and BMPD) are within 0.5 mm yr$^{-1}$ of each other and that the observed velocity on Flores Island is within 0.3 mm yr$^{-1}$ of that on nearby Colvo Island (not used here; Fernandes et al., 2004). We find that Iceland’s Westfjords move distinctly westward in our net rotation frame, an observation already evident in the results by Árnadóttir et al. (2009), but the 1–2 mm yr$^{-1}$ relative to the ITRF2005’s definition of North America plate motion (Altamimi et al., 2007) was dismissed as being insignificant at the time.

Furthermore, in our rotation-free reference frame the GPS motions at stations near the MAR are directed away from the MAR and are much larger than would be predicted from horizontal thermal contraction. For the slowly spreading MAR, strain rates from thermal contraction are expected to be much lower than the $\sim 1 \times 10^{-9}$yr$^{-1}$ predicted very near the fast-spreading East Pacific Rise (Kreemer & Gordon, 2014), and even lower still with increasing distance away from the MAR. As a corollary, far-field horizontal GIA motions should be considered in thermal contraction studies.

### 4.3. Vertical Velocity Field

The vertical velocity field we derive using GPS imaging is shown in Figure 6 and the standard deviations in Figure 7. The highest uplift rates can be seen across Hudson Bay and northern Quebec with a maximum, up to $\sim 13$ mm yr$^{-1}$, near the southeastern Hudson Bay. We also see up to $\sim 16$ mm yr$^{-1}$ in parts of

![Figure 6. Vertical velocities (positive is up) inferred from Global Positioning System imaging technique (Hammond et al., 2016). Results are shown for evaluation points on a 0.5° grid. Superimposed is the location of the transect shown in Figure 8.](image-url)
Greenland. Outside the area of uplift, the crust is subsiding nearly everywhere. A clear belt of significant subsidence can be seen immediately south of the Canadian uplift area, stretching from the Canadian Prairies and northern Great Plains in the west to the U.S. Midwest in the east. Subsidence rates are typically around \(-2\) \(\text{mm yr}^{-1}\), but a local bowl of nearly \(-6\) \(\text{mm yr}^{-1}\) is found centered on western Wisconsin. Our vertical velocity field contains the same large-scale lateral features as in previous results (Alinia et al., 2017; Argus & Peltier, 2010; Calais et al., 2006; George et al., 2012; Goudarzi et al., 2016; Mazzotti et al., 2011; Park et al., 2002; Peltier et al., 2015; Sella et al., 2007; Snay et al., 2016; Tiampo et al., 2012), but our result suggests that the belt of maximum subsidence has a larger extent and more contiguous nature.

In the far field, our results are consistent with the subsidence seen by others along the Atlantic Coast (Karegar et al., 2016). Our results also confirm the well-documented very fast (~30 \(\text{mm yr}^{-1}\)) and localized subsidence near Mexico City (Cabral-Cano et al., 2008; López-Quiroz et al., 2009; Ortega-Guerrero et al., 1999; Osmanoğlu et al., 2011; Yu & Wang, 2016). There is also significant subsidence along the coast of the Gulf of Mexico from southeastern Texas to the mouth of the Mississippi River. This conforms to previous geodetic studies (Dokka et al., 2006; Ivins et al., 2007; Kearns et al., 2015; Yu & Wang, 2016). Dokka et al. (2006) and Yu and Wang (2016) show that the subsidence in Louisiana may be associated with crustal extension due to coastal sites moving toward the Gulf. Although we do find crustal extension there, especially in southeast Texas, southeast Louisiana, and in the Florida Panhandle (Figure S6), it is not significantly resolved.

5. Discussion

We derive a self-consistent vertical velocity field, horizontal strain rate model, and horizontal intraplate velocity field. Although they are derived independently, they are interconnected and reflect the same process. This is illustrated in Figure 8, which shows a transect across the Great Lakes along the 84°W meridian (see Figure 6 for location) between latitudes 38°N and 53°N. It shows that the “hinge line” for both the vertical and northward velocity is located at \(\approx 45°\)N (Figures 8b and 8c). This hinge line is near the maximum N-S gradient in the horizontal strain rates, and the area of substantial contractional strain rates (between 41–47°N) coincides with the main gradient in horizontal velocities (Figures 8a and 8b). Strain rates become dilatational north of \(\approx 48°\)N where the vertical rates become large (i.e., >4 \(\text{mm yr}^{-1}\)). These self-consistencies give confidence in our combined interpretation below.

5.1. GIA

The observed intraplate strain rate and 3-D velocity field are consistent with the expected signal of ongoing GIA. For the strain rate field, this includes the observations that (1) the contractional belt follows the edge of the ice sheet during the Last Glacial Maximum, (2) extension lies underneath areas of maximum glaciation during the Last Glacial Maximum, (3) most areas of significant strain rate are nearly entirely dilatational (i.e., do not exhibit shear strain), and (4) principal contractional strain rate in the aforementioned belt, as well as in the far-field regions, is predominately oriented toward the former ice sheet. Observations in the horizontal intraplate velocity field in the North America no-net rotation frame include (1) the orientations of far-field velocities exhibit a radial (inward) pattern toward the former ice sheet, (2) the zero-velocity line follows the contractional strain rate belt very closely, and (3) velocities for places underneath the eastern extensional center are radially outward. Features of the vertical velocity field include the observations that (1) maximum uplift is in the areas of maximum glaciation, (2) subsidence around the former ice sheet can be related to forebulge collapse, and (3) the hinge line of zero vertical velocity follows the line of zero horizontal velocity almost exactly (at least in continental North America).

To quantitatively assess the similarity between our result and models of GIA, we compare our results with the publicly available ICE-6G_C(VMSa) model (the “GIA model”; Argus et al., 2014; Peltier et al., 2015). The GIA
model's vertical/horizontal velocity and horizontal strain rate are shown in Figure 9 (with the strain rates calculated from MELD using ~1° gridded velocities). For a fair comparison with our results, we also interpolate the GIA model's vertical and horizontal velocities to the station locations used by us, and estimated the GIA model's vertical velocity image and horizontal strain rate field (using the same parameters for MELD as was used in our model; Figures 10a and 10b). The same predictions are shown in the transects of Figure 8.

5.1.1. Vertical

The GIA model gives three uplift maxima (east, west, and north of Hudson Bay; Figure 9a), which could be recovered with the station distribution, although the latter two blur together and the area of maximum uplift west of Hudson Bay may appear smaller (Figure 10a). Large predicted offshore subsidence along the Baffin Bay and Davis Strait (Figure 9a) cannot be recovered by the network (Figure 10a), but the subsidence bowl in the Gulf of St. Lawrence can be resolved and is indeed confirmed by our result. Figures 10c and 10d show our vertical rate (Figure 6) and horizontal strain rate (Figure 4a) results minus the predictions shown in Figures 10a and 10b, respectfully. The agreement in the vertical rate is generally very good (see also transect in Figure 8c); median difference based on a station comparison is +0.3 mm yr$^{-1}$. This consistency is expected given that the GIA model was constrained by a large subset of the GPS data presented here. The difference of 0.3 mm yr$^{-1}$ could be related to the uncertainty in the connection between the IGS08 frame and the center of mass, which we have ignored, and which for ITRF2014 was most recently put at 0.30 ± 0.18 mm yr$^{-1}$ in the $z$ component (Altamimi et al., 2017).

One major discrepancy between observed and expected vertical rates exists for Greenland, where large uplift due to present-day melting (Bevis et al., 2012; Jiang et al., 2010; Khan et al., 2007) is not predicted by the GIA model (median station difference there is +5.5 mm yr$^{-1}$, less than the 8.1 mm yr$^{-1}$ difference reported by Bevis et al. (2012) using the ICE5G (VM2) model). Indeed, the observed and predicted vertical rates agree only for northeastern-most Greenland, where ice loss is minimal, as inferred from the Gravity Recovery and Climate Experiment (GRACE; Velicogna et al., 2014).

The GIA model underestimates the subsidence observed south of the former ice sheet, from the Canadian Prairies (west of Lake Winnipeg) to the Great Lakes, with a maximum in western Wisconsin. Similar discrepancies were found by similar comparisons (Peltier et al., 2015; Snay et al., 2016). Using GRACE data, several reports conclude that the larger-than-predicted subsidence is due to a post-2005 increase in terrestrial water storage following a major drought in the Canadian Prairies (Lambert et al., 2013; H. Wang et al., 2013, H. Wang et al., 2015). Simon et al. (2017) conclude that it is unlikely that this observed subsidence would arise from any of the assumed ice-history and mantle viscosity models. While terrestrial water storage increase may be an important factor in explaining subsidence in the Canadian Prairies, we point out that south of the U.S. Canada border, where no drought occurred, the observations and GIA model see subsidence as well, with two maxima in eastern Montana and western Wisconsin. Although there is a spatial correlation between model and observations, the observed subsidence, particularly in western Wisconsin, is much larger than predicted. Possibly most pertinent to the question whether the significant subsidence, particularly south of the border, could be explained by GIA, is the spatial coincidence of this belt with a belt of elevated contractional strain (discussed below).

5.1.2. Horizontal

The GIA model predicts horizontal contraction in the far-field and extension underneath the former ice sheet, with the former directed radially inward and the latter mostly oriented NE-SW (with small biaxial strain in both). To first order, our results agree with that pattern, but some major differences exist as well. Most importantly, the GIA model does not predict the relatively narrow belt of elevated contractional
strain rates we observe, with the exception of the contraction we see in the Maritime Provinces. The absence of the contractional belt in the GIA model is due to an absence of significant inward horizontal motion of the far-field. The small, southward increasing, northward motion in the GIA model of areas south of the ice sheet is actually almost entirely due to rotational feedback (W.R. Peltier, personal communication, 2017), which is the process that predicts the surface motion related to the shifting of Earth’s rotation axis during a glacial cycle.

The GIA model predicts significant contraction between Baffin Island and Greenland when based on predicted velocities on a regular grid (Figure 9b), but not when using predicted velocities at the GPS locations (Figure 10b). The former also shows extension strain rates along Greenland’s west coast, which, together with the offshore contraction seem be caused by the rapid change in velocity from Baffin Bay to Greenland in the GIA model. For now we will only consider the GIA strain rates based on the velocities predicted at the GPS stations which do not show the significant contractional strain rate between Baffin Island and Greenland (Figure 10b). We do not think that the observed contraction between Baffin Island and Greenland is due to anomalous motion of Greenland related to present-day ice melting, as is known to explain the discrepancy in vertical rates on Greenland. If that were the case, we would expect to see a contractional strain rate belt all around Greenland, and extension of Greenland itself, but that is not observed.

On the other side of the ice sheet from Baffin Bay is the other main strain rate feature not predicted by the GIA model. The GIA model only predicts some elevated strain rate near the Wisconsin-Minnesota border, but the observed contractional strain rate belt south of the former ice sheet is otherwise not predicted by the GIA model. This anomalous feature coincides with the significant subsidence that is underpredicted by the GIA model. The localized contraction is associated with the significant inward motion of the far-field south of the belt. While the subsidence, or at least part of it, could be caused by increased water storage, it is hard to understand how that would also cause such a large horizontal response in the far-field, because horizontal displacements from elastic (un)loading should decrease to zero in the far-field (Bevis et al., 2016; Wahr et al., 2013).

Figure 9. (a) The colors and vectors are vertical and horizontal velocities, respectively, predicted by the ICE-6G_C(VM5a) model (Argus et al., 2014; Peltier et al., 2015); (b) the colors indicate the dilatational strain rates, and the arrows are principal strain rate axes for ICE-6G_C(VM5a) inferred from using median estimate of local deformation on an ~1° grid of predicted horizontal velocities.
The absence of the contractional belt through New England, seen in both the observed and predicted fields, may be the result of former ice sheet, and consequently the hinge line in vertical rates, to be closer to the coast there. This limits the ability to see the horizontal strain rate because farther outward from the New England coast lies only one location with GPS constraints: Bermuda. Bermuda seems to move slightly slower than

**Figure 10.** (a) Map of vertical rates for ICE-6G_C(VMSa) inferred from Global Positioning System imaging using predictions at the same stations used to obtain the observed vertical rates in Figure 6; (b) same as Figure 9b, but now for using predicted horizontal velocities at the same stations as used to obtain observed dilatational field in Figure 4; (c) differential vertical rate field, that is, Figure 6 minus Figure 10a) differential strain rate field, that is, Figure 4a minus Figure 10b.
expected in the context of the rest of the plate. But even if Bermuda would move slightly faster, the lack of any other stations between Bermuda and New England would smear the contraction out over a large distance. The motion of northernmost Alaska fits the radial inward pattern observed for the rest of the plate. It could be argued that the motion of northern Alaska is a result of the long-lasting and far-reaching effect of postseismic relaxation shown after the Great 1964 Alaska earthquake (Suito & Freymueller, 2009). However, because the observed motions there are consistent with those in the northwestern part of the plate, which are in turn consistent with large-scale inward gradients in horizontal velocities across the plate, viscous rebound due to GIA rather than postseismic relaxation may be a more likely explanation. As a corollary, given our overall results, postseismic deformation models should consider that not all seemingly anomalous far-field motion can be contributed to postseismic relaxation.

Although our vertical rate field does not convincingly reveal two uplift domes west and east of Hudson Bay, our strain rate field does show two distinct extensional centers that coincide with the uplift and that are predicted by the GIA model. However, this is only clear in Figure S3 because the dilatation rate in the western center is barely significant at the 2σ level. It is, however, significant and regionally consistent at the 1σ level. The reason that the strain rate field may see the western center better than the vertical rate field is because strain rate is less sensitive to station sampling; the uplift can only be seen if you measure right above the uplift, whereas the strain rate estimate relies on having stations around the signal of interest. This is a clear example of how horizontal strain rates may bring unique constraints to modeling GIA. In any case, the two extensional maxima are consistent with the GRACE observations of there being two uplift domes (Paulson et al., 2007; Peltier et al., 2015; Tamisiea et al., 2007; Zhao, 2013). It has, however, been questioned to what extent that result is affected by the data period considered and the effect of present-day water storage changes (Sasgen et al., 2012; van der Wal et al., 2008).

5.1.3. Outlook

While the vertical velocity field for ICE-6G-C(VM5a) is generally supported by the observed vertical rate field, the horizontal field is not, even though we argue that the horizontal field reflects the same underlying cause as the vertical field, and at some places, it does so with less ambiguity. Most importantly, VM5a, which assumes a rather isoviscous mantle, does not predict significant inward horizontal motions (and therefore also not the contractional strain rate belt). However, several studies have shown that significant horizontal inward motions are predicted for GIA models that have a relatively strong lower mantle (James & Morgan, 1990; Mitrovica et al., 1993, 1994; Sella et al., 2007). Lateral variations in mantle viscosities and/or lithospheric thickness predominantly affect horizontal motions as well (D’Agostino et al., 1997; Gasperini et al., 1991; Giunchi et al., 1997; Latychev et al., 2005; Wu, 2005). Earth models with a stronger lower mantle and a weaker upper mantle than in VM5a can still fit the vertical rates as well (Mitrovica & Forte, 1997, 2004), but having a high-viscosity lower mantle would be more consistent with various independent geodynamic inferences (King & Masters, 1992; Mitrovica & Forte, 2004; Nakada et al., 2015; Panasyuk & Hager, 2000; Richards & Hager, 1984).

We argue that our plate-wide 3-D results can provide unprecedented constraints on the GIA process in North America. Far-field horizontal velocities and the location of reversal in sign of horizontal motion relative to the ice sheet have a particular sensitivity, which is lacking in the vertical rates, to constrain the lower mantle viscosity as well as lateral viscosity and lithospheric thickness variations (Wu et al., 2010). However, a detailed analysis of the implications of our results for Earth’s viscosity structure is beyond the scope of the present study.

5.2. Seismicity

Even though our strain rate model has an ~100-km resolution in about all of the conterminous United States within intraplate North America, few coherent strain rate features of that (or larger) size are resolved beyond 2σ uncertainty. We resolve no strain accumulation at most of the notable areas of localized (instrumental and/or historic) seismicity such as the NMSZ (Johnston & Schweig, 1996; Kelson et al., 1996; Page & Hough, 2014; Tuttle et al., 2002), the Eastern Tennessee Seismic Zone (Powell et al., 1994; Warrell et al., 2017), and the Charleston, South Carolina area (Obermeier et al., 1985; Talwani & Cox, 1985). The only exception may be the Meers fault in Oklahoma. It is a reverse fault (Crone & Luza, 1990; Madole, 1988; Myers et al., 1987) and is located at the northern end of the only significant contractional strain feature that we find in the conterminous United States. For the NMSZ, our result of insignificant strain agrees with that of, for example, Calais et al. (2005) and Craig and Calais (2014) and disagree with that of, for example, Smalley et al. (2005).
and Frankel et al. (2012). It is possible, that the findings of the latter studies were influenced by one or more outlier velocity data. In any case, it may possible that there exist significant strain rate anomalies at scales less than 100 km, which our model was not designed to detect, although we note that uncertainties increase with decreasing spatial scale.

To explore the relationship between strain rates and seismicity more quantitatively, we compare the geodetic moment (or potency) and the number of events. For active boundaries those two parameters are strongly correlated (Kreemer et al., 2002; Shen et al., 2007), and this is expected when the seismicity in the area follows the same Gutenberg-Richter relationship (i.e., same $b$-value and maximum expected magnitude; Molnar, 1979). For such an analysis the number of events needs to be independent and the catalog therefore needs to be declustered. The construction of the catalog we use is described in Appendix B. We convert the strain rates to geodetic potency by multiplying the local area of each estimate with the minimum strain rate: $\max(|\hat{\varepsilon}_1|, |\hat{\varepsilon}_2|)$ (Ward, 1994), which is equivalent to $0.5|\hat{\varepsilon}_{ee} + \hat{\varepsilon}_{nn}| + \sqrt{0.25(\hat{\varepsilon}_{ee} - \hat{\varepsilon}_{nn})^2 + \hat{\varepsilon}_{en}^2}$ as independently proposed by Holt et al. (1995). By considering the geodetic potency we do not have to assume the seismogenic thickness or shear modulus needed for calculating the geodetic moment, except to assume that those parameters are constant throughout our study area, which is an approximation. Figure 11a shows the geodetic potency superimposed with the 66 identified independent earthquakes with $M_w \geq 5.0$ since 1964.

**Figure 11.** (a) Geodetic potency (i.e., area times $\max(|\hat{\varepsilon}_1|, |\hat{\varepsilon}_2|)$ plus epicenters of independent earthquakes with $M_w \geq 5.0$ since 1964 within our study area, excluding events near mid-Atlantic Ridge and Cayman Trough; (b) normalized cumulative geodetic potency (red) and number of events (blue) after evaluation points are ordered from lowest to highest potency, and then this sorted sequence is divided in five bins with equal area. The gray dashed line shows unity, which implies that geodetic potency is equal everywhere or that the events are homogeneously distributed, respectively; (c) normalized cumulative number of events versus normalized cumulative geodetic potency (red) compared to unity (dashed gray), which would be expected if the two quantities are correlated, as is observed in active plate boundaries.
We then sort all evaluation points from lowest to largest geodetic potency and then divide this distribution in five parts, all with equal area size. For the five bins we then cumulatively add the geodetic potency as well as the number of events, and then normalize them (Figure 11b). If geodetic potency is constant throughout the study area, then the graph for the cumulative geodetic potency relative to an increasing area should be a straight line. For any heterogeneous strain rate distribution, this graph would be concave upward, which we observe. The graph with the cumulative number of events, on the other hand, closely follows the unity line, indicating a homogeneous spatial distribution in earthquakes. When plotted against each other (Figure 11c), geodetic potency and number of earthquakes thus do not indicate a one-to-one relationship. Instead, we see that slowly deforming areas are relatively more seismically active than areas with higher deformation rates.

We find quantitatively for the entire plate what Calais et al. (2016) had suggested based on a visual inspection of the central and eastern United States that the levels of strain accumulation and seismic activity seem anticorrelated. This suggests that (1) intraplate seismicity does not reflect the release of geodetic strain, and (2) the largest, GIA-controlled, strain rate does not load faults, except perhaps in zones of weakness such as continental margins. These results are consistent with a growing consensus that much of intraplate seismicity can be explained by transient stress loading, triggering the release of strain on weak stress-localizing faults that has been stored over a much longer time span (Calais et al., 2016; Craig et al., 2016; M. Liu et al., 2011; M. Liu & Stein, 2016; Mazzotti & Townend, 2010; Stein et al., 2009; Talwani, 2014). Our results suggest that the transient stress loading likely does not come from GIA. Moreover, the transient and long-term stress fields may only interfere constructively at some places and not at others (Keiding et al., 2015). We note, for instance, that the only area around the former ice sheet that shows predominant shear strain (i.e., the area between the Great Lakes and the Canadian Maritimes; see Figure 2) is also the area where the state of stress is anomalous (i.e., optimal for thrust rather than strike-slip faulting; Hurd & Zoback, 2012) and seismicity the most active (Adams & Basham, 1989). We therefore emphasize that some seismicity, such as that between the Great Lakes and the Canadian Maritimes or in Baffin Bay, may be controlled by GIA, but that it may require some optimal relationship between orientation/style of stress/strain and the orientation of fault (or zones of weakness), and our results suggest that the areas for which this applies are relatively few.

We note that we only investigate a relationship with earthquakes with $M_w \geq 5.0$, which may require larger stress perturbations to occur than smaller earthquakes. In any case, the anticorrelation between seismicity and strain rate seems to persist for lower magnitudes, as it was noted for the central and eastern United States based on smaller events (Calais et al., 2016). Finally, given the spatial resolution of our results, it is possible that seismicity is related to strain concentrations at a scale $<100$ km; however, we make the case that the detection of significant strain rates at such small scales will be difficult.

5.3. Plate Motion

From the model presented above, we obtain a counter clockwise net rotation of North America in the IGS08 reference frame of $0.2010 \pm 0.0154$° Myr$^{-1}$ and an Euler pole located at 2.3°S and 86.0°W. The associated Cartesian rotation rates and standard deviations are $\omega_x = 0.0139 \pm 0.0082$° Myr$^{-1}$, $\omega_y = -0.0204 \pm 0.0154$° Myr$^{-1}$, and $\omega_z = 0.0080 \pm 0.0129$° Myr$^{-1}$. The standard deviations for each component are defined by

$$\sigma_i = \frac{\sqrt{\sum A_i^2 (\hat{A}_i - \hat{A}_{NR})^2 / \sum A_i^2}}{\sqrt{\sum A_i^2}}$$

where $\sigma_{NR}$ is the net rotation and $A_i$ and $\hat{A}_i$ are the area around and median rotation for each evaluation point $i$ that has a model estimate. Our net rotation estimate is nearly identical for cases where we use different spatial density for grids of evaluation points, different values for $N_{min}$ and $\sigma_{max}$ or different velocity data fields from the one preferred (e.g., an MSF version or outliers included).

A 3-D comparison of the 95% confidence ellipsoid with those of other estimates is shown in Figure 12. Our estimate is very close to that of Kreemer et al. (2014), which was based on the velocities of 10 long-running stations distributed roughly equally across the entire plate. They chose that geometry to not favor any particular part of the plate and essentially average any systematic intraplate motions. In doing so, they indeed approximated the net rotation that is more formally calculated here. Our result is also consistent with that of Argus et al. (2010), which was given relative to TRF2005 and which we assume here to be approximately equal to IGS08. In fact, for consistency, we plot the GEOD2005 version of GEODVEL, because GEOD2005 allowed for no drift in the frame origin (although we find consistency with GEODVEL as well). Our result is inconsistent with that of Altamimi et al. (2012) and Blewitt et al. (2013), which is likely because they chose
only stations they thought were unaffected by GIA, that is, only stations south of the former ice sheet, which results in a systematic difference between their result and ours. At this point it is not possible to compare these various results directly with long-term estimates, which are based on sea-floor spreading rates that average over many glacial cycles, but which only reflect relative plate motions.

The rotation estimates of the studies emphasized in the comparison above are frequently used in other studies, in particular to provide a “fixed-plate” reference frame in which to interpret geodetic velocities in the plate boundaries surrounding the North American plate. Those velocities differ depending on which plate rotation is assumed. Moreover, those differences will depend on where the velocities are evaluated (e.g., Mexico versus Alaska versus Iceland). A detailed comparison is beyond the scope of this study. Such a comparison is complicated by the fact that our rotation vector does not represent a rigid-body rotation, but a net rotation, such that points inside the plate can still have nonzero motions (as we show), and, if GIA is the cause of them, sites outside the North America plate will reflect not only active tectonics but also GIA.

6. Conclusions
We present a new horizontal strain rate field, horizontal intraplate velocity field, and vertical velocity field for almost the entire intraplate area of the North American plate. The horizontal and vertical results complement each other in that they both reveal a systematic deformation pattern that is most likely controlled by GIA. Most notably, we observe a semianular belt of significant contractual strain rate surrounding the area formerly covered by ice, with the latter area undergoing extension. The contractual strain rate belt is more the result of the far-field moving toward the former ice sheet than it is the result of a radial outward motion of the area of the former ice sheet.

The contractual strain belt and far-field motion are not predicted by the ICE-6G_C(VMSa) GIA model. This suggests that the Earth model used in this particular GIA model can be improved to better fit the horizontal data. Moreover, geodetic measurements of the North American plate as a whole do not show that it rotates rigidly, rather, its rotation vector varies with location. Finally, we find no spatial correlation between strain rate and seismicity.

Appendix A: Outlier Detection
Horizontal outlier velocities are removed with a new iterative outlier detection algorithm. In this algorithm we identify for each station its direct neighbors as defined by a Delaunay triangulation (Renka, 1997). The velocities at those neighboring stations are used to estimate the associated strain and rotation rate in a least squares sense (weighted by the velocity data variances) using the same formulation as presented in section 2. From these model parameters we then estimate the predicted velocity at the location of the station being tested and compare it with the observed value. This is repeated for all data points (except those on the convex hull of the set of points, for which this approach does not work), and we sort the velocity misfits from large to small. In that order, any station with a misfit larger than a threshold (discussed below) is flagged, except if that station’s neighbors include one of the stations already flagged prior in that iteration. A next
Acknowledgments


Appendix B: Earthquake Catalog

We create our declustered earthquake catalog by taking the ANSS catalog from 1964 until 2008, not later because of the onset of significant induced seismicity in the United States (Ellsworth, 2013). Because some of the events in this time period could be aftershocks from large 19th century main shocks, we added to the Advanced National Seismic System catalog the U.S. Geological Survey catalog for the central and eastern United States (Petersen et al., 2014). The combined catalog is then declustered with the algorithm of Zaliapin et al. (2008). To exclude possible plate boundary earthquakes, we did not consider any events within 3° of the MAR or near the Cayman Transform fault. Because the declustered catalog needs to be complete for our analysis of the entire intraplate area, we only used events from 1964 and with a moment magnitude of at least 5.0. This completeness magnitude was taken from Teza et al. (2017) for the Atlantic Ocean, where the completeness is arguably the largest for all of our study area. The resulting catalog has 66 events.

References


