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Improving total field geomagnetic secular variation modeling from a new set of cross-over marine data

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ABSTRACT

A new set of cross-over marine data has been used to generate a regional model for the secular variation of the total geomagnetic field, showing the potential of the suggested approach for gaining a better knowledge of the field over oceanic regions. The model, which is valid for the Northern Atlantic region during the temporal interval 1960–2000, was obtained using spherical cap harmonic analysis (SCHA) in space and penalized splines in time. The maximum spatial expansion is equivalent to degree 9 in ordinary spherical harmonic analysis. Annual mean intensity data from different geomagnetic observatories have been used to improve the spatial and temporal resolution of the original dataset. Results indicate that the regional model improves, in terms of the root mean square error, the prediction given by the 11th generation of IGRF and CM4 global models, especially for the geomagnetic observatories considered. We also provide the uncertainty of the model coefficients and the secular variation prediction given by a bootstrap algorithm. The model is available in the EarthRef. org Digital Archive at http://earthref.org/ERDA/1728/.

1. Introduction

Studies of geomagnetic field models covering the past the century have revealed regions of particularly weak geomagnetic field intensity. One of these regions corresponds to the so-called South Atlantic Anomaly (SAA) located over the South Atlantic Ocean, with the lowest magnetic minimum in southeastern Brazil. In this region the magnetic field intensity is about 30% of the values observed over the polar regions, and around 75% of the field in the equatorial regions. At the top of the outer core, in a region currently located under South Africa, there is evidence of magnetic flux patches of reversed polarity (Barraclough et al., 1975, 1978; Gubbins and Bloxham, 1985; Hulot et al., 2002) that can be related to the SAA at the Earth's surface. In the South Atlantic and other oceanic basins (e.g. the westernmost part of the North Atlantic Ocean) the first time-derivative of the geomagnetic field, known as secular variation (SV), is also strongly decreasing.

At present, snapshots of the intensity of the geomagnetic field over the Earth's surface are given by global models such as the IGRF every 5 years (Finlay et al., 2010), although continuous models such as the CM4 comprehensive model (Sabaka et al., 2004) are also available. Errors computed from such global models vary considerably with position, reflecting the poor distribution of observatory data upon which models rely onto predict SV, in comparison with satellite data (Langel, 1987). Most of the information concerning SV of the geomagnetic field comes from magnetic observatories and through global networks of sites, known as repeat stations, which are observed periodically. They yield datasets with a distribution that is severely limited in the oceans. Recently, satellite data has become available that covers oceanic regions, but between 1980 and 2000 precise three-component magnetic field measurements over oceans were rare. It is also worth noting that although continuous measurements are now available for these regions for the last decade, they cannot enhance our knowledge of the SV over periods of several decades (Ravat et al., 2003). Here, we suggest a possible approach to refine our knowledge about SV in oceanic regions. In addition to the use of observatory and satellite data, we propose the use of pairs of total magnetic field observations from marine cruises at cross-over points which, in theory, should differ only by their time-varying components (Verhoef and Williams, 1993).

In an attempt to characterize the geomagnetic field structure and evolution in greater detail over those particular regions, we have examined the possibility of developing new SV models over the oceans using regional analytical techniques. This includes the

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use of spherical cap harmonic analysis (or a recently revised version of this technique) in space and penalized B-splines functions in time, by which the SV is modeled by taking differences relative to the means over each observing location. The potential of this approach has been demonstrated by producing a SV model of the total field over the entire North Atlantic, which is perhaps the basin best covered with marine magnetic data and which includes a rapidly evolving flux feature.

2. Data

The publication of the second version of the World Digital Magnetic Anomaly Map (WDMAM) involved the collection of more than 2400 cruises in a period of time which spans from 1960 to 2008 (Quesnel et al., 2009). All this information was carefully checked profile by profile and cleaned by removing spikes and other spurious data. A line leveling method was applied to reduce some inconsistencies between different surveys. The root mean square of the global magnetic anomaly differences at cross-overs was reduced from 179.6 to 35.9 nT which highlights the improvement in quality and coherence of this global marine magnetic data set (Quesnel et al., 2009). We used a subset of the original (i.e. with no core field correction applied) total intensity data set, and took advantage of the cleaning and editing process already carried by Quesnel et al. (2009). This allowed us to infer SV over the North Atlantic as it not only provides a better spatial and wider temporal coverage than that used twenty years ago by Verhoef and Williams (1993), but it is superior in terms of precision. Nevertheless, we limited the time span of our model to 2000.0 since the data density decreases dramatically after this date.

In order to isolate the main field from the total field measurements we subtracted the external magnetic field contributions using the CM4 model. Lithospheric contributions were extracted using the final WDMAM's oceanic magnetic anomaly database, which already included those corrections derived for each marine track-line after leveling the magnetic anomaly data set. Although the process of obtaining the main field differences must ensure that the SV model coefficients are not contaminated by crustal bias (since it is assumed that is independent of time), the subtraction of its estimation was an effective means to reduce noise (the positioning is usually inaccurate at sea so that the magnetic field might have been measured in non-coincident places but, in practice, be referred to the same cross-over point). Once corrected from external fields, the total field magnetic readings at tie-points, lithospheric contribution and long and mid-wavelength errors should represent a good approach to the main field, although they still suffer from two main different sources of error: incompletely canceled external field contributions and navigation errors. It is not easy to quantify their contribution to the error budget. To estimate it in the area under study (North Atlantic Ocean) we computed the rms total field cross-over difference, yielding a value of 43.8 nT. As we used the CM4 model to reduce each pair of magnetic readings to the same epoch, most probably the true uncertainty is smaller than that estimation, and we should consider it as a worst case scenario.

In addition to the marine data, observatory total field annual mean values were added to constrain the model in the peripheral areas. This also provided robustness to our model, because of the high quality of these data in comparison with the marine data. A selection of these data was performed. We used annual mean values taken from (a) WDC for Geomagnetism, Edinburgh, (http://www.wdc.bgs.ac.uk), or (b) derived from monthly mean values from the IPGP database (http://obsmag.ipgp.fr/wmmE.html). We compared those annual mean values relative to the mean (from (a) and (b)) with the corresponding CM4 main field differences. Those time series were thus individually checked to disregard possible outliers from (a) or (b), which were removed but using (a) as master. We substituted some (a) values by (b) values whenever the series were more in agreement with CM4 predictions in terms of rms differences. We considered the two segments separately whenever an observatory change of site took place. Fig. 1 shows the spatial and temporal distribution of both types of data: cross-over marine (41946 pairs) and observatory (752) data.

In order to reject any possible outliers present in the database, we performed an initial model using all the cross-over marine and observatory data. The standard deviation of the residual data (i.e., the difference between the input and modeled data) provided by the initial model was used as the outlier rejection criterion. All data lying outside three times the standard deviation (36.0 nT) were rejected. A total of 1620 pairs of cross-over marine data were rejected using this procedure, and the final number of input data was given by 40326 pairs of cross-over marine data and 752 observatory data.

3. Technique and model parameterization

Given the size of the explored region in the present work we have chosen the spherical cap harmonic analysis technique (SCHA, Haines, 1985). When the region is small and/or data is analyzed at different altitudes, the revised spherical cap harmonic analysis (R-SCHA, Thébault et al., 2006) should behave better. In terms of the SCHA, the potential of the internal geomagnetic field can be established as (Haines, 1988):

$$V(r,\theta,\lambda,t) = a\sum_{k=0}^{K}\sum_{m=0}^{k}\sum_{q=1}^{Q}\left(\frac{a}{r}\right)^{n_k+1} P_{n_k}^{m}(\cos\theta)(g_{n_k,q}^{m}(t)\cdot\cos m\lambda + h_{n_k,q}^{m}(t)\cdot\sin m\lambda) \tag{1}$$

where a is the average radius of the Earth, $P_{n_k}^{m}(\cos\theta)$ the associated Legendre functions with real degree $n_k(m)$ and order m, $\cos m\lambda$ and $\sin m\lambda$ the Fourier functions, and the K and Q indices are the maximum degrees of the spatial and temporal expansions respectively. $g_{n_k,q}^{m}(t)$ and $h_{n_k,q}^{m}(t)$ are the time-dependent *SCH* coefficients.

The total field or intensity F of the geomagnetic field cannot be obtained by taking the derivative of the potential in Eq. (1), contrary to the case of the north, east and vertical components. For this reason, the inverse problem aimed at modeling the intensity data is non-linear and we need an iterative approach. A linear inversion is also possible when using the Taylor series applied to the intensity data. For a certain location $\vec{r}$ and epoch t, the intensity data F can be expressed as a linear function of the *SCH* coefficients using the first order of the truncated Taylor series (see Appendix A):

$$F(t,\vec{r}) = F(t,\vec{r})|_{m=m_0} + \left.\frac{\partial F(t,\vec{r})}{\partial m}\right|_{m=m_0}\cdot(\vec{m}-\vec{m}_0) + \delta(\vec{r}) \tag{2}$$

where the vector $\vec{m}$ contains the *SCH* coefficients and $\partial F(t,\vec{r})/\partial m$ is the Frechet derivative vector for the intensity. The subindex 0 corresponds to an initial reference geomagnetic field model, denoted by the vector $\vec{m}_0$. The last function $\delta(\vec{r})$ represents the crustal field bias (i.e., the magnetic anomaly for that position). The initial reference model $\vec{m}_0$ was set as an axial dipole field with a spherical harmonic coefficient g_1^0 equal to − 30420 nT at 1960.0 with a constant rate of variation of 20 nT/yr. These values were established according to a linear fitting of the first Gauss coefficient of the IGRF model for the time interval 1960–2000.