GEOMAGNETIC FIELD IN CROATIA – THE NEW RESULTS

Eugen Vujic, Giuliana Verbanac
University of Zagreb
Faculty of Sciences, Geophysical Department
Horvatovac bb, 10000 Zagreb, Croatia
eugvujic@ gfz.hr, verbanac@irb.hr

ABSTRACT

After more than 50 years the measurements of the geomagnetic field in Croatia have been renewed. During the time interval 2003-2005 a ground survey of total magnetic field intensity in the middle northern part of Croatia has been performed. We analysed in detail these data by carefully undertaken data reduction using the following observatories: AQU, FUR, NCK and THY, which are placed relatively close to the investigated region. The measurements were reduced to the epoch 2004.5 using the one-minute observatories data. We follow an approach of comparison between the two or more observatory time sequences, when one of the observatories is used as a reference one, with known annual mean values of the total field at all observatories. For the secular variation over the investigated area we have used the first order Taylor polynomial over geographic coordinates, and the secular variations at observatories were calculated by using the monthly mean values.

Keywords: ground survey, data reduction, normal magnetic field, normal field anomalies

1 INTRODUCTION

After more than 50 years the measurements of the geomagnetic field in Croatia have been renewed. During the time interval 2003-2005 a ground survey of total magnetic field intensity in the middle northern part of Croatia has been performed. We analysed in detail these data by carefully undertaken data reduction using the following observatories: AQU, FUR, NCK and THY, which are placed relatively close to the investigated region. The measurements were reduced to the epoch 2004.5 using the one-minute observatories data. We follow an approach of comparison between the two or more observatory time sequences, when one of the observatories is used as a reference one, with known annual mean values of the total field at all observatories. For the secular variation over the investigated area we have used the first order Taylor polynomial over geographic coordinates, and the secular variations at observatories were calculated by using the monthly mean values.

The normal geomagnetic field which is the combination of the main field, remanent and induced crustal field has also been estimated by means of first-order Taylor polynomial as function of the geographic coordinates. It filters out the small scale field variations produced by the near-surface magnetic rocks (see [1], [2]). The polynomial coefficients were calculated by regression with three methods of adjustment: simple and weighted least squares fit and adjustment according to the most frequent value. Each adjustment was tested using the Monte Carlo-type stability tests. In order to obtain the distribution of the anomaly field, the core contributions as predicted by the global model were subtracted from the reduced field values.

2 MEASUREMENTS

Figure 1 shows the distribution of measurement sites and town positions within the investigated region. The measurements were performed using Overhauser effect proton magnetometers, in base or mobile mode (in the interval 2003.72-2005.94). The accuracy of instruments is 0.2 nT, and the mean error of measurements was 1.5 nT.

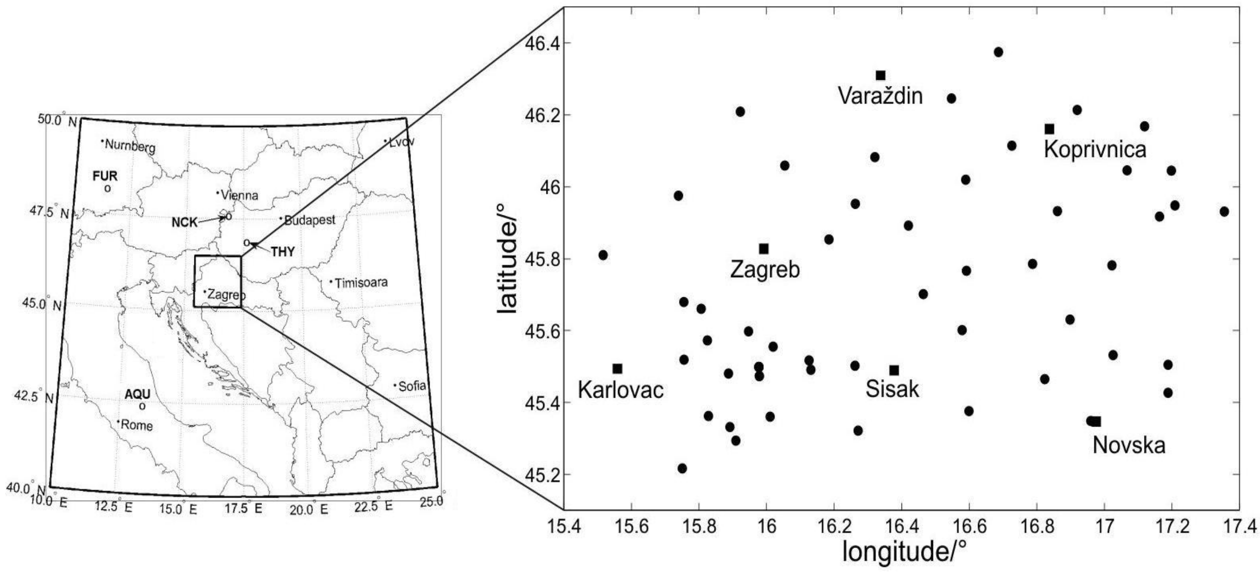

Figure 1: Locations of measurement sites (filled circles), positions of towns (filled squares) and observatories (o). The measurements were performed during autumn 2003, summer 2004, spring, summer and autumn 2005; overall 53 sites.

3 DATA ANALYSIS

3.1 Data reduction

To obtain the true magnetic field value at some epoch, it is necessary to eliminate the transient variations from records. These variations arise from ionospheric and magnetospheric currents, and depend on the geographical position, seasons, solar activity, etc. The variation of the core field on yearly and longer time scales are known as secular variation.

If we assume that the secular variation values are the same at both the observatory and the measurement site, we can apply the following relationship (see [1], [2], [3]):

$$F^{2004.5} = F(t) - F_o(t) + F_o^{2004.5}, \tag{1}$$

where $F^{2004.5}$ is the annual mean of the total field at the measurement site, $F(t)$ is the total field at the measurement site at the time of measurement, $F_o(t)$ is the total field at the observatory site at the time of measurement, and $F_o^{2004.5}$ is the annual mean of the total field at the observatory.

If the values of the secular variation are not the same, Eq.(1) can be corrected (see [1]):

$$F^{2004.5} = F(t) - F_o(t) + F_o^{2004.5} + \int_t^{t+\Delta T} \left(SV(t') - SV_o(t')\right) dt', \tag{2}$$

where SV is the secular variation at the measurement site, SV_o is the secular variation at the observatory, and ΔT is the time difference between the mid-year epoch and the time of the measurement. SV_o can be calculated as the centered difference between two consecutive total field values at some epoch (see [4]):

$$SV_0(t_{i+\frac{1}{2}}) = \frac{F_o^{epoch}(t_{i+1}) - F_o^{epoch}(t_i)}{t_{i+1} - t_i}. \quad (3)$$

The total field values at given epoch have been calculated by applying a 12-month moving average on the monthly mean values. These values are assigned to the epoch obtained by applying a 12-month moving average on the epochs each centered at the middle of the month.

We used the time-independent SV relation which resulted from the Hungarian repeat station measurements at 2003.5 and 2006.5 (*Kovács P., ELGI, personal communication, 2007*):

$$SV(\lambda,\varphi,\boldsymbol{p}) = p_0 + p_1(\varphi - 45.5)\cdot 60 + p_2(\lambda - 16)\cdot 60, \quad (4)$$

where $\boldsymbol{p} = (p_0, p_1, p_2)$ is a parameter vector, p_0=30.30 nT/year, p_1=-0.00862 nT/(year·1'), p_2=-0.00052 nT/(year·1'), φ and λ are the geographic latitude and longitude (in degrees) of the measurement site, respectively.

3.2 The normal field

In order to describe the normal field over the investigated region, the Taylor polynomial of the form (see [2], [5], [6]):

$$F(\varphi,\lambda,\boldsymbol{d}) = \sum_{j=0}^{M}\sum_{k=0}^{M-j} a_{jk}(\varphi - \varphi_0)^j(\lambda - \lambda_0)^k, \quad (5)$$

was used. Here, $F(\varphi,\lambda,\boldsymbol{d}) = F_{normal}$ is the normal field, a_{jk} are constant coefficients, $\boldsymbol{d} = (a,b,c)$ is a parameter vector and the reference values are: $\varphi_0 = 45.8°$, $\lambda_0 = 16.4°$. Since the dimension of the region is approximately 18900 km^2, the first order polynomial (M=1) has been used, with $a_{00} = a$, $a_{10} = b$ and $a_{01} = c$. The three adjustment methods to ground survey data were employed, two least squares: simple (E-fit) and weighted (WE-fit), and the adjustment according to the most frequent value (M-fit).
For the E and WE fits, the multilinear regression of the form:

$$\sum_{i=1}^{N}\left(F_i^{2004.5} - F_{normal}\right)^2 = \min., \quad (6)$$

was used to obtain unknown coefficients a, b and c. N is the number of stations that participate in regression, and $F_i^{2004.5}$ are the reduced total field values at i-th station.

Furthermore, Chauvenet's criterion of rejection has been applied in WE-fit, as follows (see [6]). After the normal field coefficients were computed, the standard uncertainty δ may be obtained from:

$$\delta = \sqrt{\frac{\Sigma}{N-3}}, \quad (7)$$

where Σ is the sum of the squares of the residuals ($rsd = F_i^{2004.5} - F_{normal}$) of the normal field, and 3 stands for the number of unknown parameters. The weight equals to 0 for the sites where the *rsd* was more than 2δ, and equal to 1 otherwise. The procedure was iterated until all of the *rsd* were below 2δ. In this manner some sites from the original data set have been rejected.

In the case of the M-fit, the weighting is ensured by a symmetrical and unimodal function of residuals. This method requires the following condition to be fulfilled (see [6]):

$$\sum_{i=1}^{N} \varepsilon^2 \cdot \ln\left[\varepsilon^2 + \left(F_i^{2004.5} - F_{normal}\right)^2\right] = \min. \tag{8}$$

The quantity ε is called dihesion and it characterizes the scattering of the reduced data around the computed normal field. The dihesion and the *rsd* determine a weight function for each station, and all stations participate in regression.

4 RESULTS AND DISCUSSIONS

The nearest observatory THY was used as reference one, while AQU, FUR and NCK were used for error limits estimation. This estimation is based on a comparison between the two or more observatory time sequences, when one of the observatories is used as a reference one and another as a variometer station, with known annual mean values of the total field at both observatories. In Table 1 the absolute values of the maximal and minimal differences between reduced data of 53 sites are given, when Eqs.(1) and (2) and different pair of observatories were used. Presuming that data reduction errors depend on distance from THY, the results displayed in Table 1 suggest that the errors for $F^{2004.5}$ are not greater than 14 nT.

Table 1: Absolute values of the maximal and minimal differences between reduced data (out of 53 sites). The first two columns are values when (1) has been used, and the last two columns when (2) has been used. In each row is combination of observatories.

Absolute values of the differences in nT	Max. with eq. (1)	Min. with eq. (1)	Max. with eq. (2)	Min. with eq. (2)
$\Delta F^{2004.5}$ with THY and FUR	10.7	0.03	11.3	0.1
$\Delta F^{2004.5}$ with THY and NCK	6.9	0.02	5.3	0.01
$\Delta F^{2004.5}$ with THY and AQU	13.6	0.07	12.1	0.1

Table 2: The results of regression obtained using different methods of adjustment (53 sites). The input data have been reduced values with THY recordings.

Adjustment Method	E-fit	WE-fit	M-fit
Coefficient			
a/nT	47555	47555	47557
b/(nT/°)	321.0	326.1	327.0
c/(nT/°)	59.2	48.9	50.5

Table 2 shows the results of different adjustment methods. They were determined using the data obtained with Eq.(2) and THY recordings. Latitude gradients are higher than longitude ones, as expected in mid-latitude regions. The differences of *b* and *c* in Table 1, respectively, between E-fit and/or M- and WE-fit are so high due to present anomalies in the investigated region. The difference of 2 nT in coefficient *a* (Table 1) between M-fit and/or E- and WE-fit is of the same order as estimated measurement errors, and is negligible compared to spatial changes of the total field

We performed also the Monte Carlo-type stability tests (see [6]), by observing the behaviour of the mean absolute deviations between the original normal field values, and the values of the normal field produced by randomly reduced sets (N=5, 10, 15,…, 50 sites out of 53). The original normal field (Table 1) and the normal field based on the randomly reduced data, gave the deviations of the normal field in the 53 original sites. These deviations are averaged over the 2000 various cases, for each of the randomly reduced data set. The results are shown in Fig. 2. The deviations increase with decreasing the number of data points in obtaining the normal field coefficients. For the randomly reduced number of points at half the mean deviation is less then 8 nT. In this case the M-fit is the most sensitive on how the data set is reduced.

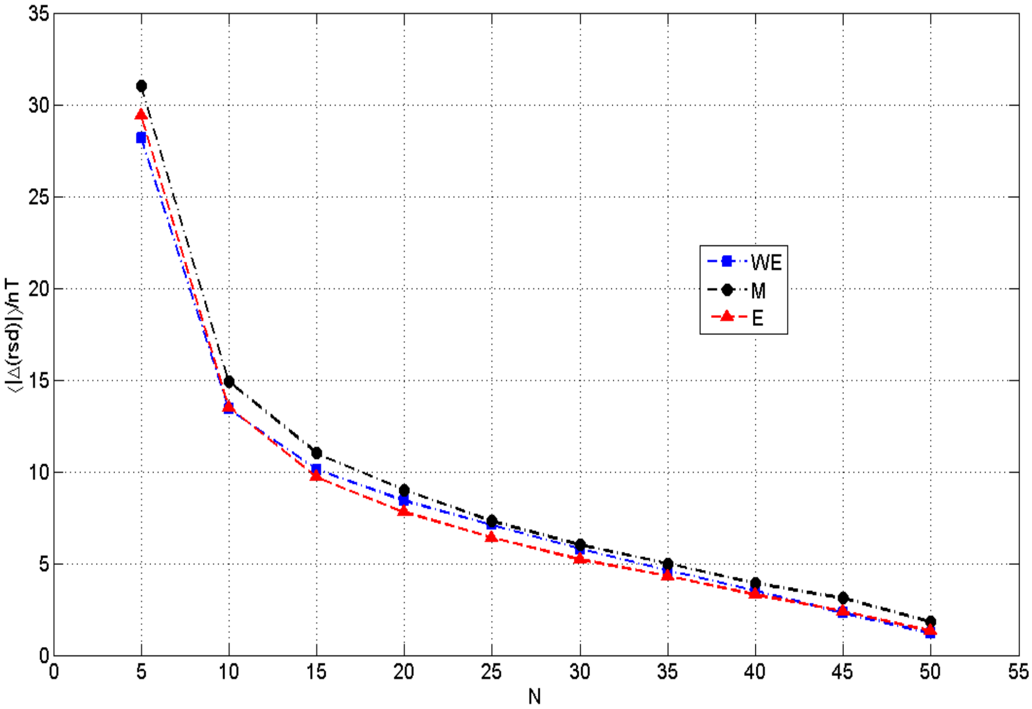

Figure 2: The results of Monte Carlo-type stability tests of the three used adjustment methods: WE-, M- and E-fit. The graph shows the mean absolute deviations between the original normal field values and the values of the normal field produced by randomly reduced sets (N=5, 10, 15,…, 50 sites out of 53).

Figure 3 shows the map of *rsd* of the normal field obtained from 53 values, using the coefficients from the last column in Table 2. Using the other coefficients from Table 2 we obtained the maps which are very similar to shown one, namely the high *rsd* anomalies appeared at the same locations. The strongest anomalies (-106 nT, negative pole of A; 74 nT, positive pole of anomaly D; 57 nT, anomaly B) are caused by near surface occurrence of volcanic rocks.

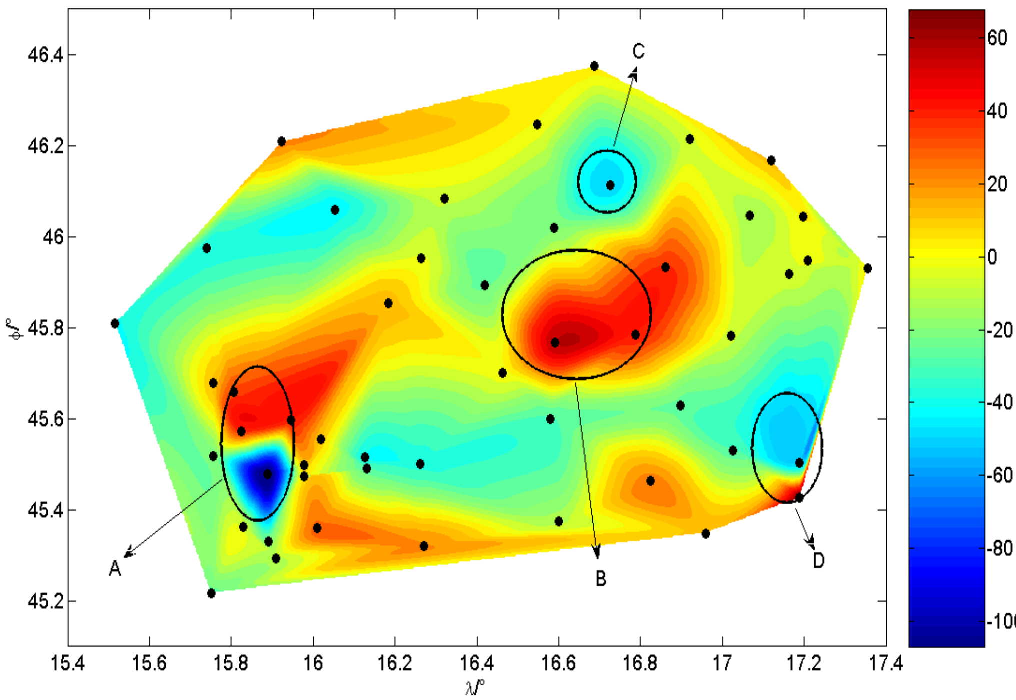

Figure 3: The residuals of the normal total field. The bar scale unit is nT. Survey sites are denoted with filled circles and letters A-D denote the major anomalies of residuals. Geographic latitude and longitude are denoted with Φ and λ, respectively.

The core contributions have been predicted by the IGRF global model (see [7]). In Figure 4. the IGRF *rsd* versus the normal field *rsd* are shown. They were calculated at 53 sites, respectively, and are highly correlated. To obtain the anomalous field we have removed the trend from the survey data: the long-wavelength part of the core field in the case of the IGRF *rsd*; the core field with long- and intermediate-wavelength part of the litospheric field in the case of the normal field *rsd*. These removals are exact only for the X, Y and Z components.

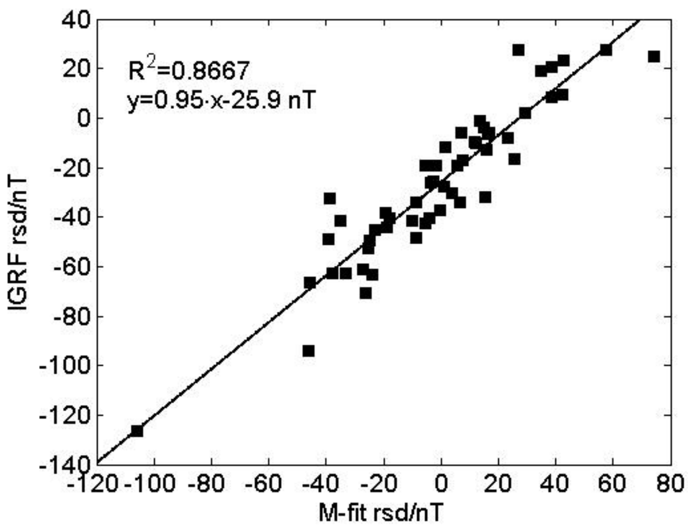

Figure 4: The IGRF residuals versus the normal field residuals at 53 sites. The IGRF field values have been calculated at 157 m above the sea level at 2004.5 epoch.

To obtain the near-surface crustal field over the region we have used the NGDC-720 model (see [8]). This model provides such an expansion for the crustal field from spherical harmonic degree 16 to 720, corresponding to the waveband of 2500 km to 56 km. Figure 5. shows the map of the model over the region. There is obvious west toward east increase in magnitude of the crustal field.

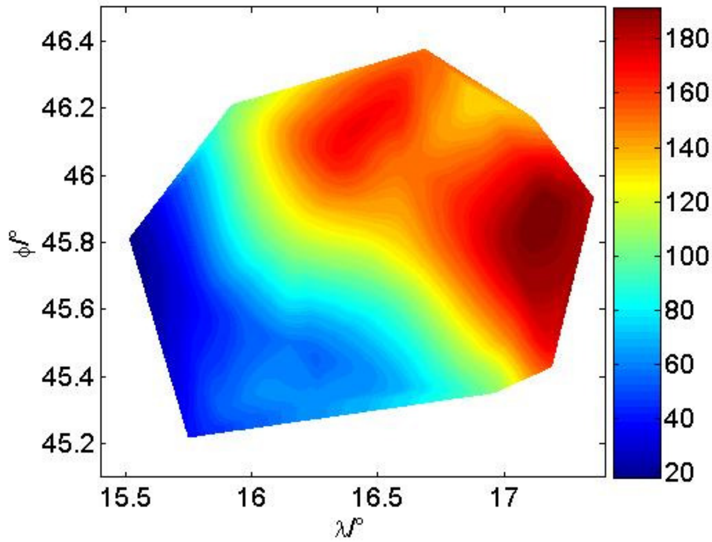

Figure 5: The near-surface crustal field over the region, computed from the NGDC-720 model. Geographic latitude and longitude are denoted with Φ and λ, respectively.

The correlation coefficient between the IGRF field values and the NGDC-720 *rsd* is *CC*=0.809. The results are displayed in Figure 6. These *rsd* comprise approximately the core field and the short-wavelength part of the litospheric field.

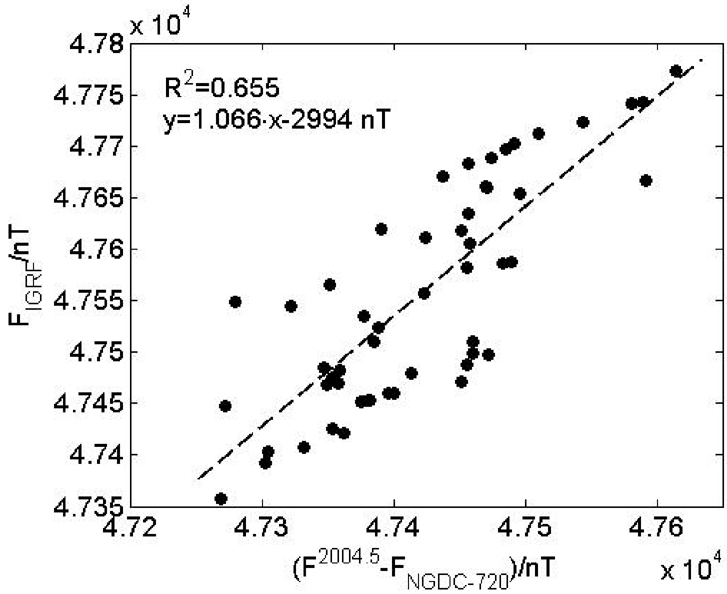

Figure 6: The IGRF field values versus the NGDC-720 residuals at 53 sites.

5 CONCLUSION

The ground magnetic survey was conducted from autumn 2003 to winter 2005 at 53 measurement sites over northern Croatia. We present the method to determine the error limits in the annual mean values of the total field. The nearest observatory THY was used as reference one, while AQU, FUR and NCK were used for error limits estimation. Presuming that data reduction errors depend on distance from THY, we estimate the maximal reduction errors to be one order in magnitude greater than measurement errors (thus one order in magnitude smaller than the local, short-wavelength, anomalous total field values).

Also, the normal field over the investigated Croatian region was obtained. Three methods of adjustment of the first order Taylor polynomial as function of geographic coordinates were used. The values of ratios of the normal field latitude and longitude gradients are as expected, in mid-latitudes. The differences between the free coefficients of the normal field are of the same order as the estimated ground survey measurement errors. The standard deviations of the normal field residuals, obtained by different adjustment methods and input data derived with THY recordings, and those obtained by M-fit and input data derived with AQU, FUR, NCK and THY recordings respectively, have approximately the same values. The corresponding mean deviations of the normal field in these two cases are less than five nT.

Further, an each adjustment was tested with respect to decreasing the input data into a regression (Monte-Carlo type test). We found that if the input data are reduced to a half, the normal field is not stable anymore. Also, we deduced that the M-fit is the most sensitive on reducing data set. The main reason why the mean deviations of the normal field due to the different adjustment methods are not smaller, is a presence of the local short-wavelength magnetic anomalies. These anomalies are announced with the mentioned mean deviations and stability tests, and qualitatively confirmed with a comparison between the geological map and the map of the normal field residuals. However, this result has to be taken with caution, because of the interpolation errors and edge effects. The three strongest anomalies (-106 nT, 74 nT and 57 nT) are caused by near surface occurrence of volcanic rocks. The anomalous field values obtained by the IGRF model and the normal field, respectively, are highly correlated (CC=0.809).

The present study gives the first detail distribution of the total field intensity, as well as the anomaly field over middle northern part of Croatia.

REFERENCES

[1] L.R. Newitt, C.E. Barton, J. Bitterly, *Guide for magnetic repeat station surveys*, International Association of Geomagnetism and Aeronomy, Warszawa, 1996, pp. 53-68.

[2] A. De Santis, M. Chiappini, G. Dominici, A. Meloni, *Regional geomagnetic field modelling: the contribution of the Istituto Nazionale di Geofisica*, Annal. Geophys., **40**, 5, 1997, pp. 1161-1169.

[3] M. Korte, E. Thébault, *Geomagnetic repeat station crustal biases and vectorial anomaly maps for Germany*, Geophys. J. Int., **170**, 2007, pp. 81-92.

[4] A. Chambodut, M. Mandea, *Evidence for geomagnetic jerks in comprehensive models*, Earth Planets and Space**, 57**, 2, 2005, pp. 139-149.

[5] W.J. Xu, G.H. Xia, Z.C. An, G.X. Chen, F.Y. Zhang, Y.H. Wang, Y.G. Tian, Z.G. Wei, S.Z. Ma, H.F. Chen, *Magnetic Survey and ChinaGRF 2000*, Earth Planets Space, **55**, 2003, pp. 215-217.

[6] P. Kovacs, A. Kormendi, L. Hegymegy, M. Chiappini, G. Dominici, *The Hungarian magnetic network and the processing of the obtained data*, Phys. Chem. Earth (A), **24**, 5, 1999, pp. 439-443.

[7] http://www.ngdc.noaa.gov/IAGA.

[8] http://www.ngdc.noaa.gov/geomag.