Wellington Gravity Survey
Part I: Gravity Data Description

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Introduction

The Wellington gravity survey was designed to utilize every 3rd receiver location along the east-west seismic lines of the Wellington 3D seismic survey. Using the surveyed locations of the seismic receivers minimized costs, and using every third receiver location provided a nominally square survey grid having measurements every 495 feet, as the receiver locations were 165 feet apart, and the receiver lines 495 feet apart. All the surveying was conducted by a subcontractor to the seismic company, Paragon Geophysical. When receiver locations were in areas difficult for gravity data acquisition (*e.g.*, in streams or in swampy ground), measurements were taken on either side of the nominal location or, occasionally, at a nearby (surveyed) seismic source location.

Data Acquisition

Two LaCoste-Romberg Model G gravity meters (serial numbers 442 and 562) were used for the survey. Each of these meters has been fitted with electronic levels and an Aliod feedback system. With the feedback system, the meter's counter, which turns the calibrated screw, is never moved, reducing any errors due to screw miscalibration. Data were recorded using Palm PDAs which receive the serial output of the Aliod system via Bluetooth. At a given station, the output of the Aliod system was sampled at 1 Hz until the output stabilized. Typically stabilization takes 40–60 s, although longer recording periods are needed at noisy sites. The last 15 s of the time series is averaged to obtain a gravity value.

Several local gravity bases were used during the survey, mostly to account for meter drift or possible meter tares. These bases were on concrete bridge abutments, which afforded very stable platforms for measurements. All the gravity data were ultimately referenced to a relative base station in Wellington at the former Wellington Post Office (DOD designation 5758-0).

The survey began February 7, 2010 and was conducted in two phases. The first period of data acquisition was from Feb, 7–18, 2010, and the second period was from March 30–April 8, 2010. The break in acquisition was due to unusually wet weather that occurred in Februry and March.

Data Processing

On a given day, drift corrections for gravity stations are calculated by using replicate measurements to estimate a linear drift occurring during the day. The drift estimation procedure does not require that all such replicate measurements be made at a single location: replicate measurements occurring at any stations during the day are utilized to estimate the instrument drift.

The gravity measurements are reduced to Free-air and Bouguer anomalies using the reduction procedures recommended by Hinze et al. (2005), except that, rather than using ellipsoidal elevations as recommended, traditional (orthometric) elevations were used. However, in the Wellington area, there is actually very little difference between the two.

Close-in terrain corrections are calculated using procedures very simlar to those described by Cogbill (1990), and outer-zone corrections are calculated using Plouff's method (Plouff, 1977). The actual terrain data used were the 0.33-sec and 1-sec data from the National Elevation Dataset of the U. S. Geological Survey. However, terrain corrections for the stations acquired during the Wellington survey are extremely small, as the survey area has very low relief.

Estimated Bouguer Density

Ideally, the optimal Bouguer density to use for gravity reductions is estimated by regressing the geometric terrain effect against the free-air gravity anomaly (Appendix I).This method is essentially the same as the Nettleton profiling method (Nettleton, 1939). However, in practice this method rarely works out well, either because (a) the range of elevations encountered in the survey area is too small to provide a reliable regression or (b) the basic assumptions underlying the Nettleton method are not met. The latter typically results when topographic variations are related to underlying geologic changes. The range of elevations encountered in the Wellington survey was only 100 ft, making the use of a Nettleton-type method problematic.

A different method that we have found reliable is to estimate the Bouguer reduction density that minimizes the residuals resulting from fitting a plane to the gravity data. The idea is that, when the areas involved are not too great, the Bouguer gravity field will be approximately planar. Finding the density that provides the best-fitting plane should provide a good estimate of the actual near-surface density. Figure 1 show the results of such a calculation, using all the gravity data acquired during the survey. Using this methodology, a reduction density of 2.45 g/cc would seem best.

Data Display

Figure 2 shows the locations of the gravity stations acquired during the survey, displayed on top of the U.S. Geological Survey topographic base. The nominal station spaing was 495 feet in both the east-west and north-south directions.

Figure 3 is a contour map of the Bouguer gravity anomaly, calculated using a reduction density of 2.45 g/cc. The contour interval used is 0.2 mGal. Figure 4 is a contour map of the residual Bouguer gravity anomaly; the contour interval used is 0.05 mGal. The residual map is calculated by subtracting a least-squares plane fit to the Bouguer data from the Bouguer gravity values. The residuals displayed on this map form the basis for the estimation of static corrections for the seismic data.

Near-surface Densities and Velocities

We have also attempted to estimate the spatial variation in near-surface densities using a variant of the method we used to estimate the overall Bouguer reduction density. In this method, an automatic procedure is used to calculate the reduction density that minimizes gravity residuals from a planar fit to the gravity data **within a window much smaller than the size of the entire gravity survey**. The window is moved

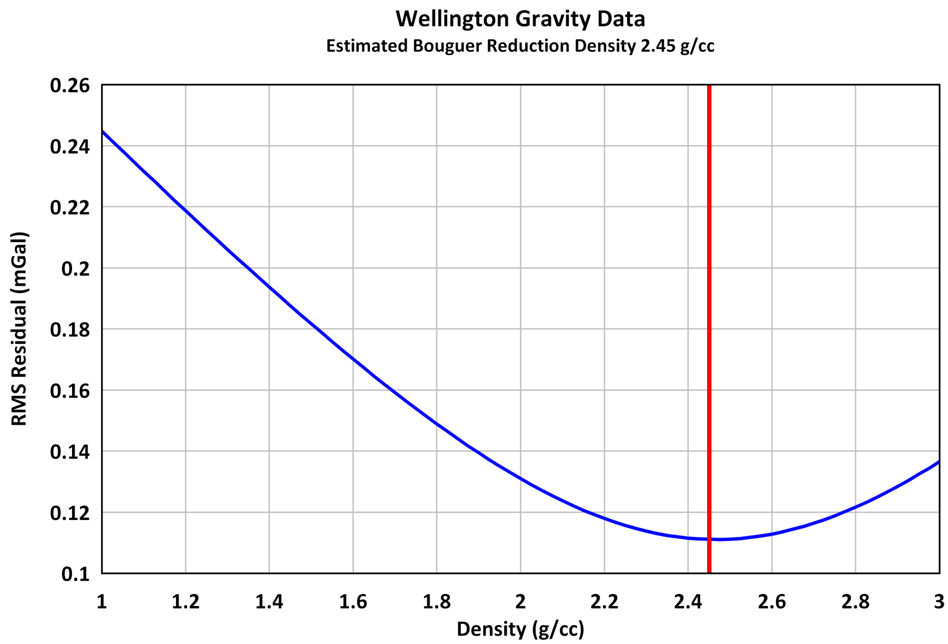

Figure 1: Plot of the residuals from a planar fit to the the observed gravity as a function of the Bouguer reduction density used.

across the entire survey area, and a density estimate at the center of each window is calculated. Near-surface velocities are calculated from the estimated densities using Gardner's equation (Gardner et al., 1974).

Figure 5 is a plot of the estimated near-surface densities, calculated using the methodology described above. A moving window 1220 m (4000 ft) in radius was used in the calculation.

Figure 6 is a plot of the estimated near-surface velocities, calculated from the estimated densities using Gardner's relationship between velocity and density.

Note that Figures 2–6 are supplied electronically as scalable PDF figures.

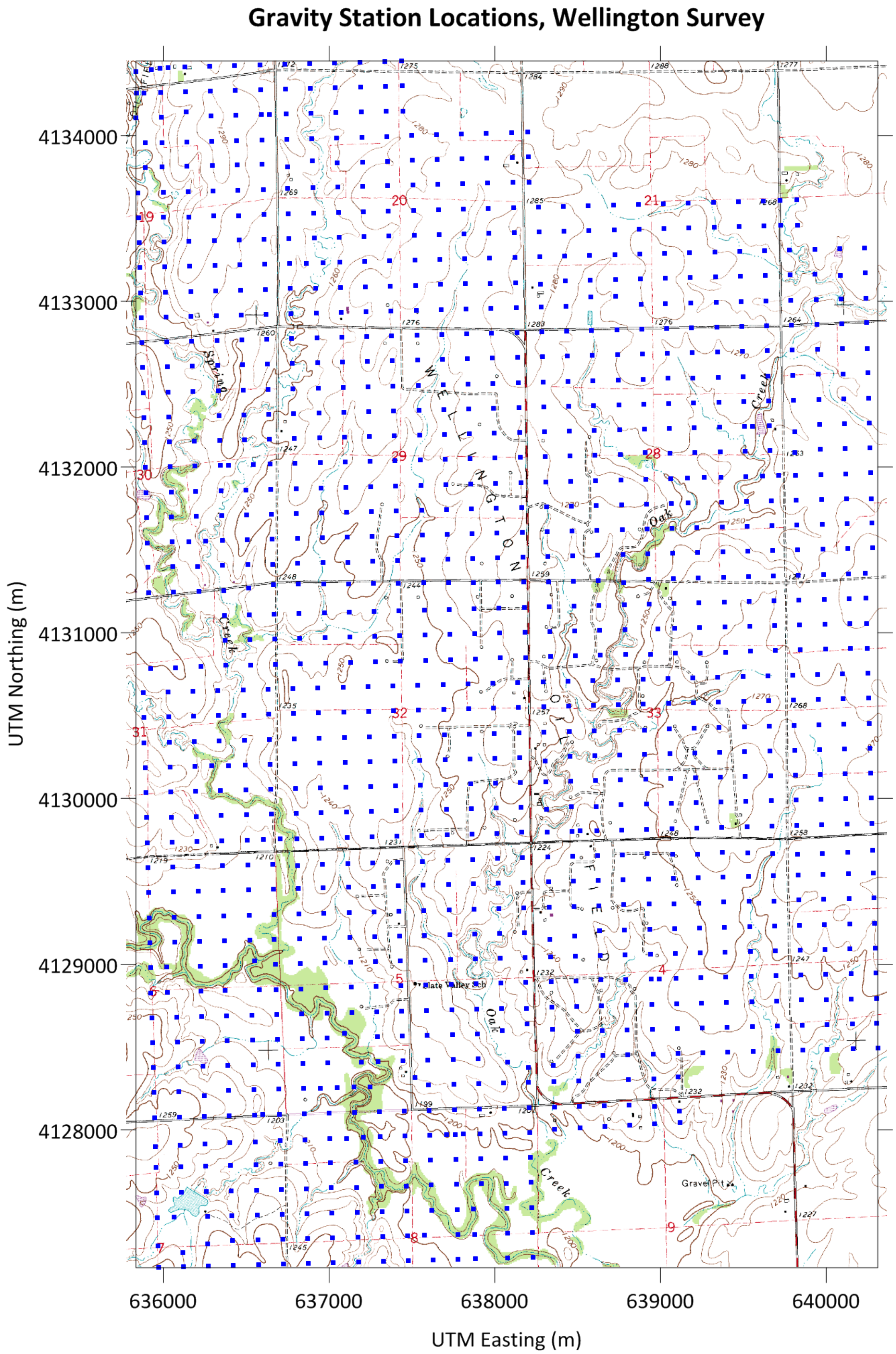

Figure 2: Map showing the locations of the 1290 gravity stations acquired during the Wellington survey.

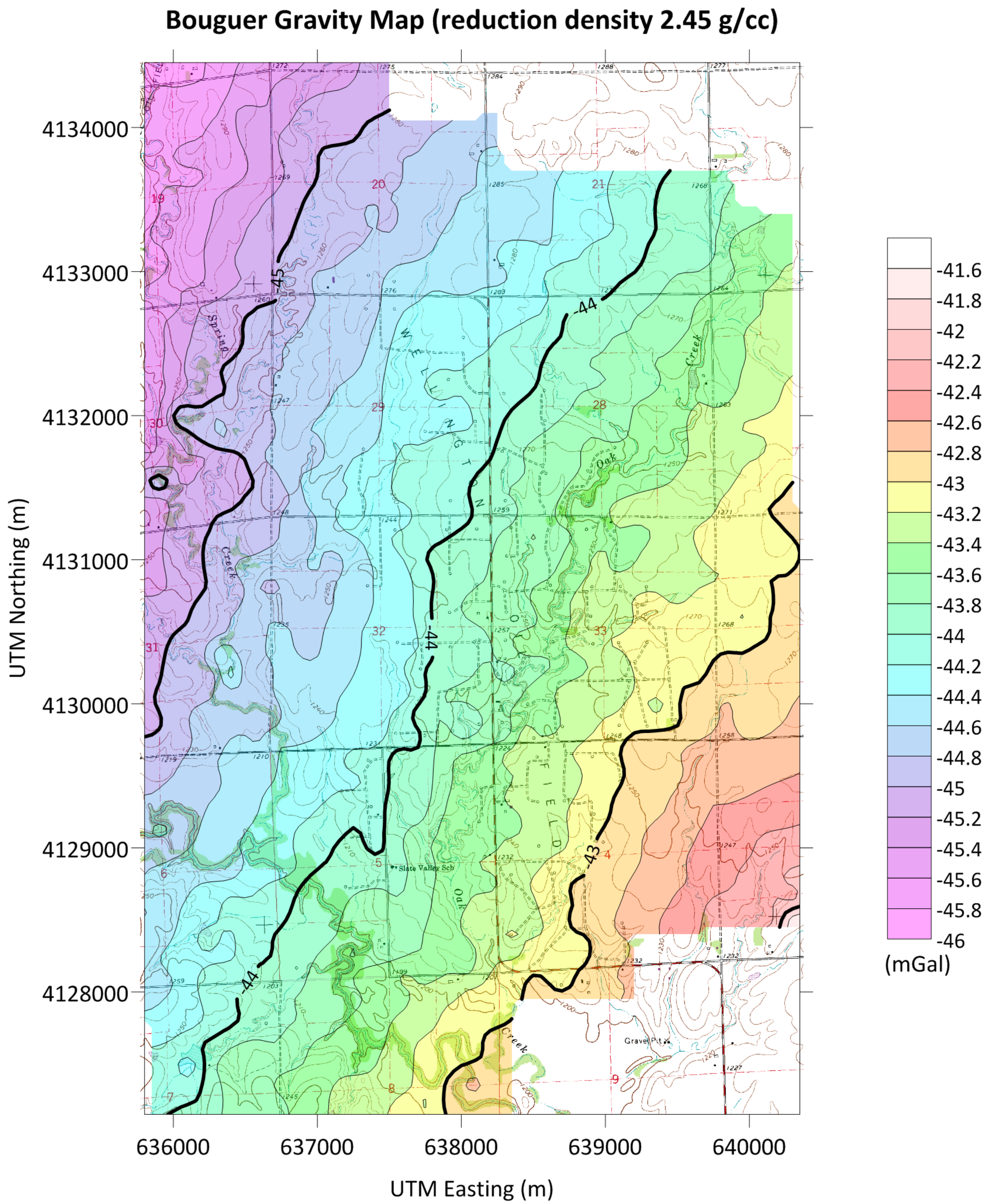

Figure 3: Contour map of the Bouguer gravity anomaly (contour interval 0.2 mGal).

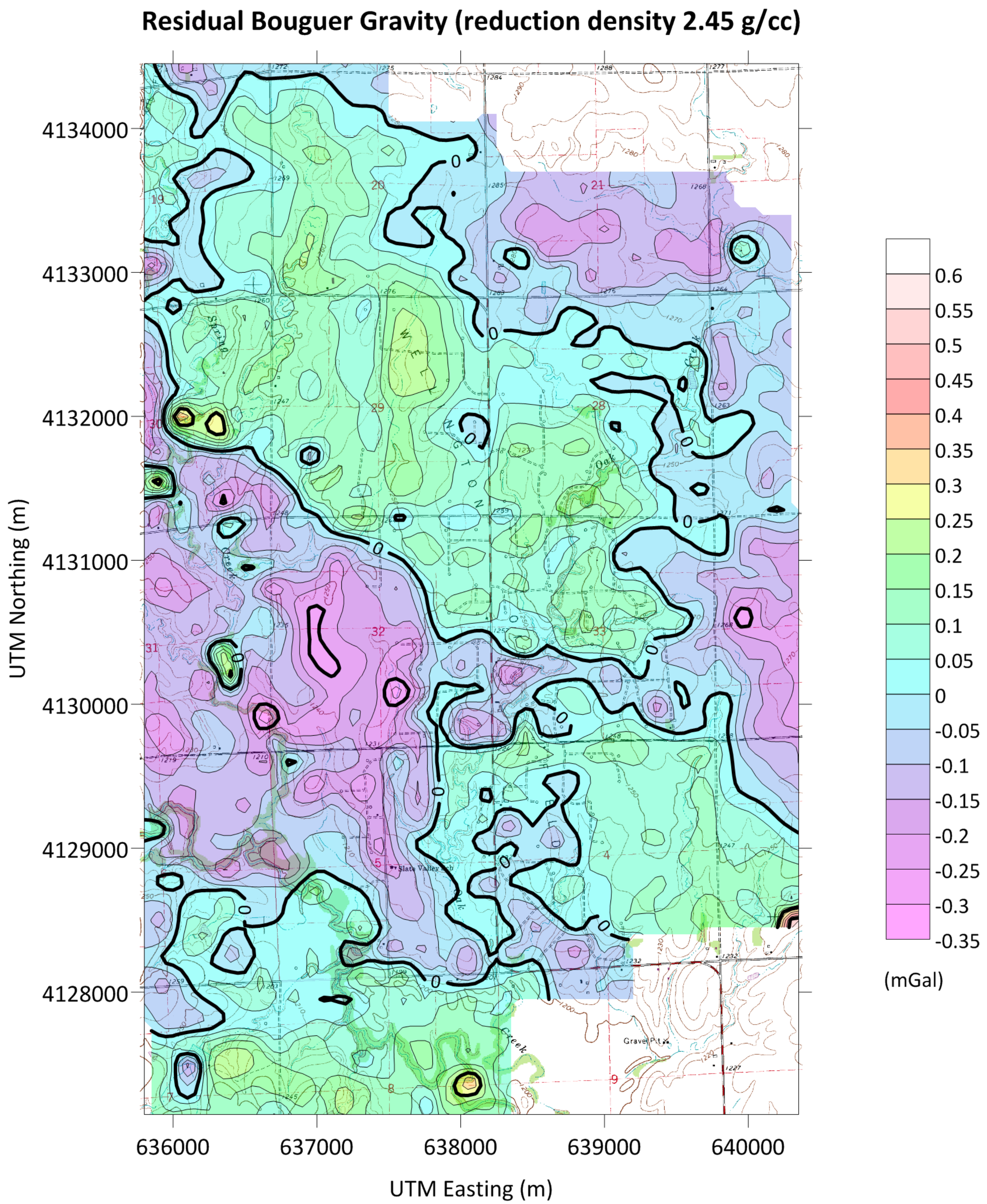

Figure 4: Contour map of the residual Bouguer gravity anomaly (contour interval 0.05 mGal), calculated by subtracting a least-squares plane from the observed Bouguer gravity anomaly.

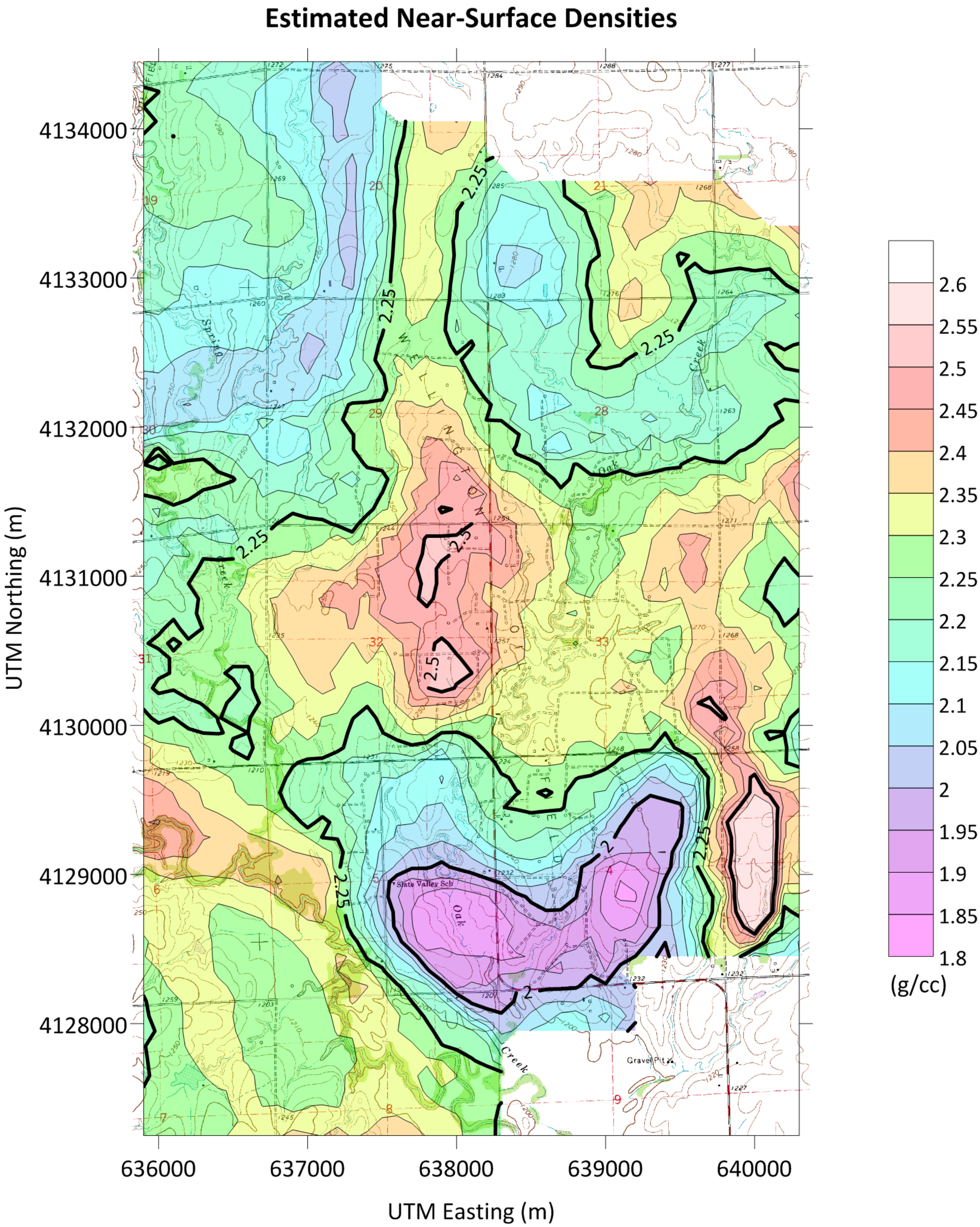

Figure 5: Color contour map of the estimated near-surface densities for the Wellington survey. Contour interval 0.05 g/cc.

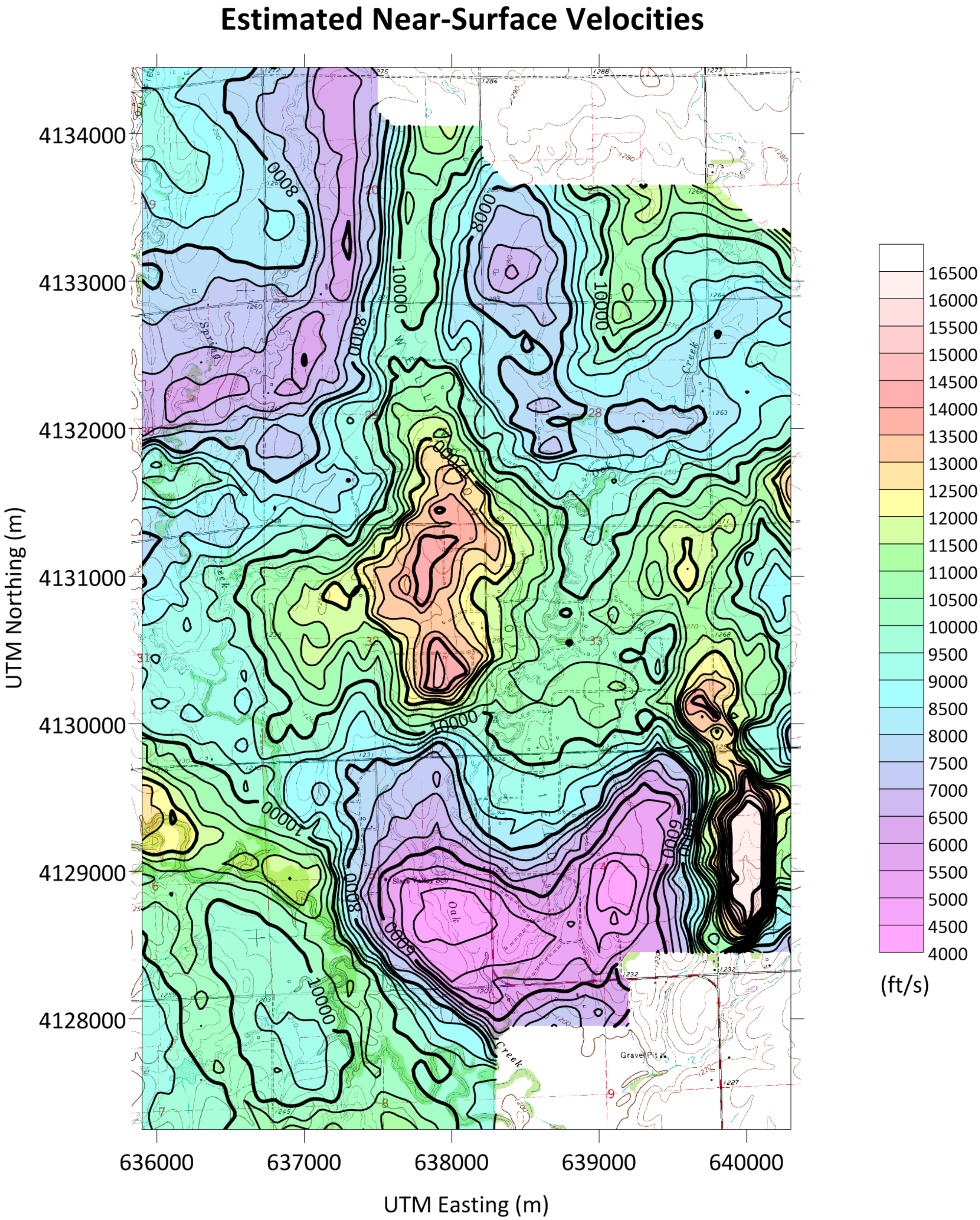

Figure 6: Color contour map of the estimated near-surface velocities for the Wellington survey. Contour interval 500 ft/s.

Appendix I: Classical Nettleton Profiling Method

The estimation of Bouguer density by correlating Bouguer gravity anomalies with elevation changes is typically termed the Nettleton profiling method, after Nettleton (1939). In the classical Nettleton method, profiles of Bouguer gravity are plotted along with topography for various values of the reduction density, which is treated as a parameter. Because the purpose of the application of the Bouguer correction is to eliminate the effect of topography, the reduction density that minimizes the correlation of Bouguer gravity with topography is selected as the best density for the topography.

Implicit in the Nettleton method is that the primary effect on the observed Bouguer gravity comes from topographic variations. If significant geologic changes occur, it may be difficult or impossible to infer a topographic density using Nettleton's method. In fact, significant correlation of geologic changes with topography is a common occurrence, which often means that one cannot apply the Nettleton method in practice.

Modern Implementation

Simple Bouguer Anomalies

In the absence of lateral geologic changes, if the topography is assumed to have constant density ρ, the simple Bouguer gravity anomaly Δg_{sba} is given by

$$\Delta g_{sba} = \Delta g_{fa} - 2\pi\gamma\rho h \tag{1}$$

where γ is the gravitational constant, h the station elevation, and Δg_{fa} the free-air gravity anomaly. The whole idea of the Nettleton method is to select a reduction density such that the calculated simple Bouguer anomaly is constant. Thus, (1) has the canonical form $\alpha + \beta x = y$, with α corresponding to Δg_{sba}, β corresponding to $2\pi\gamma\rho$, x corresponding to the observed station elevation h, and y corresponding to the observed free-air anomaly at the station. Linear regression will provide estimates of the parameters α and β; the value of α is not of particular interest, but clearly

$$\rho = \frac{\beta}{2\pi\gamma} \tag{2}$$

which is the desired estimate for the Bouguer reduction density. Note that in this formulation, one is not required to have measurements along a profile, as the correlation is done numerically rather than visually. Simply having sufficient measurements in an area with significant topographic relief is all that is required, provided the basic assumptions of the method are applicable.

Complete Bouguer Anomalies

If significant terrain corrections are needed, the simple relationship of (1) must be modified, as the Bouguer anomaly is no longer a simple function of station elevation. (1) must be modified to

$$\Delta g_{cba} = \Delta g_{fa} - \rho Q \tag{3}$$

where Q is the terrain effect evaluated for unit density (that is, just the geometrical portion of the terrain effect). Specifically,

$$Q = 2\pi\gamma h - \Delta g_{tc} \tag{4}$$

where Δg_{tc} is the traditional gravity terrain correction (always positive) evaluated for unit density and h is the station elevation.

(3) can be recast as

$$\Delta g_{cba} + \rho Q = \Delta g_{fa} \tag{5}$$

Clearly (5) is in a form suitable for linear regression. Plotting the terrain effect Q versus the free-air anomaly should provide an estimate of the correct gravity reduction density, again provided that the basic assumptions of the method are met.

References

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