GRAVITY ANOMALIES are the difference between a standard Earth model for gravity and what we actually observe. see PPT comments

The standard Earth model is based on the **HYDROSTATIC FIGURE** of a rotating Earth with the same average radial distribution of density. We derive the average radial distribution of density from seismology, and it does not need to be known very well at all in order to give a reasonable value for the flattening of the Earth, because mostly it is the rotation rate and mass of the Earth that controls the flattening.

A good review of modern practice in gravity corrections is

Hinz, W.M. et al 1984. New standards for reducing gravity data: The North American Gravity database, Geophysics **70** No 4 124-132.

When the rotating Earth comes into hydrostatic equilibrium with its own gravitation, and the centrifugal force, it assumes a shape called an ELLIPSOID (that is mathematically almost an ellipse) OF REVOLUTION. Its shape can be approximately described by the following equation

$$r = a(1 - f cos^2\theta)$$

where θ is the colatitude, that is measured southwards from the North pole, and f is called the flattening.

$$f = \frac{a - c}{a} \quad \text{FIGURE FLATTENING}$$

where a is the equatorial radius, and c is the polar radius. This general shape when $a > c$ is called an oblate spheroid.

The same theory that gives that equation for the shape also predicts a variation in gravity that follows a similar equation.

$$g = g_e(1 + \beta cos^2\theta)$$

where g_e is the equatorial gravity, g_p is the polar gravity, and β is called the gravity flattening.

$$\beta = \frac{g_p - g_e}{g_e} \quad \text{GRAVITY FLATTENING}$$

The most recent results require

$$f = \frac{1}{298.254} \qquad \beta = \frac{1}{187.48}$$
$$= 0.0033528 \qquad = 0.005347426$$

Incidentally, f and β are related through CLAIRAULT'S THEOREM.

$$f + \beta = \frac{5}{2}m$$

where

$$m = \frac{\omega^2 a}{g_e} \approx \frac{\omega^2 a}{GM/a^2} =\approx \frac{\omega^2 a^3}{GM}$$

which is the ratio of centrifugal acceleration at the equator to (Newtonian) gravity at the equator.

so M, a, ω control the flattening f. How density is distributed radially has only a very small influence on f, and therefore on g. In other words, **the shape of the Earth can be defined by gravity measurements.** Of course the connection between hydrostatic shape and gravity is much more subtle than simply these last few formulae, and the theory extends to much finer detail.

Although it does not conform to the most recent results for the flattening parameters much of the worlds gravity data base was referenced to the INTERNATIONAL GRAVITY FORMULA (1930).

$$\gamma(\theta) = 978.04900(1 + 0.0052884 cos^2\theta - 0.0000059 sin^2 2\theta) cm/s^2$$

This is based on a flattening $f = 1/297.0$.

We see from this equation that gravity varies from a minimum of

$$\gamma_e = 978.04900 \text{ Gal at the equator}$$

to a maximum of

$$\gamma_p = 983.22131 \text{ Gal at the poles}$$

This is about a half a percent variation.

The standard for gravity on the ellipsoid has undergone several iterations

$$\gamma(\theta) = 978.04900(1 + 0.0052884 cos^2\theta - 0.0000059 sin^2 2\theta) cm/s^2 \quad \text{IGF 1930}$$

$$\gamma(\theta) = 978.031846(1 + 0.0053024 cos^2\theta - 0.0000058 sin^2 2\theta) cm/s^2 \quad \text{IGF 1967}$$

$$\gamma(\theta) = 978.03267714 \frac{1 + 0.00193185138639 cos^2\theta}{\sqrt{1 - .00669437999013 cos^2\theta}} \quad WGS1984$$

Note that the differences between the first two is about 10 mgal, and between the second two about 1 mgal. For the most part the network of absolute stations is only good to a few mgal, but the absolute network is being redefined with modern equipment and will eventually be good to a hundred microgal or perhaps even a few tens of microgal. It can never be much better than this. Things - like air pressure, groundwater and so on effect local gravity at the few tens of microgal level. In the meantime data reduced with the IGRF(1930) will disagree with data reduced with WGS(1984) by about 10 mgal. You would need to be aware of this if you tried to combine old and new data.

If the various absolute stations are only good to say 1 mgal, then if your survey is tied to one absolute station, and you try to patch it onto data from an adjoining area, which was tied to a different absolute station, then there will be a mismatch of a mgal along the boundary between the two data sets.

The paper by Grauch describes similar problems with mag data sets.

V.J.S Grauch, 1993. Limitations on digital filtering of the DNAG magnetic data set for the conterminous US. Geophys. **58,** # 9 1281-1296.

WGS 1984 includes the mass of the atmosphere (the earlier IGF definitions do not). That is, to construct WGS 1984, the atmosphere was condensed on the surface of the ellipsoid, so that WGS 1984 is actually gravity at the surface of the ellipsoid, but outside the atmosphere. The reason for this is that WGS 1984 is derived mostly from satellite observations and is used in satellite gravity. To account for the addition of the attraction of the atmosphere to WGS 1984, the recommendation is to subtract it from the WGS 1984 before doing the latitude correction, or equivalently, add it to observed gravity before doing the latitude correction. The effect is:

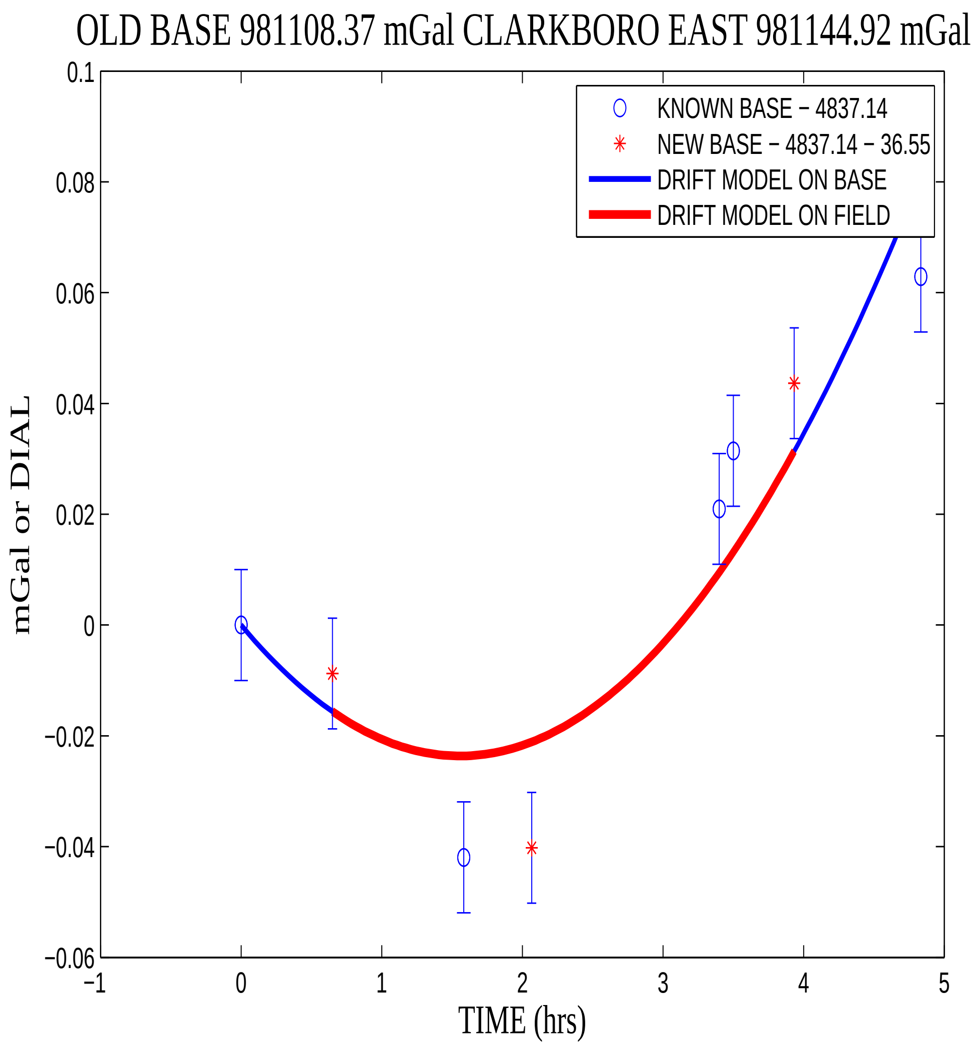
$$\delta g_A = 0.87 e^{-0.118 h^{1.047}} mGal \quad \text{for } h \geq 0$$

$$\delta g_A = 0.87 mGal \quad \text{for } h < 0$$

where h is in km. Subtract this from WGS 1984, so the latitude correction does not include the atmosphere, or add it to observed gravity, to effectively move your observation outside the (condensed) atmosphere.

Field gravimeters are relative reading instruments. They don't tell us what gravity is at some point, but only the difference between gravity at that point and some other point. If there is one station at which absolute gravity has been measured, then we can use the relative meter to tie each station to an absolute scale.

There are two approaches. The best is to run sequential repeats at a known absolute gravity station and the survey base station. At least three back and forth repeats are required and perhaps more if the two stations are far apart and travel time is excessive. Here is an example from a gravity survey performed by Mike Hartley in the Clarkboro Ferry area. The tie is between the vault and the Clarkboro ferry, a distance of about 20 km, with nearly an hour travel time. We did five vault measurements and three base station measurements. A quadratic in time was fit to all the observations with the quadratic and linear terms forced to be the same for both stations and a constant offset between the curves for the two stations was calculated. The result was a difference of 36.55 mGal, which needed to be added to the known absolute gravity at the vault to get absolute gravity at the base.



The second way is to take the average of all differences. For the Clarkboro case this results in 36.54 mGal, just a little less than before. The first method however handles the drift in a more sophisticated way and is preferred.

Because field instruments are relative, they must be calibrated somehow. This is done by transporting them over a calibration range, in which absolute gravity has been measured.

Most airports have an absolute station, and most cities and towns. Saskatoon has a few, and there is a new one at the vault. If the nearest absolute station is inconveniently far away from your survey area, then the absolute value must be transferred to your base station, by repeated loops back and forth from the absolute station to your base. The number of loops depends on the accuracy of the absolute station, and the drift of the instrument. Basically, the transfer must be as good as the original absolute value.

CONVERTING DIAL READINGS TO CALIBRATED GRAVITY

1) read the counter eg 4654.3
2) read the dial 4654.36
3) using the table obtain the value in mgals for the reading in the table which is nearest this value but less than it (ie 4600 gives 4817.47)
4) obtain the difference in counter reading ie 54.36
5) multiply this by the interval factor 1.04845
6) add 4817.47+1.04845 x 54.36=4874.46 calibrated value

PART OF G267 CALIBRATION TABLE

COUNTER READING	VALUE IN MILLIGAL	FACTOR FOR INTERVAL
4300	4502.91	1.04853
4400	4607.77	1.04853
4500	4712.62	1.04848
4600	4817.47	1.04845
4700	4922.31	1.04844
4800	5027.16	1.04855
4900	5132.01	1.04855

LaCoste, L. 1991. A new calibration method for gravity meters, Geophys. **56,** # 5 701-704.

LaCoste, L. 1991. Gravity meter calibration at LaCoste and Romberg, Geophysics, **56,** # 5 705-711.

THE LATITUDE CORRECTION

When a survey is done some station will be selected as the base station and readings will be corrected according to their positions North or South the base. The rate of change of the IGF30 with co-latitude is

$$\frac{\partial \gamma}{\partial \theta} = -978.04900 \times (0.0052884 sin2\theta + 0.0000118 sin4\theta)$$

At colatitude 45^o this corresponds to

$$\Delta\gamma = -5.1722 gal/rad \Delta\theta$$

or

$$\Delta\gamma = -.8118 mgal/km \times S$$

where S is the distance south of the base station in Km. At S'toon this is -.7805 mgal/Km.

The derivative for WGS1984 (in gal/rad) is

$$\frac{\partial \gamma}{\partial \theta} = -\,978.03267714 sin2\theta \Big[\frac{0.00193185138639}{\sqrt{1 - 0.00669437999013 cos^2\theta}} + 0.00669437999013 \frac{(1 + 0.00193185138639 cos^2\theta) cos\theta sin\theta}{(1 - 0.00669437999013 cos^2\theta)^{3/2}}\Big]$$

or in mgal/km

$$\frac{\partial \gamma}{\partial S} = -978.032667 \Big[\frac{0.00527904138145}{6378.137} sin2\theta + \frac{1.293254726 \times 10^{-5}}{6378.137} sin2\theta cos^2\theta\Big]$$

There are two ways to do the latitude correction. If the latitude of every station is known, then WGS84 is simply subtracted from every station. If the Northing in m or km of every station is known wrt some reference latitude, then subtract WGS84 at the latitude of the reference station from observed gravity at each station, and then use the horizontal N-S gradient at the reference station to adjust for the position north or south of the reference station. This method agrees with the former to within $\pm 0.01 mgal$ as long as the furthest station from the reference latitude is no more than 25 km N or S of the reference station.

The latitude effect is large, and may be significant on even small gravity surveys. For example, at mid-latitudes a horizontal NS movement of only 12 m requires a latitude correction large enough for a Lacoste G meter to see (0.01 mgal).

Gravity surveys are often done with a planned accuracy of 0.01 mgal, this is the sensitivity of the usual exploration instrument, so that the nominal variation of gravity with latitude is important even for small scale surveys. In most cases the horizontal control on station position is also determined by the latitude correction. If the planned accuracy of the survey is 0.01 mgal, then the above equation says that a Saskatoon latitudes the horizontal control must be $\pm 12m$.

Because the latitude effect is North South, we really mean the horizontal control on NS position must be $\pm 12m$, because EW horizontal gradients are much smaller, we could get away with horizontal positions much less accurate than this EW. To be a little more precise, the accuracy of horizontal control is really determined by the maximum horizontal gradients in gravity. If we had maximum horizontal gradients in gravity of 1 mgal/100 m, and we want accuracy of 0.01 mgal in the survey, then we would need horizontal control of $\pm 1m$. Gradients this large are quite unusual, especially in the Bouguer anomaly, but it the gradient in the *observed* gravity that is important, and topography can sometimes produce large gradients.

The advantage in subtracting $\gamma = -0.8118 mgal/km \times S$ from each reading, rather than do the proper latitude correction (that is subtract the IGF from every reading) is minor, and in fact it is absolutely necessary to subtract the IGF properly if you want to produce a final Bouguer anomaly that can be tied to someone else's survey. If you do take a shortcut and subtract a linear trend, then you will be left with residual that has a large offset (sometimes this is called a DC value). This is no problem, and would be removed as part of the regional/residual separation later.

FREE AIR CORRECTION

Gravity is also influenced by the relative height of stations, and so to separate these effects from more interesting effects due to subsurface structure we correct all readings to the same level. There is only one level which is globally definable and this is called the **GEOID** which is roughly speaking mean sea level.

We can approximate the variation of g with height by looking at the gravitational field of a sphere. Then

$$\begin{aligned} g =& \frac{GM}{r^2}\hat{\mathbf{r}} \\ \frac{\partial g}{\partial r} =& -\frac{2GM}{r^3} \\ \Delta g =& -2g\frac{\Delta r}{r} \\ \Delta g =& -0.3086 \Delta r mgal/m \end{aligned}$$

where Δr is the height in m above the geoid. since g decreases with height we subtract this amount (that is add 0.3086 mgal/m) from every reading. This is a pretty good approximation, but what we should really have done is differentiate vertically on the international ellipsoid rather than radially on a sphere. The IGF actually includes a term for the first derivative of gravity above the IGF ellipsoid. It is

$$\begin{aligned} \frac{\partial \gamma}{\partial h} =& -0.308791[1 - 0.001426 cos^2\theta] mgal/m \\ & + 1.442 \times 10^{-7} mgal/m^2 \times h \end{aligned}$$

In this area the above is

$$= -0.308422 + 0.0000432 \times h mgal/m \quad h \approx 600m \text{ here}$$
$$= -0.308465 mgal/m$$

so the gradient is a bit smaller at 600 m elevation than it is at sea level.

In WGS1984 (GRS1980) the vertical gradient is

$$= -(0.3087691 - 0.0004398 * cos^2\theta) + 7.2125 \times 10^{-8} h \quad mgal/m$$

The vertical gradient is defined on the ellipsoid, but the free air correction (and the Bouguer correction) commonly use orthometric height, or height above the geoid, not height above the ellipsoid. The difference can be as much as 100 m. There is currently a recommendation to define free air and Bouguer anomalies with ellipsoidal height because GPS can give ellipsoidal heights to a cm, but in many places the geoid is not known to the cm level.

The free air gradient is important in determining the vertical control needed in a gravity survey. The free air gradient means that if the elevation of the meter is **increased/decreased** 3 cm gravity will **decrease/increase** by 0.01 mgal. Since this is the reading sensitivity of the most common meters, survey elevation control must be $\pm 3cm$.

If you actually measure the vertical gradient of gravity at the vault, it is about 0.27 mgal/m. Why is this different from the theoretical, and what should we do about it?

In practice gravity does not vary exactly as the IGF defines it - there would be no gravity anomalies if it did- but rather in a complex way. Thus, if you were to measure a gradient locally, it would probably be different from what the IGF says it is in the same way that measured gravity is not the same as the IGF. This does not mean that you should measure a local free air gradient, and use this to do the free air correction, the correct one to use is the IGF. Why? Because everyone must do their corrections a standard way.

Rather, because the local gradients are due to local anomalies, but the free air correction is a model for gravity above a uniform ellipsoidal Earth.

CONTROL ON STATION POSITION

The maximum horizontal and vertical gradients supply the controls on station positions.

Horizontal gradient in IGF	1 mgal/km	$\rightarrow$ 0.01 mgal = 10 m
Vertical gradient in IGF	0.3 mgal/m	$\rightarrow$ 0.01 mgal = 3 cm

Compare this with magnetics

Horizontal gradient in IGRF	5nT/km	$\rightarrow$ 0.1 nT = 20 m
Vertical gradient in IGRF	0.03nT/m	$\rightarrow$ 0.1 nT = 3 m

These results are for gradients on the standard models. Local gradients may be greater, especially near steep topography, in which case the position control may become more stringent.

Station elevation effects gravity (and mag) in another way. Suppose a buried sphere of some positive density contrast was located under a hill. Then the increase in gravity from the source would partially offset the free air change from elevation. The hill moves the gravity meter further away from the center of the Earth, this is the standard free air, but the hill also moves the gravity meter further from the source.

There are two ways to deal with this.

1) Some modeling packages allow the modeling to take place on an arbitrary surface. In this case a free air correction would be done using the standard value, and the modeling package would adjust for the attraction of the hill, and the source. This is the prefered method.

2 The data could be corrected to a level surface, and ~~the~~ interpreted. I will show how to do this later, but transforming potential fields from one surface to another is possible either as continuation - level surface to level surface, or drape to level, or level to drape.

Xia, J., D.R. Sprowl, and D. Adkins-Heljeson, 1993. Correction of topographic distortions in potential-field data: A fast and accurate approach, Geophysics, **58,** # 4 515-523.

Grauch, V.J.S., and Campbell, 1984. Does draping aeromagnetic data reduce terrain induced effects? Geophysics **49,** 75-80.

THE BOUGUER CORRECTION

Local gravity is also influenced by the attraction of material between the gravity station and the geoid. Most of this attraction is due to the average density of the rock between the station and the geoid.

In the Bouguer correction the attraction of the topography is replaced by the attraction of the BOUGUER SLAB, which is a slab of uniform density and thickness equal to the station height above the geoid.

To compute the attraction of a slab consider first the attraction of an infinitesimally thin sheet of surface density σ at point P above it

Imagine shining a flashlight beam from P down to the sheet, then the attraction of the illuminated are on the sheet is

$$\begin{aligned}\Delta g =& G \int \frac{cos\phi}{r^2} \sigma dA \\ =& G\sigma \int \frac{cos\phi}{r^2} dA \\ =& G\sigma \int \frac{cos\phi}{r^2} \frac{r^2 d\Omega}{cos\phi} \\ =& G\sigma \int d\Omega \\ =& 2\pi G\sigma\end{aligned}$$

4π is the solid angle subtended by a sphere, we are only integrating over half of a sphere. Notice that the height above the sheet does not enter into this equation.

We can construct a slab by integrating over thin sheets

$$\Delta g = 2\pi G\rho t$$

where t is the thickness of the slab (not the height above the slab) ρ is the density of the slab. If P is above the slab the attraction is down, if P is below the slab the attraction is up, with the same amplitude. With a mean crustal density of $\rho = 2.67gm/cc$ this becomes

$$\Delta g = 0.1119mgal/m \times h$$

This is a very interesting and useful formula. The attraction of a Bouguer slab of density x is just

$$\Delta g = 0.1119mgal/m \times \frac{x}{2.67} \times h$$

For example, what is the gravity anomaly that you would expect from a thin sill at shallow depth? If the depth is shallow compared to the radius of the sill, then near the center of the sill, the attraction will be given by the Bouguer formula. Let say a target is 4 m wide, 3m deep. If the till is 2 gm/cc then

$$\Delta g = 0.1119mgal/m\frac{2}{2.67} \times 3m = .25mgal$$

Since the attraction of the slab increases g we subtract this amount to correct for it.

In the above integration, we could have integrated over less than half a solid angle, in which case we would have found gravity at the apex of a cone.

The solid angle is now $2\pi(1 - cos\alpha)$ where α is the semiangle of the cone, and the attraction of a cone at its apex is

$$2\pi G\rho h(1 - cos\alpha)$$

So at the top of a mountain the Bouguer correction would overestimate the true correction by a factor $(1 - cos\alpha)$ or $g = g_b(1 - cos\alpha)$. At the base the attraction is less than 1/20 of the attraction at the apex (and up instead of down) whereas the Bouguer attraction here would be zero. Everywhere else on the cone the Bouguer attraction is closer to the true attraction than it is at the apex, so we know the maximum error in the Bouguer correction.

For grades of 1 in 10 $\alpha = 94^o$ and the cones attraction is only 5% smaller than the slabs. 1 in 10 is a pretty steep gradient (steeper than most roads in the mountains) so the Bouguer correction is pretty good. So topography is unimportant unless gradients are much steeper than 1 in 10, and more than a few m in amplitude.

Note, however, that a 1-m thick slab gives a ~1 mGal Bouguer correction, 5% of which is ~0.05 mGal. This is comparable to ~15 cm uncertainty in station elevation mentioned above.

The Bouguer slab correction is sometimes also called the Bullard A correction. Refinements are the Bullard B and Bullard C corrections. The Bullard B correction adjusts for the difference between the slab and a spherical cap of radius 166.735 km, and thickness the station height above sea level.

The formula is complicated, so the values are usually gotten from a table. The correction is not large, about 0.001 mgal if h is less than a Km. This is less than the normal survey accuracy, so the Bullard B is not normally done, except at high elevation, or for microgravity surveys. An approximate equation is

$$BB \approx Ah + Bh^2 + Ch^3$$

$$A = 1.464 \times 10^{-3}$$

$$B = -3.533 \times 10^{-7}$$

$$C = 4.5 \times -14$$

Also power of 10

where BB is in mgal and h is elevation in m. At elevations below 4 km the cap has more downward attraction than the slab, so the Bullard B adds something to the Bouguer attraction. At elevations greater than about 4 Km the attraction of the cap is less than the attraction of the slab, (because now material at distances of greater than 167 km makes a sensible angle with the vertical) so Bullard B reduces the Bouguer correction. At Saskatoon it is only 0.006 mgal, so not a concern.

I think what is meant here is that for variations
in elevations of ~500 m by about 100 m,
the Bullard B correction can be included in the
adjustment of terrain density, with an error below 0.01 mGal

SUMMARY

FREE AIR ANOMALIES are the differences between observed gravity and the International Gravity Formula, for that latitude and height.

$$\Delta g_{FA} = g_o - \gamma_o \quad \gamma_o \text{ contains the free air correction}$$

Free air anomalies are never used in exploration geophysics, but they are useful in geodynamics, because they are almost isostatic anomalies.

BOUGUER ANOMALIES are Free Air anomalies with a Bouguer correction

$$\Delta g_b = g_o - \gamma_o - 0.1119 mgal/m \times h + BB + BC$$

Note that in both these formulas g_0 is total gravity that is 98?.????cm/s. How do we get this if we have only relative reading instrument? We reference to an absolute station.

CONTROL ON TERRAIN DENSITY

How much accuracy do we need in the terrain density? ± ?? gm/cc. There is of course a limit beyond which it makes no sense to try for better accuracy. To determine this you ask the question "How much error in density will produce a gravity signal from topography which is less than the survey target?

Suppose you have topography h and your survey plan is to get errors in gravity below $\pm 0.01 mgal$. If the error in density is $\Delta\rho$ then the error in the Bouguer terrain correction is

$$\Delta g \approx 2\pi G \Delta\rho h$$

and

$$\Delta\rho = \frac{\pm\Delta g}{2\pi G h} = \frac{0.01 \times 10^{-3} cm/s^2}{2\pi \times 6.67 \times 10^{-8} dyne\, cm^2/gm^2 h}$$

$$= \pm\frac{23.9}{h} gm/cm^3/cm$$

If you have topography of say $\pm 10m$ this is

$$\Delta\rho = \pm 0.02 gm/cm^3$$

This exercise should be repeated for every survey, because it depends on the survey design accuracy, and the maximum topography.

The maximum horizontal gradient determines the accuracy you will need in horizontal positioning.

The maximum vertical gradient determines the accuracy you will need in elevations.

The maximum elevation change determines the accuracy you will need in terrain density.

DATA REDUCTION IN MAGNETICS

Data reduction in magnetics is much simpler than in gravity. The reason for this is the relative scale of anomalies and nominal variations. Anomalies of several hundred nT, or even thousands of nT in a field area of several km are quite common, but the main field itself varies by only $0.03nT/m$ vertically and 5 nT /km horizontally. Compared to anomalies of hundred's of nT from susceptibility contrasts these are negligible, and so a free air correction is rarely done. The latitude correction is done, but only because an International reference field is subtracted from observations, and this has a latitude correction in it, not because the latitude correction is important. These gradients also mean that we do not have to level a mag survey, or at least not as accurately as a gravity survey. Horizontal positioning is also not as critical in a mag survey (5nT/KM and $\pm 1nT \sim 1/5km$) However, with large local gradients ie for very small targets with large fields, as in an environmental survey, positioning may be very important.

Terrain distortions of the field are generally smaller than in gravity as well. To first order the magnetic field above a terrain is proportional to the induced magnetization, and not to thickness of the terrain. On the shield this may sometimes be important, but particularly in a sedimentary environment, there is no mag equivalent of the Bouguer and terrain corrections. In addition, it is much more common to do a quantitative interpretation of gravity, but mag modeling is much less certain due to the possibility of two separate sources (induced and remanent) magnetization, and more complicated shape effects.

Drift in magnetometers is not the problem it is with gravity meters but there can be large changes in the inducing field. The magnetic field is variable on all time scales, so that the Geomagnetic reference field is updated every five years, and interpolated between these epochs. There are diurnal variations of about 50 nT caused by tidal motions in the ionosphere, and magnetic storms lasting for days/weeks with several hundred nT variations. At higher latitudes, storms are more severe, so that it is common to have mag surveys in the arctic interrupted by Mag storms, although this is probably more for their effect on navigation than the field) than it is in the south. At high latitudes, storms are also likely to have smaller horizontal scales, so that base station variations may not accurately reflect whats happening a few Km away. Moderate storms, and the diurnal variation can be handled with a drift correction, the same as is done for gravity ie with repeat readings. Because the drift is essentially in the magnetic field itself, and not in the instrument it is also possible to use an automatic base station to monitor changes in the field with time and then subtract these from field readings.

The drift correction applied by the OMNI is

CORRECTED = (UNCORRECTED - DATUM) - (BASE - REFERENCE)

CORRECTED = (UNCORRECTED - BASE) - (DATUM - REFERENCE)

where REFERENCE and DATUM are constant values keyed into the field and base instruments.

THE DRIFT CORRECTION IN GRAVITY

The drift is due to 1) Tidal variations (on a good instrument this effect dominates) 2) stretching of the spring due to temperature variations, mechanical stress, or simply a type of curing, 3) atmospheric pressure, 4) operator error.

The first is usually the largest, tidal variations can be 0.1mgal in 6 hours. Sometimes stretching of the spring is important. Atmospheric pressure is only about 0.3 mgal max.

The tidal variations can be predicted to within 0.01 mgal at most stations, so it is best to compute this and subtract from all readings (that is all drift and field readings) before doing the drift correction. It is also possible to include the tidal correction with a general drift correction.

The atmospheric pressure effect is about 0.3 microgal/mb (if pressure increases there is more mass overhead decreasing gravity). This is usually so small that a few pressure readings per day are sufficient. Often this step is not done at all, and the pressure drift is incorporated into a general drift correction.

Each days drift is likely to be unrelated to the drift during the previous or following day, so it is best to interpolate a drift correction on each separate day. This is why the first and last readings are always taken at the base station.

The purpose of the drift correction is to reduce the observations so that they look like they were all taken at the same 'time'. Therefore, you select an arbitrary time, the time of the first base station reading is a convenient choice. Any base station reading that reads higher than this indicates the meter was reading high at this time, so whatever that amount is should be subtracted from whatever field reading was taken at that time.

The repeats are sometimes useful as auxiliary drift information. Suppose the first of a pair of repeats was taken at 9 am on day 2, and the second at 2 pm on day 1 (repeats can happen before the original!). Because they were at a station which may be considerably above or below the base both may read higher or lower than any of the base readings. However, if we add to both repeats a constant to bring one of them into agreement with the base reading at that time, then we can tell by the remainder if the drift between the time of the two repeats was the same as the drift of the base between these times. In fact it is possible to do the drift correction entirely with repeat readings, although this is not normally done unless there is a compelling reason.

Repeats should be taken well separated in time and by different operators, if different operators were used in the survey. This is because the drift and operator bias are among the largest sources of error, and we use the repeats to assess the overall repeatability.

AEROMAGNETIC DATA PROCESSING

Aeromagnetic data is flown as a sequence of **traverse lines,** or **flight lines** or simply **lines,** and **tie lines,** or **ties** for short, which are orthogonal to the lines. The tie line spacing is coarser than the traverse line spacing. The old standard used to be the tie spacing was six to ten times the line spacing, but current practice is three to five times, especially for high resolution surveys. The purpose of the tie lines is to aid in the drift correction, and in the removal of the aircraft signature, heading, elevation etc.

Aeromagnetic data is routinely acquired at elevations from 50 m to 800 m above ground surface, and increasingly at the lower end of this range. Aeromagnetic surveys are flown either at constant elevation (constant barometric elevation until GPS control became available), and at constant terrain clearance or Mean Ground Clearance (also called drape flying). Most of the aeromagnetic map of Canada has been flown at Mean Ground Clearance, except for the interior of the Cordillera, where the topography is too steep. Flying at constant barometric height is the cheapest, but in areas of steep topography, a constant barometric height survey is a very poor choice because over valleys the aircraft may be very high above ground, and miss fine details in the magnetic field. Drape flown surveys, on the other hand, are more difficult to fly and process, but the data quality is better. Most light aircraft have climb rates of less than 100 m /km, so if the topography is steeper than this the aircraft may not be able to maintain constant height above the topography. Also, since adjoining lines are flown in opposite directions, there may be considerable difference in ground clearance over the same feature on neighbouring lines, because the aircraft can descend much faster than it can ascend. In most cases this is not a major concern, because the vertical gradient of the field is small. Differential GPS would provide a position for the aircraft good to less then 10 cm, or .1 nT in the main field. The anomalous field would have similar gradients.

It is becoming normal for the flight lines to be planned in advance with the aid of a digital terrain model, and the known flight characteristics of the aircraft. The direction of the flight lines may in fact be chosen in part based on the climb rate of the aircraft, and in part based on geologic strike if known (otherwise E-W).

The choice of traverse line spacing (the main flight lines), the tie line spacing, and the MGC are based on rules of thumb which relate to the target depth. In a detailed study of, for example, a sedimentary basin, the MGC might be on the order of 100 m, the traverse line spacing might be between 400 and 800 m, and the tie line spacing might be between 1200 and 2400 m. That is, the tie line spacing is about three times the traverse line spacing. In a reconnaissance survey the altitude might be greater than this, perhaps as much as 300 m, the traverse line spacing 800 m to 2000 m, and the tie line spacing again three times the traverse line spacing. There has been a trend to tighter tie line surveys in the last few years, the old standard being more like 6 to 10 times the traverse line spacing.

With the above figures, a target at 300m depth (ie 400 m below aircraft) would be detected, and a target at 1200m depth would be adequately resolved. If the plane flies at say 100 km /hr and samples every .1 sec, then samples are taken every 3 m or so. Thus, there is considerable data redundancy along the flight line. This is of some marginal use in gridding to a smaller size than the traverse and flight line spacing. However, note that wiggles in the line data with horizontal scale less than twice the aircraft height have an apparent source depth above the ground! Therefore the lines and ties are often individually filtered to remove wavelength with an apparent source depth less than the MGC of the aircraft.

There are several problems which are unique to aeromag, or more difficult as opposed to ground mag. First of all, the plane itself has a magnetic signature, which is different depending on the heading of the plane wrt the total field vector. Thus, the sensor may read several tens of nT higher when flying in one direction compared to another. This is called **heading noise.** In addition, there may be pitch, yaw and roll signatures in the recorded field. These are called **maneuver noise.** Small signals may also arise from instrumentation on the airplane, as well as from currents induced in the airframe by the changing magnetic field the airplane sees, ie apparent changes in a static field due to the motion of the aircraft, and temporal changes in the field itself. Over the oceans, currents induced in ocean waves by the static magnetic field may produce magnetic fields of a few nT.

The process of getting rid of the aircraft dependent effects is called **compensation.** Compensation used to be handled by enclosing the mag sensor in a three component Helmholtz coil. The currents in the Helmholtz coils were adjusted in real time to compensate for the aircraft heading and movement. To do this, the pilot flew a square course over magnetically flat terrain and made pitch yaw and roll maneuvers on each leg of the flight. The recorded magnetic field was used to program an analogue compensator such that just the right amount of current was sent to each coil to cancel the change in field produced by the maneuver. This method of compensation resulted in errors of about $\pm 5nT$. In modern systems, the Helmholtz coils are replaced by a three component fluxgate magnetometer. The process of flying a compensation square is still done, but the various maneuvers are cross correlated with the three component fluxgate to produce a total field compensation signal which is subtracted from the reading of the main magnetometer. Modern systems correct for maneuver noise to about $\pm 0.15nT$ and heading noise to about $\pm 0.25nT$.

Noise from instrumentation on the aircraft is more difficult to cancel and the crew takes great pains to keep everything on the aircraft the same, including all positioning of equipment and baggage.

The speed of the aircraft, say 100km/hr, and the sampling rate perhaps 0.1 samples/sec, means that samples are taken every 3 m, along the flight lines. A 1 sec period micropulsation would then become a 30 m scale variation along the flight path.

The information is very detailed along flight and tie lines, but remember flight lines might be a hundred m apart, so the question arises as to what the resolution is. There is no easy way to answer this. Certainly dikes as narrow as a few m can be adequately resolved if they strike perpendicular to flight lines, but even a large dike might not be resolved if it was parallel to flight lines and in between two of them. For most purposes, it is safest to consider the flight line spacing is a measure of the resolution. Another way to look at this question is through the flight height. The full width at half height for a blocky source will be about half the elevation of the airplane above the source. Therefore, any anomalies with a full width less than this have an apparent source depth of less than the aircraft height! This is clearly impossible, and flight lines might be individually filtered to remove features this narrow.

Imagine a survey flown at constant barometric level. Because of the finite time taken to fly each traverse there is a progressive shift in the level of the mag intensity along the line, due to the 'diurnal' variation, plus an offset wrt neighbouring lines, due to flight direction and 'diurnal', as well as attitude signals, variations in altitude (above mean sea level) and in ground clearance. This results in **in-**

tersection errors at the crossover point of ties and lines. Subtracting a base station reading eliminates most of the 'diurnal' except if the survey area is very large, and compensation removes most of heading and maneuver noise. Residuals from these corrections, as well as all uncompensated variations, altitude, ground clearance etc. are removed by **leveling.** Leveling involves adding a constant, or a slowly varying function (in time) to each traverse line and tie line to minimize the intersection errors between the traverse lines and the tie lines. For example, suppose the diurnal and maneuver noise and all other source of noise were perfectly compensated, except for the heading error. Then all NS lines would read say ten nT higher than they should, and all SN lines would read ten nT lower. EW lines would read say 2 nT higher and WE lines 2 nT lower. In this idealized situation it would be a relatively simple matter to find the constant needed to be added to each line to minimize the intersection errors. Leveling refers to the magnetic level of each line, not a height, so the leveling correction is not intended to correct for altitude changes, although it will do this if each line differs from its neighbours by a constant elevation. However, since the free air effect in mag is so small, 0.02nT/m typically, this is not important in any case. The effect of constant offsets in ground clearance are not removed by leveling, because the effect of ground clearance on a magnetic anomaly depends on the horizontal scale of the anomaly as well as its size. At the present, ground clearance effects are not routinely corrected.

Because the diurnals as recorded by the field sensors will not correlate perfectly with the diurnal recorded by the base station, there is some difference of opinion about whether the diurnal correction should be done at all. Some people believe that the leveling process described below is a better way to handle the diurnal, compared to even several base stations.

In a drape survey individual traverse line may of course be flown at a different mean elevation, so the recorded field for that line may have an offset wrt a neighbouring line. While this offset may be due to the flight level of the aircraft, it is not a leveling error, and is handled in mapping the draped survey to a level surface.

AEROMAGNETIC DATA PROCESSING

Aeromag data processing is a lengthy and complicated process involving a number of steps.

Pre-processing

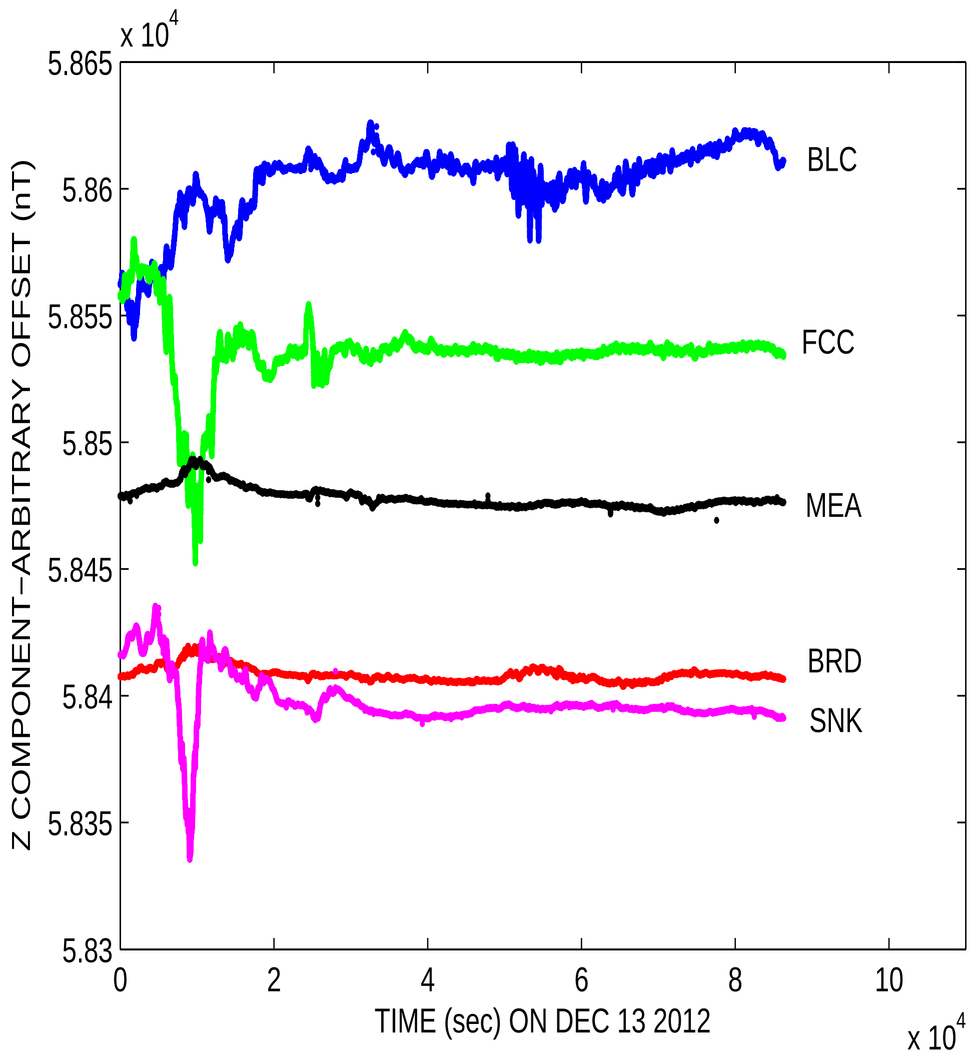
- **Verifying and editing raw data** Single station anomalies, gaps in the primary data (field magnetic) or secondary data (navigation, or base station)
- **locating the data in x and y** Modern navigation is real time differential GPS, which gives the position of the GPS antenna to within a few meters. Before real time differential GPS, it was common to adjust the horizontal position of the crossovers. GPS horizontal positions are accurate enough that this should not be done.

Processing

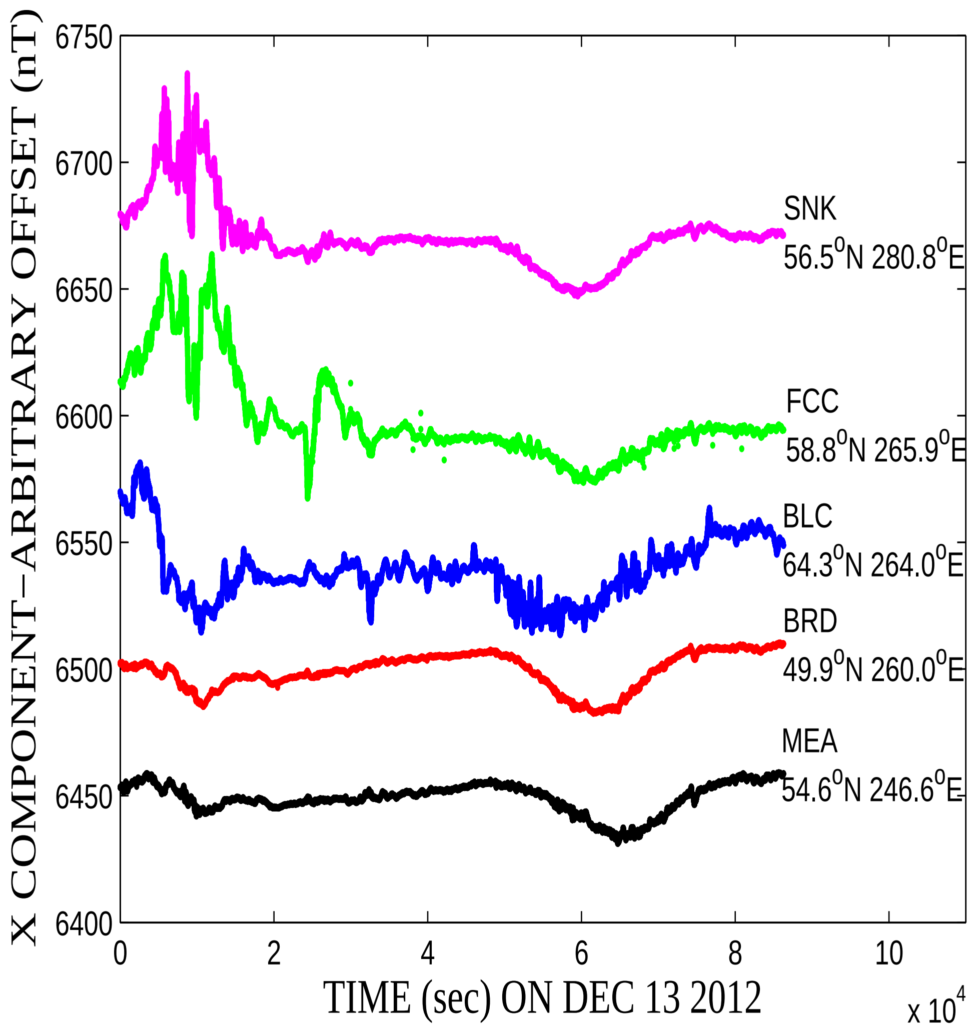
- **parallax corrections** The flight camera, or GPS antenna is normally near the front of the airplane, while the mag sensors are in a stinger off the tail. This means there is a ten m or more difference (called the *cable length*) between defined position and the measurement point. Since flight lines are flown alternately in opposite directions, the same feature is located twice the cable length apart on successive lines. The navigation is interpolated to the measuring point rather than the other way around.
- **removing diurnals** This is done by subtracting a (usually only one) base station series. Residual diurnals (the difference between the base station time series and a time series at a field station) are removed as part of the leveling process. There may also be differences between the flight line height and the tie line height, although these can be kept small with modern (GPS) navigation. Alternatively, one could install several base stations, and build a model for the variation of the diurnal over the survey area. I am not aware of this as a common practice.
- **removing the component due to the regional** IGRF, or a local national version is subtracted from field data.
- **leveling** Leveling removes residual diurnals, (the difference between the diurnal at the base station and at every other point in the survey), tie and sleight height differences, any instrumental drift (not in the main magnetometers but possibly in the secondary fluxgates used for compensation), and residual compensation including heading.

- **micro-leveling** is applied after the data is gridded to remove any artifacts of the flight line-tie line pattern.

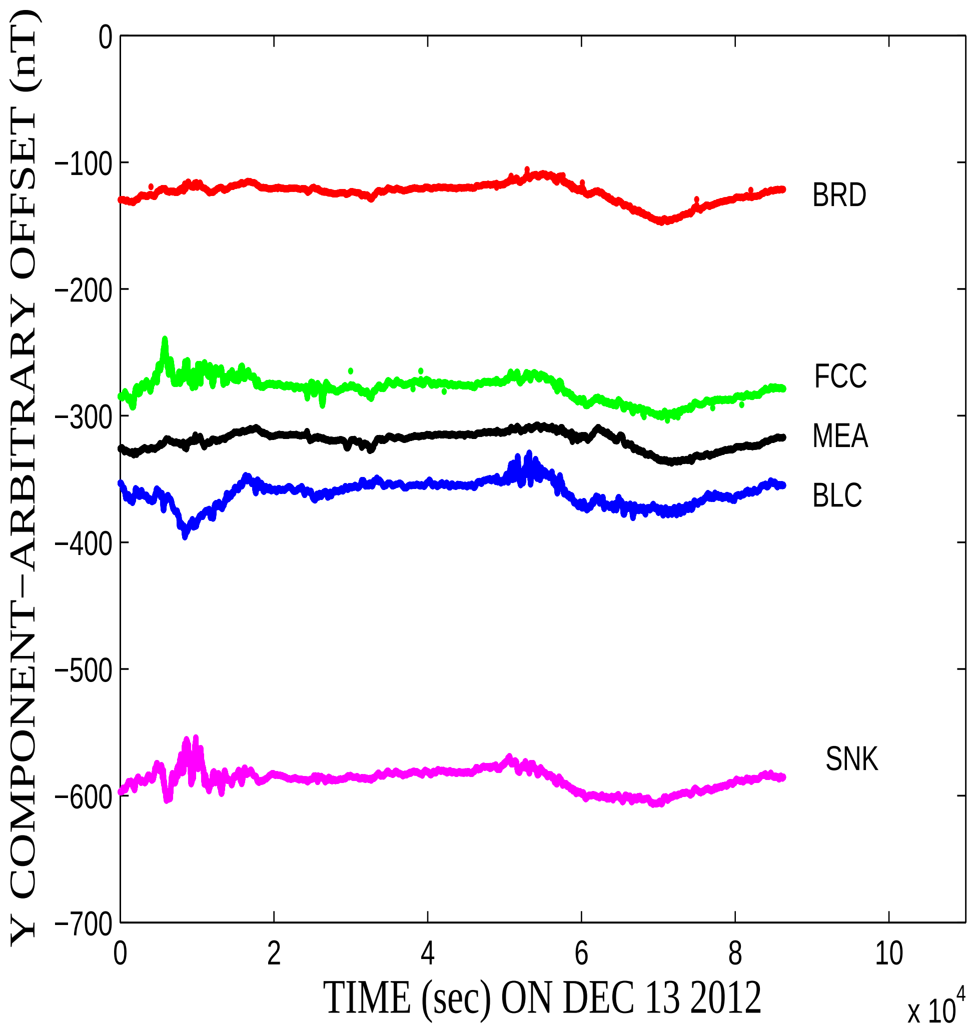
The first step is to subtract the base station readings from the field readings. This assumes that the temporal change at the field station is exactly the same as at the base. How good an approximation this is depends on how far apart the base is from the furthest survey point, and how much spatial variation there is in the field. The following figures show the temporal variations (minus an arbitrary offset) in five western Canadian stations.



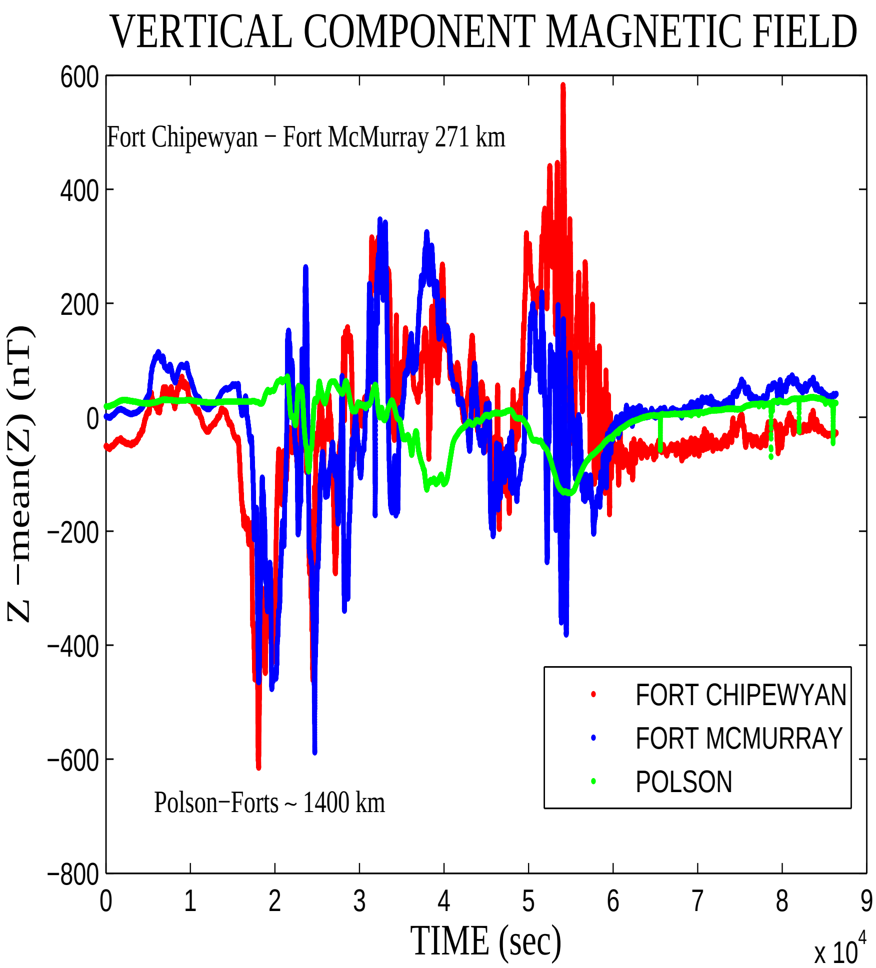
The vertical component. Note that the short period terms correlate but sometimes positively, or negatively and with a delay.



The X component (north). Note that the long period variations highly correlate.



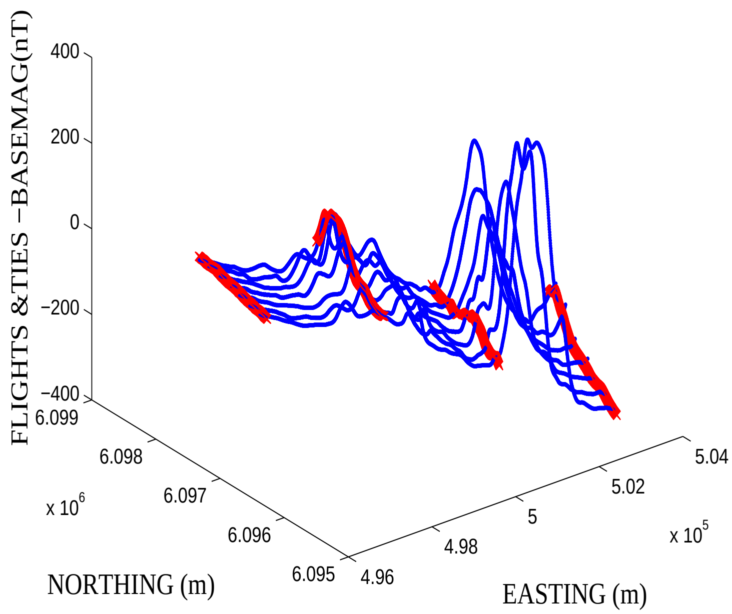
The Y component (east).



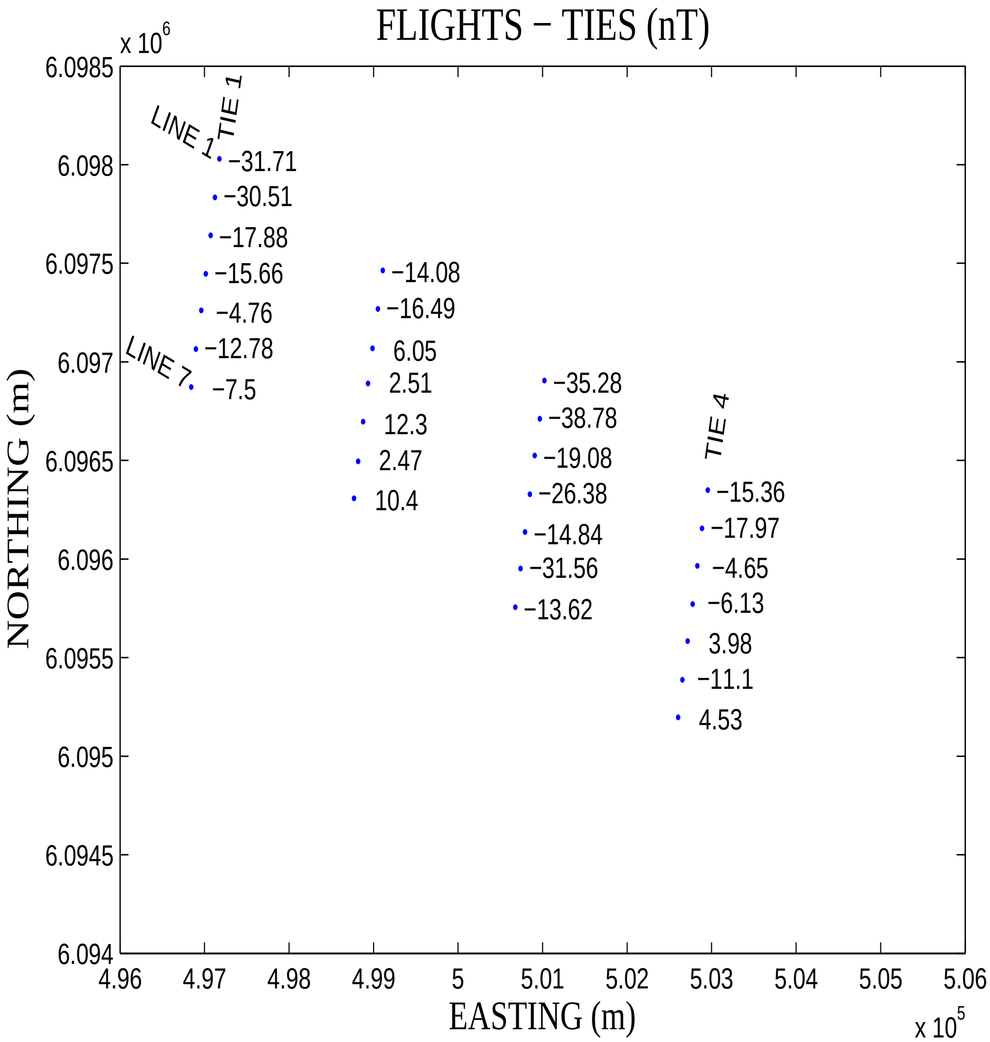
Here are three stations in Montana, Alberta and NWT. Shown is the vertical component minus the mean value at each station for the day. Polson to the two Forts is about 1400 km and the two forts are about 270 km apart. During the middle of the day there was a fairly intense magnetic storm. Before and after the storm the two Forts correlate to within about 10 nT. During the storm even these two relatively close stations can see quite different variations in magnetic field. Notice that the variation in the field at the two stations in the North is much greater than at the one station in the south.

LEVELING

There is no standard leveling procedure and each operator has his own proprietary procedure. However they do share a lot in common. The crossover points of the lines and ties rarely produce the same values due to, residual diurnals, or residual compensation etc. Leveling is the process by which these *intersection errors* are systematically apportioned between the ties and lines.

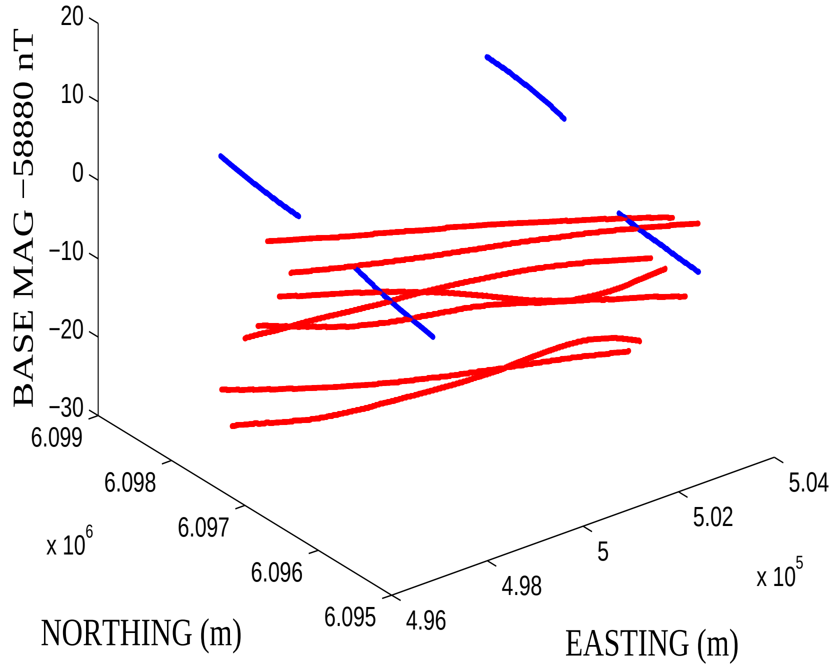


This is a small section (7 flight lines and 4 tie lines) of a much larger aeromag survey flown in Manitoba. The base station was located about 40 km SW of the survey area. The base corrected flight line data are shown as blue dots and the base corrected tie line data are shown as red crosses. The differences are only a few nT and would require small leveling changes.



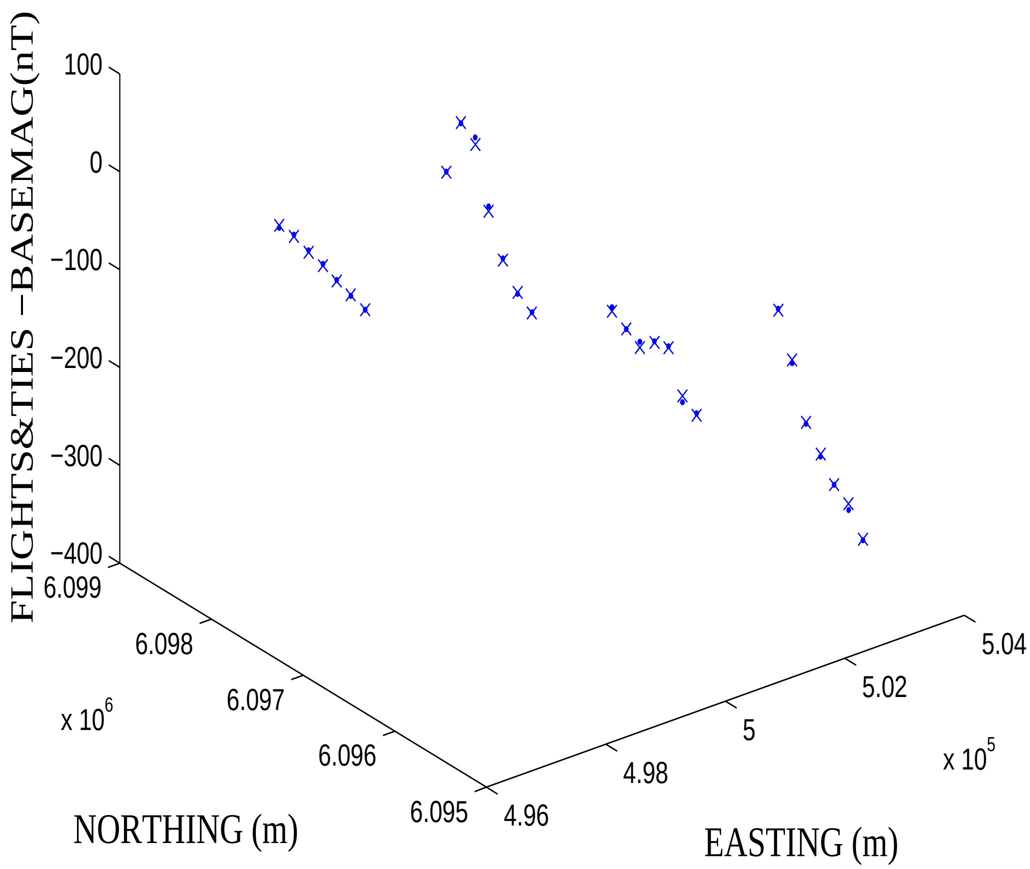
These are the intersection errors in the raw data. The range is about 50 nT. Most of this is due to the diurnal.

The tie lines are typically shorter, so these can be considered to be drift free. This is not strictly true, but typically the ties are flown over a much shorter time span than the flights, so less variation during a tie line is expected.

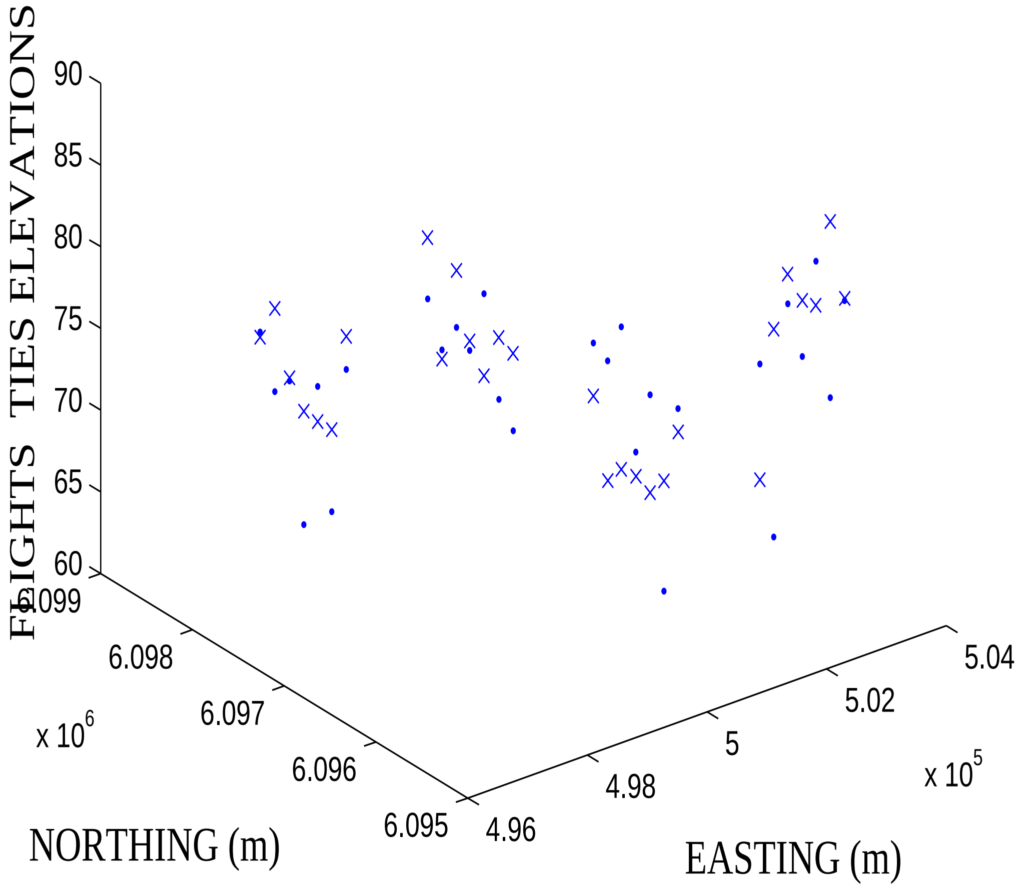


Here are the base observations minus 58880 nT at the flight (red) and tie (blue) times and coordinates. This is what the data would look like minus the static field and assuming that the temporal variations over the survey area are exactly the same as at the base station. Note that the tie lines are short, so during a tie line flight there is very little variation. It looks like two tie lines were flown per day. The flight lines are longer, so there is some variation (a few nT) during each flight.

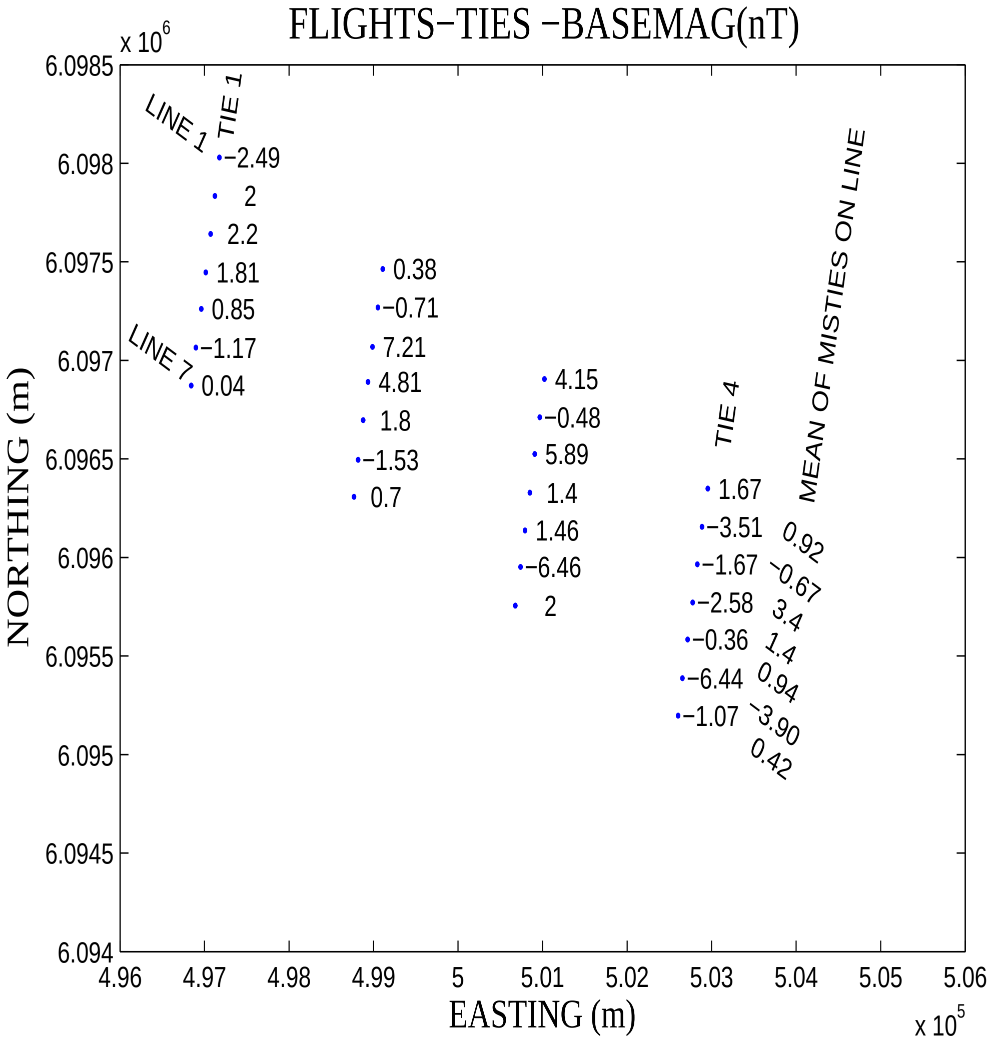
If significant drift during the tie line collection is suspected, or significant elevation differences are known a rough leveling procedure may be done on the tie lines.



This is the collection of crossovers on the base corrected flights (dots) and ties (crosses).

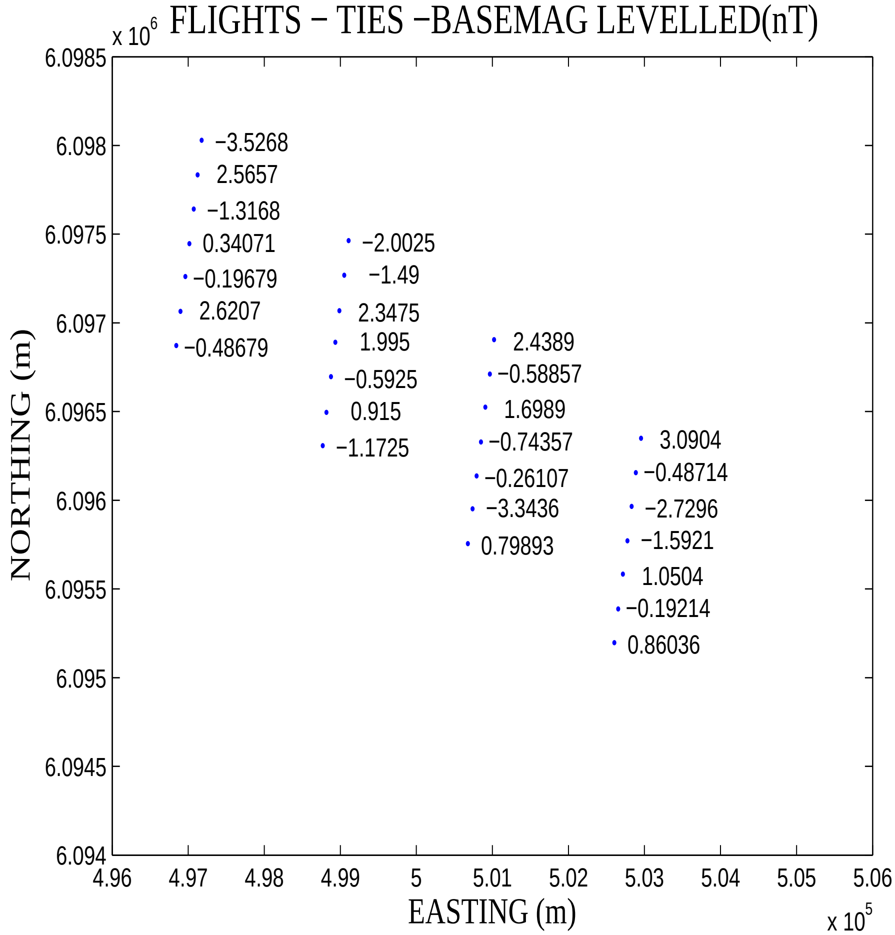


These are the flight line sensor elevations (dots) and tie line elevations (crosses) at the crossovers. There are no large height differences, so no concern with elevation differences.



This is the crossovers with the intersection errors printed. Without doing any leveling of the tie lines (that is assuming there is no residual diurnal drift, or elevation difference). The standard deviation of all the intersection errors is only 3.1 nT. The last line on the right are the mean differences along each line. These indicate the flight lines are reading high (or low) by this amount on average. Therefore this amount should be subtracted from each line. The standard deviation of the corrected crossovers is 2.34 nT, so a small improvement was made.

A second approach is to ask what constant value must be added to each flight line and each tie line to give the smallest possible standard deviation for the intersection errors? The result of this procedure is a standard deviation for corrected intersection errors of 1.81 nT, so a small improvement on the previous method was made. A further improvement could be to assume a linear variation with time along each individual line and solve for that.



Intersection errors after first order leveling that added a constant to each flight and tie. There are still errors due to second order effects like spatial variability of the temporal changes and differences in sensor height.

microleveling

When the data are gridded, a banding along the flight lines with a wavelength similar to the tie line spacing, and narrow in the tie line direction, and a similar banding in the tie line direction with wavelength about twice the line spacing is sometimes observed. The details of microleveling are proprietary, and probably differ greatly from company to company. Typically, a 2D band pass filter is applied to the gridded data. Power with wavelengths near twice the line spacing in the tie line direction and very short wavelength in the line direction, and twice the tie spacing in the line direction and very short in the tie direction is cut.

GRAVITY		MAGNETICS	
WHAT'S GOOD	WHAT'S BAD	WHAT'S GOOD	WHAT'S BAD
	Instrument drifts	instrument does not drift	
	complicated reductions	no free air or Bouguer reductions	
quantitative interpretation, only source density contrast is important			qualitative interpretation mostly, induced and remanent magnetizations are possible
directly related to lithology			sometimes directly related to lithology
	slow data collection leveling and horizontal control more important	rapid data collection looser controls on vertical and horizontal	