

Supplementary material 1

THEORY

Summary of formulae and comments on their physical meaning

The theory is mostly inspired by Pedersen and Rasmussen (1990), Beiki and Pedersen (2010) and comes from our own papers Kalvoda et al., (2013) or Křeločnik et al., (2014, 2017a, 2020a). These last two references are the proper source of information for this *Supplement*.

The *disturbing static global gravitational potential* outside the masses of a celestial body (planets, moons) in the spherical harmonic expansion can be written as:

$$T(r, \varphi, \lambda) = \frac{GM}{r} \sum_{l=2}^{\infty} \sum_{m=0}^l \left(\frac{R}{r}\right)^l (C'_{l,m} \cos m\lambda + S_{l,m} \sin m\lambda) P_{l,m}(\sin \varphi), \quad (\text{A1})$$

where GM is a product of the universal gravitational constant G and the mass M of the planet (also known from satellite analyses as the geocentric gravitational constant in the case of the Earth), r is the radial distance of an external point where T is computed, R is the radius of the planet (which can be approximated by the semi-major axis of a reference ellipsoid), $P_{l,m}(\sin \varphi)$ are the Legendre associated functions, l and m are the degree and order of the harmonic expansion, (φ, λ) are the (planeto)centric latitude and longitude, and $C'_{l,m}$ and $S_{l,m}$ are the *harmonic geopotential coefficients* (also known as *Stokes parameters*); fully normalized, $C'_{l,m} = C_{l,m} - C_{l,m}^{el}$, where $C_{l,m}^{el}$ belongs to the reference ellipsoid. The word “*disturbing*” here means the difference between the total gravitational potential of the actual body and the gravitational potential of a reference body, i.e. the reference ellipsoid, usually taken as a rotational ellipsoid with some flattening on the poles due to the rotation of that body.

A set of numbers $C'_{l,m}$ and $S_{l,m}$, presented to a maximum degree L_{max} , is called the *gravity/gravitational (field) model of the Earth*. (“Gravity” means gravitational effect plus effect of centrifugal force of the studied body).

There is $l \times (l-1)$ terms in such a model, if is complete to the maximum degree and order l, m (or d/o) and if a few first (lowest degree) terms are

not omitted (sometimes these terms are set at zero, due to reasons which will not be discussed here).

The gravity/ gravitational models are usually based on a great amount of diverse satellite and terrestrial data collected from around the world over a long time; then such a result is known as a high-resolution “*combined model*” (e.g., GEM 2008, Pavlis et al., 2012; EIGEN 6C4, Förste et al., 2014) in a contrast to “*satellite-only models*”.

Let us recall that $C'_{l,m}$ and $S_{l,m}$ are considered to be constants (excluding a few lowest degree zonal harmonics, which have often been published with a secular trend and semi/annual or other time variable components). We speak about *static gravity/gravitational models*.

There are also *variable gravity/gravitational field solutions*, derived from the global satellite data (mainly from the GRACE mission). They are based on short arc solutions (from observations gathered for one month or a shorter interval), so they are available for a much lower L_{max} than the static models (say to $d/o = \sim 100$ instead of ~ 2000). This is not our topic.

The *gravity/gravitational aspect* is a functional/function of the gravity (gravitational) field potential T . It can be its derivative or any other function, often non-linear. We work with the following *gravity aspects* (descriptors): the gravity anomaly (or disturbance) Δg , the Marussi tensor (Γ) of the second derivatives of the disturbing potential (T_{ij}), two gravity invariants (I_j), their specific ratio (I), the strike angles (θ) and the virtual deformations (vd) – Klokočník et al., (2017a, 2020a).

All the gravity aspects together provide thorough information about the density anomaly due to the causative body that is more complete than, for example, information that the traditional and usual gravity anomalies themselves could yield. The set of gravity aspects informs about location, shape, orientation, a tendency to a 2D or 3D pattern, and stress tendencies and may partly simulate “dynamic information” although the input data are always the same – those harmonic geopotential coefficients $C'_{l,m}$ and $S_{l,m}$ of a static gravity field model. The whole theory is arranged in such a way that we cannot use any input other than the harmonic coefficients of a gravity model.

The spherical approximation of the *gravity anomaly* Δg (free air, without any geophysical model) is computed as the first radial derivative of T by

$$\Delta g = -\frac{\partial T}{\partial r} - 2\frac{T}{r}. \quad (\text{A2})$$

Instead of (A2), one can use the *gravity disturbance*, which is as (A2), but without the second term (often small). The gravity anomalies/disturbances are computed from measurements by ground gravimeters, airplane or marine gravimeters or derived from measurements performed by means of satellite altimetry.

The gravity gradient tensor Γ (the *Marussi tensor* or simply the *gravity tensor*) is a tensor of the second derivatives of the disturbing potential T of the gravity field model. The Marussi tensor was considered the centerpiece of traditional differential geodesy; up to the second order this tensor systematically synthesizes all the dynamical and geometric properties of the Earth's gravity field (see Klokočník et al. (2020a) for references).

The tensor Γ is given in the local north-oriented reference frame (x, y, z), where z has the geocentric radial direction, x points to the north and y is directed to the west (Pedersen and Rasmussen, 1990):

$$\Gamma = \begin{bmatrix} T_{xx} & T_{xy} & T_{xz} \\ T_{yx} & T_{yy} & T_{yz} \\ T_{zx} & T_{zy} & T_{zz} \end{bmatrix} = \begin{bmatrix} \frac{\partial^2 V}{\partial x^2} & \frac{\partial^2 V}{\partial x \partial y} & \frac{\partial^2 V}{\partial x \partial z} \\ \frac{\partial^2 V}{\partial y \partial x} & \frac{\partial^2 V}{\partial y^2} & \frac{\partial^2 V}{\partial y \partial z} \\ \frac{\partial^2 V}{\partial z \partial x} & \frac{\partial^2 V}{\partial z \partial y} & \frac{\partial^2 V}{\partial z^2} \end{bmatrix} \quad (\text{A3})$$

Outside of the body masses Γ satisfies Laplace's differential equation, i.e. the trace of the Marussi tensor (A3) is zero. The tensor Γ is symmetric ($T_{yx}=T_{xy}$, $T_{zx}=T_{xz}$, $T_{zy}=T_{yz}$) and harmonic ($T_{xx}+T_{yy}+T_{zz}=0$); it contains nine components, but just five linearly independent components.

Gravity gradiometry is the measurements of T_{ij} . The gravimeters measure the first derivative of T , i.e. the accelerations Δg , the gradiometers measure the second derivatives of T . From the actual formulae for their computation one can see that the gravity gradients are more sensitive to the close-by mass distribution (density anomalies) than the gravity accelerations. We explain in Sect. 2.3 of (Klokočník et al. 2020a).

Terrestrial gradiometers (torse balances) to measure T_{ij} are known from geophysics (see any textbook), but they were not too successful in practical use (too noisy). Now, owing to technical progress, they can successfully be used for measurements on board of airplanes and are used for prospection at local scales.

The *GOCE* mission (Gravity and Ocean Circulation Explorer, ESA) was the first (2009) and until now (2022) still the last gradiometric instrument working successfully in space on low orbit (measuring six of

T_{ij} components million times during its ~5-year lifetime at a carefully selected orbit fulfilling special criteria on orbital resonances), based on microaccelerometers (a pair of them in each spatial direction x,y,z).

The Marussi tensor has already been used locally (this means in areas of a few per few kilometres) for petroleum, metal, diamond, groundwater, etc., explorations (for more information and for further references see Klokočník et al. 2020a).

The Marussi tensor is a rich source of information about density anomalies; it provides useful details about the target objects shallowly located beneath the Earth's surface. This extra information can be used by tensor imaging techniques to enhance the source anomalies; it has been tested for local features (economic minerals, oil and gas deposits, fault location, etc.). The tensor components are used at local scales to identify and map the geological contact information, either the edges of the source targets or the structural/stratigraphic contact information. The horizontal components identify the shape and the geological setting of a responsible body. The quantity T_{zz} is best suited for target body detection; T_{zz} helps to define the isopath/density relationships of body mass with relation to its geological setting, see e.g. Saad (2006).

Under arbitrary coordinate transformation, any gravity field and any Γ have just three global *gravity invariants* which remain constant. Here they are labelled I_0 , I_1 , and I_2 :

$$I_0 = \text{trace} (\Gamma)$$

and this one is zero outside the masses of the studied body (known also as the Laplace equation). The remaining two invariants read in general:

$$I_1 = \frac{1}{2} [\text{trace} (\Gamma)^2 - \text{trace} (\Gamma^2)],$$

$$I_2 = \det (\Gamma).$$

This can be transformed (using the components of Γ in Eq. 3) to:

$$\begin{aligned} I_0 &= T_{xx} + T_{yy} + T_{zz} \\ I_1 &= (T_{xx}T_{yy} + T_{yy}T_{zz} + T_{xx}T_{zz}) - (T_{xy}^2 + T_{yz}^2 + T_{xz}^2) = \\ &\quad \sum_{\{i,j\} \in \{x,y,z\}} (T_{ii}T_{jj} - T_{ij}^2) \\ &= - (T_{xx}^2 + T_{yy}^2 + T_{zz}^2 + T_{xy}^2 + T_{yz}^2 + T_{xz}^2), \end{aligned} \tag{A4}$$

$$\begin{aligned} \mathbf{I}_2 &= \det(\mathbf{\Gamma}) = \\ &= T_{xx}(T_{yy}T_{zz} - T_{yz}^2) + T_{xy}(T_{yz}T_{xz} - T_{xy}T_{zz}) + T_{xz}(T_{xy}T_{yz} - T_{xz}T_{yy}) \end{aligned} \quad (\text{A5})$$

The invariants are mathematically independent of the coordinate system chosen, so invariant (“resistant”) with respect to any rotation. The invariant $\mathbf{I}_0$ is useful for numerical checks of the actually measured T_{ii} . The invariant $\mathbf{I}_1$ is the sum of the six products of two tensor coefficient matrix elements, a nonlinear functional model with regard to the geopotential harmonics. The invariant $\mathbf{I}_2$ is the determinant “det” of $\mathbf{\Gamma}$.

The invariants can be looked upon as non-linear filters enhancing sources with big volumes (Pedersen and Rasmussen 1990). They discriminate major density anomalies into separate units. It is useful and helpful that the resultant computed anomaly response retains the same shape and orientation, i.e. it is independent of the observer’s choice of axes; this is significant for interpretation when mapping geological structures.

Pedersen and Rasmussen (1990) showed that the ratio I of the invariants $\mathbf{I}_1$ and $\mathbf{I}_2$, defined as

$$0 \leq I = -\frac{(\mathbf{I}_2/2)^2}{(\mathbf{I}_1/3)^3} \leq 1, \quad (\text{A6})$$

always lies between zero and unity for any potential field. If the causative body is strictly 2D (flat), then $I=0$. Thus, the ratio can be an indicator of two-dimensionality, sometimes called the “2D factor”. If $I=0$, then we have the necessary but not sufficient condition for two-dimensionality. If the causative body – as seen from the observation point – looks more “3D-like” (for example, a volcano), then I grow and eventually approach 1.

The gradient tensor $\mathbf{\Gamma}$ contains information about subsurface strike (stress) directions. Pedersen and Rasmussen (1990) defined the *strike angle* θ (strike lineaments, strike direction) as follows:

$$\tan 2\theta_s = 2 \frac{T_{xy}(T_{xx}+T_{yy})+T_{xz}T_{yz}}{T_{xx}^2-T_{yy}^2+T_{xz}^2-T_{yz}^2} \quad (\text{A7})$$

where θ is estimated within a multiple of $\pi/2$; and only one value represents the main direction of $\mathbf{\Gamma}$. Provided that the ratio I in (A6) is small, the strike angle may indicate a dominant 2D structure. If one were able to rotate with the structure in such a way that the elements of the first row and first column of $\mathbf{\Gamma}$ were identically equal to zero, then one would reach a

“correct” direction of “stress fields” described by θ (Beiki and Pedersen, 2010).

Mathematically, θ is the main direction of $\mathbf{\Gamma}$. Geophysically, it is an important direction for the ground structures; it may indicate areas with a lower porosity or “stress directions”.

The strike angles usually show chaotic directions. Sometimes, they are oriented dominantly (aligned) into one direction; we say they are “combed” (Figs. 2-5, 9a, b). The theory of combed θ has been presented for the first time by Klokočník et al. (2019). The combed values, mostly for small I ($I < 0.3$) may predict, among others, possible oil or gas fields (e.g., Klokočník et al. 2020a).

The situation remains, however, not unambiguous when solely using the gravity data. The reason is that not only can oil and gas fields be detected by the combed θ but also groundwater reservoirs, water-filled depressions, paleolakes or stress fields after impact at and near the impact craters. The combed θ probably relates to changes of porosity, for example possibly revealing the porosity increase due to an impact pressure deformation. It is evident that we need additional information to the gravity aspects; this means geological or geophysical information, namely magnetic anomalies, archaeological data, etc.

Now we define “*virtual deformation*” (vd), introduced for the first time in Kalvoda et al. (2013) and Klokočník et al. (2014). It is analogous to the tidal deformation from geodesy and geophysics; one can imagine the directions of such a deformation due to “erosion” brought about solely by gravity.

If there were a tidal potential represented as in our case by $T(A1)$, then horizontal shifts (deformations) would exist due to this and they could be expressed in the north-south direction (latitude direction) as

$$u_{\Phi} = l_s \frac{1}{g} \frac{\partial T}{\partial \varphi} \quad (A8)$$

and in the east-west direction (longitudinal direction) as

$$u_{\Lambda} = l_s \frac{1}{g \cos \varphi} \frac{\partial T}{\partial \lambda} \quad (A9)$$

where g is the gravity acceleration $9.81 \text{ m}\cdot\text{s}^{-2}$, l_s is the elastic coefficient (called the Shida number) expressing the elastic properties of the Earth as a planet (generally $l_s = 0.08$), φ and λ are the geocentric latitude and

longitude of the point P where we measure T ; and the potential T is expressed in $[\text{m}^2 \text{s}^{-2}]$. In our case, T is represented by Eqs. (A1), (A8) and (A9). The practical problem is that the actual values of the Shida parameters l_s for the Earth's surface (for the specific locations) are not known; thus, we will know (A8) and (A9) and subsequent quantities only as relative values.

The formalism of continuum mechanics was applied to derive the main directions of the deformations, meaning to transform the horizontal shifts to a small deformation (e.g., Brdička et al. 2000). The tensor of a small deformation $\mathbf{E}$ is defined as a gradient of the horizontal shifts (A8) and (A9):

$$\mathbf{E} = \begin{pmatrix} \epsilon_{11} & \epsilon_{12} \\ \epsilon_{21} & \epsilon_{22} \end{pmatrix} = \begin{pmatrix} \frac{\partial u_x}{\partial x} & \frac{\partial u_x}{\partial y} \\ \frac{\partial u_y}{\partial x} & \frac{\partial u_y}{\partial y} \end{pmatrix}. \quad (\text{A10})$$

The tensor $\mathbf{E}$ has two parts: $\mathbf{e}$ is the symmetrical, and $\mathbf{\Omega}$ is the anti-symmetrical part:

$$\mathbf{E} = \mathbf{e} + \mathbf{\Omega} = (\mathbf{e}_{ij}) + (\mathbf{\Omega}_{ij}) \quad (\text{A11})$$

The symmetrical tensor $\mathbf{e}$ reads:

$$\mathbf{e} = \begin{pmatrix} \mathbf{e}_{11} & \mathbf{e}_{12} \\ \mathbf{e}_{21} & \mathbf{e}_{22} \end{pmatrix} = \begin{pmatrix} \epsilon_{11} & (\epsilon_{12} + \epsilon_{21})/2 \\ (\epsilon_{12} + \epsilon_{21})/2 & \epsilon_{22} \end{pmatrix}, \quad (\text{A12})$$

The parameters of deformation are:

$\Delta = \mathbf{e}_{11} + \mathbf{e}_{22}$	total dilatation	(A13)
$\gamma_1 = \mathbf{e}_{11} - \mathbf{e}_{22}$	pure cut	
$\gamma_2 = 2\mathbf{e}_{12}$	technical cut	
$\gamma = (\gamma_1^2 + \gamma_2^2)^{1/2}$	total cut	
$a = \frac{1}{2} (\Delta + \gamma)$	major semi-axis of the ellipse of deformation	
$b = \frac{1}{2} (\Delta - \gamma)$	minor semi-axis of the ellipse of deformation	
$\alpha = \frac{1}{2} \text{atan}(\gamma_2 / \gamma_1)$	direction of the main axis of deformation.	

Thus, we can write

$$vd = (a^2 + b^2)^{1/2}$$

$$\text{and} \quad vd = f_1(\gamma, \Delta, e_{ij}) = f_2(A10) = f_3(A9, A8, A1). \quad (A14)$$

Note that different specialists make use of different terminology for the same or similar quantities quoted in (A13).

The derivatives of the disturbing geopotential obtained originally as the directional values (A8), (A9) were transformed to small positional deformations, or shifts. But basically, the information content of the Marussi tensor and of vd is the same, only exposed in different ways.

To illustrate vd , the semi-axes $\mathbf{a}$, $\mathbf{b}$ of the deformation ellipse are computed. As we already know, the local values of l_s are not known, and, in turn, only the main directions of vd (and not their amplitudes) can be computed.

It is very interesting and may sounds unusual that vd provide dynamical information, even though they are, as well as all the gravity aspects mentioned here, computed from static gravity models (represented by a set of $C'_{l,m}$ and $S_{l,m}$).

As already mentioned, the vd is analogous to the tidal deformation and characterizes the “tensions” (directional compression and dilatation) generated by the causative body (Kalvoda et al. 2013). We can understand the vd as a principal axis transformation from the horizontal gradients of the deflections of the vertical (A8) and (A9). Since the potential is forward modelled from the topography, it is also related to curvature of topography (Klokočník et al. 2014, p. 93).

Table. 1. Physical and mathematical significance of the gravity aspects.

Gravity aspects [unit]	Name reference to formula in this Supplement	meaning
Δg [mGal]	gravity anomaly gravity perturbation SM1-A2	traditional quantity used in geodesy and geophysics, radial derivative of the disturbing potential
T_{zz} [E]	radial component of the Marussi tensor SM1-A3	Marussi tensor, tensor of the second derivatives of the disturbing potential
I_1 [s ⁻⁴]	gravity invariant SM1-A4	one of the gravity invariants
I_2 [s ⁻⁶]	gravity invariant SM1-A5	one of the gravity invariants
I [-]	ratio of the gravity invariants SM1-A6	indicator of shape of the causative body
θ [deg]	strike angle SM1-A7	stresses, tensions θ can be the main direction of the Marussi tensor
vd [-]	virtual deformation SM1-A13	dilatation, compression

Combed strike angles

The *combed strike angles* are strike angles θ oriented roughly in one and the same direction. (Fig. A1a). We define the combed coefficient *Comb* for θ as a measure or degree of θ being combed; it is a relative value in the interval $(0,1)$, where 0 means to be “not combed” (the vectors of θ are in diverse directions) and 1 means to be “combed” (perfectly aligned, the vectors of θ are oriented into one prevailing direction). They are different ways how to arrange such a tool. We comment on them.

The following are the input data to the statistics:

$$\theta_i \in (-90^\circ, 90^\circ), i = 1, \dots, n$$

for n pixels in the studied area or zone. We compute the main direction of the combed θ as the mean value of θ_i ; let us denote it as θ_{Comb} :

$$\theta_{Comb} = \frac{\sum_{i=1}^n \theta_i}{n}$$

by choosing the angles θ_i either in the interval $(-90^\circ, 0^\circ)$ or in the interval $(0^\circ, 90^\circ)$. We use the following important condition:

$$\forall (|\theta_i - \theta_{Comb}| > 90^\circ): \theta_i = 180^\circ - |\theta_i|$$

which means that even two angles θ_i in opposite directions are counted as one direction. For example: for $\theta_{Comb} = 80^\circ$ and $\theta_i = -80^\circ$, a deviation from the main *Comb* direction is 20° .

Let us define a root mean square value of scatter (variance) of θ_i for n pixels as:

$$rmsv = \sqrt{\frac{\sum_{i=1}^n (\theta_i - \theta_{Comb})^2}{n}}$$

Then the required looked-for value of the main *Comb* direction can be defined as

$$Comb = 1 - \frac{rmsv}{90^\circ}$$

As a measure of the degree of θ “being combed”, we make use of the relative values of θ_i :

$$\theta_i^{relat} = 1 - \frac{abs(\theta_i - \theta_{Comb})}{90^\circ} \quad (A14)$$

So the *Comb* value is the main local direction of the tested set of θ_i in the given region; the departures of the individual θ_i from *Comb* are plotted in a preselected optimum size of n rectangular pixels in a relative scale; these are the values of (A14); if some of θ_i fit in the main direction of *Comb*, then the pixel has the value 1; if not, then the values (A14) are in the interval $\langle 0,1 \rangle$. This serves as a simple statistical evaluation to compare areas with combed strike angles to those with “non-combed” strike angles.

If *Comb* is smaller than 0.55, we say that θ_i of the given region are “not combed”; if *Comb* > 0.65, we say θ_i are “combed”, for *Comb* > 0.99 highly or perfectly combed.

a *combed strike angles aligned in the studied area =*
one-way oriented strike angles

b *dishevelled strike angles*
=strike angles are in diverse directions

comb_factor → 1

comb_factor → 0

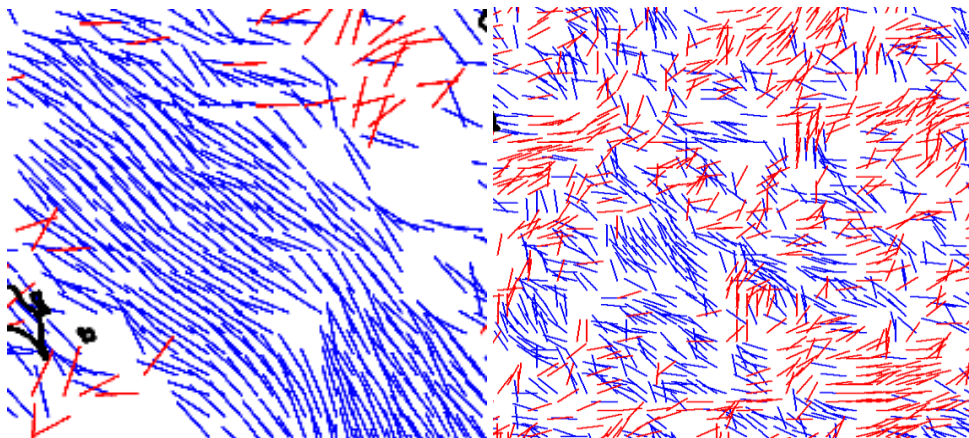


Fig. A1-1a,b: The strike angles θ [deg]:

(a) combed (aligned into one prevailing direction)
and (b) chaotic

We can work with the *Comb* factor according to (Kletetschka et al. 2022). For the point P in question, in the studied area, we select points B_i around P . The “comb coefficient” in P is then equal to the weighted mean value of the scalar product of the unit vector in P with the unit vectors in B_i . The *Comb* factor lies always in the interval $<0,1>$. For perfectly aligned (combed, correlated, parallel) vectors from the surroundings of the point P considered in the relevant analysis, we have *Comb* = 1.

Alternatively, we can compute the *Comb* factor by the following way. Given: the unit vector $\mathbf{X}$ (x_x, y_x) in the point of interest P and the unit vectors $\mathbf{B}_i$ (x_{Bi}, y_{Bi}) in the surroundings points. We work in a local planar coordinate system x, y , the origin is in P (see Fig. A1-2 on the next page).

The scalar product of the unit vectors $\mathbf{X}$ and $\mathbf{B}_i$ is by definition

$$\mathbf{X} \cdot \mathbf{B}_i = x_x \cdot x_{Bi} + y_x \cdot y_{Bi} = \cos(ai)$$

where a_i is an angle between $\mathbf{X}$ and $\mathbf{B}_i$.

The *Comb* factor is then defined as “weighted means” of the sum of the scalar products above. Weight w_i is defined as

$$w_i = 1/(dsi)^2$$

where (dsi) is distance between P and surrounding points (see Figure.). Influence of points far from P is then lower. We define

$$Comb = \frac{\sum_1^n w_i \cos(ai)}{\sum_1^n w_i}$$

where n is number of the surrounding points. From this formula it is clear, that *COMB* depends on *cosine* function. (Remember that *cos* is very flat around zero). Example: If all $a_i = 0$, then *COMB* = 1, to be read: the strike angles are perfectly combed.

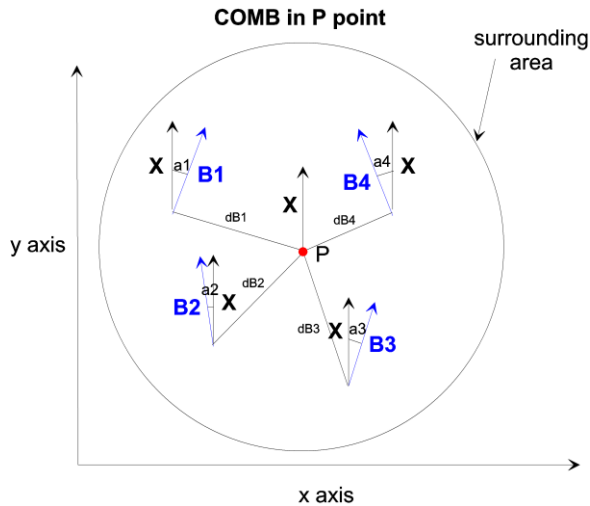


Fig. A1-2: how we actually compute the *Comb* factor to characterise degree of alignment of the strike angles in a predefined polar zone

Tutorial – more about the strike angles

The strike angles – physical meaning and interpretation possibilities

The combed (aligned) strike angles θ bear information about a higher porosity, about changes of porosity/density and stress fields connected with various geological features and processes. There is a plenty of such observations and applications discovered till now.

Hallo–Alignments around the impact craters. After the impactor's explosion and during the post-impact evolution, the strike angles may have changed their original pre-impact direction. The original direction was given by properties of the local “material” and previous geo-evolution. Due to the impact event, there is a special change revealed by the strike angles while the other gravity aspects exhibit nothing surprising (e.g., negative Δg or T_{zz} inside the crater, both positive in the rim; fragmented extremes of I_1 and I_2 around the crater, negative vd inside). The strike angles inside and around the crater become significantly combed. They may provide evidence about the direction of deformations. They seem to

be *parallel to the weakness in the strength of the rock*, e.g., direction schistosity and or presence of faults or micro-fault zones. The impact event energy is imprinted into the gravity aspects. There may be even no crater (an airburst may happen), but the release of the energy of the impactor triggers the stress in the rock along the cleavage planes while the stress perpendicular to these planes is preserved. This internal anisotropy is what is being detected by the strike angles.

Diverse post-impact influences and changes may reach large distance from the source crater. The bigger impacting body, the more energy and thus the bigger crater, the more intensive and extensive consequences which remain written into the strike angles. The impacts enhance porosity and permeability in rocks of all parts of the impact structures and around it, sometimes to large distances. The consequences can be, for example, a triggering mechanism to move rocks/minerals from deeper layers to closer to the surface (e.g., Vredefort or Sudbury craters). The impacts may lead to crustal deformation and deposit extensive blankets of material local-dependent ejecta with economic minerals and hydrocarbons.

It was recognized that the combed θ might be tested and used as a *preliminary diagnostic tool* for further exploration of sites with target deposits (Klokočník and Kostecký, 2014; Klokočník et al., 2021a). Important to note that no specific digging is required to apply this method. The method can help to decide to exclude or include the region from more detailed later investigation, based on the traditional data and procedures, thus potentially saving money supporting or avoiding drilling on the spot.

Linearly combed strike angles – Rivers, canyons, faults, groundwater or ice, permafrost on the Earth (regolith on the Moon), paleo-lakes, oil and gas deposits, sedimentary layers, mud – their appearance is signalled by the combed (aligned, straight) strike angles θ , and of course, in a specific way, by the other gravity aspects. This is not only a reaction to a higher porosity of that place in comparison with its surrounding, but also a reaction to certain flows, such as slope movements or land-sliding, tensions, stresses, a force connected with a genesis and evolution of such features.

Reservoir source rocks (TD) are porous and absorbent, and can be saturated with water, oil and gas in various combinations. They should thus provide contrasting (higher) porosity, and relatively lower density, with the relevant changes in the gravity signal in a form of the gravity aspects as they are known for lower density causative bodies on the Earth (depending on the background beneath the sedimentary layers). A similar situation can be expected on Mars (Klokočník et al., 2023b) as for the

ground water or sediments, possibly bearing hydrocarbons. Potential hydrocarbons are more likely to be found in places with ordered combed structures of huge size, which probably represent places with linear (uniform) water flow (as on the Earth) that would carry (water lift) and concentrate hydrocarbons in suitable places, e.g., under impermeable clay layers.

Ground Resolution

A usual simple estimate of the smallest representable feature of the gravity field or the **shortest half-wavelength L_{half} (as a distance on a sphere) that can be resolved with all C_{lm} , S_{lm} to max. degree L_{max}** , is

$$L_{half} = \pi R / L_{max}$$

or equivalently, taking the circumference $2\pi R = 40\,050$ km for the Earth:

$$L_{half} = 2 \pi R / (2 L_{max}) = 40\,050 / (2 L_{max}).$$

This L_{half} is what we call the “ground resolution”. For $L_{max} = 300$ and 2190 , we have $L_{half} = 67$ and 9 km, respectively. The former limit roughly corresponds to the resolution of the GOCE data alone, and the latter belongs to the combined models, EGM 2008, EIGEN 6C4 or SatGavRET2014 (in the case of the Earth).

Body	R [km] rounded	max d/o of model used	resolution [km]
the Earth	6380	2190	9
the Moon	1740	600	9
Mars	3400	80-120	130-90
Venus	6050	120	160

Tutorial – a simple model of the signal falling with the increasing distance from its source

To illustrate in a simple example, how the various gravity aspects behave (decrease) with increasing distance from the causative body (in fact a density anomaly), it is assumed that the source body is a mass point, with a selected value of GM (for its definition see Eq. A1).

It is obvious from the theory that in this simplified case, with arbitrary $GM=Gm$ for a sphere, some of the gravity aspects do not exist or equal

zero (θ and vd). The remaining aspects are shown in Fig. A1-3; their spherical simplification leads to $I_1 = -3 (Gm)^2 r^{-6}$ and $I_2 = -2 (Gm)^3 r^{-9}$.

A diverse speed of decrease of the “gravitational signal” can be seen with the increasing depth of the source mass body for various aspects. The slowest decrease turns out for the gravity anomaly (with r^2). The invariants decrease quickly (with r^6 and r^9).

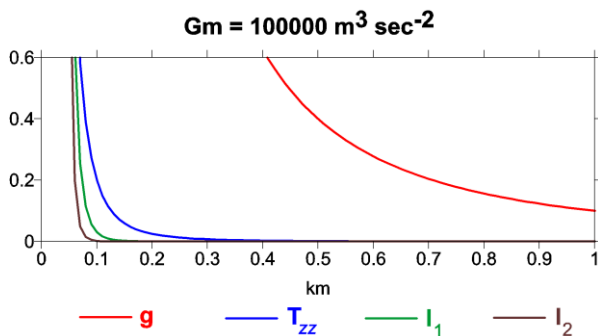


Fig. A1-3. An example of a decrease in the values of the gravity aspects of the model potential with increasing distance (depth) from the source. On the x axis there are depths in kilometres, on the y axis there is an arbitrary quantity (for a simple intercomparison of all aspects). This is an illustrative case for a mass point with a randomly selected value of Gm .

Fig. A1-3 tells us that aspect I_1 or I_2 best describes the density anomaly in shallow depths under the surface while Δg and T_{zz} are also related to deeper structures. While all of these aspects have the same “common mother” (a recent gravity field model in terms of Stokes parameters), they have different properties (behaviour, sensitivity, relationship to various effective depths of the causative body) and describe the same density anomaly from different views. Thus, their entire set provides much more information, more complex, more complete information (but also more complicated to compute) about the density anomalies than only the traditional Δg could provide. This is the crucial reason to use the gravity aspects instead of Δg and to apply them all together.

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(for more references see the main text)

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