Chapter No.: 1
Title Name: Misra

1.1 Seismic Reflection Method

Whenever a reflection seismic section is mentioned, something similar to Figure 1.1 comes to mind. The process leading to the generation of such a section is briefly discussed in this chapter. There is a large volume of literature detailing all the processes and their variations (e.g., Sheriff and Geldart, 1995; Yilmaz, 2001; Liner, 2004; Ashcroft, 2011; Herron and Latimer, 2011; Onajite, 2014). Only a brief account is given here to build the platform for the following chapters.

Seismic waves propagate through the Earth at velocities that depend on the acoustic impedance and density of the medium through which they travel. The acoustic impedance, $Z$, is expressed by (Liner, 2004):

$$ Z = V \rho $$ (1.1)

where $V$ is the seismic wave velocity and $\rho$ is the rock density. If the rock varies in density in several directions, one can work with the "effective density" deduced in Mukherjee (2017, 2018, in press).

When a seismic wave propagating through the Earth encounters a boundary between two materials of different acoustic impedances, a part of the energy reflects off the interface while the remainder refracts through it. Seismic reflection prospecting involves generating seismic waves at the surface, which propagate into the subsurface, and capture the reflected wavefronts from the different interfaces while propagating. At each layer most of the energy is transmitted or refracted and a part reflects back (Sheriff and Geldart, 1995; Yilmaz, 2001; Liner, 2004; Ashcroft, 2011; Herron and Latimer, 2011; Onajite, 2014).

To generate the disturbance, a ‘shot’ or a vibration is made on the sea surface or on Earth’s surface on land. As the wave propagates into subsurface, each layer reflects the wave at multiple incidence angles and these reflected waves are measured at the surface by receivers, which are hydrophones on water and geophones on land (Figure 1.2). The distance between the source and the receiver is termed the ‘offset’. The data from receivers near the source are called ‘near offset’ and those far away as ‘far offset’. The near receivers receive the reflected signal quicker than those further away from the source, so the response of the same boundary will appear progressively later (Figure 1.3). There are two types of seismic waves: (i) P-waves (longitudinal/compressional body waves), where the particle motion is parallel to the direction of wave propagation, and (ii) S-waves (shear/transverse waves), where particles move perpendicular to the wave propagation direction. P-waves convert into S-waves and vice versa when they transmit or reflect across a boundary, where there is a phase change i.e. solid to liquid/gas or liquid/gas to solid. Pore spaces have liquid/gas and thus this conversion is very common. Both P and S waves follow Snell’s law of reflection and refraction (Yilmaz, 2001). The angle of incidence equals the angle of reflection; the incident ray, the reflected ray, and the normal to the plane of incidence are co-planar (Figure 1.4). The refracted seismic waves also follow Snell’s law, which states:

$$ \frac{\sin \theta_1}{V_{PR}} = \frac{\sin \theta_2}{V_{PT}} = \frac{\sin \phi_1}{V_{SR}} = \frac{\sin \phi_2}{V_{ST}} $$ (1.2)

where $\theta_1$ is the angle of incidence, $V_{PR}$ the velocity of reflected P-wave, $\theta_2$ the angle of transmitted P-wave, $V_{PT}$ the velocity of transmitted P-wave, $\phi_1$ the angle of the reflected S-wave, $V_{SR}$ the velocity of the reflected S-(converted from P-) wave, $\phi_2$ the angle of the transmitted S-wave and $V_{ST}$ the velocity of the transmitted S-wave.

1.2 Seismic Data Acquisition

Earth’s interior can be imaged by reflection seismic data much like remote sensing satellites image the Earth surface. Rock layers, for example sands and shales, differ
Figure 1.1 A seismic section showing reflections from sedimentary boundaries. Seismic data courtesy Reliance Industries Ltd. Reproduced with permission from the Directorate General of Hydrocarbons (DGH), India.

Figure 1.2 (a) Schematic diagram showing layout of transmitted and reflected energy from the shot point and to the individual receivers. At every boundary, a part of the energy is transmitted and another part is reflected. The latter reach the receivers at the surface. Note there are two reflectors, numbered (1) and (2). Rn: receiver number; (b) shows the corresponding simplified seismic wriggle traces.

Figure 1.3 Schematic seismic response of reflector (1) on the receivers as shown in Figure 1.1.
in density. The acoustic impedance (Equation 1.1) to seismic wave velocity passing through the layers thus differs and it is reflections of the wave that are imaged. These layers appear as reflections in a seismic record and are interpreted to reconstruct the geological architecture. This reconstruction is challenging and takes time. To understand the data properly, the data acquiring process has to be understood.

Seismic data acquisition parameters are guided as per requirements. For example, if a large NE trending anticline of 100 km² area at 3000 m below sea level is to be imaged, the spacing between individual lines in a 2D seismic survey should be 10–20 km. On the other hand, when a 50–100 m long reservoir fault demands mapping, a high resolution 3D volume will be acquired (see Section 1.3.4 for a discussion on resolution). The trend of the structure will determine the orientation (azimuth) of the receiver lines, for example the NE trending structure will need NW oriented receiver lines. Again, for crustal architecture studies (Misra et al., 2015, 2016) a larger record length (Box 1.1), a high-energy source, a larger offset etc. yield a good image.

Three primary environments can be envisaged for seismic data acquisition: (i) land, (ii) marine, and (iii) transition zones.

### 1.2.1 Land Acquisition

Seismic land acquisition involves acquiring seismic data in any part of the Earth exposed aerially: deserts, forests, swamps, mountains, inhabited areas, tundra, permafrost etc. Acquiring data may involve logistic problems and labour. 2D data are commonly acquired in logistically difficult areas. 3D seismic volumes are normally acquired before firming up a certain play or some particular prospect(s) prior to drilling in hydrocarbon exploration.

The energy sources commonly used on land for acquisition of seismic data are (Dobrin and Savit, 1988; Onajite, 2014; Schlumberger, 2016):
- Explosives: the most commonly used explosives are dynamite and primacord. The source can be placed below the weathered rock layer. This reduces the noise of the topmost layer. However, it often requires drilling of expensive 3–100 m deep shot holes and is difficult to handle. It can also have a significant cultural and environmental impact.
- Vibrators: one of the most commonly used sources presently. They have low cultural and environmental influences, high signal-to-noise ratio and it does not involve drilling holes. However, the topmost soil layer can cause noise and, thus, recording and processing can complicate.
- Gas or air guns: these can be used in rough terrains. Shot holes are not required and the guns are not expensive. However, energy penetration is limited and noise is high.
- Weight drop: these are also relatively inexpensive and holes need not be drilled. Unfortunately, different sources in an array cannot be synchronized and noise is high.

The receivers, i.e. geophones are the mechanical/analogue devices that convert mechanical ground movement into electrical energy. Single geophones are often connected together to form geophone arrays, which maximize signal-to-noise ratio. Recording stations gather the response from the receivers and perform basic processing to render the data readable. The recording process further controls parameters such as record length, sample rate etc.

### 1.2.2 Marine Acquisition

This involves acquiring seismic data in shallow (~10–40 m) to ultra-deep (>1500 m) water by specialized vessels with arrays of hydrophones (receivers) towed by one or more cables, called streamers. A seismic source comprising an air gun or an array of air guns is also towed together with the geophones.

Two types of energy sources are used in marine seismic surveys (Games and Wakefield, 2014; Onajite, 2014; Schlumberger, 2016):

Figure 1.4 Schematic diagram showing mode conversion of incident P wave in P- and S-waves at a boundary of two lithologies with different velocities $V_1$ and $V_2$. The refracted (transmitted) waves follow Snell’s law. $S_R$: Reflected mode converted S-wave; $P_R$: reflected P-wave; $S_T$: Transmitted mode converted S-wave; $P_T$: transmitted P-wave.
Introduction to Seismic Data

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- Water guns: these use a piston that pushes water very fast to generate a bubble, which collapses to produce a large vibration/seismic energy. Water guns require specialized processing due to pressure variations that occur before the main pressure pulse.

- Air guns: these are the most commonly used sources presently. Air is kept at very high pressure within chambers. The compressed air is released causing a large vibration and forming an air bubble. The bubble continues to oscillate and generates a series of smaller decaying vibrations. A number of air guns are used in an array to minimize the size of the bubble and to make the shape of the bubble nonspherical. The oscillating air bubble minimizes the decaying vibration.

Receivers are (array of) hydrophones attached inside a jacketed streamer filled with a liquid lighter than water. A combination of weights balances the streamer at a predetermined depth below the water surface. The length of a streamer, depth below the water surface and spacing between two consecutive streamers are critical as it constrains the depth of the target/objective that can be imaged. The streamer length is usually the same as the depth of the objective.

Many times ocean bottom cable surveys are used as a cost-effective method to obtain repeated (of the order of few months) or 4D seismic data over producing fields. 4D surveys observe the depletion patterns of reservoirs and identify target areas. Permanently deployed cables used to get numerous 4D seismic data are known as “life of field seismic” (Barley and Summers, 2007).

1.2.3 Transition Zones

There are varied types of landforms like shores/beaches, deltas and estuaries, reefs, river mouths, swamps, marshlands etc. In such zones the water depth prevents seismic land acquisition techniques. Conventional seismic acquisition vessels cannot work. A mix of both land and marine techniques are used; for the latter, smaller boats than those used in open oceans are employed.

Box 1.1 Important terms in seismic acquisition and processing.

1) **Acoustic impedance (AI):** A product of density and velocity. The greater the AI, the stronger the reflection. Acoustic impedance, \( Z = V \rho \).
2) **Amplitude:** Reflection strength.
3) **Fold:** Not the fold in structural geology! For the one in structural geology, see Chapter 4. The geophysical fold is the numbers of traces in a CMP gather, so if data has 10 traces, the “foldage” of the data is 10 and when it has 60 traces, the foldage is 60. The more the foldage of the data, the clearer will the imaging, due to a higher signal-to-noise ratio.
4) **Migration:** Restoring a dipping reflector to its correct subsurface position.
5) **Minimum phase wavelet:** Front loaded energy, that is at time zero minimum energy and elsewhere maximum.
6) **Offset:** Distance between the source and the receiver.
7) **Receiver:** Geophone (on land) or hydrophone (offshore); devices to measure ground movement or sound from the shot.
8) **Record length:** Predetermined time after which the receiver measures the reflected seismic waves. A larger record length indicates greater depth of imaging.
9) **Reflection coefficient:** Type and size of acoustic impedance change.
10) **Reflections:** Acoustic waves reflected from an interface of contrasting lithologies. If the boundaries are gradational, the reflections may be chaotic or comprise a number of reflections, depending on the frequency of the data.
11) **SEGY:** or SEG-Y, is a popular format for storing seismic data. Controlled by the Society of Exploration Geophysicists. Associated file extensions: .seg y and .sgy.
12) **Seismic reflector:** Boundary across which the competence changes. Also called **acoustic-impedance boundary**.
13) **Shot:** Initial disturbance/sounding/explosion; generated by explosives or seismic vibrators on land and by pressurized-air guns offshore.
14) **Shot point:** Geographic location of the shot, measured precisely by positioning systems.
15) **Source pulse or wavelet:** Resulting sound wave from the shot. This wavelet is commonly used for zero-phasing the seismic data (see “Zero-phase wavelet”).
16) **Streamer feathering:** Only in marine acquisition; the deviation of the streamer array away from the linear towing direction. This may happen due to water currents or change in towing direction.
17) **Streamer length:** Length of the cable on which the hydrophones are attached. The length of the streamers depends on the depth of the objective. Typically, the streamer length is equal to the target depth.
18) **Trace:** Stream of reflections recorded by geophone.
19) **Two-way time:** Time from shot to recording. It is the time required for the wave to reach the reflector from the source and the reflected wave to reach the receiver.
20) **Zero-phase wavelet:** The wavelet has maximum energy at time zero.
The acquired seismic data have to be processed because they contain noise related to acquisition problems. Moreover, the geophysical data have to be modified to represent the geology of the subsurface. Seismic data processing is neither a one-step nor an entirely computer-based automatic process. Rather, it requires a good understanding of the data, especially their limitations, and exploration or production objectives. A good resource for studying processing is Yilmaz (2001). Only key methods/features in seismic data processing, namely (i) noise, (ii) gathers, (iii) migration and stacking and (iv) resolution, are mentioned here so that the geologist understand the data before interpretation.

### 1.3 Noise

The signal part of the data is termed as **primaries**. The most important processing workflow deals with the removal of seismic noise, because noise is expected in all the seismic data. Any amplitude or reflection event not generated from a geological boundary is termed **noise** in the seismic data. Noise is inevitable in seismic data, as numerous sources of noise exist even in the most optimal acquisition conditions. The most basic noise may result from faulty equipment; human error may turn out to be the most erratic one. The wind blowing, flowing water during marine acquisition, cable vibration, lightning or even a dog bark can induce noise! Seismic noise may also arise when the lithological parameters, such as density and velocity, are presumed incorrectly (Tarantola, 1986). Sometimes, a high amplitude reflection – commonly the sea bed – may appear repeatedly in the seismic record below it; these are referred to as **multiples** – reflected energies from other interfaces, as in

![Figure 1.5](image-url) Formation of multiples in seismic acquisition for one shot point and receiver set. Originated receiver set. Dotted lines: primaries; solid lines: multiples.

![Figure 1.6](image-url) Multiples in a seismic record. P: primaries; M: multiples. Note that the multiples appear at approximately twice the depth of the primaries. Seismic section courtesy US Geological Survey, Department of the Interior/USGS.
to understand the form of noise and whether it can be removed easily or whether additional processes will be required. For example, sporadic and random noise is difficult to identify and remove. Yilmaz (2001) describes noise removal comprehensively.

As will be discussed in Section 1.4, certain seismic attributes (coherency, dip, azimuth etc.) can be used to quickly interpret/detect complex structures, for example faults and fractures. It is important to understand the effect of noise in the seismic data, which may manifest itself as geological structures.

In spite of the best efforts in seismic data acquisition and processing, these data contain both signal and noise. Unless the seismic interpreter recognizes the characteristics of seismic noise, there is a real danger that noise in the seismic data may be misinterpreted as a structural feature. Knowledge about the source and nature of nongeological features that appear on attribute maps is, therefore, crucial (Hesthammer and Fossen, 1997).

Thus, it is possible that noise can masquerade as interesting structural features; only an experienced structural interpreter can differentiate between these artefacts and a real feature. Certain tools can be useful to understand the origin and effect of noise on seismic data; one such tool is seismic forward modelling. A conceptual geological model can be converted to an acoustic model and can be forward modelled to get synthetic seismic data, interpretation of this synthetic seismic data in conjunction with the real seismic data can discriminate the nongeological features present in the real seismic data.

The other tool that can be used is a measurement of the quality of seismic data. Signal-to-noise ratio (SNR) is one quantity that can provide a measurement of the quality of the seismic data; it does not directly help in structural interpretation but can be quite useful in laying down the expectation from structural interpretation. The interpreter will have an idea of the level of noise present in the data and will, thus, be warned against overinterpretation of the structural features.

A simple definition of the SNR was introduced by Yang and Chengchu (1997). SNR is a ratio of the power of signal to the power of noise, represented by the equation:

\[ \text{SNR} = \frac{(S_a + n - Sn)}{Sn} \]  

(1.3)
where SNR is the signal-to-noise ratio, ‘Sa + n’ the seismic data with noise, ‘Sn’ the noise in the seismic data, ‘a’ the seismic signal and ‘n’ the seismic noise.

1.3.2 Gathers

Gathers are the raw data form and are very important for the processing geophysicists. An interpreter must also have some knowledge about gathers to make a sound interpretation. Gathers are used to visualize the seismic data for errors, noise and geological signatures. These are displays of seismic trace with a common acquisition parameter. The most common types are:

- Common shot gather: the common acquisition parameter is the shot point in this type of a seismic gather. However, different receiver locations are considered. It is the result of a single shot and contains information about signal and noise produced from one specific shot (Figure 1.9a, 1.9b).
- Common receiver gather: as the name suggests, the common acquisition parameter is the receiver. These gathers contain information for all shot points for a particular receiver. It contains signal and noise associated with a particular receiver (Figure 1.9c, 1.9d).
- Common mid-point (CMP) gather: in this sort of gather, the incidence point of the seismic wave is same for all the traces. So, data from several shot gathers are combined to create a CMP gather. These gathers are the starting point for most of the steps in a processing workflow. A ‘fold’ in seismic data processing is the number of traces in a CMP gather (Figure 1.9e, 1.9f).
- Common offset gather: the common parameter here is that the offset is always kept equal for all the traces in the gather. It is effectively a single fold section for a specific offset. It is useful in some processing workflows, for example migration (Figure 1.9g, 1.9h).
- Common depth point (CDP) gather: the constant parameter is the depth of the reflector. In case of a horizontal bed, CMP = CDP. However, for a dipping bed they are unequal. Here the traces are reflections arriving from a particular depth. However the offsets for each shot point receiver pair differ (Figure 1.9i, 1.9j).

1.3.3 Migration and Stacking

After removing certain types of noise from gathers other steps in data processing follow, two of which are migration and stacking. Migration and stacking are important steps because they lead to the geophysical data representing the geology of the subsurface. In the case of horizontal layers in the subsurface, the boundaries are imaged at their correct positions in a zero-offset seismic section. However, dipping layers are not imaged at their correct position in a zero-offset seismic profile, because the reflected rays sway laterally (Figure 1.10) when they travel from the reflector to the receiver and their response is perceived at an apparent reflection point. This apparent reflection point is not located on the reflector itself but shifts laterally and above it. The shift increases with depth, dip of the reflector and velocity. Migration aims to position the reflection correctly.

Thus, in the case of complex geometries such as antiforms, synforms, salt bodies (e.g., Mukherjee et al., 2010; Mukherjee, 2011) and etc. present in the subsurface, the image becomes unusable due to the above mentioned phenomena. In such situations, migration must inevitably be carried out to position the boundaries perfectly in the subsurface (Figure 1.11). Migration has additional benefits:

- its mimics the wave propagation more accurately, thus producing realistic images of the subsurface;
- prestack migration generates more precise velocities;
- prestack migration is essential for amplitude versus offset (AVO) analysis (Aki and Richards, 1980; Ostrander, 1984; Shuey, 1985);
- migration also enhance spatial resolution.

Migration can be carried out either in the time domain or the depth domain. In the time domain, a root mean square velocity for the entire section is used. The depth domain, on the other hand, uses interval velocities for each layer. Time-migrated data have a smooth velocity model that does not allow much lateral variation in velocity, whereas the depth-migrated data employ a complex velocity model with strong lateral variation. Therefore, time-migrated data are less sensitive to erroneous velocity than the depth-migrated data. However, when the subsurface has strong lateral variation in velocity, time-migrated data cannot position correctly the reflectors in the subsurface.

The traces in a gather (Section 1.3.2), most commonly the CMP gather, are added together or ‘stacked’ after migration. This generates an approximate zero-offset trace for the gather. This process sums up the response for all the traces in the gather, resulting in nullifying some noise and amplifying signals.

1.3.4 Resolution of Seismic Data

Features identifiable on the seismic data are controlled by its resolution, which provides the ability to distinguish two separate features/events. Thus, resolution of the seismic data limits observation or interpretation of small-scale structures. Resolution in seismic data is considered in both the vertical and horizontal perspectives. In the vertical scale, wavelength (λ) is the measure of the resolution. If two subhorizontal events or reflections are to be distinguished, they must differ by at least λ/4.
Figure 1.9  Top row: schematic seismic ray path diagram; bottom row: gather for the set-up above. (a, b) Common shot gather; (c, d) common receiver gather; (e, f) common offset gather. Green triangles, dark to light: Receivers, near to far. Red stars: shot; Green triangles: receivers. (g, h) Common midpoint gather; (i, j) common depth point gather. Originated point point gather. Red stars: shot; Green triangles: receivers.
In seismic data, velocity increases and frequency drops with depth and, thus, resolution falls. Consider a shallow event with the following parameters:

Velocity ($v$): 2000 m s$^{-1}$
Frequency ($f$): 50 Hz

Therefore, wavelength ($\lambda$) = $v/f = 2000/50 = 40$ m

Thus, the resolution is $\sim \lambda/4 = 10$ m.

Now, consider a deep event with the following parameters:

Velocity ($v$): 3000 m s$^{-1}$
Frequency ($f$): 20 Hz

Therefore, Wavelength ($\lambda$) = $v/f = 3000/20 = 150$ m

Thus, the resolution is $\sim \lambda/4 = 37.5$ m.

So, what can be identified on a seismic image? Figure 1.12 shows an approximately 10-m high outcrop with
observable deformation on an outcrop scale. However, comparing that with a seismic wavelet, it can be noted that the deformation at the outcrop scale is far below the seismic resolution. Thus, it is nearly impossible to ‘see’ such features using seismic data. Figure 1.13 shows another outcrop, approximately 200 m high. Here, the beds may be resolved but the deformation will not. Looking at global landmarks of mentionable height, only the tallest of them can be successfully resolved in seismic data (Figure 1.14). The 53-m high leaning tower at Pisa would barely be resolved; the same is true for the Taj Mahal (Agra, India) or the Statue of Liberty (New York, USA). The other towers in Figure 1.14 will be successfully resolved.

Resolution considered in the horizontal sense, known as lateral resolution, is also important during seismic interpretation. The seismic wave propagates through the subsurface in the form of a spherical wavefront as shown in Figure 1.15. As the wavefront propagates away from the source, it spreads out to a larger area. Reflections from a reflector or surface are not from a point but from a region depending upon the dimensions of this wavefront. Thus, signals received from this region (AA’) at the same interval of time are indistinguishable.

This region is fairly large, a roughly circular area of the reflecting surface, and is called the first Fresnel zone. The radius of this zone is often taken as the horizontal resolution for unmigrated seismic data. 3D migration tends to collapse this Fresnel zone and, thereby, increase the horizontal resolution of the seismic data.

The horizontal resolution is given by the width of this Fresnel zone:

$$r = \frac{v}{2} \sqrt{\frac{t_0}{f}}.$$  \hspace{1cm} (1.4)

where $r$ is the radius of the Fresnel zone/Horizontal resolution, $v$ the velocity, $t_0$ is the two-way travel time and $f$ the frequency.

Newer techniques such as shear (S-) wave reflection seismic profiles can provide very high resolution. See Chapters 3.14, 3.15, 5.3 and 5.4, where S-wave reflection profiles are shown to give resolutions as small as 0.5–1 m. Such techniques require further studies to be used widely.
Visualization and identification of many structures are sometimes easily achieved using seismic attributes. Seismic attributes are measured, computed or implied derivatives from seismic data. There may be various inputs for computing a seismic attribute, such as:

- Single trace: attributes derived from a single migrated and stacked trace are called trace attributes (Taner and Sheriff, 1977; Taner et al., 1979).
- Set of pre-stack traces: attributes derived from these are known as multitrace attributes (Taner and Sheriff, 1977; Taner et al., 1979).
- Surface: a surface for input is usually an interpreted horizon gridded and interpolated to a surface. Attributes derived from surfaces are called surface attributes.
- Volume: a small part of a 3D seismic volume or an entire volume is sometimes used as an input to derive seismic attributes. Attributes thus derived are called volume-based attributes.

There are numerous seismic attributes and they are used for a number of requirements. There are some, for example, that help in highlighting specific features, be it geomorphological or related to structural deformation, in the subsurface. The selection of the attributes depends on the feature of interest that is to be analysed. The most common seismic attributes used to interpret deformation are mentioned here. There are several attributes for interpreting faults in seismic data (Chen and Sidney, 1997). Seismic attributes are employed primarily for interpreting deformation along a surface and in volume.

### 1.4.1 Coherency

This seismic attribute (Figure 1.16) measures the similarity between neighbouring seismic traces. It can be calculated along a horizon or within a volume. It is most commonly used to identify faults, channels or other
1.4.2 Dip and Azimuth

Dip is an attribute that computes, for each trace, the best fit line with 2D data or plane with 3D data between neighbouring traces. Dip is computed on an interpreted horizon and the magnitude of dip is that of the said plane in degrees. This indicates paleogeography. However, when the changes in dip are sharp, faults may be indicated.

Azimuth is another attribute that is employed together with dip. Azimuth calculates the direction of maximum dip of the plane from the dip attribute. Azimuth is measured in degrees, clockwise from north. Thus, dip and azimuth attributes together compute dip and dip direction of an interpreted horizon. An example is shown in Figure 1.17.
1.4 Curvature

In general, curvature is a measure of the tightness of a curve at particular points. Thus, a larger curvature means the area is more deformed than an area with a lower curvature (Figures 1.18 and 1.19). For a 2D curve, curvature is the reciprocal of the radius of curvature (Chopra and Marfurt, 2007). The most important curvature attributes are the maximum ($k_{\text{max}}$) and minimum curvatures ($k_{\text{min}}$). Together, they are called principal curvatures. They are combined to derive other curvature attributes such as:

- **Mean curvature** $k_{\text{mean}} = 0.5 \times (k_{\text{max}} + k_{\text{min}})$ (1.5) – indicates the overall deformation of the area of interest.
- **Gaussian curvature** $k_{\text{g}} = k_{\text{max}} \times k_{\text{min}}$ (1.6) – a good indicator of whether a surface is warped.
- **Curvedness** $r = (k_{\text{max}}^2 + k_{\text{min}}^2)^{1/2}$ (1.7) – a good measure of folding intensity.
- **Shape index**, $s = (2/\pi) \tan^{-1} [(k_{\text{max}} + k_{\text{min}})/(k_{\text{max}} - k_{\text{min}})]$ (1.8) – classifies the shape of the deformed geomorphology of the area: dome, ridge, saddle, valley and so on.

**Figure 1.17** Volumetric dip–azimuth along a picked horizon. Note the NE striking and N-dipping faults in the lower part of the image. Reproduced from Paes and Marfurt (2016). AAPG © 2016 whose permission is required for further use.

**Figure 1.18** Sketch to show the idea of curvature along a 2D line. Anticlines have a positive curvature and synclines have a negative one. The anticline on the left has a higher curvature than the one on the right. That is because the one on the left is more tightly folded than the right.

**Figure 1.19** Calculated maximum curvatures for a domal surface. Note the linear faults that can be easily interpreted on this image. Reproduced from Bergbauer et al. (2003). AAPG © 2003 whose permission is required for further use.
References


