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GEOPHYSICS FOR SLOPE STABILITY

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Abstract. A pre-requisite in slope stability analyses is that the internal structure and the mechanical properties of the soil or rock mass of the slope, are known or can be estimated with a reasonable degree of certainty. Geophysical methods to determine the internal structure of a soil or rock mass may be used for this purpose. Various geophysical methods and their merits for slope stability analyses are discussed. Seismic methods are often the most suitable because the measurements depend on the mechanical properties that are also important in the mechanical calculation of slope stability analyses. Other geophysical methods, such as electromagnetic, electric resistivity, self-potential, and gravity methods, may be useful to determine the internal structure, but require a correlation of found boundaries with mechanical properties.

Keywords: Geophysics, seismic, electromagnetic, geo-electric, resistivity, self-potential, gravity, boundary, property, slope stability

1. Introduction

Many geophysical tools exist to investigate slopes and to establish material inhomogeneities, boundaries, and properties of materials. Most methods exist for many years, but certainly, in the last decades most methods have undergone a large change due to the availability of cheap computer power. As a result geophysical surveys have become easier to do by non-specialists, interpretations are more reliable and more accurate, and last but not least, have become considerably cheaper. Still regrettably, geophysical methods are seldom used and often only boreholes or soundings are made to investigate the sub-surface. The geophysical methods described are readily available and are helpful to the engineering geologist and geotechnical engineer in obtaining the material properties and boundaries of sub-surface materials.

Boundaries between different materials can be obtained by all geophysical methods. It is, however, to be noted that the boundary measured is often not based on a difference in properties (a ‘contrast’) that is mechanically interesting. In, for example, electromagnetic methods the properties measured are the dielectric constant and/or conductivity of the materials. If a change in either of these coincides with a boundary in mechanical properties, the found boundary is of interest for slope stability. However, if the boundary measured is only a local enrichment of

Surveys in Geophysics 21: 423–448, 2000.

the slope material with, for example, manganese or iron, the boundary is of none or of little interest. The methods that are most frequently used in slope stability investigations are discussed: seismic, geo-electric, electromagnetic, and gravity methods.

2. Seismic Methods

Seismic methods are based on the measuring of an elastic wave (also: seismic, shockwave, or acoustic wave) traveling through the sub-surface. The wave is reflected or refracted on boundaries characterized by different densities and/or deformation properties. Seismic methods can nearly always be used to determine the internal structure of materials in a slope. Sometimes logistics and practical problems as how and where geophones and sources can be placed, may make the method impractical. Refraction seismic studies have been the standard tool for geotechnical work for years. However, state-of-the-art computerized seismographs for use in geotechnical work handle 24 or more channels each connected to one geophone and, hence, measure the signal of many geophones in one round. This reduces the quantity of sources necessary, but more important, has opened the option to do seismic reflection surveys in geotechnical work with relatively low costs. A problem often encountered is the frequency content of the source signal. For many slope stability problems it is important to investigate the structure with a high resolution. This means that high frequencies should be available in the seismic signal.

2.1. Type of Waves

Seismic waves may be compression (P-waves), shear (S-waves), or surface waves, such as Raleigh and ground waves. Different properties of the soil or rock mass material have different influences on the behavior of the wave depending on the type of wave. Traditionally compression (P-) waves were used, as these are easy to generate. Compression waves are most sensitive to changes in the normal (compression) stiffness of the materials while shear and surface waves are more influenced by the shear stiffness of the materials through which the wave passes. Shear stiffness is, however, often of more interest than normal stiffness because shear stiffness can often be related to the shear strength of sub-surface materials (Helbig and Mesdag, 1982). Sources for shear waves have been cumbersome, but new devices have been developed recently (Ghose et al., 1996; Peeters et al., 1998).

2.2. High-Resolution Seismic Sources

Traditionally, sources in seismic surveys for geotechnical work consist of a hammer blow on a metal plate. The contact between plate and soil or rock, often via a weathered topsoil layer, does not allow the introduction of a seismic signal into

Surveys in Geophysics 21: 423–448, 2000.

the ground with a high energy content in high frequencies. Explosives, and in particular, fast burning explosives, such as, caps or fuses, give a far better ('spiked') signal with more energy in the high frequency components of the signal. Obviously, the use of explosives is often forbidden or otherwise problematic. Alternatively, sources that emit a controlled signal ('vibro-seis') into the ground can be applied (Ghose et al., 1996, 1998). The energy level per unit of time of these sources is considerably less than the energy per unit of time released by explosive sources, but the controlled signal can be correlated far better with the received signals and allows for an increase in the noise/signal ratios of the received signals. In addition, all energy is concentrated in the required high frequencies and no energy is lost in low frequencies that are not of interest.

Apart from applying a source with high frequencies in the signal, also the geophone and line spacing influences the resolution. It makes not much sense to obtain a resolution in very high detail in depth and have a very low resolution in the directions over the plane of the measurements. It should be noted that time and costs increase dramatically if the resolutions have to be high in all directions.

2.3. Seismic Refraction

Seismic refraction is based on the first arrival of a signal that travels through a layer with a higher velocity. Table I gives some characteristic seismic P-wave velocities for sub-surface materials. The method has been standard used for years (Stötzner, 1974; Telford et al., 1990; Williams and Pratt, 1996). Figure 1 shows a simple situation of a more-or-less regular topsoil layer on top of a slightly weathered rock mass on a slope. The first arriving signal at the geophones is for the first two geophones the direct signal traveling through the topsoil layer and for geophones 3 to 12 the refracted wave traveling through the slightly weathered rock mass. For the refracted waves applies that the angle of incidence and angle of refraction equal θ. The topsoil has velocity $V1$ and the slightly weathered rock mass velocity $V2$. The angle θ is given by:

$$\sin(\theta) = \frac{V1}{V2}. \tag{1}$$

The thickness of the residual soil layer can be calculated easily from Figure 1b and Equation (2).

$$\text{depth} = \frac{1}{2}\frac{V1^{*}t_i}{\cos(\arcsin(V1/V2))}, \tag{2}$$

where t_i is the intercept time (see graph Figure 1b).

For an inclined plane boundary the survey should be repeated with the source position at the other end of the geophone spread (Figure 2). If the inclination between boundary and surface is relatively small, the velocities and depth become:

Surveys in Geophysics 21: 423–448, 2000.

TABLE I
Characteristic P-wave velocities ranges (modified after Anon., 1995).

Material	P-wave velocity (m/s)	Material	P-wave velocity (m/s)
Air	360	Weathered sedimentary rock	300–3000
Dry sand	400–1000	Metamorphic rock	1000–6000
Clay	300–1800	Unweathered basalt	1000–4300
Weathered igneous and metamorphic rock	450–3700	Limestone	500–6700

Note that the material descriptions are crude and do not account for variations in, for example, water content, number of discontinuities, or whether discontinuities are open, filled, or closed, etc. These factors influence the velocity values far more than most of the material constituents.

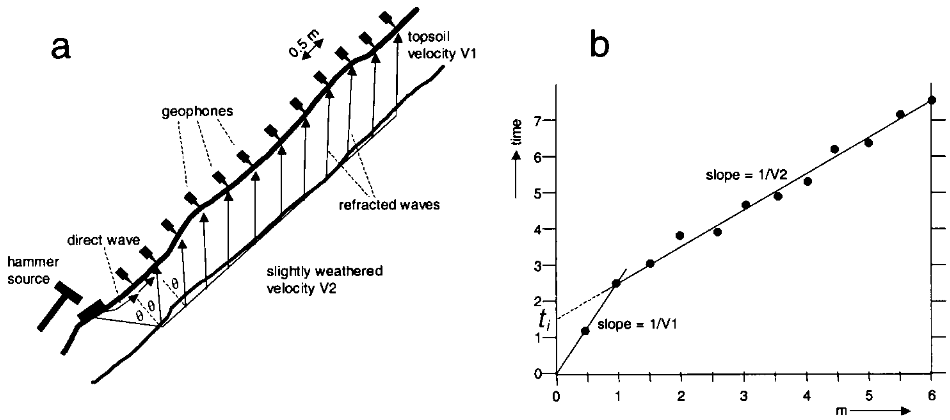

Figure 1. Seismic refraction survey for a boundary parallel to surface. (a) Ray paths, (b) travel time versus distance for the first arrived signal.

$$\frac{1}{V2} \approx \frac{1}{2}\left(\frac{1}{V\text{down}} + \frac{1}{V\text{up}}\right)$$
$$ti_{\text{down}} = \frac{2z_{\text{down}}}{V1}\cos(\theta) \quad ti_{\text{up}} = \frac{2z_{\text{up}}}{V1}\cos(\theta), \tag{3}$$

where z_{down}, z_{up} are the depths below down-dip respectively up-dip source point.

Non-computerized interpretation of refraction seismic studies is easy for simple 2 and 3-layer situations with plane boundaries, but becomes very difficult and inaccurate if the boundaries are irregular or if the number of layers increases. State-of-the-art computerized interpretation techniques include programs based on wavefront methods (Sandmeier, 2000; Telford et al., 1990; Tomo, 2000). In these programs for each arrived signal, a hypothetical path through the sub-surface is

Surveys in Geophysics 21: 423–448, 2000.

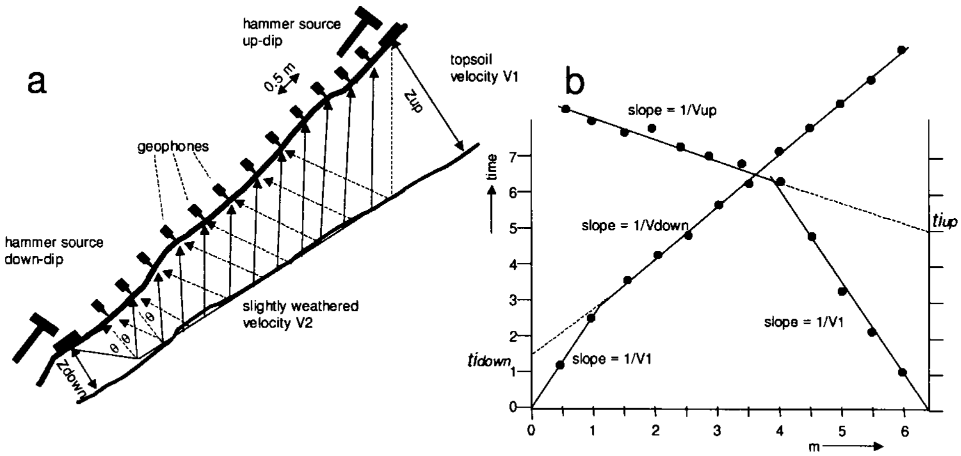

Figure 2. Seismic refraction survey for interface inclined to surface. (a) Ray paths, (b) travel time versus distance for the first arrived signal.

constructed. The travel times through each layer and the velocities of the layers are optimized to the arrived signal by an algorithm in the computer program, and the hypothetical travel paths are adjusted. This is repeated until a best fit of the data on the travel times is obtained. The best travel-times/velocity section is then converted to a depth section. Interpretation of irregular boundaries and multi-layer situations are usually no problem with these programs. The resolution that can be obtained is dependent on the frequency content of the source signal and the spacing of the geophones.

2.4. Seismic Reflection

Traditionally the equipment and the computers to analyze the measured signals in reflection seismic surveys were too expensive to be used in geotechnical work. In recent years, the availability of cheap and powerful computers allows the introduction of reflection seismic in geotechnical work (Bruno et al., 1998; Kurahashi et al., 1998). Reflection seismic has a similar set up of source and geophones as for refraction seismic surveys, however, not only the first-arrivals are considered, but the complete received signals are incorporated in the interpretation (Figure 3).

Reflection occurs on an interface where the deformation characteristics of soil or rock mass are different on both sides of the interface. This is governed by the so-called 'acoustic impedance' (Z), which equals the product of density and seismic velocity. The 'impedance contrast' (δ) is the ratio of the acoustic impedances on both sides of the interface:

$$\text{acoustic contrast} = \delta = \frac{Z_1}{Z_2} = \frac{\text{density}_1 \times \text{velocity}_1}{\text{density}_2 \times \text{velocity}_2}. \tag{4}$$

Surveys in Geophysics 21: 423–448, 2000.

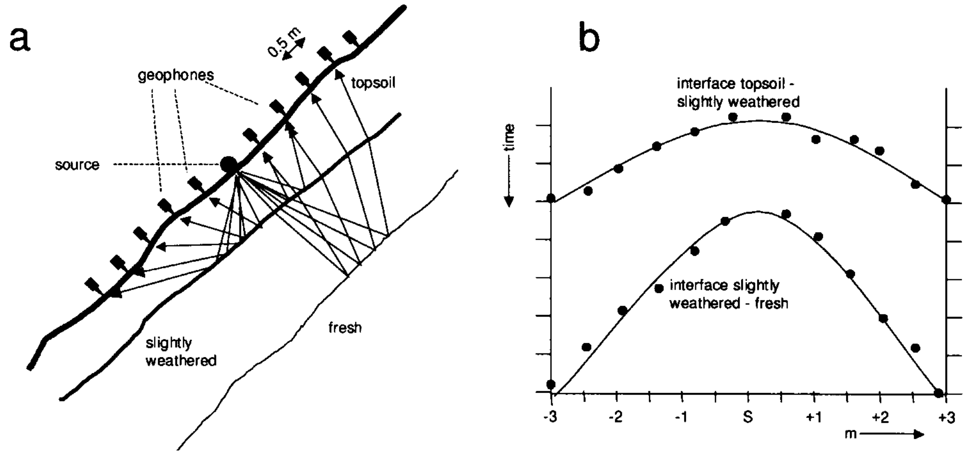

Figure 3. Seismic reflection survey. (a) Ray paths, (b) travel time versus distance for the arrival of the reflected signals from the two reflectors.

The acoustic impedance and the angle of incidence of the incidence seismic wave govern the energy in the transmitted and reflected waves. The relations between reflected and transmitted energy contents are complicated, but the reflection coefficient (R) is simple for normal incidence (90°) of the incidence seismic wave:

$$\text{reflection coefficient} = R = \left(\frac{\delta - 1}{\delta + 1}\right)^2. \tag{5}$$

Table II gives values for velocities, densities, acoustic contrast, and reflection coefficients for some material boundaries. The reflection coefficient is independent whether the seismic wave passes from medium one to medium two or vice versa. Note the very high reflection at weathering horizons, interfaces with water, and the virtual 100% reflection on interfaces with air. The higher the reflection coefficient the more energy is reflected and the more easily the reflector boundary is detected with reflection seismic surveys.

Processing the received signals is extensive. Commonly used processing techniques in geotechnical work are ‘Common Depth Point’ gathering (CDP), and digital filtering. Computerized processing programs are available for use in the seismograph or can be run in laptop computers.

2.5. Seismic Tomography

Seismic tomography works on the bases of a series of geophones and source positions on the surface or in boreholes. Figure 4a gives an example of a seismic tomography survey between two boreholes. Geophone and source are lowered in two different boreholes. At regular intervals, a measurement of the travel time

Surveys in Geophysics 21: 423–448, 2000.

TABLE II

Characteristic P-wave velocities, densities, acoustic contrasts, and reflection coefficients for some soil and rock mass materials).

Material	First medium		Second medium		Acoustic contrast	Reflection coefficient
	P-wave (m/s)	Density (kg/m^3)	P-wave (m/s)	Density (kg/m^3)	$\delta = Z1/Z2$	R
Fresh sandstone to fresh limestone	2000	2400	3000	2400	0.67	0.040
Fresh limestone to fresh sandstone	3000	2400	2000	2400	1.50	0.040
Highly weathered sandstone to slightly weathered sandstone	500	2200	1200	2400	0.38	0.20
Fresh sandstone to open discontinuity (air)	2000	2400	360	1.2	11111	0.999
Fresh sandstone to water	2000	2400	1500	1000	3.2	0.27
Clay to slightly weathered sandstone	400	1500	1200	2400	0.21	0.43
Granitic residual soil to fresh granite	600	2000	3500	2500	0.14	0.57
Dense sand to slightly weathered limestone	1000	1800	2500	2400	0.3	0.29

between the source and geophone is made. Thereafter the source and geophone are swapped and the same procedure is repeated. The result is a set of travel times from all source positions to all geophone positions from borehole 1 to borehole 2 and vice versa. A computer program optimizes the velocities of the materials in-between the two boreholes on the measured travel times. The results are normally presented as velocity contours (Figure 4b), which give an idea about the boundaries between different areas in the soil or rock mass between the boreholes. The velocities can be correlated with actual material boundaries or with the quality of the soil and rock mass in-between the boreholes, for example, with the number of discontinuities in a rock mass, degree of weathering, presence of karst holes, etc.

2.6. ANISOTROPY

Soil and rock masses are often anisotropic. Reasons for anisotropy are orientated minerals or sets of orientated discontinuities, such as bedding planes, joints, fractures, etc. The anisotropy causes that deformation properties of soil and rock mass are not the same in every direction. This also influences the behavior of seismic waves in the mass. For example, a rock mass with one set of discontinuities will have higher seismic velocities parallel to these discontinuities and lower velocities perpendicular to the discontinuities. Figure 5 shows a so-called seismic refraction fan shooting and the resulting seismic velocities measured in different directions. Velocity differences can be 50% or more in different directions.

Surveys in Geophysics 21: 423–448, 2000.

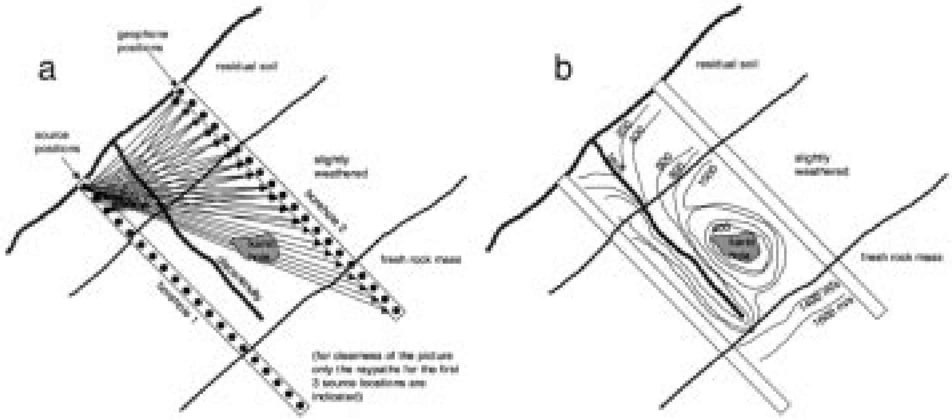

Figure 4. Tomography to determine rock mass quality. (a) Source and geophone positions, (b) velocity contours in m/s.

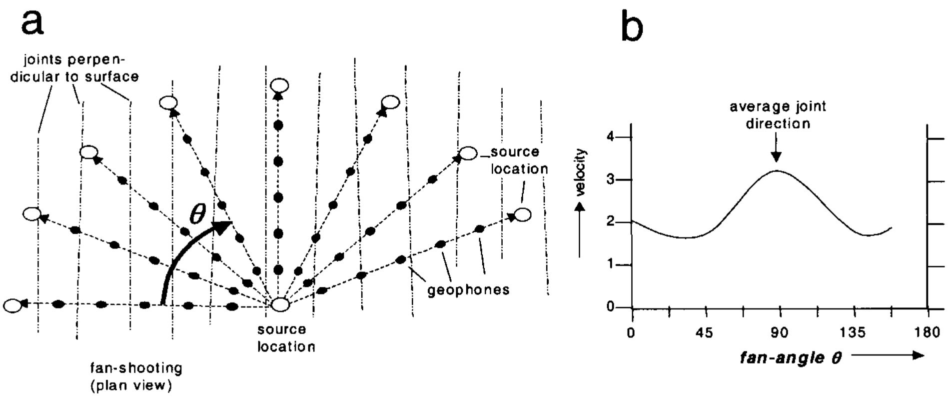

Figure 5. Fan-shooting. (a) Surface layout of sources and geophones, (b) seismic velocity versus 'fan-angle θ' (after Hack and Price, 1990).

Anisotropy in the depth direction may cause erroneous interpretations. Figure 6 shows a slope with a series of discontinuities that are open at surface but become closed at a certain depth. The intact rock material in-between the discontinuities is the same near the surface as deeper in the rock mass and has seismic velocity $V2$. The difference between the surface rock mass and the deeper rock mass is, hence, only the presence of the discontinuities. A seismic refraction survey is done on the slope perpendicular to the direction of the discontinuities. The seismic waves have to cross or go around the discontinuities. However, the energy of a seismic wave crossing the open discontinuities will be reduced strongly because most energy is reflected on the open discontinuity surfaces (Table II). The energy is often so far reduced that the arrival of this signal is not noticed and the signal measured is from the wave that travels around the open discontinuities. This causes the time-distance graph to show an apparent two-layer case of seismic refraction: an apparent first

Surveys in Geophysics 21: 423–448, 2000.

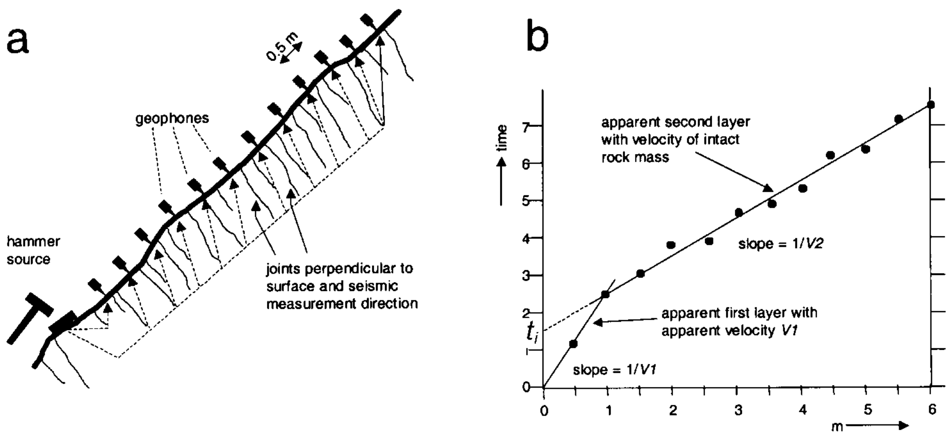

Figure 6. Anisotropy in vertical direction. (a) Ray paths, (b) travel time versus distance for the first arrived signal.

layer with a low velocity ($V1$) being the layer with open discontinuities, and a second layer with the velocity of the rock mass without discontinuities which is the velocity of intact rock $V2$. Note that parallel to the discontinuities the waves do not need to cross the discontinuities and only direct waves are measured with velocity $V2$.

2.7. Attenuation and Absorption

The energy contained in a seismic wave spreads spherically with distance from the source point. The energy flowing through an area of 1 m^2 is related to the distance from the source point by:

$$\frac{E_2}{E_1} = \left(\frac{r_1}{r_2}\right)^2, \tag{6}$$

where E is the energy flowing through 1 m^2, r, the distance from source point.

The 2nd power relation in Equation (6) causes a rapid drop of the energy with distance. Another reason for reduction of energy with distance is absorption. Absorption of energy takes place by loss of energy due to non-ideal elastic behavior of the soil and rock masses. The absorption is related to the frequency of the wave. Higher frequencies are more absorbed than lower frequencies. This causes a change in the overall model of the wave with distance. Further energy losses occur due to reflection, refraction and diffraction, and change of wave type at interfaces, for example, from P-wave to S-wave and vice versa.

The rapid loss of energy flow per m^2 and in particular, the rapid loss of energy of higher frequency components in a seismic signal cause that high resolution seismic studies for geotechnical work are difficult to do. The energy emitted in

Surveys in Geophysics 21: 423–448, 2000.

high frequencies by a seismic source is normally relatively low and the rapid loss of energy causes that with larger distances the high frequency components are lost in noise.

2.8. MEASUREMENT OF SOIL AND ROCK PROPERTIES

Measurement of mechanical properties of soil and rock masses is to a certain extent possible with seismic methods. If P and S waves are measured, a so-called ‘seismic E modulus’ and ‘seismic Poisson’s ratio’ can be determined:

$$\begin{aligned}
\text{P-wave velocity} &= V_p = \sqrt{\frac{\lambda + 2\mu}{\rho}} \\
\text{S-wave velocity} &= V_s = \sqrt{\frac{\mu}{\rho}} \\
\text{Seismic Young's modulus} &= E = \frac{\mu(3\lambda + 2\mu)}{(\lambda + \mu)} \\
\text{Seismic Poisson's ratio} &= \sigma = \frac{\lambda}{2(\lambda + \mu)},
\end{aligned} \tag{7}$$

where λ, μ are Lamé’s constants, ρ, the density of material.

It should be realized that these properties are not the same as the Young’s modulus and Poisson’s ratio obtained from static or dynamic laboratory or field tests. This is because the deformation behavior of most soil and rock masses depends on frequency and is non-linear. The frequencies in seismic waves are normally not the same as frequencies used in dynamic testing and, more important, the stress and strain in seismic waves are very small compared to stress and strain in laboratory or field tests.

The relation between the frequency and the wavelength is:

$$V = f \times \lambda, \tag{8}$$

where V is the velocity of seismic wave, f, the frequency, and λ, the wavelength.

Inhomogeneities in a soil or rock mass, such as boulders in a soil, joints in a rock mass, will not individually be detected if the wavelength of a seismic wave is low compared to the dimensions of the inhomogeneities. A rule-of-thumb is: if the dimension of the object is more than half the wavelength, it can be recognized as individual object (whereas the object will not be determined individually if the dimension is less than half the wavelength).

The influence of discontinuities on seismic P-wave velocity has been established empirically by many authors (among others Deere et al., 1967; Merkler et

Surveys in Geophysics 21: 423–448, 2000.

al., 1970). The relation by Deere et al. relates Rock Quality Designation to seismic velocities measured in the field and in the laboratory:

$$\left(\frac{V_{\text{field}}}{V_{\text{laboratory}}}\right)^2 \times 100\% \approx RQD$$
$$V_{\text{field}} = \text{seismic velocity measured in the field}$$
$$V_{\text{laboratory}} = \text{seismic velocity measured in the laboratory}, \qquad (9)$$

where RQD is the Rock Quality Designation.

The seismic velocity measured in the laboratory is done on intact rock. The velocity measured in the field is the velocity of a signal passing through intact rock but also through or around discontinuities and will hence result in a lower velocity.

Although very little work has been done in slope stability studies on relating material properties to attenuation of seismic signals (Luijk, 1998; Pyrak-Nolte and Shiau, 1998) it is expected that features such as discontinuities have a marked influence on the amplitude of the seismic signal of particular frequencies. Directly applicable methods are not yet established but are expected in the near future.

3. Electromagnetic Methods

Electromagnetic methods have been used in slope stability investigations for a long time (Bogoslovsky and Ogilvy, 1977; Bruno et al., 1998; Stötzner, 1974). A differentiation should be made between low frequency (for slope stability work typically so-called EM31, EM34) equipment and high frequency (geo- or groundradar) equipment. Penetration of an electromagnetic field in the sub-surface depends on the electric conductivity and dielectric constant of the materials in the sub-surface and on the frequency of the transmission field. The higher the conductivity or frequency the less penetration is obtained. Table III gives electromagnetic properties of some sub-surface materials.

Electromagnetic methods do not result directly in a position of the measured features in space. A conversion of the electromagnetic measurements to distance has to be made. This is straightforward if electromagnetic properties of the materials are known in detail. However, mostly these are unknown and electromagnetic properties have to be determined from samples or the measured profile has to be correlated with borehole information.

3.1. Geo- or Groundradar Surveys

Geo- or groundradar uses high frequency electromagnetic waves. A transmission antenna transmits an electromagnetic pulse of high frequency. Conductive materials in the sub-surface reflect the pulse signal and the reflected pulse is received by a receiving antenna, similar to ship or airplane radar. In some systems, only one

Surveys in Geophysics 21: 423–448, 2000.

TABLE III

Characteristic electromagnetic properties of rock and soil masses (after Anon., 1995).

Material	Dielectric constant	Electric velocity (for frequency 100 MHz) (m/ns)	Attenuation (for frequency 100 MHz) (dB/m)
Air	1	0.3	0
Metal			Infinite
Fresh water	80	0.33	2×10^{-1}
Seawater	80	0.01	0.1
Dry sand	3–5	0.15	0.01
Wet sand	20–30	0.06	0.03–3
Limestone	4–8	0.12	0.4–1
Clay	5–40	0.06	1.0–300
Granite	4–6	0.13	0.01–1
Rock salt	5–6	0.13	0.01–1
Shale	5–15	0.09	1.0–100

Note that the material descriptions are crude and do not account for variations in, for example, water content, clay content due to weathering, number of water filled discontinuities, or quantity of minerals in solution in water, etc. These factors influence the property values far more than most of the material constituents.

antenna is used alternatively as transmission and receiving antenna. CDP and filter techniques similar to those used in reflection seismic surveys may be applied to enhance the measurements.

Geo- or groundradar works with electromagnetic waves with frequencies between 10 and 1200 MHz, which gives resolutions in the order of 0.1–3 m. Higher frequencies result in more detail and in a higher resolution. The absorption of the energy of the waves is higher for higher frequencies and hence less penetration is obtained. High conductivity of the sub-surface materials strongly reduces the penetration. In conductive materials, for example, salt water, the penetration is reduced to centimeters and a penetration of not more than a few meters is obtained in clay or peat. Penetration is normally limited to about 40–60 m in most materials, in optimum circumstances a penetration of more than 300 m may be obtained, for example in dry unweathered granite.

The field setup is comparable to single-channel seismic surveys (Figure 7). The transmission antenna is the source and the receiving antenna is the geophone. Analyses of the measurements of a georadar are also comparable to seismic single-channel reflection studies. The transmission and receiving antennas can be built into boxes on skids or on wheels, or the antennas are lowered in one borehole (reflection study) or in two boreholes (electromagnetic tomography). In contrary to seismic surveys, the antennas do not need to be in direct contact with the ground;

Surveys in Geophysics 21: 423–448, 2000.

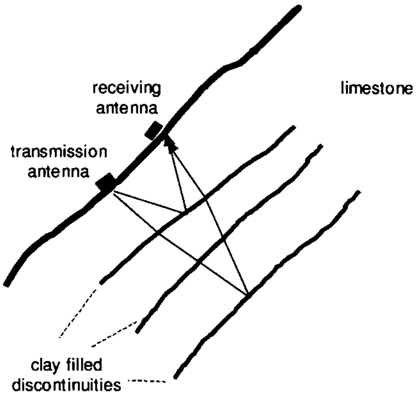

Figure 7. Groundradar survey.

however, the distance between the ground and the antenna should be as small as possible to avoid loss of energy. Wheel mounted antennas allow for the antennas to be towed by hand or car while making measurements at regular intervals.

Special care should be taken that the surface layers (topsoil) or the road pavement (if the survey is done on a road) does not consist of too conductive materials as otherwise the signal will not penetrate into the deeper sub-surface materials. Groundradar signals are very sensitive to local variations in the sub-surface. In particular, man-made fills often prohibit the use of groundradar as the scattering of the signal in the man-made fill distorts the signal from deeper materials.

Surveys to determine the presence of individual inhomogeneities in a soil or rock mass have been done successfully. For example, the location of karst holes in limestone, or open or clay filled discontinuities in a rock mass can be determined with high accuracy. The contrast measured is dependent on the dielectric constant and conductivity of the materials on both sides of the boundary. The measured boundary is normally only of interest in slope stability if the electromagnetic contrast coincides with a mechanical boundary.

Interpretation of groundradar data is often treacherous. For example, Figure 8 shows a groundradar survey done to establish the location of karst holes in a limestone slope. The groundradar measurements show the contrast between air and limestone as the electromagnetic contrast between air and limestone is large. However, the contrast between calcareous silt and limestone is small, if any, and the reflections from the filled karst hole may be vague and missed in the interpretation. The interpretation may be even more treacherous because the zone with enriched manganese and iron will have a large electromagnetic contrast with the surrounding non-enriched limestone and will result in a clear reflection on the records. In the filled karst hole, the enriched zone is continuous through the silt because the enrichment has taken place after filling of the karst hole. The continuous reflection through this karst hole may erroneously be interpreted as a reason to believe that indeed there is no karst hole present.

Surveys in Geophysics 21: 423–448, 2000.

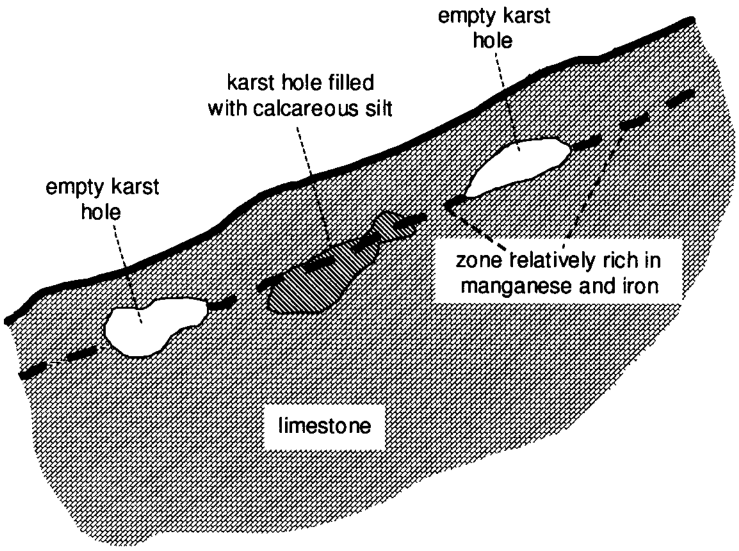

Figure 8. Groundradar survey to establish karst holes.

3.2. Low-Frequency Electromagnetic Surveys

The low-frequency electromagnetic methods are based on the transmission of an electromagnetic field from a ‘transmission coil’. The normally used frequencies are in the range from 800 Hz to 10 kHz. This transmission field (the primary field) will cause a secondary induced field in the materials in the sub-surface. A ‘receiver coil’ receives the primary electromagnetic field together with the secondary field. The measuring equipment allows for comparison of amplitude and phase shift of the primary and secondary fields. The intensity of the secondary field depends on the conductivity of the materials in the sub-surface. The form of the coils and the distance between the coils depend on the frequencies used and the required depth of the investigation. This also allows for two types of investigation: vertical profiling or also called ‘depth sounding’, and horizontal profiling. In vertical profiling, the distance between the coils is increased with regular steps and consequently the received signal is more influenced by deeper buried materials. In horizontal profiling, the distance is kept constant but the whole array of coils is moved and at regular distances, measurements are made. As rule-of-thumb can be used that the depth penetration is not more than about half the spacing between the coils. Vertical profiling or depth sounding can also be achieved by using different frequencies at the same location. Deeper buried materials will have less influence on higher frequency transmissions and vice versa.

Low frequency EM surveys are very simple to do, fast, and the equipment is easy to operate. The method will virtually never be able to determine boundaries with enough accuracy, as the resolution is low. Non-the-less the method works very well for determining the extent of a (thick) clay filled discontinuity in limestone or

Surveys in Geophysics 21: 423–448, 2000.

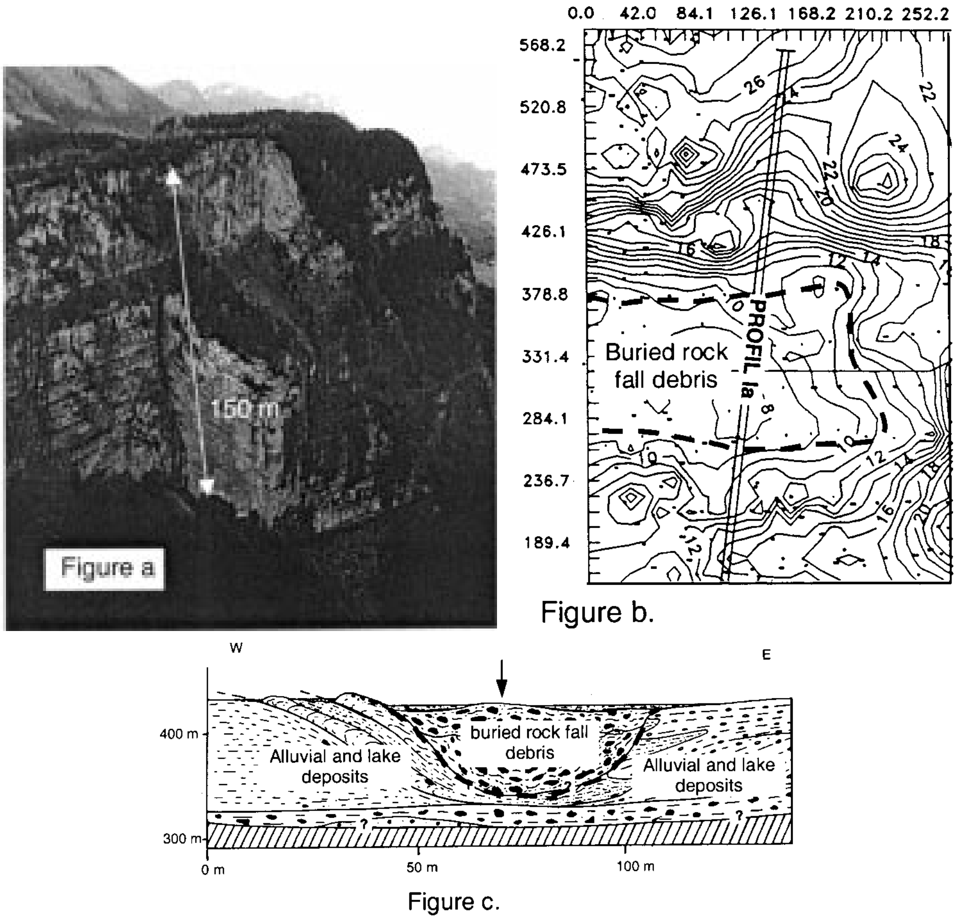

Figure 9. EM34 survey to establish the run-out of the rock fall of the left part of the cliff 300 years ago. The run-out is important, as the right part of the cliff is deemed unstable. (a) Cliff where rock fall originated, (b) EM34 surface contours, contour values in mS/m, distances in meters, (c) interpretation based on EM34 survey a long profile 1a (after Graaff and Rupke, 1999).

the presence of rock boulders in clay or sand. Figure 9 shows an example of an EM34 survey to determine the run-out distance of a past rock fall from a cliff. The cliff is situated about 600 m above a small village. The left side of the cliff failed some 300 years ago. The right side of the cliff is deemed unstable, and when it fails, it would partially fall in the village. It is therefore important to know the run-out distance to evaluate the hazard for the village. The rock fall debris from the past is buried in alluvial and lake deposits consisting of clay and peat. The EM34 survey established very accurately the extent of the rock fall debris, which was later confirmed by trial pits and trenches.

Surveys in Geophysics 21: 423–448, 2000.

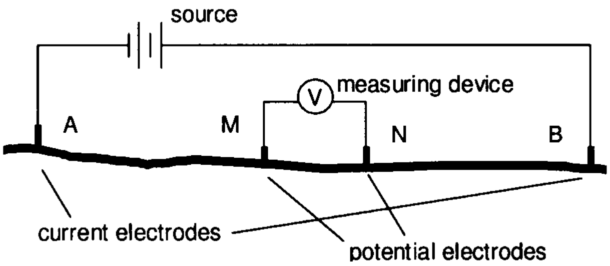

Figure 10. Array of current and potential electrodes.

4. Geo-Electrical or Resistivity Methods

Geo-electrical or resistivity measurements have been done for a long time in slope stability (Bogoslovsky and Ogilvy, 1977; Stötzner, 1974). The restriction was that the interpretation of the measurements was mathematically highly complicated and that only standard situations could be solved, normally with the help of standard graphs. Lately improvements in commercially available equipment and computer programs allow for easier analyses that have led to new methodologies as two- and three-dimensional 'resistivity-imaging' or 'resistivity tomography' (Dahlin and Bernstone, 1997; Griffiths and Turnbull, 1985; Griffiths et al., 1990; Li and Oldenburg, 1992; Loke and Barker, 1996; Vogelsang, 1994; Ward, 1990).

Geo-electrical or resistivity measurements are based on the difference in resistivity between different sub-surface materials. Table IV gives a list of characteristic resistivity values for sub-surface materials. The measuring equipment consists of two current electrodes and two measuring electrodes, a DC current source, and a measuring device. The two current electrodes are inserted in the ground and between the electrodes, a DC potential is maintained that causes a DC current to flow through the ground. The two measuring (= potential) electrodes measure the potential at locations at the surface (figure 10). The potential measured between the two potential electrodes is given by Equation (10). The potential electrodes are normally a copper wire in a $CuSO_4$ solution in a porous pot. The porous pot is in contact with the ground (Figure 13). As it costs only little time to do a self-potential measurement SP surveys are normally combined with resistivity surveys (see below).

$$\begin{aligned} \Delta V &= V_M - V_N = \frac{\rho I}{2\pi} G \\ G &= \frac{1}{\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN}}, \end{aligned} \tag{10}$$

where ρ is the apparent resistivity, I, the current, G, the geometric factor, AM, BM, AN, BN are the distances between electrodes.

Surveys in Geophysics 21: 423–448, 2000.

TABLE IV

Resistivity of soil and rock masses. Left columns show ranges and right columns show dependency on water content (values marked with (1) after Telford et al., 1990, with (2) after Vogelsang, 1994).

Material	Resistivity range (Ohmm = Ωm)	Material (Ohmm = Ωm)	Resistivity
Dry sand (2)	800–5000	Coarse grained sandstone (0.18% water) (1)	1×10^8
Clay (2)	3–150	Coarse grained sandstone (0.39% water) (1)	9.6×10^5
Slate (1)	6×10^2–4×10^7	Dolomite (0.96% water) (1)	8×10^3
Limestone (2)	500–3500	Dolomite (2% water) (1)	5.3×10^3
Sandstone (2)	300–3000	Granite (0.0% water) (1)	1×10^{10}
Granite (1)	300–1×10^6	Granite (0.31% water) (1)	4.4×10^3
Debris and dumped soil (2)	200–350	Basalt (0.0% water) (1)	1.3×10^7
Domestic garbage (2)	12–30	Basalt (0.49% water) (1)	9×10^5
Natural water in sediments (1)	1–100	Basalt (0.95% water) (1)	4×10^4
Sea water (1)	0.2	Siltstone (0.38% water) (1)	5.6×10^8
Scrap metal (2)	1–12	Siltstone (0.54% water) (1)	1.5×10^4

Note that the material descriptions are crude and do not account for variations in, for example, clay content due to weathering, number of water filled discontinuities, or quantity of minerals in solution in water, etc. These factors influence the resistivity values far more than most of the material constituents.

The depth to which the sub-surface materials influence the measured potential depends on the distance between the electrodes, the used array, the current introduced in the ground, and the sensitivity of the measuring equipment. The method can be used for vertical and for horizontal profiling. Deeper materials will influence the potential on the potential electrodes if the distance between the electrodes is larger or if the current is larger. For vertical profiling the spacing between the electrodes is increased with regular steps while the center of the array is fixed. For horizontal profiling, the array of potential electrodes and current electrodes is moved over the surface.

Apart from surface surveys as described above, geo-electric surveys can also be done in one or more boreholes. If one borehole is used for current and potential electrodes, the measurement procedure and interpretation is similar to surface surveys with only the orientation being different. If the potential electrodes are in different boreholes, a form of ‘tomography’ can be achieved (this ‘tomography’ should not be confused with the electrical imaging discussed below which in some literature is also denoted as ‘tomography’). Note the overruling influence of the presence of fresh water (Table IV). The presence of saline water is even more

Surveys in Geophysics 21: 423–448, 2000.

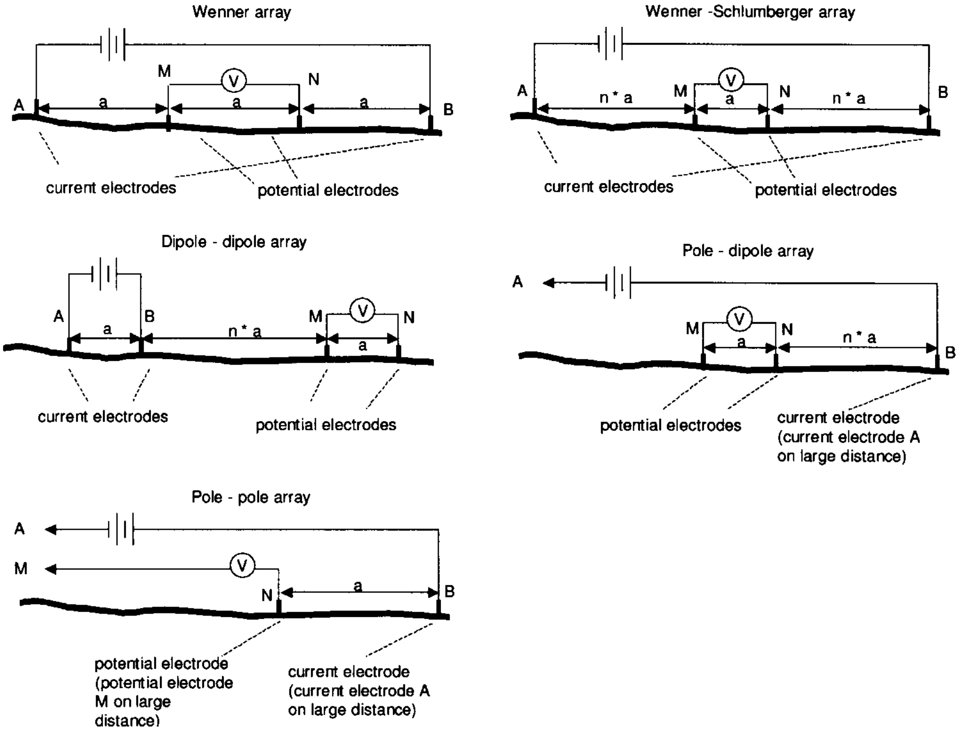

Figure 11. Various arrays for current and potential electrodes.

pronounced. Geo-electrical resistivity surveys are therefore for a very large part governed by the quantities of water present in the sub-surface materials.

4.1. Electrode Arrays

Various electrode arrays are possible in resistivity surveys (Figure 11). The maximum sensitivity of all arrays is obtained near the measuring electrodes. Table V gives the median depth of investigation for the different arrays for a homogeneous sub-surface model. The median depth roughly indicates the depth to which a particular array can be used.

The choice of the array for a field survey depends on the type of feature to be surveyed (e.g., the sensitivity of the array to vertical and horizontal changes in the subsurface resistivity and the depth of investigation), the sensitivity of the resistivity meter, the background noise level, and the signal strength. The advantages and disadvantages are briefly discussed.

4.1.1. *Wenner Array*

The Wenner array is relatively sensitive to vertical changes in the subsurface resistivity below the center of the array and less sensitive to horizontal changes in the subsurface resistivity. The Wenner array is best used for horizontal structures, but

Surveys in Geophysics 21: 423–448, 2000.

TABLE V

Median depth (Z_e) of investigation for different arrays. L is total length of array. For configuration of electrodes see Figure 11 (after Edwards, 1977).

Array	n	Z_e/a	Z_e/L	Array	n	Z_e/a	Z_e/L
Wenner		0.52	0.173	Pole-dipole	1	0.52	
Wenner-Schlumberger	1	0.52	0.173		2	0.93	
	2	0.93	0.186		3	1.32	
	3	1.32	0.189		4	1.71	
	4	1.71	0.190		5	2.09	
	5	2.09	0.190		6	2.48	
	6	2.48	0.190	Pole-pole		0.867	
Dipole-dipole	1	0.416	0.139				
	2	0.697	0.174				
	3	0.962	0.192				
	4	1.220	0.203				
	5	1.476	0.211				
	6	1.730	0.216				

is relatively poor in detecting narrow vertical structures. The Wenner array has a large signal strength.

4.1.2. *Dipole-Dipole Array*

The array is suitable for vertical structures, vertical discontinuities and cavities, but less for identifying horizontal structures. The array is most sensitive to resistivity changes between the electrodes in each dipole pair. The depth of investigation is smaller than for the Wenner array. The signal strength becomes small for large values of the 'n' (Figure 11). The equipment should therefore be of good quality and the resistivity meter should have a high sensitivity. A good contact between the electrodes and the ground should be maintained.

4.1.3. *Wenner-Schlumberger Array*

This array is moderately sensitive to both horizontal and vertical structures. The median depth of investigation for this array is larger than that for the Wenner array for the same distance between the outer electrodes. The signal strength for this array is smaller than that for the Wenner array, but it is higher than for the dipole-dipole array.

Surveys in Geophysics 21: 423–448, 2000.

4.1.4. *Pole-Pole Array*

An array, with only one current and one potential electrode pole cannot exist. However, it can be simulated if one current and one potential electrode are placed at a distance more than 20 times the distance between the *N* and *B* electrodes. This requirement may give practical problems if the distance between the *N* and *B* electrodes is more than a few meters as there may be no space to place the electrodes. The array is sensitive for noise due to the large distance between the potential electrodes.

4.1.5. *Pole-Dipole Array*

The pole-dipole array is asymmetrical and results in asymmetrical apparent resistivity anomalies in the pseudo section for surveys over symmetrical structures. This effect can be removed by repeating the measurements with the electrodes reversed. The *A* electrode must be placed sufficiently far from the survey line. The error caused by neglecting the effect of the *A* electrode in the calculations is less than 5% if the distance to the *A* electrode is more than 5 times the *N*–*B* distance. The pole-dipole array has a higher signal strength compared with the dipole-dipole array. The array is not as sensitive to noise as the pole-pole array because the distance between the potential electrodes is not as large. The signal strength is lower compared with the Wenner and Wenner-Schlumberger arrays but higher than the dipole-dipole array.

4.1.6. *High-Resolution Electrical Surveys with Overlapping Data Levels*

A technique similar to CDP gathering in seismics can be used to improve the data quality for resistivity surveys, particularly in noisy areas. This is done by using overlapping data levels with different combinations of ‘*a*’ and ‘*n*’ values for the Wenner-Schlumberger, dipole-dipole and pole-dipole arrays.

4.2. RESISTIVITY IMAGING

The standard methodology as explained above does not take into account horizontal and vertical changes in resistivity in the same series of measurements. Lateral changes are neglected in vertical profiling and vertical changes are neglected in horizontal profiling. This is mostly no problem for a layered sub-surface. If the sub-surface materials change in vertical and in horizontal direction, such as, for example, karst holes or discontinuity surveys, the standard method is less suitable. In recent years two- or three- dimensional (2D or 3D) electrical imaging surveys to map areas with lateral and vertical changes in resistivity have been developed (Dahlin and Bernstone, 1997; Li and Oldenburg, 1992; Loke and Barker, 1996).

Rather than using one set of current and measuring electrodes, a series of electrodes (20 or more in a 2D survey or 256 or more in a 3D survey) is used. The electrodes are alternating used as current or potential electrodes. The electrodes are connected to an automated (computer operated) switch box that selects the

Surveys in Geophysics 21: 423–448, 2000.

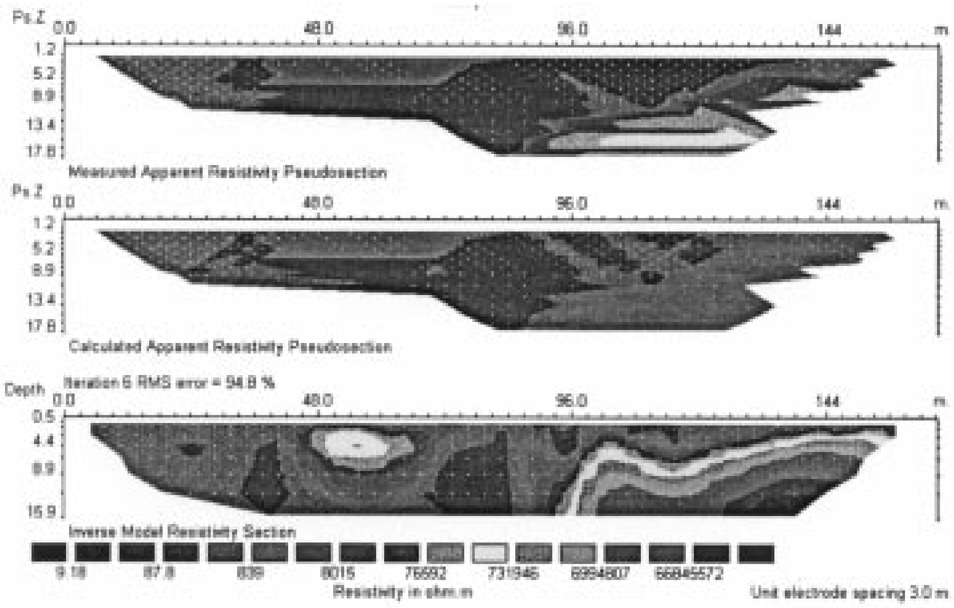

Figure 12. Geo-electrical imaging survey (after Graaff et al., 2000).

4 electrodes to be used. A computer controls the switch box and the measuring device, and runs a program that selects the electrodes, makes the measurement, and stores the measurement. The same or a different program can do the processing of the measured signals.

The measured resistivity data is normally plotted by the pseudo section contouring method (Figure 12). The horizontal location of the point is placed at the midpoint of the set of electrodes used to make that measurement. The vertical location of the plotting point is placed at a distance that is proportional to the separation between the electrodes. Another method is to place the vertical position of the plotting point at the median depth of investigation (Edwards, 1977). The pseudo section gives an approximate picture of the true subsurface resistivity distribution. Various computer programs are available that calculate the apparent resistivity pseudo section for a user defined 2D or 3D subsurface model. The 2D and 3D resistivity imaging or tomography is a very promising investigation method.

Figure 12 shows an example of a geo-electric imaging survey for establishing soil and rock mass movement due to the presence of anhydrite or gypsum in a slope. The high resistivity values at the right of the line correspond with the presence of rock, the low resistivity values correspond with unconsolidated sediments (a road and pipelines cause the high resistivity values between 50 and 70 m).

Surveys in Geophysics 21: 423–448, 2000.

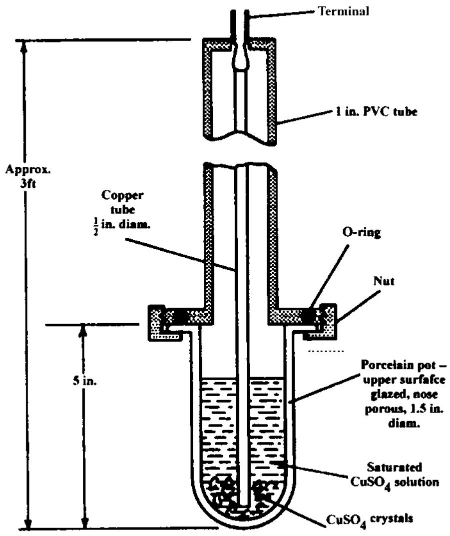

Figure 13. $CuSO_4$ electrode for SP and electrical surveys (after Telford et al., 1990).

5. Self-potential (SP)

Self-potential (SP) surveys are based on certain electrochemical or mechanical processes in the sub-surface that create spontaneous electrical potentials (Bogoslovsky and Ogilvy, 1977; Bruno et al., 1998; Telford et al., 1990). Groundwater is always the controlling factor in these processes. The spontaneous potentials are related to, for example, weathering of sulphites, variation in rock properties, geologic contacts, corrosion, etc.

The measuring equipment is simple. Two electrodes in contact with the ground and a potential meter connected to both electrodes. The electrodes consist of a copper wire in a $CuSO_4$ solution in a porous pot (Figure 13). This is done to prevent erratic self-potentials between the electrodes and the ground. SP surveys are normally combined with an electrical resistivity survey.

Interpretation of SP survey data in geotechnical work is normally only qualitative. The distribution of SP data may be related to a weathering or soil profile, to changes in weathering profiles, or to the extent of certain geological boundaries. Quantitative interpretation is very difficult as the SP potentials for most sub-surface materials important for geotechnical work are unknown.

6. Micro-Gravity

Gravimetry surveys investigate the difference in densities between different sub-surface materials. In geotechnical work, the name micro-gravity is used to indicate that the differences measured are very small. Table II gives some densities of

Surveys in Geophysics 21: 423–448, 2000.

different soil and rock masses. The gravity measured at the surface of the earth is compared with a theoretical value (normal gravity) corresponding to an earth model in which only radial density variation is present. It is given by:

$$\gamma = \gamma_e(1 + 0.0053024 \sin^2 \varphi - 0.56 \times 10^{-5} \sin^2 2\varphi), \tag{11}$$

where γ is the gravity at sea level, φ, the latitude, γ_e, the gravity at sea level at the equator = 978.03267715 cm/s^2.

The difference between observed and normal gravity (anomalies) are interpreted in terms of lateral variation of density. The standard unit in gravity surveys is cm/sec^2 and is called the 'gal'. Micro-gravity meters measure with an accuracy up to about 10^{-6} gal or 10^{-3} mgal (= milli gal).

The gravity measurements are influenced by a whole series of factors, such as, tidal effect, elevation, topographic relief, and instrument drift, that need to be compensated for with proper corrections. In micro-gravity surveys for geotechnical work, most of the influences may be less critical as only relative measurements are made on short distance. Interpretation is carried out by comparing the observed anomaly with a numerical model of the bodies that cause them. Micro-gravity surveys have been used with success to establish karst holes and differences in groundwater levels. The applications to slope stability studies are rare and require an accurate topographic map to correct the effect of irregular topographic relief on the gravity measurements. Potentially, gravimetry can give 'in-situ' estimate of the density of slope material using methods that correlate elevation with gravity differences (Nettleton, 1939; Parasnis, 1962).

7. Discussion

None of the geophysical methods is better than another method. The success with which a method is applied fully depends on the circumstances at the site and on the sub- surface materials. Unsuccessful surveys are nearly always due to a lack of proper preparation. Often a survey is done only based on a vague article describing a similar type of survey, or on just the recommendation of an assumed 'expert'. To avoid disappointing results it is therefore important to establish on forehand whether it is likely that the required structures or properties can be measured with success in a particular situation. This should not be done only qualitative by estimation based on experience, but calculations should be done that simulate reality at the actual site as best as possible. Many surveys are unsuccessful because this has not been done properly.

A combination of different methods for the same site leads often to successful geophysical surveys because different features of the sub-surface structure are detected by different methods (Anon., 1995; Bruno et al., 1998; Williams et al., 1996). Secondly, different methods may confirm the existence of a vague feature

Surveys in Geophysics 21: 423–448, 2000.

TABLE VI

Tentative indication of suitability of various geophysical methods. It is assumed that basic boundary conditions have been fulfilled; for example, no highly conductive materials present in topsoil where a ground radar survey is to be done of the under laying rock mass.

Method		Artifacts, pipes, foundations, etc.	Property determination for geotechnical purposes	Structure			
				Low contrast*		High contrast*	
				Simple**	Complex**	Simple**	Complex**
Seismic	Refraction	−−	+	−	−−	++	−
	Reflection	−	+	+	−	++	+
	Tomography	−	+	++	++	++	++
Electro-magnetic	Low frequency	++	−−	−	−−	−	−−
	groundradar	++	−−	++	+	++	++
Geo-electrical	Normal	−	−−	−	−−	++	+
	2/3D imaging	−	−−	++	++	++	++
Self-potential		−−	−−	−−	−−	−	−
Gravity		−	+	−	−−	++	+

−− = not suitable, − = marginal, + = good, ++ = very good.

* Low and high contrast refer to the contrast in property values measured between the different materials that define the structure. ** Simple and complex structure refer to the complexity of the structure to be measured, for example, simple should be something like two horizontal or slightly inclined layers, e.g., a topsoil layer on a rock slope, complex should be a series of irregular layers and objects, e.g., a debris flow deposit.

Surveys in Geophysics 21: 423–448, 2000.

that may be missed if only one method is used. Table VI lists the various methods and whether a method is more or less suitable for a particular task. It should be realized that the table is very crude and is in no-way conclusive.

8. Conclusions

The internal structure and the mechanical properties of a soil or rock mass have to be known or estimated with a reasonable degree of certainty to assess slope stability. Direct observation of the internal structure and testing of properties of soil or rock mass is always preferred, but this requires boreholes or trenches that are often impossible or too costly to be made. Geophysical methods can be cost effective means to establish the internal structure and/or the properties of the soil or rock mass materials where boreholes or trenches are not possible or impractical. Seismic methods are most suitable because the measured properties directly depend on the mechanical properties of the materials. Modern seismic equipment allows for high-resolution surveys and interpretation. Developments in electromagnetic and geo-electrical methods are promising and these techniques may well become standard tools in geotechnical practice in due time. Provided that the measured boundaries can be correlated with boundaries of mechanical properties of the soil or rock mass.

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