VELOCITIES IN REFLECTION SEISMOLOGY pdf

1: Seismic imaging and inversion -

Reflection seismology has reached and consolidated this position because it has shown itself to be capable of adapting to the increasing complexity of the requirements of exploration. Initially directed towards geometric mapping of the sub-surface, it became the means of detection of structural traps in geotectonically quiescent regions, and.

The seismic refraction method has a long history with the oil industry, and as such the equipment is easy to obtain and deploy. Because this method uses travel time to measure the seismic properties of materials it is highly precise in the measurement of seismic velocity and quite accurate in the measurement of material thicknesses in many instances. Because seismic velocity is diagnostic for different types of material, and generally increases with degree of induration or hardness, the seismic refraction method not only can measure depth to a hard layer but can be used to non-invasively classify the type of material e. In addition, seismic layers often correlate closely with geologic contacts. Spectrum utilizes the Seistronix RAS twenty-four channel signal enhancement seismograph along with geophones and associated cabling to collect seismic refraction data. HOW IT WORKS The seismic refraction method is based on the fact that when a wave reaches a boundary between two materials having different seismic velocities, that seismic wave will be refracted or bent either toward the normal to the interface or away from the normal to the interface, depending on whether the velocity increases or decreases at the boundary. In the special case where layer velocity increases with depth at the boundary, critical refraction occurs where seismic waves travel along the interface between the two materials. The angle at which the seismic waves are critically refracted the critical angle is uniquely determined by the ratio of the velocities of the two materials: In practice, a linear array or spread of geophones is established along the ground surface and connected to a multi-channel seismograph. The first arrivals at each geophone location on the seismic record from a given shot are plotted on a time vs. The slope of the line segments created by the first arrivals is the inverse of the velocity of the material through which the waves have traveled, and the intercept time, crossover distance or critical distance can be used to calculate the depth to the target layer. During a typical seismic refraction survey, shots at several different locations on the spread are made in order to obtain a measure of the two-dimensional variation of seismic velocity with depth along the line. Several different software programs are currently available to process seismic refraction data, and the end product is generally a two-dimensional profile indicating the variation of seismic velocity along the line. Spectrum uses the generalized reciprocal method GRM to acquire, process, and interpret seismic refraction data. GRM is a seismic refraction interpretation method designed to accurately map undulating refractor surfaces from in-line refraction data using both forward and reverse shots. The method is related to the Hales and the reciprocal methods of seismic refraction interpretation. In this method it is crucial to acquire at least 7 shots per spread: Assuming a channel acquisition system, this requires one center shot between geophones 12 and 13, one shot between each of geophone pairs 6 and 7 and 18 and 19, a near-offset shot from each end geophone geophones 1 and 24 , and one far-offset shot off each end of the spread. In practice, Spectrum often uses 13 shots per spread for high resolution investigations. In order to map the bedrock surface, one condition that must be met is that of having an off-end shot at each end of the spread such that each first arrival on the geophone spread is a bedrock arrival. Blind zones and hidden layers may be encountered. A hidden layer is an intermediate velocity, intermediate depth layer whose thickness or velocity is such that rays from a deeper, higher velocity layer arrive at the ground surface sooner than rays from the hidden layer. Standard seismic refraction interpretation schemes will yield significant errors in calculated refractor depths in the presence of blind zones or hidden layers. GRM is the only method capable of overcoming this problem. Data processing includes the selection of first break arrival times, the generation of time-distance plots for each line, the assignment of selected portions of the travel time data to individual refractors, and the phantoming of travel time data for the target lower refractor. Once this preprocessing work has been done, GRM processing begins. Once this is done, layer thicknesses and velocities are calculated and a geophysical interpretation of the geological parameters may be made. The end product is a seismic refraction profile that indicates the seismic layers detected, the depths to the interfaces between layers as they vary along the line and the seismic velocities encountered. Delineation of faults or other geologic structures.

VELOCITIES IN REFLECTION SEISMOLOGY pdf

2: Seismic wave - Wikipedia

6 Physics and chemistry of the Earth's interior - Seismic reflection Velocity determination The interval velocities cannot be determined very well from stacking.

Higher frequencies Data Processing Processing is typically done by professionals using special purpose computers. These techniques are expensive but technically robust and excellent results can be achieved. A complete discussion of all the processing variables is well beyond the scope of this manual. However a close association of the geophysicist, the processor and the consumer is absolutely essential if the results are to be useful. Well logs, known depths, results from ancillary methods, and the expected results should be furnished to the processor. At least one iteration of the results should be used to ensure that the final outcome is successful. One important conclusion of the processing is a true depth section. The production of depth sections requires conversion of the times of the reflections to depths by derivation of a velocity profile. Well logs and check shots are often necessary to confirm the accuracy of this conversion. Advantages and Limitations It is possible to obtain seismic reflections from very shallow depths, perhaps as shallow as 3 to 5 m. Variations in field techniques are required depending on depth. Containment of the air-blast is essential in shallow reflection work. Success is greatly increased if shots and phones are near or in the saturated zone. Severe low-cut filters and arrays of a small number of geophones are required. Generally, reflections should be visible on the field records after all recording parameters are optimized. Data processing should be guided by the appearance of the field records and extreme care should be used not to stack refractions or other unwanted artifacts as reflections. Subbottom Profiling A variant of seismic reflection used at the surface of water bodies is subbottom profiling or imaging. The advantage of this technique is the ability to tow the seismic source on a sled or catamaran and to tow the line of hydrophones. This procedure makes rapid, continuous reflection soundings of the units below the bottom of the water body, in other words, the subbottom. This method and significant processing requirements have been recently developed by Ballard, et al. The equipment, acquisition, and processing system reduce the need for over-water boring programs. The developed WES imaging procedure resolves material type, density, and thickness Ballard, et al. Basic Concepts The acoustic impedance method may be used to determine parameters of the soft aqueous materials. The acoustic impedance z for a unit is the product of its pb and VP. Since the E ref l , can be determined, and E inc , and z 1,water are known, z i , 2 may be determined. V P , 2 may be assessed from the depth of the boundary, and thus? The material properties of lower units can be found in succession from the reflections of deeper layers. Data Acquisition A variety of different strength sources are available for waterborne use. By increasing strength, these sources are: Although there is some strength overlap among these sources, in general, as energy increases, the dominant period of the wave increases. For the larger source strength, therefore, the ability to resolve detail is impaired as period and wavelength become larger. The conflicting impact of energy sources is the energy available for penetration and deeper reflections. The greater energy content and broad spectrum of the boomer allow significantly greater depth returns. Some near-bottom sediments contain organic material that readily absorbs energy. Higher energy sources may allow penetration of these materials. Data collection is enormous with a towed subbottoming system. Graphic displays print real-time reflector returns to the hydrophone set. Recording systems retrieve the data for later processing. The field recorders graph time of source firing versus time of arrival returns. Data Processing Office processing of the field data determines the subbottoming properties empirically. The processing imposes the Global Positioning System GPS locations upon the time of firing records to approximately locate the individual shot along the towed boat path. The seismic evaluation resolves the layer VP and unit depths. From the firing surface locations and unit depths, the field graphs are correlated to tow path distance versus reflector depths. Figure shows cross sections of the Gulfport Ship Channel, Mississippi. These are fence diagrams of depth and material types once all parallel and crossing surveys are resolved. Advantages and Limitations The subbottoming technique can be applied to a large variety of water bodies. Saltwater harbors, shipping channels, and river waterways were the original objective for the technique. The method is now used on locks, dams, reservoirs, and engineering projects such as the location of pipelines. Reflected subbottoming signal amplitude cross section, 3. Ballard, McGee and Whalin, Figure 9. Density cross section in Gulfport ship canal, Mississippi. McGee and Whalin Fathometer Surveys Basic Concepts Fathometers are also called Echo Sounders and are similar to reflection seismic profilers in that they also employ an acoustic source and receiver placed immediately beneath the surface of the water. However, fathometers differ from reflection seismic profilers in that they use higher frequency acoustic source pulses varying from less than 10 kHz up to about kHz. Some of this energy transmitted by the source is reflected from the sediments at the water bottom, and the reflections are recorded by the receiver and stored digitally. Data Acquisition Fathometers determine water depth by repeatedly transmitting seismic energy through the water column and recording the arrival time of the reflected energy from the water bottom. The time required for the seismic signal to travel from its source to a reflector and back is known as the two-way travel time, and it is measured in milliseconds ms equal to 1 x seconds. The Fathometer calculates the depth to a water bottom by dividing the two-way travel time by two and multiplying the result by the velocity of sound through water: Fathometer surveys are conducted while traveling at a moderate speed in a boat. Typically, the transducer is mounted on the side of the boat and placed in the water. Data recording is essentially automatic with a t recorder plotting providing a hard copy of the data or a computer screen may be used for the display. The data may also be stored on magnetic tape for further processing and plotting. Fathometers are calibrated by adjusting the value of V, which may vary slightly depending on water type. Most Fathometers use a narrow-bandwidth kHz seismic signal. These Fathometers provide accurate depth data, but little or no information about the subbottom. Fathometers that use a lower frequency signal, such as 20 kHz, can detect reflected energy from subbottom interfaces, such as the bottom of an infilled scour hole. Fathometer systems come with black and white t recording systems and in color systems. Colors are often assigned according to the different amplitudes of the reflected signals. Fathometers color step sizes as low as four dB are now available, allowing quite small changes in the reflected signals to be observed. An event marker button is often available allowing vertical line marks to be placed on the records when specific locations are selected by the operator. Sometimes the data can be downloaded to a computer allowing digital processing to be done along with data plots. A color Fathometer can be calibrated to measure and display in color the amplitude of the reflected signal, which, in constant water depths, can be related to characteristics of the bottom material. Estimated seismic interval velocities can be used to transform the time-depth profile into a depth profile. However, water velocities are a function of suspended sediment load, and can vary appreciably. The data are interpreted by viewing the plotted data. The response of specific objects may be used if these were noted on the records using a button marker. Figure 10 shows Fathometer data recorded with a kHz transducer. Note that only the water depth is observed in these data. Because of the high frequency, little energy is transmitted into the bottom sediments, and thus no reflections are observed from within the sediments. Fathometer data recorded with kHz transducer. Because of the lower frequency, some of the energy is transmitted into the sediments and reflections are seen. Fathometer data recorded using a 3. The tool can provide an accurate depth-structure model of the water bottom if acoustic velocities are known. Post acquisition processing migration can be applied. The main disadvantages of Fathometers in continuous mode are as follows: The source and receiver need to be submerged. Profiles cannot be extended across emerged sand bars or onto the shore. The equipment is relatively expensive hardware and software. Post acquisition processing migration may be required in areas where significant structural relief is present. Fathometers are also employed in a spot survey mode. In this type of survey, sounding data single reflection traces are acquired at irregularly or uniformly spaced intervals typically on the order of meters at the water surface. The first high-amplitude reflected event is usually interpreted to be the water bottom reflection. Note, that spot data usually cannot be accurately migrated because of aliasing problems. The pages found under Surface Methods and Borehole Methods are substantially based on a report produced by the United States Department of Transportation:

VELOCITIES IN REFLECTION SEISMOLOGY pdf

3: Spectrum Geophysics Utility Locating and Geophysical Services

Get this from a library! Velocities in Reflection Seismology. [Jean-Pierre Cordier] -- Although considera bIe efforts are now being made to find new sources of energy, all the experts are agreed that hydrocarbons will have to provide the greater part of our energy needs for a.

Stress and Strain A force applied to the surface of a solid body creates internal forces within the body: Stress is the ratio of applied force F to the area across which it is acts. Strain is the deformation caused in the body, and is expressed as the ratio of change in length or volume to original length or volume. Triaxial Stress Stresses act along three orthogonal axes, perpendicular to faces of solid, e. Pressure Forces act equally in all directions perpendicular to faces of body, e. Strain Associated with Seismic Waves Inside a uniform solid, two types of strain can propagate as waves: Axial Stress Stresses act in one direction only, e. Change in volume of solid occurs. Associated with P wave propagation Stresses act parallel to face of solid, e. No change in volume. Fluids such as water and air cannot support shear stresses. Associated with S wave propagation. At low to moderate strains: At very high strains: Constant of proportionality is called the modulus, and is ratio of stress to strain, e. Seismic Body Waves Seismic waves are pulses of strain energy that propagate in a solid. Two types of seismic wave can exist inside a uniform solid: P waves have highest speed. Volumetric change Sound is an example of a P wave. B S waves Secondary, Shear, Shake Particle motion is in plane perpendicular to direction of propagation. If particle motion along a line in perpendicular plane, then S wave is said to be plane polarised: SV in vertical plane, SH horizontal. No volume change S waves cannot exist in fluids like water or air, because the fluid is unable to support shear stresses. Near the surface the particle motion is retrograde elliptical. Rayleigh wave speed is slightly less than S wave: Love waves Occur when a free surface and a deeper interface are present, and the shear wave velocity is lower in the top layer. Particle motion is SH, i. Seismic Wave Velocities The speed of seismic waves is related to the elastic properties of solid, i. Constraints on Seismic Velocity Seismic velocities vary with mineral content, lithology, porosity, pore fluid saturation, pore pressure, and to some extent temperature. Sedimentary Rocks In sedimentary rocks, effects of porosity and grain cementation are more important, and seismic velocity relationships are complex. Various empirical relationships have been estimated from either measurements on cores or field observations: If the velocities of pore fluid and matrix known, then porosity can be estimated from the measured P wave velocity. Nafe-Drake Curve An important empirical relation exists between P wave velocity and density. Crossplotting velocity and density values of crustal rocks gives the Nafe-Drake curve after its discoverers. Only a few rocks such as salt unusually low density and sulphide ores unusually high densities lie off the curve. Waves and Rays In a homogeneous, isotropic medium, a seismic wave propagates away from its source at the same speed in every direction. The wavefront is the leading edge of the disturbance. The ray is the normal to the wavefront. Reflection and Refraction at Oblique Incidence When a P wave is incident on a boundary, at which elastic properties change, two reflected waves one P, one S and two transmitted waves one P, one S are generated. Angles of transmission and reflection of the S waves are less than the P waves. Amplitude of Reflected and Transmitted Waves At oblique incidence, energy transformed between P and S waves at an interface. Amplitudes of reflected and transmitted waves vary with angle of incidence in a complicated wave given by Zoeppritz equations. Example P wave reflection amplitude can increase at top of gas sand. Same formulae apply to S waves at normal incidence. Refracted wave travels along the upper boundary of the lower medium. The ray associated with this head wave emerges from the interface at the critical angle. This phenomenon is the basis of the refraction surveying method.

4: An Introduction to Seismic Refraction Theory and Application

Page 60 Reflection seismic 1 Version () (t2-x2)-Analysis The (t2-x2)-Analysis is based on the fact, that the Moveout-expression for the square of t and x result in a linear event.

These traces were recorded at different offset distances, and the travel times for seismic waves traveling to and from a given reflecting horizon varies with that distance Fig. This moveout is expressed in terms of a velocity and represents the seismic velocity that the entire overburden, down to the point of each particular reflection, would have to result in the idealized hyperbolic shape observed. This velocity analysis is usually conducted by examining the semblance or some other measure of similarity across all the traces, within a moving time window, and for all reasonable stacking velocities Fig. The seismic processor then selects the best set of velocities to use at a variety of reflectors and constructs a velocity function of two-way travel time. At the top of the figure is a schematic ray diagram, showing an earth model with four reflecting interfaces; rays are drawn from three source locations to three receiver locations, as they are reflected from two of the interfaces the other source-receiver rays and reflections from other interfaces are not shown. The lower part of the figure shows the seismograms that would be recorded from this scenario, ignoring the direct wave in the upper layer, multiples, and noise. Notice that the distance used to label the seismic gathers is the total sourcereceiver distance. Synthetic seismic gather taken from Yilmaz. The gather is analyzed over narrow time windows for the values of semblance or some other measure of similarity according to a range of stacking velocities. The contours indicate the level of semblance, and the processing geophysicist selects the values deemed to be because of primary reflections and not events reflected multiple times. The direct wave straight-line arrivals seen at the upper edge of the arrivals on the seismic gather are not considered in the analysis after Yilmaz [1]. The final stacked traces exhibit a considerably better signal-to-noise ratio than the individual seismic traces recorded at zero-offset, but the improvement is better than just the square root of the number of traces that might be expected because of the systematic removal of coherent noise. Much of the noise present in individual seismic traces is not random but represents unwanted events, including surface waves or ground roll and multiply-reflected arrivals from shallow horizons; both of these can usually be identified in the velocity analysis and selected against. The stacking process then removes most of the unaligned energy associated with these types of coherent noise. The velocities obtained in the analysis previously described are not true seismic velocitiesâ€”they are simply those velocities which provided the best stack of the data and may or may not truly reflect the actual root mean square RMS velocities that approximate the accumulated effect of the stack of layers above the reflector the name RMS is derived from the arithmetic used to define this overall velocity. If we assume, however, that the stacking velocities do in fact provide a reasonable approximation to the aggregate effect of the layers overlying each reflector, the actual velocities of each layer can be obtained through a set of equations because of Dix [2] see Fig. The stacking velocities are those obtained from the velocity analysis see Fig. Time and depth migration Even after accounting for normal moveout and stacking the gathered traces to a common zero-offset equivalent set of traces, the locations of the reflected events are not usually correct because of lateral variations in velocity and dipping interfaces. Each seismic trace is plotted directly beneath the respective midpoint or bin location used for stacking, but the reflection from any given interface may not have come from that location. The events have been shifted downdip to deeper locations, and the dip of the interface is less steep. In the simple case shown in the figure, we need only know the velocity of the one overlying layer, but in more realistic cases, the velocity function may be quite complex and is derived through a trial-and-error approach guided by statistical tests of lateral coherence, knowledge of expected geologic structure, and other constraints such as interval velocities and well log data. The problem can become quite difficult in complicated 3D data sets, and software has been developed to manage and visualize the velocity volume. The result of this model-driven 3D migration can be somewhat subjective, and, although it is possible to create structures where none really exist through this process, migration should be performed on all seismic data sets for appropriate imaging of structures. The results can occasionally be quite dramatic for interpretation; for example, a locally high feature on an unmigrated data set may move to a significantly different map location after migration. In general, the more dramatic the structure, or the larger the velocity contrasts between layers, the more important 3D migration is for proper imaging. The true Earth model in two-way travel time for a simple dipping interface is shown at the top, with normal-incidence zero-offset seismic rays drawn to two surface source-and-receiver locations. Because the seismograms are plotted directly beneath the surface locations, a seismic section will display the dipping interface at the incorrect location, as shown at the bottom. Notice that the seismic images the event downdip of its true location and with a less-steep dip. The processing step of migration attempts to correct for this, displacing the events back to their true locations. These three panels show the same cross section of the Earth. The panel on the left was imaged as a 2D stack, extracted from a 3D data volume without migration. The panel in the center was imaged using 2D migration techniques. The panel on the right was imaged using 3D post-stack time migration. Note the improving quality of the data, particularly deeper in the section after Yilmaz [1]. The process of imaging through modeling the velocity structure is a form of inversion [4] of seismic data, and the term inversion is often used to imply building a velocity model which is iteratively improved until it and the seismic data are optimally in agreement. The finest results can usually be obtained from prestack depth migration, in which each sample of each trace, prior to gather, is migrated using the velocity function to a new location then stacked and compared with various tests for model improvement; the model is changed, and the process is repeated. In areas where it is important to image beneath layers of high velocity contrasts, such as beneath salt bodies, prestack depth migration is required. The example [8] shown in Fig. The process required to create the final stack is as follows: Then, the salt velocity which is fairly constant and typically much higher than the surrounding sediments, resulting in severe bending of seismic ray paths is used for the half-space beneath the top of salt. The reflections from the base of the salt body then appear, although the underlying sediments are very poorly imaged. Finally, the velocity model within these sediments is modified until an acceptable image is obtained. The upper part shows a result of imaging beneath salt using prestack time migration; the middle part uses post-stack depth migration; and the bottom uses prestack depth migration. Note the increasing ability to image sediments below the salt body after Liro et al. Trace inversion for impedance Seismic reflections at zero offset result from contrasts in acoustic impedance, involving just the P-wave velocity and density of the layers at the interface. If we can identify the seismic wavelet that propagated through the earth and reflected from the layer contrasts, we can then remove the effect of that wavelet and obtain a series of reflection coefficients at the interfaces. Then, we can simply integrate these reflection coefficients and determine the acoustic impedance in the layers between the interfaces. However, the approaches used are quite different, and the two processes should not be confused. Future research developments may tend to blur this distinction by integrating appropriate aspects of both techniques into one method. Acoustic impedance If the seismic data were noise-free and contained all frequencies, from zero frequency infinite wavelength to very high frequencies short wavelengths , the solution should be unique, but seismic data are noisy and band-limited and do not contain the very lowest frequencies nor the higher frequencies that are often of interest. In general, a calibrated and competently processed inversion volume can be of considerable use to the interpreter or the engineer, providing insight to layer properties and continuity, which may not be apparent from the traditional reflection-seismic display; in particular, the thinner beds are usually more distinctly identified through removal of wavelet tuning interference of reflections from the top and bottom of the bed and subtle changes in impedance that are not easily recognized in the reflection image that can be seen in the inverted volume. Because the inversion process results in volume properties, rather than interface properties, it is possible to isolate and image individual bodies within certain impedance ranges. An example of the results of body-capture after a sparse-spike inversion, intended to identify hydrocarbon reservoirs, is shown in Fig. The results of trace inversion can be used to identify spatially distinct bodies with specific impedance ranges. This example is of the same area of the Gulf of Mexico as shown in Fig. In this area, virtually all these bodies are hydrocarbon reservoirs, although not all are large enough to be economically produced. The two reservoirs identified by A and B in Fig. A perspective view of a single horizon containing several potential reservoirs is shown from the Teal South area of the Gulf of Mexico. The coloring is based on the amplitude of the reflected arrival at this horizon, with the hotter colors indicating larger negative amplitudes, resulting in this case from high-GOR oil in both producing and unproduced reservoirs. The reservoirs have been highlighted for increased visibility on the black-and-white version of a typically color display. Yet typical seismic data has been processed by stacking all appropriate offsets after correcting for normal moveout and muting, and the amplitude of each reflection represents a sort of average amplitude over all of the offsets used. In many cases, this distinction is not important because the amplitude normally decays slightly with offset after routine correction for geometric spreading and affects all stacked samples similarly, but for many cases, and especially those of most interest, the amplitudes vary with offset. Inverting a seismic section containing stacked data does not always yield a true acoustic impedance volume. In practice, this is true for seismic compressional waves at normal incidence but is not valid for compressional waves at nonnormal incidence in a solid material because of partial conversion to reflected and refracted shear waves. In the cases where these assumptions are not true, we must recognize that the values of acoustic impedance resulting from the inversion process are not precise; in fact, the disagreement of the acoustic inversion results, with a model based on well logs, is often an indication of AVO effects and can be used as an exploration tool. Elastic impedance In order to separate the acoustic model compressional-wave only from the elastic model including shear effects , the inversion process can be conducted on two or three different stacked seismic volumes, each composed of traces that resulted from stacking a different range of offsets. The volume created from traces in the near-offset range or a volume made by extrapolating the AVO behavior to zero offset at each sample is inverted to obtain the acoustic impedance volume. Converted-wave data can also be inverted for shear impedance.

5: seismic_refraction_versus_refl

the nature of the fluids which it contains has little influence (not detectable by seismic means) on the value of the reflection coefficient; if the sandstone has a high porosity (24%). it is theoretically possible to distinguish between deposits of hydrocarbons and those of water.

Seismic Refraction Versus Reflection Introduction The difference between seismic refraction and seismic reflection is never obvious to the non geophysicist, and rarely explained in simple terms by geophysicists. Due to the similarity of the names, many non geophysicists assume that the terms are interchangeable, or are unaware that there are critical differences between the two techniques that may make one vastly preferred or the other completely unusable given site specific conditions or project goals. General Seismic Principles Seismic techniques generally involve measuring the travel time of certain types of seismic energy from surficial shots i. In the subsurface, seismic energy travels in waves that spread out as hemispherical wavefronts i. The energy arriving at a geophone is described as having traveled a ray path perpendicular to the wavefront i. In the subsurface, seismic energy is refracted i. The refraction and reflection of seismic energy at density contrasts follows exactly the same laws that govern the refraction and reflection of light through prisms. Note that for each seismic ray that strikes a density contrast a portion of the energy is refracted into the underlying layer, and the remainder is reflected at the angle of incidence. The reflection and refraction of seismic energy at each subsurface density contrast, and the generation of surface waves or ground roll , and the sound i. The ground motion produced by a shot is typically recorded as a wiggle trace for each geophone see Example Seismic Record at right. Seismic Refraction Seismic refraction involves measuring the travel time of the component of seismic energy which travels down to the top of rock or other distinct density contrast , is refracted along the top of rock, and returns to the surface as a head wave along a wave front similar to the bow wake of a ship see Seismic Refraction Geometry below. The shock waves which return from the top of rock are refracted waves, and for geophones at a distance from the shot point, always represent the first arrival of seismic energy. Seismic refraction is generally applicable only where the seismic velocities of layers increase with depth. Therefore, where higher velocity e. In addition, since seismic refraction requires geophone arrays with lengths of approximately 4 to 5 times the depth to the density contrast of interest e. Greater depths are possible, but the required array lengths may exceed site dimensions, and the shot energy required to transmit seismic arrivals for the required distances may necessitate the use of very large explosive charges. In addition, the lateral resolution of seismic refraction data degrades with increasing array length since the path that a seismic first arrival travels may migrate laterally i. Recent advances in inversion of seismic refraction data have made it possible to image relatively small, non-stratigraphic targets such as foundation elements, and to perform refraction profiling in the presence of localized low velocity zones such as incipient sinkholes. Seismic Reflection Seismic reflection uses field equipment similar to seismic refraction, but field and data processing procedures are employed to maximize the energy reflected along near vertical ray paths by subsurface density contrasts see Seismic Refraction Geometry below. Reflected seismic energy is never a first arrival, and therefore must be identified in a generally complex set of overlapping seismic arrivals â€” generally by collecting and filtering multi-fold or highly redundant data from numerous shot points per geophone placement. Therefore, the field and processing time for a given lineal footage of seismic reflection survey are much greater than for seismic refraction. However, seismic reflection can be performed in the presence of low velocity zones or velocity inversions, generally has lateral resolution vastly superior to seismic refraction, and can delineate very deep density contrasts with much less shot energy and shorter line lengths than would be required for a comparable refraction survey depth. The main limitations to seismic reflection are its higher cost than refraction for sites where either technique could be applied , and its practical limitation to depths generally greater than approximately 50 feet. At depths less than approximately 50 feet, reflections from subsurface density contrasts arrive at geophones at nearly the same time as the much higher amplitude ground roll surface waves and air blast i. Reflections from greater depths arrive at geophones after the ground roll and air blast have passed, making these deeper targets easier to detect and delineate. Seismic reflection is particularly suited to marine applications e. Comparison The differences between seismic refraction and reflection are summarized in the table below.

VELOCITIES IN REFLECTION SEISMOLOGY pdf

6: Reflection seismology - Wikipedia

3 Physics and chemistry of the Earth's interior - Seismic refraction Reflection and refraction Seismic rays obey Snell's Law (just like in optics).

What is Seismic Refraction? One can study subsurface velocity and layer interface structure by analyzing the first arrival times of P-waves longitudinal or compressional waves at the surface of the earth. This technique is termed seismic refraction. Applications of subsurface imaging include locating buried archeological sites, assessing subsurface geological hazards, defining aquifer geometry, and exploring for fossil fuel and other natural resources. P-waves traveling through rock are analogous to sound waves traveling through air. The speed a P-wave propagates through a medium depends on the physical properties i. Spherical wave fronts emanate from a source, as well as ray paths. Ray paths travel normal to the spherical wave surface. For seismic refraction discussion, it is useful to imagine seismic waves as ray paths. When a ray encounters an inhomogeneity in its travels, for example a lithological contact with another rock, the incident ray transforms into several new rays. A reflected wave enters and exits at the same angle measured to the normal of the boundary - angle of incidence equals angle of reflection. For refraction seismology, the critical angle is the most important angle value to understand. If angle r equals 90 degrees, then the refracted wave propagates along the boundary interface. These secondary ray paths exit at the critical angle. A Simple Refraction Model: Two Horizontal Layers In the ideal world of engineering , refraction seismology is most easily understood through a horizontal two layer model. Seismic waves are generated from a source sledge hammer. Geophone receivers record seismic signals received along the survey profile. Since P-waves travel at the fastest speeds, the first seismic signal received by a geophone represents the P-wave arrival. Five P-waves are of interest in refraction seismology: The direct wave propagates along the atmosphere-upper layer 1 boundary. A transmitted wave through layer 2 is termed a diving wave. A reflected wave enters with the same angle of incidence as exit angle. If the critical angle is achieved, the critically refracted head wave travels along the layer 1-layer 2 interface. Refracted waves propagate from the interface, with exit angles equal to the critical angle. Arrival times can be represented on a travel-time graph or T-X plot, that is P-wave arrival times usually in milliseconds verses distance geophone location. What are we trying to calculate? Of interest are velocities of P-wave propagation through layers 1 and 2, and also thickness of layer 1. To obtain these values, a healthy combination of equations and interpretation from the T-X plot is required. Analysis of the direct wave yields V1. On the profile view, notice that the wave arrives at a geophone located a known horizontal distance from the source. Thus, V1 should equal geophone-source distance divided by P-wave arrival time for a given geophone. On the T-X plot, the direct wave is represented by an interpolated line for arrival time data passing through the origin. The slope of this interpolated line is time over distance, or the inverse of velocity. The slope of lines on the T-X plot is termed slowness. Another interpolated line can be observed on the T-X plot, a line representing the refracted wave. The distance between the source and first geophone to receive the refracted wave is termed critical distance. Cross-over distance is defined as the position where the refracted wave overtakes the direct wave. A common analogy to the cross-over phenomenon are the travels of a cyclist and motorist. The cyclist decides to pedal along the bike path situated on the bay shore. With regards to distance to the Marin Headlands, the bike path distance is much less than highway distance. The cyclist can pedal at a constant rate of 15 miles per hour. Think of the cyclist as a direct wave. Meanwhile, the motorist e. Eventually the motorist finds the on ramp onto Highway and heads north for the Golden Gate Bridge at a brisk 55 mph. Since the motorist is traveling at a significant speed, the cyclist can only wave to the motorist as the car speeds past the cyclist on the bridge. The motorist must wait several tens of minutes at the outcrop of Franciscan melange before the cyclist arrives at the designated field trip meeting place. In this example, the cross-over distance occurred at the southern end of the Golden Gate Bridge. The speed of the cyclist 15 mph represents the P-wave velocity of layer 1. Highway speed for the motorist 55 mph would represent the P-wave velocity for layer 2. Back to the T-X plot To determine depth of layer 1 Z1 , the time intercept ti of the refracted wave must be noted. Two Layer Dipping Model When discussing dipping layers, one wants to quantify the amount of dip. For a simple case of two dipping layers, seismic refraction can be utilized to calculated dip of the layers. For a given survey profile, sources must be located at the beginning of the profile forward shot and at the end of the profile reverse shot. P-wave arrival times for both forward and reverse shots can be plotted on a T-X plot. From the Principle of Reciprocity, time required for a ray to travel along the forward and reverse shot should be the same, since the ray pathways are the same. From the T-X plot, V1 and V2 velocities for forward and reverse shots can be calculated, as well as the time-intercepts for forward and reverse refracted waves. Small delta represents the dip of layer-1 layer-2 boundary. Why only stop with interpretation of two horizontal layers? Calculation of layer velocities and thicknesses for multi-layers requires patience with many equations chock full of algebra and trigonometry. Interpretation of T-X plots remains the same. Problems and Limitations The preceding models assume planar boundary interfaces. Conformable sequences of sedimentary rock may form planar boundaries. However, erosion and uplift easily produce irregular boundary contacts. More sophisticated algorithms can process refraction surveys where irregular interfaces might be expected. Profile length and source energy limit the depth penetration of the refraction method. Typically, a profile can only detect features at a depth of one-fifth survey length. Thus, refraction imaging of the Moho would require profile lengths of over one hundred kilometers; an unreasonable experiment. Larger sources could be utilized for greater depth detection, but certain sources e. Refraction depends on layers to increase in velocity with depth. In the hidden slow layer senario, a buried layer is overlain by a faster layer. No critical refraction will occur along the boundary interface. Thus, refraction will not detect the slow layer. All is not lost since reflection seismology could detect the slower layer. Seismograms require careful analysis to pick first arrival times for layers. If a thin layer produces first arrivals which cannot easily be identified on a seismogram, the layer may never be identified. Thus, another layer may be misinterpreted as incorporating the hidden layer. As a result, layer thicknesses may increase. References Cited These sources offer excellent discussion of theory, derivation of formulas, and practical examples of refraction seismology. Exploration Geophysics of the Shallow Subsurface. An Introduction to Geophysical Exploration.

7: Seismic Refraction Surveying

Seismic reflection amplitude is proportional to the from velocities and densities logged in the adjacent wells This process allows to tie the stratigraphic.

The method requires a controlled seismic source of energy, such as dynamite, a specialized air gun or a seismic vibrator, commonly known by the trademark name Vibroseis. By noting the time it takes for a reflection to arrive at a receiver, it is possible to estimate the depth of the feature that generated the reflection. What is Seismic Reflection? The methods depend on the fact that seismic waves have differing velocities in different types of soil or rock: The methods enable the general soil types and the approximate depth to strata boundaries, or to bedrock, to be determined. The refraction microtremor method combines the urban utility and ease of microtremor array techniques with the operational simplicity of the SASW technique, and the shallow accuracy of the MASW technique. By recording urban microtremor on a linear array of a large number of lightweight seismometers, the method achieves fast and easy field data collection without any need for the time-consuming heavy source required for SASW and MASW work. By retaining all the original seismograms and by applying a time-domain velocity analysis technique as is done in MASW, the analysis described here can separate Rayleigh waves from body waves, air waves, and other coherent noise. Transforming the time-domain velocity results into the frequency domain allows combination of many arrivals over a long time period, and yields easy recognition of dispersive surface waves. The Seismoelectric method also called the Electroseismic method or seismo-electric is based on the generation of electromagnetic fields in soils and rocks by seismic waves. Although the method is not reported to detect groundwater flow, it does measure the hydraulic conductivity, which is related to permeability and, therefore, to the potential for groundwater flow. Seismic Refraction The seismic refraction method is based on the measurement of the travel time of seismic waves refracted at the interfaces between subsurface layers of different velocity. For shallow applications this normally comprises a hammer and plate, weight drop or small explosive charge blank shotgun cartridge. Energy radiates out from the shot point, either travelling directly through the upper layer direct arrivals , or travelling down to and then laterally along higher velocity layers refracted arrivals before returning to the surface. This energy is detected on surface using a linear array or spread of geophones spaced at regular intervals. Beyond a certain distance from the shot point, known as the cross-over distance, the refracted signal is observed as a first-arrival signal at the geophones arriving before the direct arrival. Observation of the travel-times of the direct and refracted signals provides information on the depth profile of the refractor. Shots are deployed at and beyond both ends of the geophone spread in order to acquire refracted energy as first arrivals at each geophone position. Data are recorded on a seismograph and later downloaded to computer for analysis of the first-arrival times to the geophones from each shot position. Travel-time versus distance graphs are then constructed and velocities calculated for the overburden and refractor layers through analysis of the direct arrival and T-minus graph gradients. Depth profiles for each refractor are produced by an analytical procedure based on consideration of shot and receiver geometry and the measured travel-times and calculated velocities. The final output comprises a depth profile of the refractor layers and a velocity model of the subsurface. The primary applications of seismic refraction are for determining depth to bedrock and bedrock structure. Due to the dependence of seismic velocity on the elasticity and density of the material through which the energy is passing, seismic refraction surveys provide a measure of material strengths and can consequently be used as an aid in assessing rippability and rock quality. The technique has been successfully applied to mapping depth to base of backfilled quarries, depth of landfills, thickness of overburden and the topography of groundwater.

8: Seismic Reflection and Refraction surveys.

Seismic reflection uses field equipment similar to seismic refraction, but field and data processing procedures are employed to maximize the energy reflected along near vertical ray paths by subsurface density contrasts (see Seismic Refraction Geometry below).

A series of apparently related reflections on several seismograms is often referred to as a reflection event. By correlating reflection events, a seismologist can create an estimated cross-section of the geologic structure that generated the reflections. Interpretation of large surveys is usually performed with programs using high-end three-dimensional computer graphics. Sources of noise[edit] Sources of noise on a seismic record. In addition to reflections off interfaces within the subsuface, there are a number of other seismic responses detected by receivers and are either unwanted or unneeded: Air wave[edit] The airwave travels directly from the source to the receiver and is an example of coherent noise. Low velocity, low frequency and high amplitude Rayleigh waves are frequently present on a seismic record and can obscure signal, degrading overall data quality. This motion causes a disturbance in the upper medium that is detected on the surface. Multiple reflection[edit] An event on the seismic record that has incurred more than one reflection is called a multiple. Multiples can be either short-path peg-leg or long-path, depending upon whether they interfere with primary reflections or not. Cultural noise[edit] Cultural noise includes noise from weather effects, planes, helicopters, electrical pylons, and ships in the case of marine surveys , all of which can be detected by the receivers. Applications[edit] Reflection seismology is used extensively in a number of fields and its applications can be categorised into three groups, [14] each defined by their depth of investigation: This can be combined with seismic attribute analysis and other exploration geophysics tools and used to help geologists build a geological model of the area of interest. A method similar to reflection seismology which uses electromagnetic instead of elastic waves, and has a smaller depth of penetration, is known as Ground-penetrating radar or GPR. Hydrocarbon exploration[edit] Reflection seismology, more commonly referred to as "seismic reflection" or abbreviated to "seismic" within the hydrocarbon industry, is used by petroleum geologists and geophysicists to map and interpret potential petroleum reservoirs. The size and scale of seismic surveys has increased alongside the significant concurrent increases in computer power during the last 25 years. This has led the seismic industry from laboriously â€” and therefore rarely â€” acquiring small 3D surveys in the s to now routinely acquiring large-scale high resolution 3D surveys. The goals and basic principles have remained the same, but the methods have slightly changed over the years. The primary environments for seismic exploration are land, the transition zone and marine: Land â€” The land environment covers almost every type of terrain that exists on Earth, each bringing its own logistical problems. Examples of this environment are jungle, desert, arctic tundra, forest, urban settings, mountain regions and savannah. Transition Zone TZ â€” The transition zone is considered to be the area where the land meets the sea, presenting unique challenges because the water is too shallow for large seismic vessels but too deep for the use of traditional methods of acquisition on land. Examples of this environment are river deltas, swamps and marshes, [19] coral reefs, beach tidal areas and the surf zone. Transition zone seismic crews will often work on land, in the transition zone and in the shallow water marine environment on a single project in order to obtain a complete map of the subsurface. Diagram of equipment used for marine seismic surveys Marine â€” The marine zone is either in shallow water areas water depths of less than 30 to 40 metres would normally be considered shallow water areas for 3D marine seismic operations or in the deep water areas normally associated with the seas and oceans such as the Gulf of Mexico. Seismic surveys are typically designed by National oil companies and International oil companies who hire service companies such as CGG , Petroleum Geo-Services and WesternGeco to acquire them. Another company is then hired to process the data, although this can often be the same company that acquired the survey. Finally the finished seismic volume is delivered to the oil company so that it can be geologically interpreted. Land survey acquisition[edit] Desert land seismic camp Receiver line on a desert land crew with recorder truck Land seismic surveys tend to be large entities, requiring hundreds of tons of equipment and employing anywhere from a few hundred to a few thousand people, deployed over vast areas for many months. Vibroseis is a non-impulsive source that is cheap and efficient but requires flat ground to operate on, making its use more difficult in undeveloped areas. The method comprises one or more heavy, all-terrain vehicles lowering a steel plate onto the ground, which is then vibrated with a specific frequency distribution and amplitude. For a long time, it was the only seismic source available until weight dropping was introduced around , [23] allowing geophysicists to make a trade-off between image quality and environmental damage. Compared to Vibroseis, dynamite is also operationally inefficient because each source point needs to be drilled and the dynamite placed in the hole. A land seismic survey requires substantial logistical support. In addition to the day-to-day seismic operation itself, there must also be support for the main camp for catering, waste management and laundry etc. Unlike in marine seismic surveys, land geometries are not limited to narrow paths of acquisition, meaning that a wide range of offsets and azimuths is usually acquired and the largest challenge is increasing the rate of acquisition. The rate of production is obviously controlled by how fast the source Vibroseis in this case can be fired and then move on to the next source location. Attempts have been made to use multiple seismic sources at the same time in order to increase survey efficiency and a successful example of this technique is Independent Simultaneous Sweeping ISS. The cables are known as streamers, with 2D surveys using only 1 streamer and 3D surveys employing up to 12 or more though 6 or 8 is more common. The streamers are deployed just beneath the surface of the water and are at a set distance away from the vessel. The seismic source, usually an airgun or an array of airguns but other sources are available, is also deployed beneath the water surface and is located between the vessel and the first receiver. Two identical sources are often used to achieve a faster rate of shooting. By the early s, it was accepted that this type of acquisition was useful for initial exploration but inadequate for development and production, [26] in which wells had to be accurately positioned. The seismic properties of salt poses an additional problem for marine seismic surveys, it attenuates seismic waves and its structure contains overhangs that are difficult to image. This configuration was "tiled" 4 times, with the receiver vessel moving further away from the source vessels each time and eventually creating the effect of a survey with 4 times the number of streamers. The end result was a seismic dataset with a larger range of wider azimuths, delivering a breakthrough in seismic imaging. Marine survey acquisition ocean bottom seismic OBS [edit] Marine survey acquisition is not just limited to seismic vessels; it is also possible to lay cables of geophones and hydrophones on the sea bed in a similar way to how cables are used in a land seismic survey, and use a separate source vessel. This method was originally developed out of operational necessity in order to enable seismic surveys to be conducted in areas with obstructions, such as production platforms , without having the compromise the resultant image quality. Conventional OBC surveys use dual-component receivers, combining a pressure sensor hydrophone and a vertical particle velocity sensor vertical geophone , but more recent developments have expanded the method to use four-component sensors i. Four-component sensors have the advantage of being able to also record shear waves , [30] which do not travel through water but can still contain valuable information. In addition to the operational advantages, OBC also has geophysical advantages over a conventional NATS survey that arise from the increased fold and wider range of azimuths associated with the survey geometry. Time lapse acquisition 4D [edit] Time lapse or 4D surveys are 3D seismic surveys repeated after a period of time. The 4D refers to the fourth dimension which in this case is time. Time lapse surveys are acquired in order to observe reservoir changes during production and identify areas where there are barriers to flow that may not be detectable in conventional seismic. Time lapse surveys consist out of a baseline survey and a monitor or repeat survey, acquired after the field was under production. Some of these surveys are collected using ocean-bottom cables because the cables can be accurately placed in their previous location after being removed. Better repetition of the exact source and receiver location leads to improved repeatability and better signal to noise ratios. A number of 4D surveys have also been set up over fields in which ocean bottom cables have been purchased and permanently deployed. Deconvolution , Seismic migration , and Multidimensional seismic data processing There are three main processes in seismic data processing: Deconvolution operations can be cascaded, with each individual deconvolution designed to remove a particular type of distortion. CMP stacking is a robust process that uses the fact that a particular location in the subsurface will have been sampled numerous times and at different offsets. This allows a geophysicist to construct a group of traces with a range of offsets that all sample the same subsurface location, known as a Common Midpoint Gather. Less significant processes that are applied shortly before the CMP stack are Normal moveout correction and statics correction. Unlike marine seismic data, land seismic data has to be corrected for the elevation differences between the shot and receiver locations. This correction is in the form of a vertical time shift to a flat datum and is known as a statics correction, but will need further correcting later in the processing sequence because the velocity of the near-surface is not accurately known. This further correction is known as a residual statics correction. Seismic migration is the process by which seismic events are geometrically re-located in either space or time to the location the event occurred in the subsurface rather than the location that it was recorded at the surface, thereby creating a more accurate image of the subsurface. Geologic modelling The goal of seismic interpretation is to obtain a coherent geological story from the map of processed seismic reflections. The aim of this is to produce structural maps that reflect the spatial variation in depth of certain geological layers. Using these maps hydrocarbon traps can be identified and models of the subsurface can be created that allow volume calculations to be made. However, a seismic dataset rarely gives a picture clear enough to do this. This is mainly because of the vertical and horizontal seismic resolution [39] but often noise and processing difficulties also result in a lower quality picture. Due to this, there is always a degree of uncertainty in a seismic interpretation and a particular dataset could have more than one solution that fits the data. In such a case, more data will be needed to constrain the solution, for example in the form of further seismic acquisition, borehole logging or gravity and magnetic survey data. Similarly to the mentality of a seismic processor, a seismic interpreter is generally encouraged to be optimistic in order encourage further work rather than the abandonment of the survey area. In hydrocarbon exploration, the features that the interpreter is particularly trying to delineate are the parts that make up a petroleum reservoir â€” the source rock , the reservoir rock, the seal and trap. Seismic attribute analysis[edit] See also: Seismic attribute Seismic attribute analysis involves extracting or deriving a quantity from seismic data that can be analysed in order to enhance information that might be more subtle in a traditional seismic image, leading to a better geological or geophysical interpretation of the data. Attributes that can show the presence of hydrocarbons are called direct hydrocarbon indicators. It became clear that there was a lack of understanding of the tectonic processes that had formed the geological structures and sedimentary basins which were being explored. The effort produced some significant results and showed that it is possible to profile features such as thrust faults that penetrate through the crust to the upper mantle with marine seismic surveys. Land[edit] On land, conducting a seismic survey may require the building of roads , for transporting equipment and personnel, and vegetation may need to be cleared for the deployment of equipment. If the survey is in a relatively undeveloped area, significant habitat disturbance may occur and many governments require seismic companies to follow strict rules regarding destruction of the environment; for example, the use of dynamite as a seismic source may be disallowed. Seismic processing techniques allow for seismic lines to deviate around natural obstacles, or use pre-existing non-straight tracks and trails. With careful planning, this can greatly reduce the environmental impact of a land seismic survey. The more recent use of inertial navigation instruments for land survey instead of theodolites decreased the impact of seismic by allowing the winding of survey lines between trees. Marine[edit] The main environmental concern for marine seismic surveys is the potential for noise associated with the high-energy seismic source to disturb or injure animal life, especially cetaceans such as whales , porpoises , and dolphins , as these mammals use sound as their primary method of communication with one another. Conversely, the study found that male humpback whales were attracted to a single operating airgun as they were believed to have confused the low-frequency sound with that of whale breaching behaviour. In addition to whales, sea turtles , fish and squid all showed alarm and avoidance behaviour in the presence of an approaching seismic source. It is difficult to compare reports on the effects of seismic survey noise on marine life because methods and units are often inadequately documented. It is circumstantial evidence such as this that has led researchers to believe that avoidance and panic might be responsible for increased whale beachings although research is ongoing into these questions. Furthermore, the specific characteristics of seismic sounds and the operational procedures employed during seismic surveys are such that the resulting risks to marine mammals are expected to be exceptionally low.

VELOCITIES IN REFLECTION SEISMOLOGY pdf

9: Seismic Reflection Methods | Environmental Geophysics | US EPA

GEOS Spring 1 Lab 5: Reflection Seismology I (35 points) Simple models, Reflection coefficients, NMO, Dipping layers I This lab will help you think critically about what seismic reflection data represents.

Body waves travel through the interior of the Earth. Surface waves travel across the surface. Surface waves decay more slowly with distance than body waves, which travel in three dimensions. Particle motion of surface waves is larger than that of body waves, so surface waves tend to cause more damage. Body waves[edit] Body waves travel through the interior of the Earth along paths controlled by the material properties in terms of density and modulus stiffness. The density and modulus, in turn, vary according to temperature, composition, and material phase. This effect resembles the refraction of light waves. Two types of particle motion result in two types of body waves: Primary and Secondary waves. P-wave Primary waves P-waves are compressional waves that are longitudinal in nature. P waves are pressure waves that travel faster than other waves through the earth to arrive at seismograph stations first, hence the name "Primary". These waves can travel through any type of material, including fluids, and can travel nearly 1. In air, they take the form of sound waves, hence they travel at the speed of sound. S-wave Secondary waves S-waves are shear waves that are transverse in nature. Following an earthquake event, S-waves arrive at seismograph stations after the faster-moving P-waves and displace the ground perpendicular to the direction of propagation. Depending on the propagational direction, the wave can take on different surface characteristics; for example, in the case of horizontally polarized S waves, the ground moves alternately to one side and then the other. S-waves can travel only through solids, as fluids liquids and gases do not support shear stresses. They can be classified as a form of mechanical surface waves. They are called surface waves, as they diminish as they get further from the surface. They travel more slowly than seismic body waves P and S. In large earthquakes, surface waves can have an amplitude of several centimeters. Rayleigh wave Rayleigh waves, also called ground roll, are surface waves that travel as ripples with motions that are similar to those of waves on the surface of water note, however, that the associated particle motion at shallow depths is retrograde, and that the restoring force in Rayleigh and in other seismic waves is elastic, not gravitational as for water waves. In a layered medium like the crust and upper mantle the velocity of the Rayleigh waves depends on their frequency and wavelength. See also Lamb waves. Love wave Love waves are horizontally polarized shear waves SH waves , existing only in the presence of a semi-infinite medium overlain by an upper layer of finite thickness. Love , a British mathematician who created a mathematical model of the waves in Stoneley wave A Stoneley wave is a type of boundary wave or interface wave that propagates along a solid-fluid boundary or, under specific conditions, also along a solid-solid boundary. Amplitudes of Stoneley waves have their maximum values at the boundary between the two contacting media and decay exponentially towards the depth of each of them. These waves can be generated along the walls of a fluid-filled borehole , being an important source of coherent noise in VSPs and making up the low frequency component of the source in sonic logging. The scheme of motion for spheroidal 0S2 oscillation. Dashed lines give nodal zero lines. Arrows give the sense of motion. Free oscillations of the Earth are standing waves , the result of interference between two surface waves traveling in opposite directions. Interference of Rayleigh waves results in spheroidal oscillation S while interference of Love waves gives toroidal oscillation T. The modes of oscillations are specified by three numbers, e. The number m is the azimuthal order number. The number n is the radial order number. It means the wave with n zero crossings in radius. For spherically symmetric Earth the period for given n and l does not depend on m. Some examples of spheroidal oscillations are the "breathing" mode 0S0, which involves an expansion and contraction of the whole Earth, and has a period of about 20 minutes; and the "rugby" mode 0S2, which involves expansions along two alternating directions, and has a period of about 54 minutes. The mode 0S1 does not exist because it would require a change in the center of gravity, which would require an external force. The mode 0T2 describes a twisting of the northern and southern hemispheres relative to each other; it has a period of about 44 minutes. Presently periods of thousands of modes are known. These data are used for determining some large scale structures of the Earth interior. Since shear waves cannot pass through liquids, this phenomenon was original evidence for the now well-established observation that the Earth has a liquid outer core , as demonstrated by Richard Dixon Oldham. This kind of observation has also been used to argue, by seismic testing, that the Moon has a solid core, although recent geodetic studies suggest the core is still molten[citation needed]. Notation[edit] Earthquake wave paths The path that a wave takes between the focus and the observation point is often drawn as a ray diagram. An example of this is shown in a figure above. When reflections are taken into account there are an infinite number of paths that a wave can take. Each path is denoted by a set of letters that describe the trajectory and phase through the Earth. In general an upper case denotes a transmitted wave and a lower case denotes a reflected wave. The two exceptions to this seem to be "g" and "n".

VELOCITIES IN REFLECTION SEISMOLOGY pdf

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