A SOURCE STUDY OF THE AEGEAN EARTHQUAKE OF MARCH 27, 1975, FROM INVERSION OF TELESEISMIC BODY WAVES

Ioannidou E., Stavrakakis G., Drakopoulos J.

National Observatory of Athens, Seismological Institute

The source mechanism (douple-couple orientation, moment, centroid depth, source time function) of the Aegean earthquake of March 27, 1975 has been determined from an inversion of teleseismic long-period P and SH waves. The earthquake is modelled by two subevents of different focal mechanisms and seismic moments. The first one is characterized be nearly pure normal faulting on a fault plane with strike 41° and dip 46°. The second subevent is characterized by strike slip faulting on a fault plane striking 238° with dip 86°.

INTRODUCTION

Modeling of body waves either in frequency or time domain is a powerful method and provides important information on the details of the source rupture process. In time-domain analysis of body waves, the observed waveforms are modeled by a source time function, and the time constants associated with it are interpreted in terms of the source dimension and the particle velocity of the fault motion.

When the observed body waves are relatively simple, the modeling can be done using either forward or inverse methods. Langston (1976) used a time-domain inversion method to determine complexities of the source time function.

Kikuchi and Kanamori (1982) proposed an inversion technique to obtain information about asperities and barriers on the fault plane. Using this inversion procedure, the rupture process of some large earthquakes occurred in the area of Greece has been determined (Stavrakakis et al.,1986a; Stavrakakis et al., 1986b; Stavrakakis et al., 1989).

Nabelek (1984) developed an inversion technique adequate for analysing moderate earthquakes. In this modeling, the douple-couple earthquake source is parameterized by the strike, dip, and rake angles of one of its P-wave nodal planes, the centroid depth, and the amplitudes of a specified number of overlapping isosceles triangles that represent the source time function. This procedure has been applied in the present study to obtain information about the source process of the Aegean earthquake of March 27, 1975.

INVERSION AND MODELING OF TELESEISMIC BODY WAVES

(1) Data Preparation

The data set used in the present study consists of the long-period P and SH wave seismograms recorded by the WWSSN. We selected P and SH wavefroms from all available azimuths. To minimize complications due to upper mantle structure and core interaction, we limited the epicentral distance range from 30° to 90° for P waves, and 30° to 70° for the SH waves.

The seismograms were digitized at unequal time interval and interpolated at 0.5 s. The amplitudes were normalized to a peak magnification of 1500 and epicentral distance 40°. Because, generally, the SH waves as a whole have a different signal power from the P waves, the SH waves were given a weight which is proportional to the ratio of the rms amplitude of the P wave data set to the rms amplitude of the SH wave data set. Furthermore, stations with similar azimuths and takeoff angles were grouped together and given weights inversely proportional to the square root of the number of the stations in each group. In this manner, both P and SH waves and all sectors of the focal sphere are assigned equal importance.

(2) Inversion Strategy

In the present study several results of inversion are discussed. First, the fault plane solution proposed by McKenzie (1978) used to analyse the observed seismograms with a relatively simple model. Since the wave matching was poor, we introduce two complex models in order to improve the solution.

(3) Cource Parameterization

For the earthquake which is analyzed in the present study, we determined the best fitting douple-couple point source from an inversion of long-period P and SH waveforms. The inversion technique proposed by Nabelek (1984, 1985) as applied by several researchers (Bergman et al., 1984; Bergman and Solomon, 1985; McCaffrey and Nabelek, 1986; Huang et al., 1986; Nelson et al., 1987; Nabelek et al., 1987; Pacheco and Nabelek,1988; Boyd and Nabelek, 1988) is employed. The inversion yields centroid depth and source time function as well as douple-couple orientation and seismic moment. The orientation of the best fitting douple couple is determined to within about +5° formal error in each angular coordinate, and the moment to within about +30% at the 2σ (95% confidence) level. The formal error (2σ) in centroid depth is typically +0.1 to 0.3 km but because of bias introduced by factors not reflected in the estimated of a posterior parameter variances, such as the alignement in the time between the synthetic and the observed waveforms, the true uncertainty may be greater, up to +2 km in some cases.

The far-field point-source time function Q(t) is parameterized by a series of overlapping isosceles triangle functions T_{k} of duration $2\Delta\tau$ and adjustable ralative amplitudes w_k as defined by Nabelek (1984):

$$Q(t) = \Sigma w_k T_{k}(t-\tau_k) \tau_k = (k-1)\Delta\tau \quad (1)$$

where N_{*} is the number of the time function elements. It should be mentioned that N_{*} are described a priori and w_k are determined by the inversion. This parameterization is equivalent to an approximation of the source time function by the trapezoidal rule.

The triangle function $T_{\Delta\tau}$ can be also written as a convolution of two box-car functions $B_{\Delta\tau}(t)$ of equal duration $\Delta\tau$,

$T_{\Delta\tau}(t-\tau_k) = B_{\Delta\tau}(t-\tau_k) * B_{\Delta\tau}(t)$ (2)

The effect of a propagating horizontal line source is achieved by varying the duration of one of the boxcars according to the relationship

$\Delta\tau' = \Delta\tau[1-u_r p\cos(f-\theta)$ (3)

where p is the ray parameter, u_r is the rupture velocity, φ is the station azimuth, and θ is the fault strike. $\Delta\tau$ is choosen a priori as in the case of the point source.

One of the most important source parameters determined by the inversion procedure is the centroid depth. Therefore, the source resolution deserves particular attention. The inversion parameters of the best fitting point source are those which minimize the mean square residual r^2, defined as

$r^2 = \Sigma\Sigma w_j(s_{ij}-o_{ij})^2/\Sigma M_j$ (4)

where s_{ij} and o_{ij} are the amplitudes of the synthetic and observed seismograms at the j^{th} station and time sample i, w_j is the weight used for the station j, M_j is the number of samples in the time window used in the inversion for station j, and N is the total number of stations.

To investigate the uncertainty in the depth of the crustal level, a test of significance is conducted for the differences in station residuals between waveforms of the best fitting depth and those obtained when the inversion is performed with the depth fixed at nearby values. The mean square residual for the j^{th} station is

$r^2_j = (1/M_j)\Sigma(s_{ij}-o_{ij})^2$ (5)

The mean square residual is redefined as

$R^2 = (1/N)\Sigma r_j^2$ (6)

The quantities r^2 and R^2 would be equal only if all stations have the same inversion window length and weight.

In the present study, the distribution of individual station residuals r^2_{ij} is investigated with χ^2 test of goodness of fit and found that the residual can generally be considered to follow a normal distribution. Then, we tested the null hypothesis μ_D by forming the statistic

$t = (\mu_D)/\sigma_d/(N)^{1/2}$ (7)

which follows the t distribution with N-1 degrees of freedom. In the previous relationship μ_D and σ_D are the mean and standard deviation of the set of differenced

$d_j = r_{jA}^2 - r_{jB}2$, j= 1,2,...,N (8)

where r_{jA} and r_{jB} are the matched station residuals at two depths A and B. Since the null hypothesis is tested that the solution at one depth is better than the other, only the one-tailed test is used.

WAVEFORM INVERSION RESULTS

In this section, the source parameters obtained from the inversion of P and SH waveforms of the northern Aegean sea earthquake of March 27, 1975, are presented. Attenuation coefficients (t^*, s) of 1 and 4 s were assumed of P and SH waves, respectively. The calculated body-waves packets are the sum of direct P and S and reflections from pP, sS and sP. The source time function is convolved with the resulting time series of pulses to determine the final shape of the synthetic seismogram.

All inversions were performed with the same source velocity structure: a water layer of 0.5 km and a half-space with P ans S wave velocity α=6.5 km/sec and β=3.7 km/sec, respectively, and a density ρ=2.8 gr/cm^3.

First, inversion was performed for a single point source in which a source time function of 30 s and of 10 elements was allowed to be complicated but was constrained to be positive. The starting source mechanism (strike=41°, dip=60°, rake=235°) is that determined by McKenzie (1978). With this parameterization we obtained the solution illustrated in figures 1a and 1b for P and SH waves, respectively. The inferred source time function has a duration of 12.4 s. A seismic moment of 4.04x10^{18} Nm is obtained with the following T-statistics parameters:

meamn rms=1.35, mean d_j=1.82, S.D of d_j=2.47, t=3.89

The shape of the inferred source time function suggests the complexity in rupture propagation and the existence of subevents. It also indicates that the seismic moment is released at the later phase of the rupture. This might be a reason for the mismatch which is observed at all stations.

Next, we introdued a moving line source model with a rupture velocity of 2 km/sec and an azimuth of 40°N. The synthetic waveforms are shown in fugures 2a and 2b for P and SH waves, respectively with the inferred source time function. The matching of P waves is satisfactory good, however the matching of SH waves are still poor.

In order to improve the fit, we next introduce a strike-slip subevent commencing approximately 4 s from the origin time, with a distance offset of 10 km. Figures 3a and 3b show the obtained synthetic waveforms for P and SH waves, respectively. For the second strike-slip subevent the following source parameters have been obtained: strike=133°±0.73, dip=88°±0.7, rake=319°±0.7, centroid depth=15 km, seismic moment=1.92x10^{18} Nm. The inferred source time function has a duration of about 3.5 s. From its shape it is evident that the moment released at the later phase of the rupture propagation. However, the matching has been relatively improved.

Finaly, the crustal model has been changed by introducting a sediment layer of 1.5 km with P and S wave velocities α=5.2 km/sec and β=2.9 km/sec, respectively and a density ρ=2.2 gr/cm^3. For the strike-slip subevent we obtained: strike=238°±0.73, dip=86°±0.57, rake=182°±0.58, centroid depth=10 km+0.02. The duration of the inferred source time function is of about 5 s. The synthetic waveforms are compared with the observed ones in figures 4a and 4b, respectively. The matching is sufficient better corresponding to a rms = 1.32. This solution has been adopted.

DISCUSSION AND CONCLUSION

In the present study, the source process of the Aegean earthquake of March 27, 1975 has been investigated by inverting teleseismic long-period P and SH waves.

The striking feature of this earthquake is that it consists of two subevents of different focal parameters. The main component of faulting is normal, which induced a strike-slip faulting at shallower depth. For the normal subevent the following source parameters have been obtained: strike=41°±0.91, dip=46°±0.87, rake=232°±0.91, centroid depth=15 km +0.02, seismic moment 8.67×10^{17} Nm with a source time function of duration of about 4 s.

For the second strike-slip subevent which occurred at a depth of 10 km we obtained a strike of 238°±0.73, a dip of 86°±0.57, and a rake angle of 182°±0.58. The seismic moment amounts to 2.86×10^{18} Nm. The remarkable result is that the second subevent is larger than the first one.

The inferred source process of the Aegean sea earthquake of March 27, 1975, indicates that the deformation field in the investigated area is complex and cannot simply interpreted by pure extension field. However, we need to analyze more seismic events which occurred in the northern Aegean Sea to make a detailed interpretation of the stress field.

REFERENCES

BERGMAN, E.A., NABELEK, J.L. and SOLOMON, S.C., 1984. An extensive region off-ridge normal-faulting earthquakes in the southern Indian Ocean. J. Geophys. Res., 89:2425-2443.

BERGMAN, E.A. and SOLOMON, S.C. (1985). Earthquake source mechanisms from body-waveform inversion and intraplate tectonics in the northern Indian Ocean. Physics Earth and Planetary Interiors, 40, 1-23.

BOYD, T.M. and NABELEK J. (1988). Rupture process of the Andreanof Islands Earthquakes of May 7, 1986, Bull. Seism. Soc. Am., 78, 1653-1673.

HUANG, P.V., SOLOMON, S.C., BERGMAN, E.A. and NABELEK, J.L. (1986). Focal depths and mechanisms of mid-Atlantic ridge earthquakes from body waveform inversion, J. Geophys. Res., 91, 579-598.

KIKUCHI, M. and KANAMORI, H. (1982). Inversion of complex body waves. Bull. Seism. Soc. Am., 72, 491-506.

LANGSTON, C.A. (1976). A body wave inversion of the Koyna, India, earthquake of December 10, 1967, and some implications for the body wave focal mechanism, J. Geophys. Res., 81 2517-2529.

McCAFFREY, R. and NABELEK, J.(1986). Seismological evidence for shallow thrusting north of the Timor trough, Geophys. J.R.astr. Soc., 85, 365-381.

McKENZIE, D.P.(1978). Active tectonics of the

Alpine-Himalayan belt: The Aegean Sea and surrounding region, Geophys. J.R. astr. Soc., 55, 217-254.

NABELEK, J.(1984). Determination of earthquake source parameters from inversion of body waves, Ph. D. thesis, 361 pp, Mass. Inst. of Techn., Cambridge, 1984.

NABELEK, J.(1985). Geometry and mechanism of faulting of the 1980 El Asnam, Algeria, earthquake from inversion of teleseismic body waves and comparison with field observation, J. Geophys. Res., 91, 12713-12728.

NABELEK, J., WANG-PING CHEN, H.YE(1987). The Tangshan earthquake sequence and its implications for the evaluation of the North China Basin, J. Geophys. Res., 92, 12615-12628.

NELSON, M.R., McCAFFREY, R. and MOLNAR, P. (1987). Source parameters for 11 earthquakes in the Tien Shan, central Asia, determined by P and SH waveform inversion, J. Geophys. Res., 92, 12629-12648.

PACHECO, J. and NABELEK, J. (1988). Source mechanisms of three moderate California sequence earthquakes of July 1986, Bull. Seism. Soc. Am., 78, 1907-1929.

STAVRAKAKIS G., N., DRAKOPOULOS, J., and MAKROPOULOS, K. (1986a). A rupture model for the Corinth earthquake sequence of 1981, Earthquake Prognostics, Proc. from 2nd Intern. Seminar held in Berlin, June 24-27, 1986, 129-153.

STAVRAKAKIS, G.N., TSELENTIS, G-A., and DRAKOPOULOS J. (1986b). Iterative deconvolution of teleseismic P waves from Thessaloniki earthquake of June 20, 1978, Pageoph, 124, 1039-1050.

STAVRAKAKIS, G.N., IOANNIDOU, E., and DRAKOPOULOS J. (1989). A source model for the January 17, 1983 Cephalonia (W. Greece) earthquake from time domain modeling of teleseismic P waves. Bolletino Geoph. Teor. App.,Vol. XXXI,N.122, 149-157.

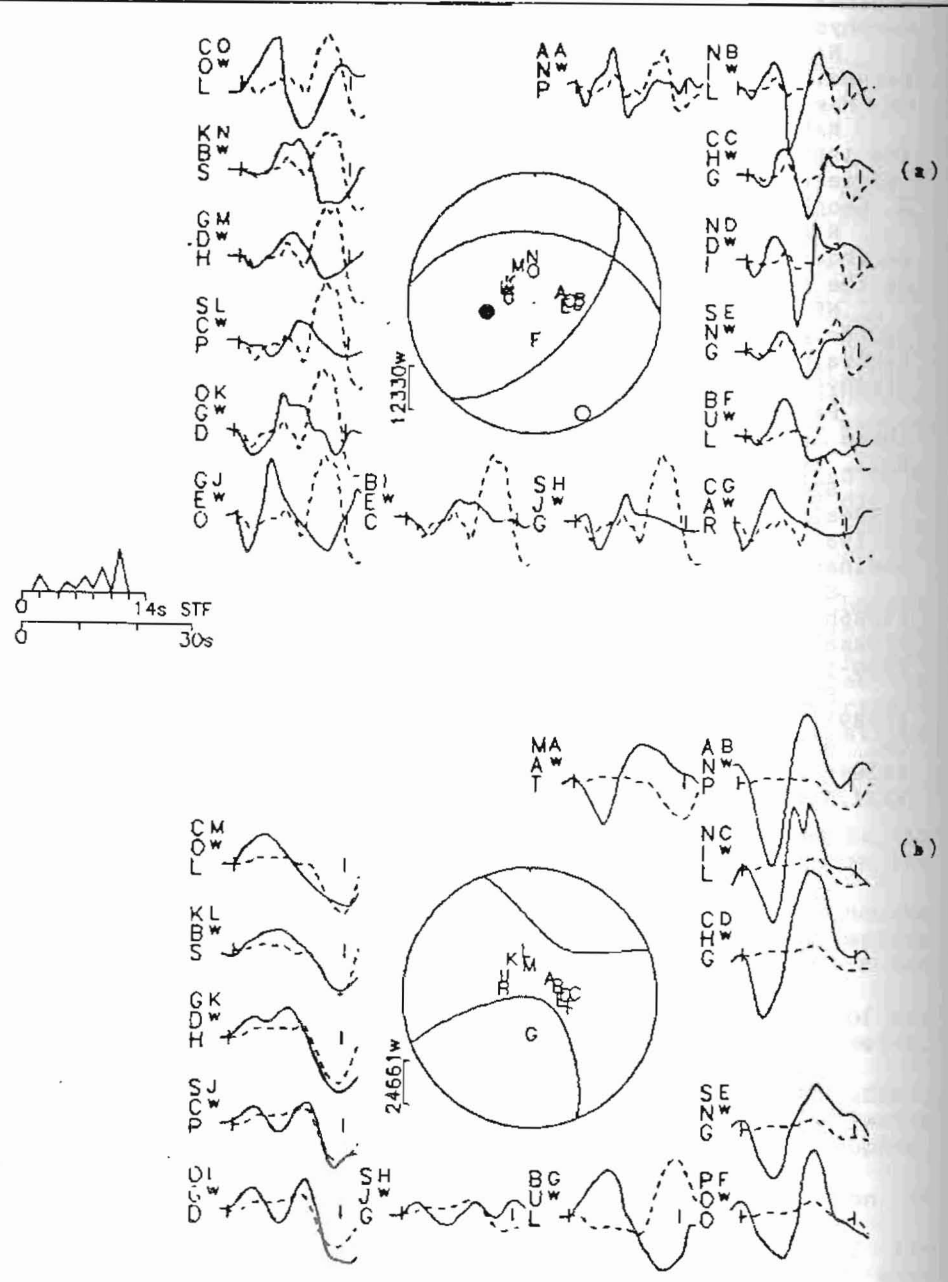


Figure 1a: Observed and Synthetic Waveforms of P waves.
Figure 1b: Observed and Synthetic Waveforms of SH waves.
(A simple model, McKenzie, 1978)

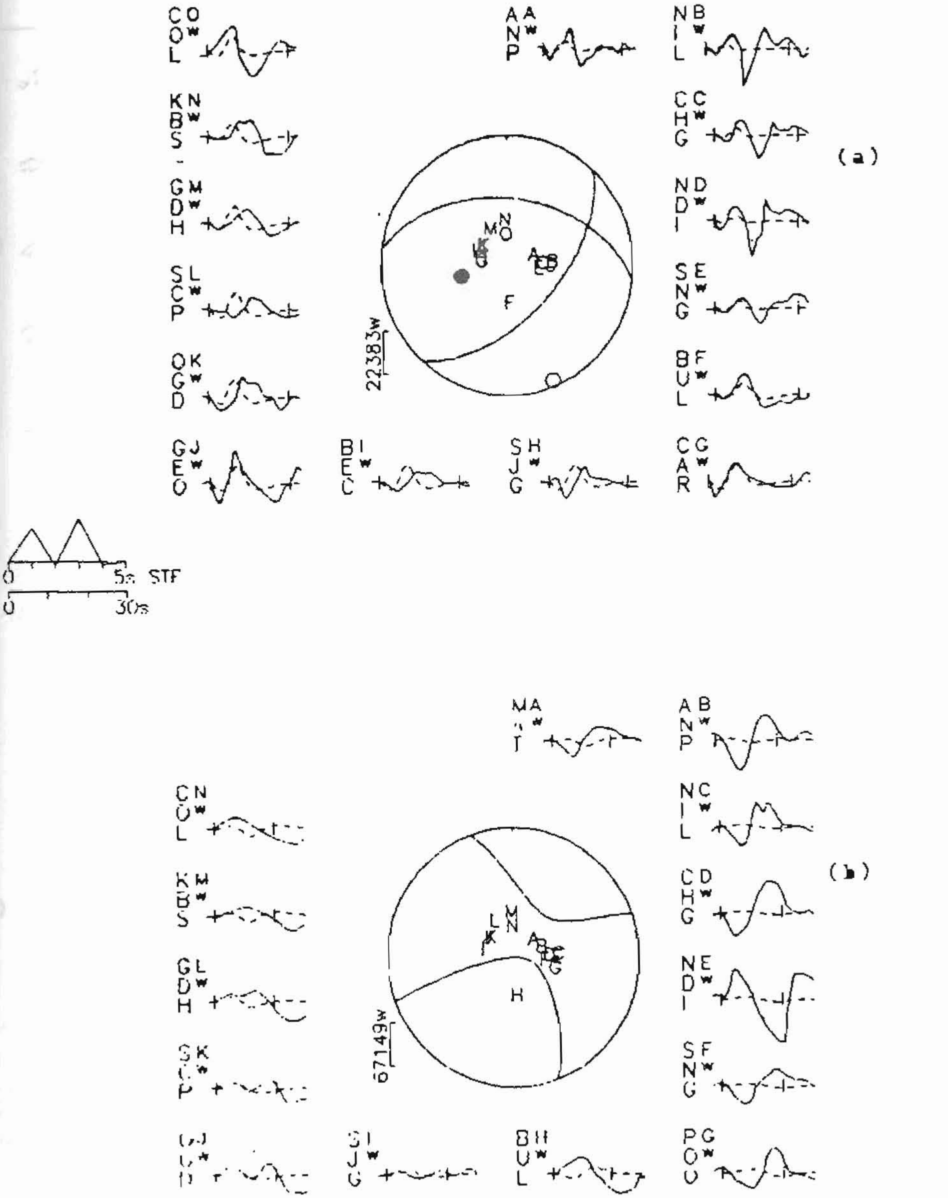


Figure 2a: Observed and Synthetic Waveforms of P waves.
Figure 2b: Observed and Synthetic waveforms fo SH waves.
(A moving line model)

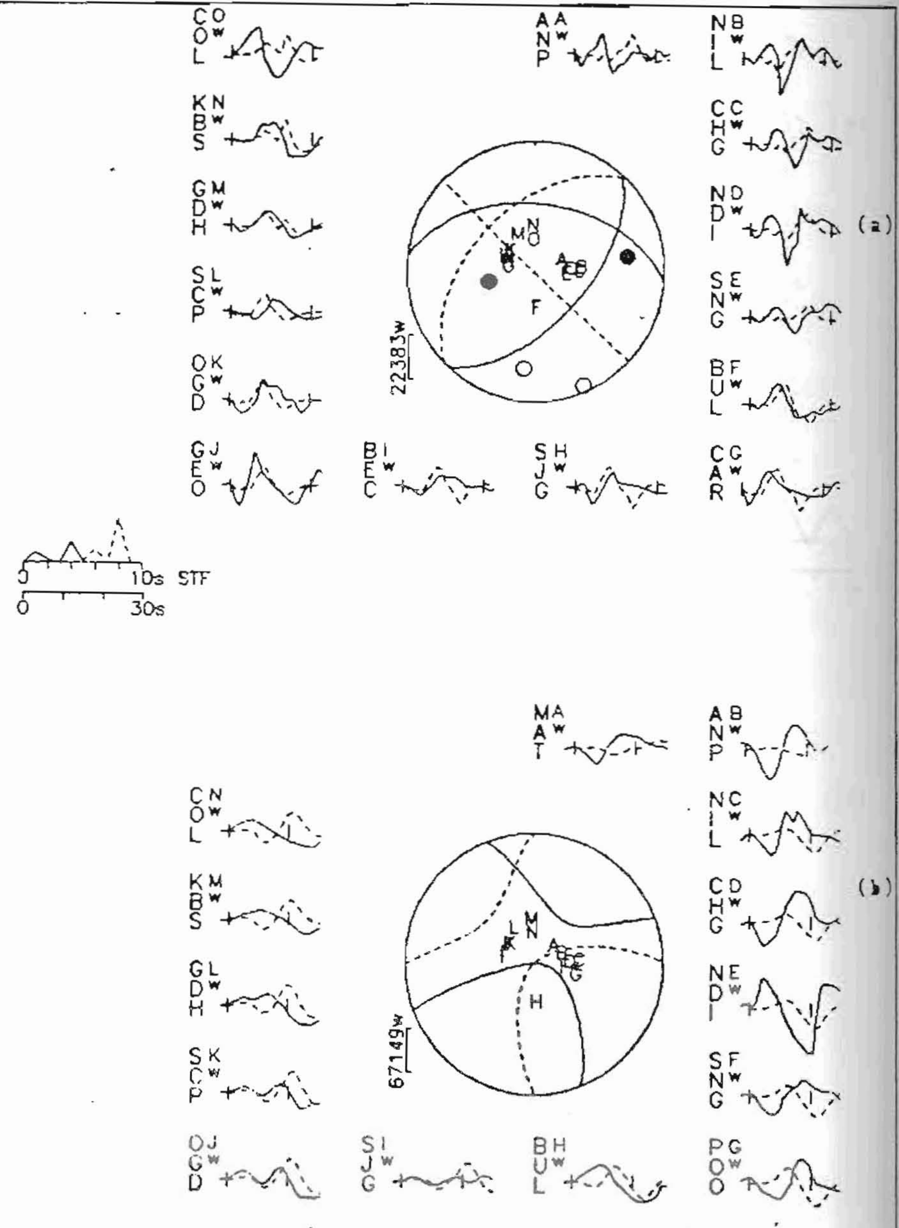
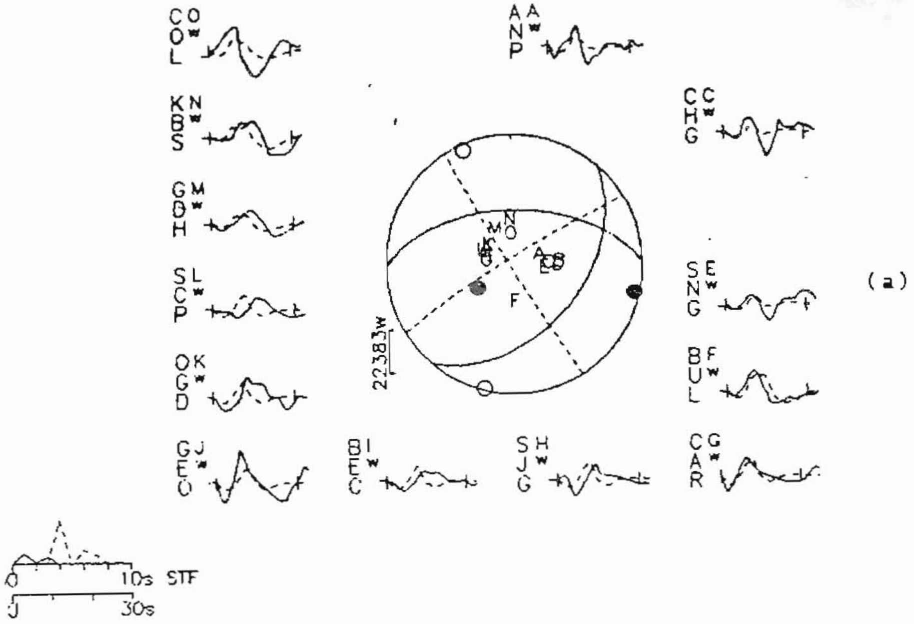


Figure 3a: Observed and Synthetic Waveforms of P waves.
Figure 3b: Observed and Synthetic Waveforms of SH waves.
(Two subevents with different focal mechanism)



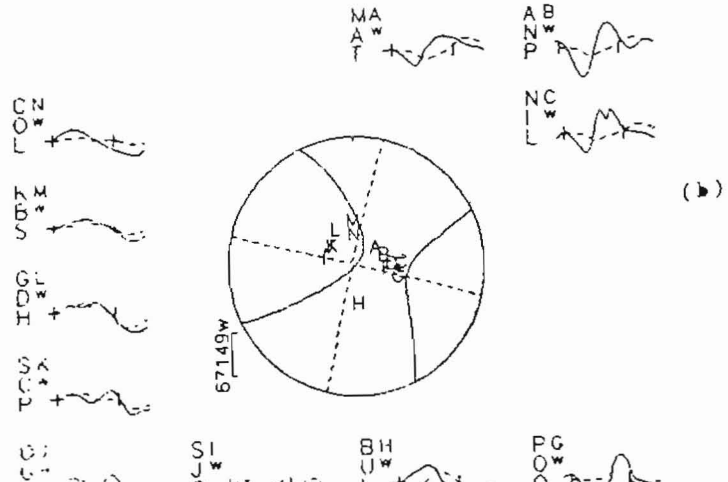


Figure 4a: Observed and Synthetic Waveforms of P waves.
Figure 4b: Observed and Synthetic Waveforms of Sh waves.
(Two subevents with different focal mechanism and with a sediment layer)