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Source Parameters of Some Large Earthquakes in Northern Aegean Determined by Body Waveform Inversion

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Abstract—Average source parameters for three large North Aegean events are obtained from body wave inversion for the moment tensor. The parameters for the events are as follows:

Date	M_s	$M_0 * 10^{25}$ dyn-cm	Strike°	Dip°	Rake°	Depth
Feb. 19, 1969	7.1	22.4	216	81	173	10 km
Dec. 19, 1981	7.2	22.4	47	77	−167	6 km
Aug. 6, 1983	6.8	12.1	50	76	177	9 km

The events exhibit dextral strike-slip faulting with the T axis striking NS and nearly horizontal, implying extension in this direction. The focal mechanisms obtained are in agreement with the seismotectonic regime of the North Aegean. It is known that the region is tectonically controlled by the existence of the strike-slip Anatolian fault and its westward continuation in the Aegean, as well as the NS extension the whole Aegean area undergoes.

The components of the moment tensor show that the region is dominated by compression in the EW direction which is encompassed by extension in the NS direction. All the events were found to be shallow (≤ 10 km) with a source time function of approximately 8 s duration and small stress drop values.

The teleseismic long period vertical P-waves exhibited distortions, that could be attributed to lateral inhomogeneities in the source structure or more probably to a nonflat water-crust interface.

Key words: Moment tensor inversion, North Aegean, Greece.

1. Introduction

The Aegean and surrounding region is an area of intense seismic activity. Focal mechanisms and Neogene-Quaternary geology indicate a very complex and rapidly varying tectonic picture.

The northernmost part of the Aegean area, the North Aegean trough, 1000–1500 m deep, is more than 500 km from the lithospheric boundary of the Hellenic

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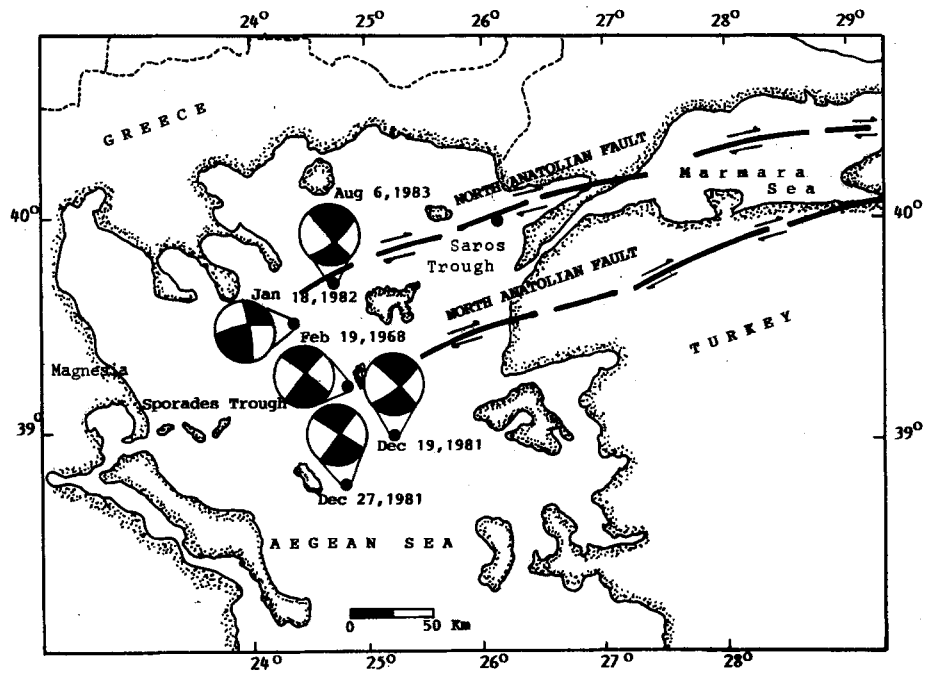

Figure 1
The North Aegean trough system with lower hemisphere projections of fault mechanisms determined in this study together with those of December 27, 1981 (KING and KIRATZI, 1985) and of January 18, 1982 (PAPAZACHOS *et al.*, 1984).

arc. This graben which extends from the Saros trough to the east as far as Magnesia to the west (Figure 1), takes up the dextral strike-slip motion of the North Anatolian fault at its western termination. The North Anatolian fault, which began in the middle to late Miocene time to absorb the relative motion between the Anatolian and Black Sea plates, consists mainly of two active strands, both which continue in the Aegean (PAPAZACHOS *et al.*, 1986).

The trough itself and the surrounding basins constitute the southern margin of the European plate which is less deformed than the Hellenic arc (MCKENZIE, 1972, 1978; LEPICHON and ANGELIER, 1979).

The area has suffered numerous earthquakes, the most recent started in December, 1981 with a series of strong events which ended in November, 1983. In this paper, moment tensor inversion analysis is performed for the following events: February 19, 1968 with $M_s = 7.1$ and epicenter 39.4°N, 24.9°E; December 19, 1981 with $M_s = 7.2$ and epicenter 39.1°N, 25.2°E; and August 6, 1983 with $M_s = 6.8$ and epicenter 40.1°N, 24.8°E. Focal mechanisms based on first motion data have already been published for these events by MCKENZIE (1972), PAPAZACHOS *et al.* (1984), and ROCCA *et al.* (1985). The need for reliable fault plane solutions, so as to understand the complicated tectonics of this area, motivated the present paper.

Moreover, the region is very important relative to seismic hazard since new activity is expected to burst there soon (KARACOSTAS *et al.*, 1986).

2. Theory and Data Processing

The moment tensor formalism of the seismic source inverse problem has received considerable attention due to its obvious simplicity and linearity (LANGSTON, 1981). The feasibility of the method is based upon the fact that an arbitrarily oriented point shear dislocation (i.e., having a body force equivalent to the double couple) can be represented by the summation of three Green's functions (LANGSTON and HELMBERGER, 1975). LANGSTON (1981) derived expressions to linearize the displacement equations, using the moment tensor formalism. Briefly, the inversion involves determining the parameter change matrix (ΔP) in the matrix equation $\mathbf{A}\,\Delta P = \Delta c$. $\mathbf{A}$ is the matrix of partial derivatives of the synthetics with respect to each parameter. The Δc matrix is the difference between the synthetic seismograms, (computed using some starting parameters) and the observed data. The $\mathbf{A}$ matrix is computed analytically using the equations for the synthetic seismograms and the Green's functions calculated for an assumed focal depth and source structure. Singular value decomposition is used to determine the inverse of the $\mathbf{A}$ matrix. For more details of the method refer to LANGSTON (1981), BARKER and LANGSTON (1982) and PAVLIN and LANGSTON (1983).

For the present work, microfiche records of both long- and short-period seismograms for the events studied were obtained from the WWSSN archives. Records were principally selected for inversion from stations in the range of 30° to 90° to avoid complexities in the waveforms caused by upper mantle and core-mantle boundary structures.

Table 1 lists the stations, from which the records were used in the inversion for each earthquake separately. As shown, the azimuthal coverage is sufficient in all cases. The long-period seismograms were digitized twice (top and bottom of the trace line) for better accuracy. The *S*-wave data were rotated into the respective station's back azimuth in order to extract long-period *SH*- and *SV*-waveforms. The use of *P*-, *SV*, and *SH*-waves, with their different radiation patterns allows the moment tensor elements to be better resolved. Moreover, effects of unmodelled structure near the source and along the ray path are decreased by inverting the *P*- and *SH*-, *SV*-seismograms from all stations simultaneously.

Green's functions for a point source with a step dislocation in a plane-layered medium were computed for different source depths, employing the ray theory technique outlined in LANGSTON and HELMBERGER (1975). The source region velocity structure listed in Table 2 was used (PANAGIOTOPOULOS and PAPAZACHOS, 1985) and the Green's functions were determined by summing rays for vertical strike-slip, vertical dip-slip and 45° dip-slip sources embedded at various

Table 1

WWSSN body wave data included in the inversions

Station	Δ°	*Az*°	*Baz*°	*P*	*SH*	*SV*
Event of February 19, 1968						
AAE	33	154	340	+		
ATL	82	307	47	+	+	
BEC	70	296	55	+	+	
BLA	77	308	50	+		
BUL	60	176	357	+		
COL	76	357	6	+		
GDH	50	332	84	+	+	
KOD	55	107	311	+		
NAI	42	162	346	+		
OGD	71	308	54	+		
PRE	65	177	357	+		
QUE	35	92	297	+		
TRN	80	276	51	+		
WES	69	308	56			
Event of December 19, 1981						
BLA	78	308	50	+		
CAR	85	279	51	+		
COL	76	357	6	+	+	+
GDH	50	333	84		+	
KEV	31	1	183	+		
NDI	44	88	298	+		
PDA	39	284	72	+		
POO	47	102	307	+		
SHL	57	83	302	+		
WES	69	308	56	+		
Event of August 6, 1983						
AAM	75	313	49		+	
AKU	35	311	115	+		
ANP	79	67	309		+	+
BLA	77	308	49	+		
BUL	60	176	357	+		
COL	75	357	6	+	+	+
DAG	41	346	127		+	
KOD	55	108	312	+		
MAT	82	48	315	+		
NAI	43	162	346	+		
NDI	44	88	299	+		
SNG	75	94	310	+		
TRN	80	276	51	+	+	+
WES	68	307	55	+		

Table 2

Crustal structure model at the source used in calculating synthetic seismograms

V_p (km/sec)	V_s (km/sec)	Density (gr/cm^3)	Thickness (km)
1.50	0.001	1.00	1.0
6.30	3.50	2.70	30.0
7.90	4.39	3.35	Half-space

depths in the particular velocity structure assumed. Receiver structures are assumed to be a homogeneous half-space. Responses from 33 *P*- and 3 *SH*-, *SV*-rays, which included all significant arrivals within the first 35 s of the signal for *P* and *SV*-waves and 25 s for *SH*-waves, were calculated and, summed up. These responses were then convolved with the WWSSN 15–100 instrument responses and Futterman's attenuation operator with $t^* = 1$ for *P*-waves and $t^* = 4$ for *S*-waves. The source time function was parameterized as a series of boxcars, each of 1 s duration. The synchronization of the initial points of the windowed Green's functions with the associated body wave data was very carefully performed since it plays a critical role in the inversions. The short-period records were primarily used in the cases of nodal stations to identify the *P* arrival. In all cases, the waveform data showed a strong interference of phases at the beginning of the records which suggested a shallow source depth, <15 km.

A different number of boxcar elements was tested, so as to avoid truncating the time function. Inversions for the moment tensor and 8–10 time function elements were performed with the source at 3, 6, 9, 10, 12 and 15 km depth. Five iterations were performed, with rapid convergence within three. In this study, the parameters were not weighted, and Wiggins covariance matrix for the data was assumed to be the identity matrix. During the test inversions a cut-off value for the maximum allowable variance of the parameter changes was arbitrarily set so that all parameters were well resolved and any singular values are assumed to be zero. The waveforms of the *S*-waves, registering larger amplitudes than those of the *P*-waves, commanded a higher importance in the inversion and were therefore preferentially fit by the synthetics. The *S*-wave data were weighted to equalize their importance, with no significant change in the synthetics.

Indicators of the performance of the inversion are the RMS error, measure of the fit of the synthetics to the data, and the least-square's error, LSE, a measure of the overall quality of the inversion. Finally, the compensated linear vector dipole, CLVD, is the remainder following the rotation of the moment tensor into its principal axes and its decomposition into a major double-couple. Plots of the variation of RMS, LSE and CLVD with depth were used to determine an optimum source depth for each case. Minima were revealed for d = 10 km for the event of February 19, 1968, for d = 6 km for December 19, 1981 and for d = 9 km for the

Table 3

Inversion results

Inversion Parameters Moment Tensor	Febr 19, 1968		Dec 19, 1981		Aug 6, 1983	
	p	$\sigma\Delta p$	p	$\sigma\Delta p$	p	$\sigma\Delta p$
$M_{11} \times 10^{25}$ dyn.cm	−21.1	0.078	−20.5	0.076	−11.2	0.005
$M_{22} \times 10^{25}$ dyn.cm	+21.01	0.077	+21.1	0.079	+11.9	0.006
$M_{12} \times 10^{25}$ dyn.cm	+6.24	0.071	−0.62	0.086	−1.97	0.005
$M_{13} \times 10^{25}$ dyn.cm	+4.72	0.026	−0.06	0.073	−2.18	0.002
$M_{23} \times 10^{25}$ dyn.cm	−0.13	0.016	−8.06	0.056	−1.91	0.002
$M_0 \times 10^{25}$ dyn.cm	22.42		22.38		12.11	
Time Function	p	$\sigma\Delta p$	p	$\sigma\Delta p$	p	$\sigma\Delta p$
S_1	0.102	0.0018	0.074	0.0014	0.152	0.0026
S_2	0.095	0.0030	0.061	0.0022	0.065	0.0041
S_3	0.168	0.0036	0.147	0.0024	0.188	0.0045
S_4	0.155	0.0039	0.154	0.0025	0.208	0.0046
S_5	0.149	0.0039	0.130	0.0024	0.151	0.0046
S_6	0.163	0.0036	0.158	0.0025	0.132	0.0045
S_7	0.066	0.0030	0.093	0.0024	0.082	0.0041
S_8	0.102	0.0019	0.085	0.0024	0.022	0.0026
S_9			0.065	0.0022		
S_{10}			0.032	0.0014		
Principal Axes	$AZ°$	Plunge°	$AZ°$	Plunge°	$AZ°$	Plunge°
Pressure Axis	78	5	270	20	273	7
Intermediate Axis	340	78	90	71	152	75
Tension Axis	170	13	280	0	6	14
RMS FIT (μm $\times 10^0$)	8.31		5.56		3.19	
LEAST SQUARE ERROR	0.26 (μm$^2 \times 10^4$)		0.20 (μm$^2 \times 10^4$)		0.11 (μm$^2 \times 10^0$)	
ERROR CLVD	4.62%		14.7%		3.7%	

event of August 6, 1983. Table 3 lists the inversions results for the above-mentioned source depths, i.e., the moment tensor elements, the time-function elements, the standard error of the parameter changes ($\sigma\Delta p$), computed assuming 10% error in the data amplitudes and the major double-couple parameters for each event.

2.1. Event of February 19, 1968

Figures 2 and 3 display the inverted *P* and *SH* waveforms and the resulting synthetics for a source depth of 10 km. The nodal surfaces of the moment tensor and its major double-couple are also plotted on a lower hemisphere equal area projection (broken lines on the figures). Both the synthetics and the waveforms are normalized to their maximum amplitude, shown to the right side of each trace (in microns). The synthetics exhibit a good fit in amplitude and overall shape, within the inversion window, though they fail to fit the distortions observed in the waveforms. LANGSTON (1977) notes that teleseismic *P*-waves radiated from sources

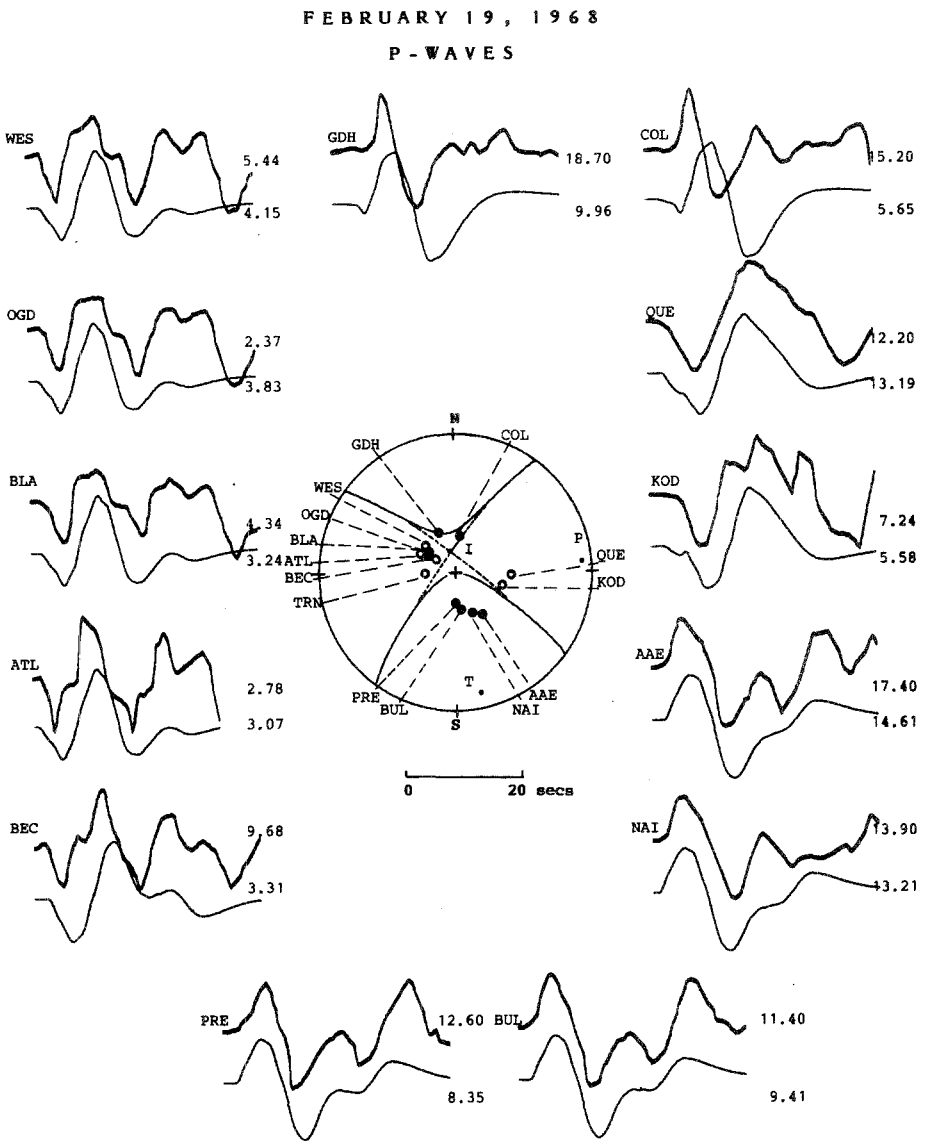

Figure 2

Observed (top) and synthetic (bottom) *P* waveforms for the event of February 19, 1968. The waveforms are normalized to their maximum amplitude, indicated to the right of each trace. The moment tensor nodal surfaces (solid lines) and its major double-couple (dashed lines) are also shown in a lower hemisphere equal area projection. The location and the polarities of *P*-wave first motions are also shown on the figure.

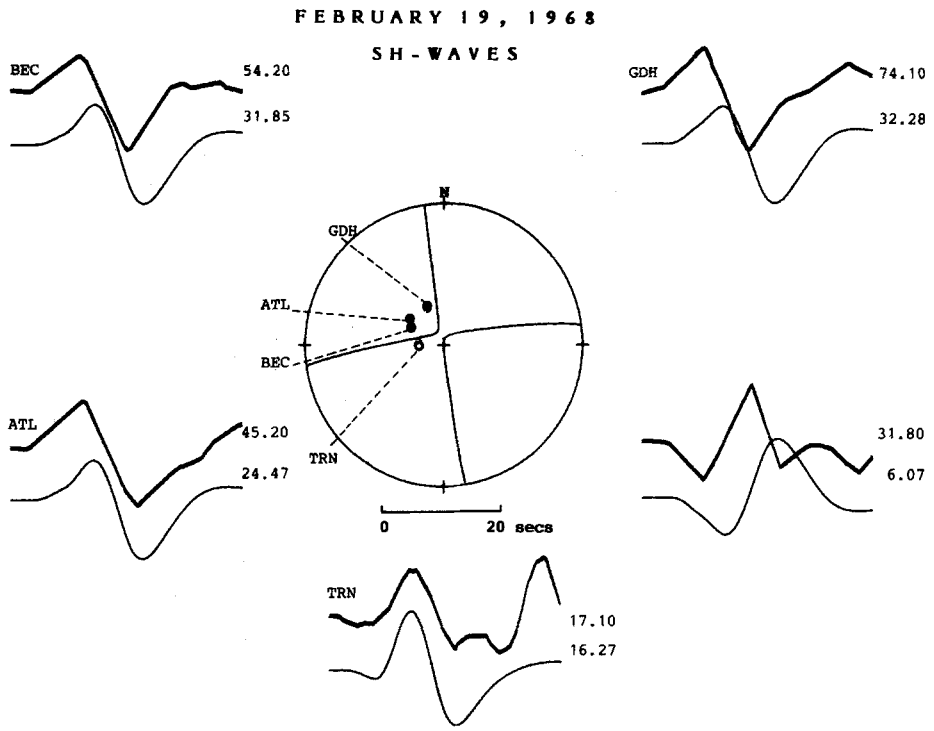

Figure 3
Observed and synthetic *SH* waveforms for the event of February 19, 1968. Notation the same as Figure 2.

whose radiation pattern varies rapidly with azimuth (e.g., strike-slip faults) can be greatly distorted by a dipping structure at the source. It is probable that the misfit observed in the long period *P*-waveforms of the events studied here is attributed to the nonhorizontal nature of the sea floor in the epicentral region, since strike-slip sources show the greatest sensitivity to bathymetric dip (WIENS, 1987).

MCKENZIE (1972) based on first motion data, presents a strike-slip solution for this event with a normal component (strike 205°, dip 70°NW and strike 114°, dip 87°SW). The fault planes obtained here with the moment tensor analysis are in favorable agreement with MCKENZIE's (1972). We favor the NE striking plane as the fault plane because it parallels the general trend of the Anatolian fault.

This earthquake occurred near the island of Agios Efstratios. Field observations and air photograph analysis of the surface fault trace, which cuts the island in a NE-SW trend, and is associated with the 1968 shock, show a dextral strike-slip motion with normal component (PAVLIDES *et al.*, 1989).

The seismic moment, determined by dividing the modulus of the moment tensor by $\sqrt{2}$, is 2.24×10^{26} dyn.cm. The average stress drop, assuming a circular crack, is of the order of 13 bars.

2.2. *Event of December 19, 1981*

The epicenter of this earthquake (39.1°N, 25.2°E) was located in the sea west of the island of Lesbos. On December 27, 1981, its largest aftershock of $M_s = 6.5$ occurred in the southwesternmost part (38.8°N, 24.9°E) of the aftershock area. On January 18, 1982, one month after the first shock, a new seismic sequence started, with an earthquake of $M_s = 7.0$ and an epicenter (39.8°N, 24.5°E) located 60 km northwest of the epicenter of December 19, 1981 (PAPAZACHOS *et al.*, 1984).

Figures 4 and 5 show the inverted *P* and *SH* waveforms and the resulting synthetics for a source depth of 6 km. The inversion yielded a dextral strike-slip fault, with strike 47°, dip 77° (ESE), rake 193° for one plane and strike 314°, dip 78° (WNW) and rake −14° for the other. This mechanism is slightly different from the one proposed by PAPAZACHOS *et al.* (1984), based on first motions of long and short period *P*-waves.

The focal mechanism for the largest aftershock of December 27, 1981, determined by KING and KIRATZI (1985) by forward modeling, manifests an almost

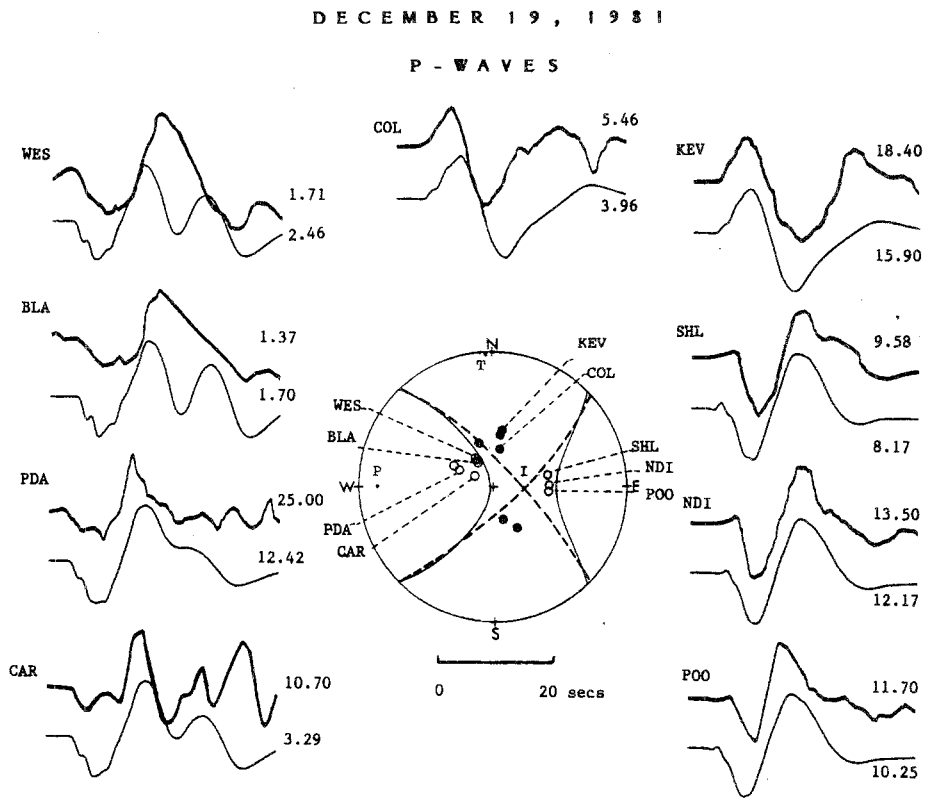

Figure 4
Same as Figure 2 for the event of December 19, 1981.

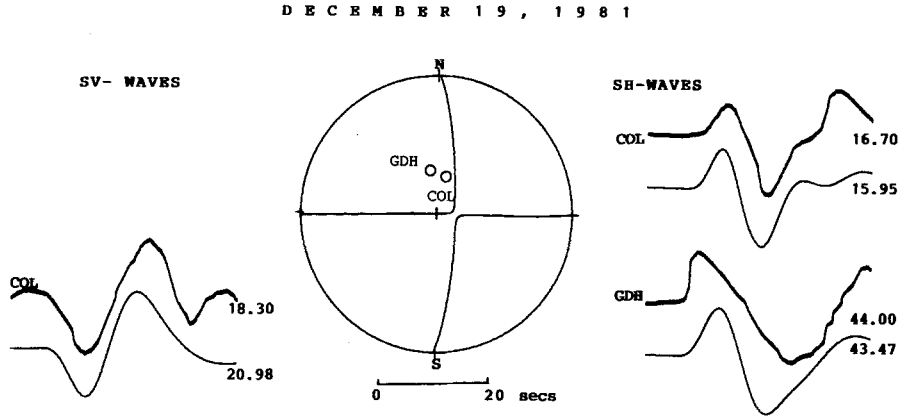

Figure 5
Observed and synthetic *SV* and *SH* waveforms for the event of December 19, 1981. Notation as in Figure 3.

pure strike-slip fault with strike 33°, dip 87° and rake 188°. As for the event of January 18, 1982 an attempt was made to model its source parameters utilizing moment tensor inversion. However, the data were few and noisy and the inversion parameters were not well resolved. Regardless, all the solutions for this event yielded strike-slip faulting with a strong thrust component. PAPAZACHOS *et al.* (1984), also using *P* polarities, determine a strike-slip fault with strike 56°, dip 50° (WNW) and rake 26°.

The seismic moment of the event of December 19, 1981, determined by the inversion, is 2.24×10^{26} dyn.cm while the average stress drop, assuming a circular crack with the fault area determined by the aftershocks (KIRATZI *et al.*, 1985), is 8 bars. Using M_0 as it was determined by the inversion and a downdip width equal to twice the focal depth divided by the sine of the dip angle, an average slip of 67.3 cm was obtained ($\mu = 3 \times 10^{11}$ dyn.cm^{-2}).

2.3. Event of August 6, 1983

This was a widely felt earthquake which occurred 30 km WNW of the island of Limnos (40.08°N, 24.78°E). Figures 6 and 7 display the inverted *P*, *SV* and *SH* waveforms and the resulting synthetics for a source depth of 9 km. In this case also, the synthetics show favorable fit in overall shape and amplitude. The focal mechanism corresponds to a dextral strike-slip fault with strike 50°, dip 76°, rake 177°, for one plane and strike 140°, dip 87° and rake 14° for the other. This mechanism corresponds well with the one determined by ROCCA *et al.* (1985), based on first motion data.

The seismic moment determined by the inversion is 1.21×10^{26} dyn.cm while the

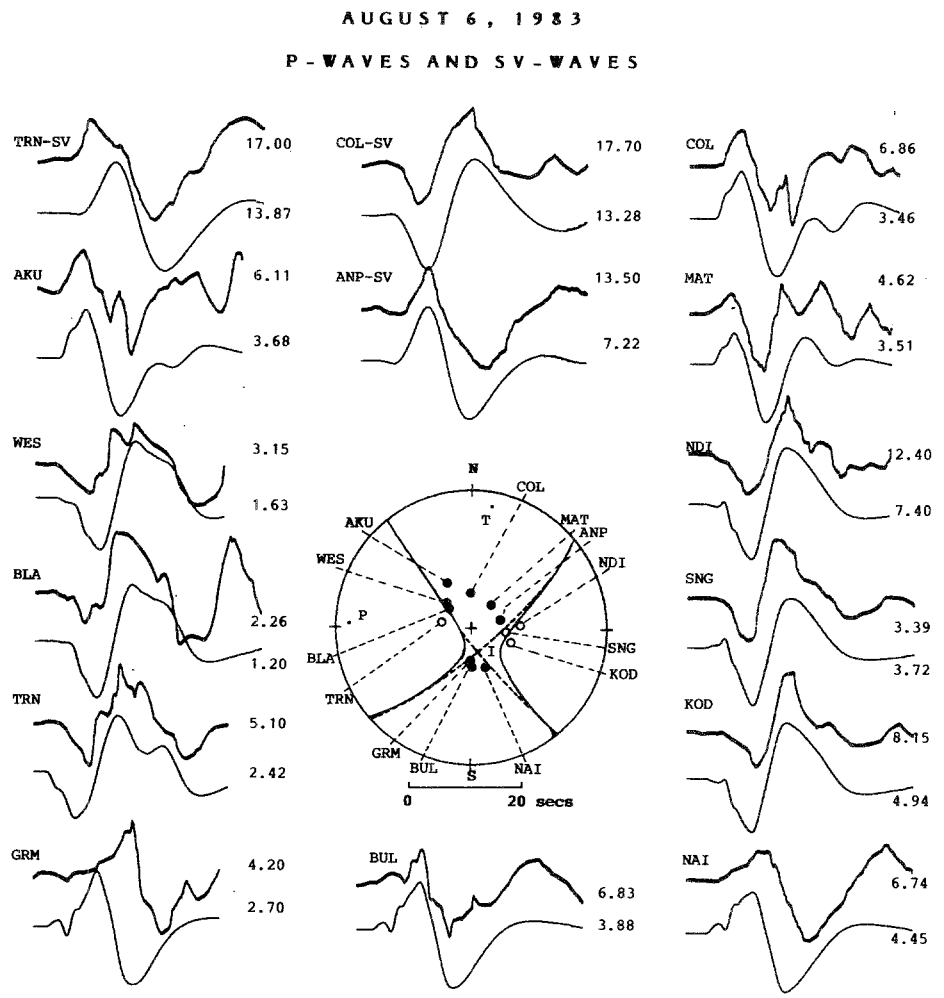

Figure 6
Observed and synthetic *P* and *SV* waveforms for the event of August 6, 1983. Notation as in Figure 2.

average stress drop, assuming a circular crack, is approximately 28 bars. The, average slip, calculated as before, was found equal to 45.5 cm.

3. Conclusions and Discussion

Source mechanisms, depths, seismic moments and source time functions have been obtained for the events of February 19, 1968, December 19, 1981 and of August 6, 1983 occurring in the Northern Aegean, employing the moment tensor inversion formalism. The observed waveforms of these events exhibited prominent

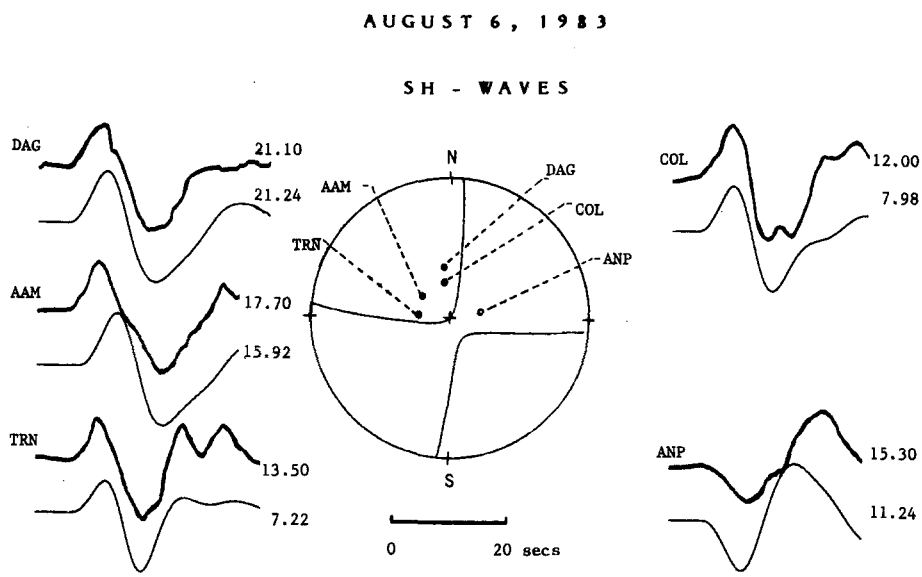

Figure 7
Same as Figure 3 for the event of August 6, 1983.

water multiples and complexity which were not modelled by the moment tensor inversion, assuming a point source and a flat lying bathymetry in the epicentral regions. Examination of the improvement of the synthetics with a structure model containing dipping layers at the source will be the scope of future work.

All focal depths determined are shallow (≤ 10 km) which places the source in the crust and intensifies previous suggestions that the strain rate in the Aegean is high enough to raise the isotherm representing the seismic-aseismic transition above its more typical continental depth of around 15 km (JACKSON and MCKENZIE, 1988).

The derived focal mechanisms of the events studied exhibit dextral strike-slip faulting. As previously stated the Northern Aegean is tectonically controlled by the NS extension the whole Aegean is undergoing and the motion of the westward part of the strike-slip North Anatolian fault. The moment tensor elements and the focal mechanisms of the events studied are in very good agreement with this tectonic regime.

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