MEASUREMENT AND LOCATION OF EARTHQUAKES IN WESTERN ALASKA, THE GULF OF ALASKA, AND THE BERING SEA

by

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SECTION I: OBJECTIVES AND PRELIMINARY CONCLUSIONS

by Cliff Frohlich

OBJECTIVES

As stated in our renewal proposal entitled "Measurement and Location of Earthquakes in Western Alaska, the Gulf of Alaska and the Bering Sea," the specific objectives of our research were:

1. To recover the strong motion OBS instruments deployed in the Gulf of Alaska in 1980, and deploy these and other instruments in the Gulf of Alaska and the Bering Sea.

2. To monitor microseismic activity rates near the Amak Fault Zone and the Port Moller Graben to determine whether these features are currently active, and if so, to determine the level of activity.

3. In collaboration with personnel of the University of Alaska, to apply recently developed sophisticated location methods to the earthquake data that have been collected by the University of Alaska land network. We intend to obtain the most accurate locations possible for events detected by the network.

4. In collaboration with personnel of the University of Alaska, to develop a velocity model and set of station corrections that allow the University of Alaska land network to determine the best possible locations for events that occur within and adjacent to their network. To determine realistically the strengths and limitations of the land network for the determination of earthquake detection and earthquake risk in the offshore area.

PRELIMINARY CONCLUSIONS, AND PLAN OF THIS REPORT

In order to reach these objectives we undertook five more or less separate research projects. Some of these projects were continuations of previous work. These projects were:

1. Recovery, deployment, and analysis of strong motion OBS instruments in the Gulf of Alaska and the Bering Sea. The instruments deployed in 1980 were recovered in June of 1981 from the Miller Freeman (see Figure I-l). In September of 1981 new instruments were deployed from the Alpha Helix. These will be recovered in July of 1982. Analysis of the data obtained in 1981 is still in progress. Persons interested in these results should contact Dr. Paul Donoho of the Institute for Geophysics at the University of Texas.

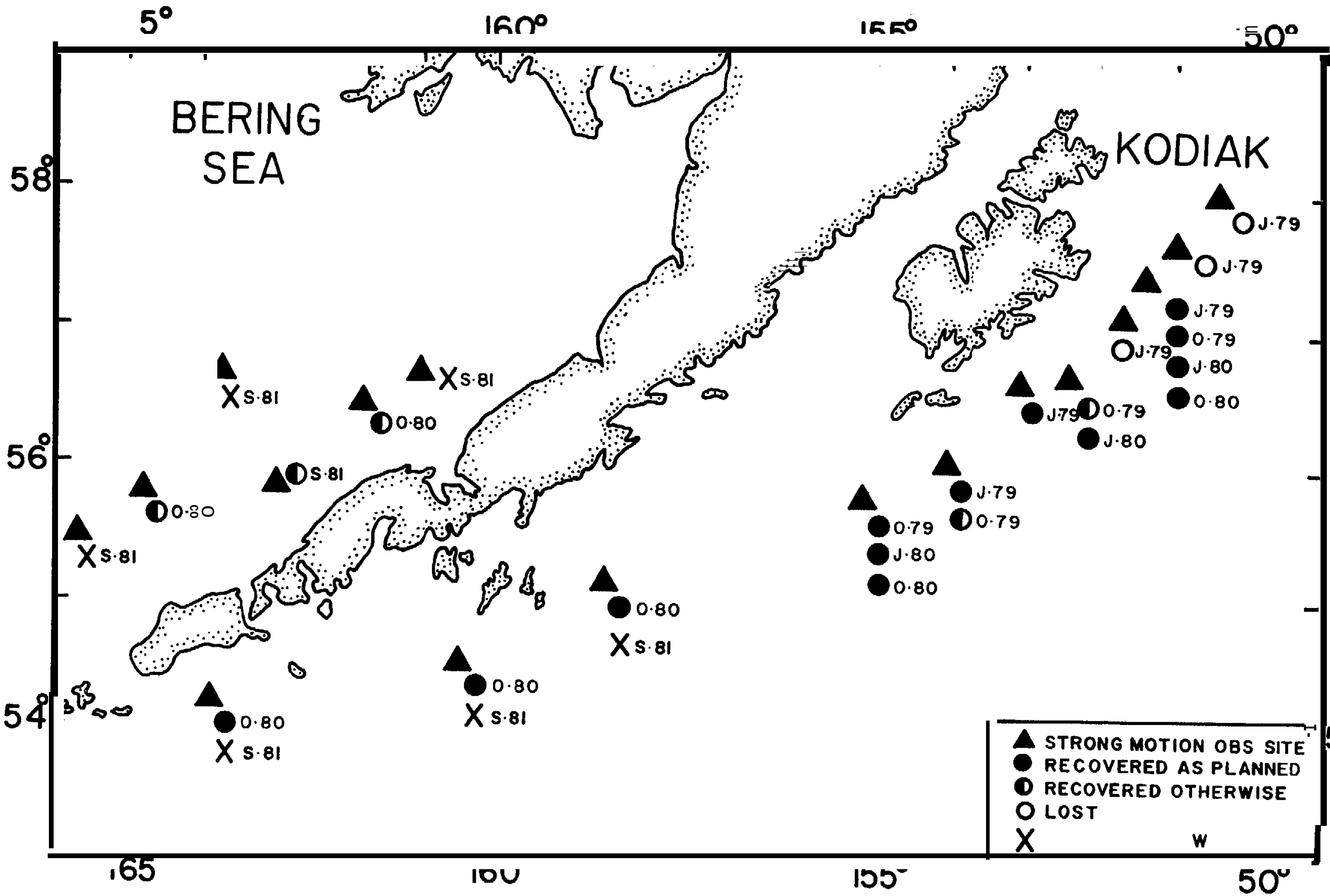

FIGURE I-1: Location map showing where strong motion OBS instruments have been deployed since 1979. Triangles are OBS sites, and symbols below exp ain recovery situation. Letter and number to the right of recovery symbol are month and year of deployment, e.g., J-79 is July 1979, 0-79 is October of 1979, J-80 is June of 1980, 0-80 is October of 1980, and S-81 is September of 1981.

2. Deployment and recovery of high gain OBS instruments in the Bering Sea near Amak Island and Port Moller. These instruments were deployed for six days in June of 1981 from the Miller Freeman. Analysis of this work is complete, and makes up section II of this report. Although the instruments worked well, no microearthquakes were recorded that occurred within the Bering Sea. Although a six day microearthquake project is by no means definitive for risk determination, the absence of seismicity is consistent with the conclusion that risk from shallow events in the study region is low. Probably risk associated with subduction zone events in the Aleutian Island arc is much greater.

3. Joint Hypocenter Determination (JHD) of regional events with focal depths of 50 km and greater in the Cook Inlet area. Using a JHD program developed specifically for this project, we relocated 178 events with focal depths beneath 50 km in the neighborhood of Cook Inlet. This work makes up section 111 of this report. These locations are the most accurate locations available at present for this area. The station corrections determined should allow more accurate determination of epicenters in this area in the future using the stations in the UA network. This work corroborates the previous work which suggests that a "double Benioff zone" may exist in the Cook Inlet area.

4. Joint Hypocenter determination of teleseismic events in the Kodiak Shelf region. Using a JHD program developed previously we relocated 34 shallow events which occurred offshore of Kodiak Island. This work makes up section IV of this report. The relocated epicenters form a much less diffuse pattern than the epicenters reported by the International Seismological Center (ISC). In addition, analysis of systematic errors in teleseismic locations suggests that these events are situated on the bathymetric shelf break, rather than about 20 km to the north as reported by the ISC.

5. Joint Hypocenter Determination of regional events in the Kodiak Shelf region. This work is still in progress. For m o r einformation, interested persons should contact Dr. Hans Pulpan of the University of Alaska or Dr. Cliff Frohlich of the University of Texas Institute for Geophysics.

SECTION II: NO LOCAL EARTHQUAKES RECORDED IN BRIEFOBS SURVEY IN THE BERING SEA NEAR AMAK ISLAND, ALASKA

by Cliff Frohlich

Recent hydrocarbon exploration in the Gulf of Alaska and the Bering Sea (Hanley and Wade, 1981) has provided motivation for studying the seismicity in these areas. For this reason, the University of Texas has been engaged in ocean bottom seismograph (OBS) research in these areas since 1978, both with strong motion instruments (Steinmetz et al. 1981) and high gain equipment (Lawton et al. 1982).

Because the NOAA ship Miller Freeman planned to visit the Bering Sea for about a week in June of 1981 to recover some strong motion OBS instruments that had been deployed in October of 1980, we decided to undertake a brief microearthquake survey of two basement features in the Bering Sea (Figure 11-1). These features were the Amak Fault zone and the Port Moller graben. Since 1960 the National Earthquake Information Service (NEIS) has reported three shallow earthquakes occurring along the Amak Fault zone, and one shallow event along the Port Moller graben. A network operated by Lament-Doherty observatory of Columbia University has detected a cluster of small shallow events-that occurred in 1980 on the peninsula south of Port Moller Bay (Klaus Jacob, personal communication). Davies (1981) presents a more detailed discussion of the historical seismicity in the St, George Basin and adjacent regions.

For this study we deployed four vertical-component high-gain Texas OBS instruments similar in design to those described by Latham et al. (1978). After about six days the 'Miller Freeman recovered three of these instruments (Table II-l). Although we detected a weak radio signal from the fourth instrument it was not recovered at this time. Nevertheless, in October of 1981 a fisherman found it and returned it to the University of Texas. No events of any kind were recorded by the fourth instrument.

Each of the three instruments recovered by the Miller Freeman recorded more than a hundred events, however, most of these events did not appear to be earthquakes. All the events looked similar to the events discussed by Buskirk et al. (1981) which are thought to be of biological origin. None of the instruments recorded any events which appear to be local earthquakes.

However, the two stations deployed near Amak Island did record two earthquakes. The first event occurred on 5 June 1981 about 380 km from the OBS instruments (NEIS location: 52.259N 165.186W 11 km 070915.8). No clear first arrival times could be read for this event, as on one OBS the arrival was extremely emergent, and on the other the event occurred during a very noisy portion of the record. A second event occurred on 7 June 1981 about 200 km from the OBS stations (NEIS location: 53.877N 165.086W 14 km 175231.6), and both stations recorded this event clearly (Figure II-2). For the location reported by NEIS, P arrivals at both stations occur about 6.0 sec later than predicted by the JB tables.

Because the OBS survey lasted for only six days, the significance of the absence of any recorded local events is not entirely clear. The fact that two distant events were observed by stations HG1 and HG2 suggests that local events would have been recorded if they did occur.

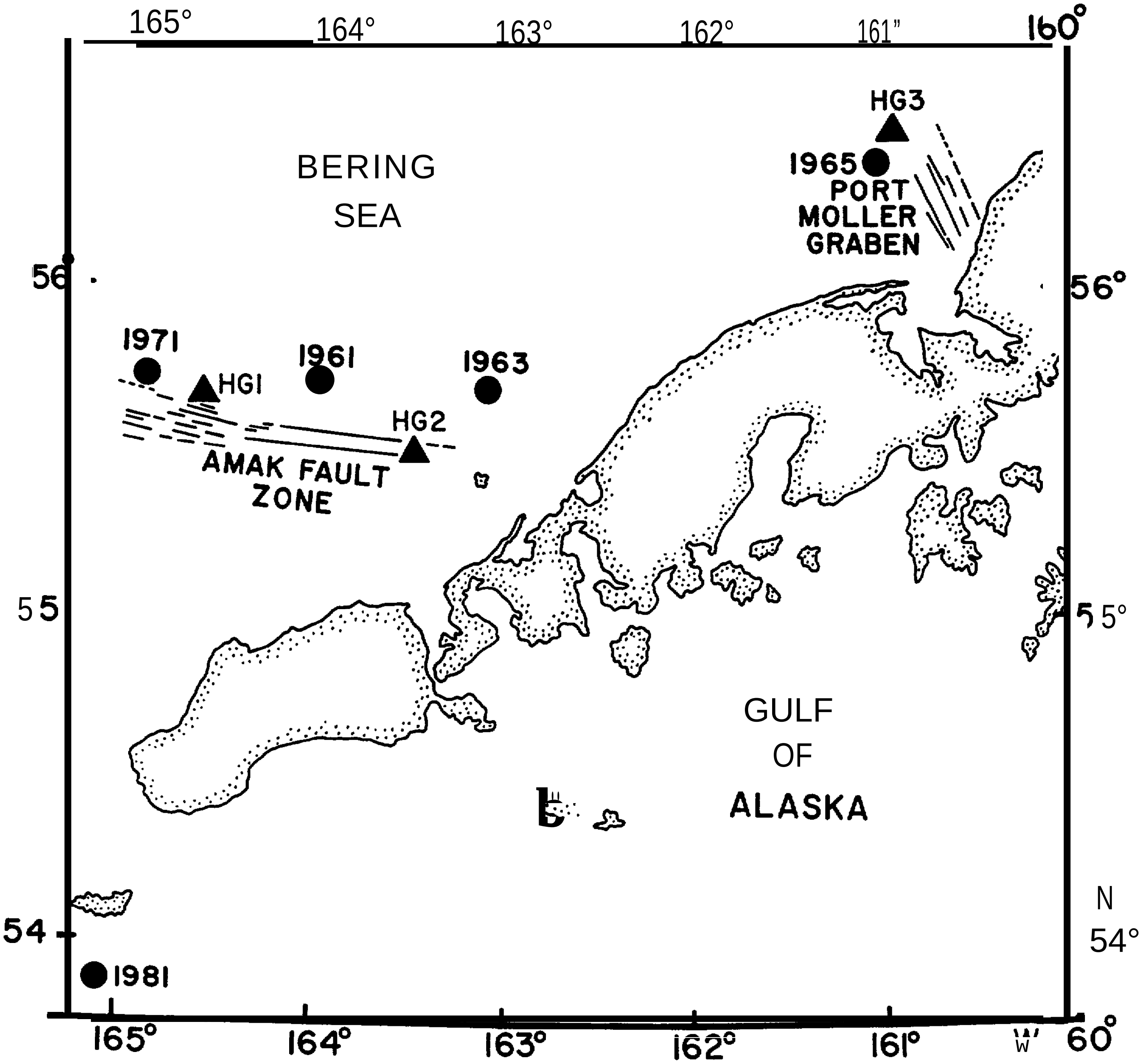

FIGURE II-1: Location map of the Amak-Port Moller area, showing the locations of the OBS stations recovered (filled triangles) and events reported by NEIS in the Bering Sea since 1960 (filled circles, with year of event adjacent). In addition, the map shows the NEIS location of one of the events of 7 June 1981-which was recorded by stations HG1 and HG2.

Three teleseismically recorded events have been reported near the Amak fault zone since 1960 (Figure II-l), one in 1961, one in 1963, and one with magnitude 5.2 in 1971. Assuming a b-value of 1, three events of magnitude 5 will occur as often as 30,000 events of magnitude 1. Thus if one observes 3 magnitude 5 events in 20 years, one should observe about 4 magnitude 1 events each day, or about 25 in the six days that the OBS instruments were deployed. Since no events were observed, either this b-value is too high, or else the events cluster in time. Near Port Moller, the only shallow event reported teleseismically since 1960 occurred in 1965, and had a magnitude of 4.2. However, this event was reported by only 8 stations. As events with focal depths beneath 70 km are known in this area, it is possible that the 1965 event was a deeper event whose focal depth was incorrectly determined.

TABLE II-1: High-gain OBS stations deployed in the Bering Sea in 1981.

STATION	DEPLOYED	LOCATION	WATER DEPTH (fro)	RECOVERED
HG1	03 JUN 0650	55 39.9N 164 39.6W	52f	09 JUN 1330
HG2	03 JUN 0230	55 30.0N 163 30.0W	39f	09 JUN 0730
HG3	02 JUN 1300	56 30.0N 160 50.1W	33f	08 JUN 1300
HG4	02 JUN 0300	58 14.9N 160 39.8W	14f	---

ACKNOWLEDGEMENTS

Paul Donoho, Jeff Lawton, and Phil Roper provided essential field and moral support for this research. We thank Capt. Jack Atwell and the crew of the Miller Freeman for their assistance and their good fellowship.

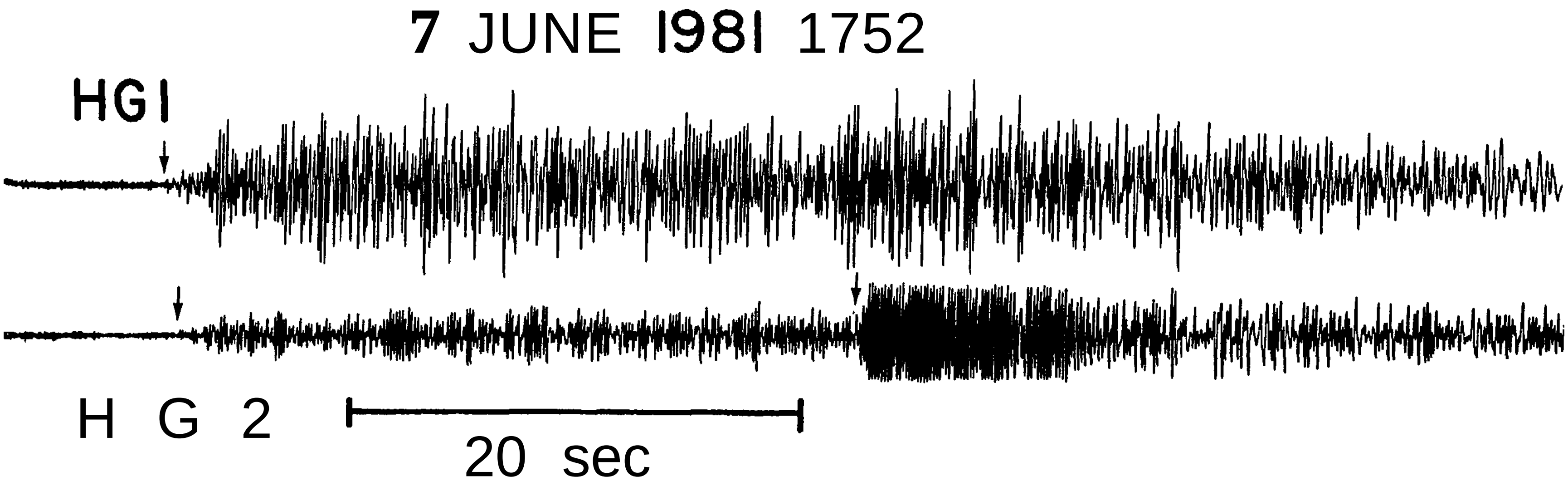

FIGURE II-2: Seismograms recorded at stations HG1 and HG2 for the event of 7 June 1981. The arrows show the author's picks for the P arrival at station HG1, and the P and S arrivals at station HG2. These arrival times were 1753:10.2 (HG1), 1753:10.7 (HG2-P), and 1753:40.7 (HG2-S).

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SECTION III: JOINT HYPOCENTER DETERMINATION OF INTERMEDIATE DEPTH EARTHQUAKES NEAR COOK INLET, ALASKA

by

Cliff Frohlich and Hans Pulpan

INTRODUCTION

Although a few seismograph stations have been operated in southwestern Alaska for many years (Lahr, 1975), there has only been a true local network in the region near Cook Inlet since about 1975. This network, capable of locating most earthquakes with magnitudes as small as 2 or below, includes about 30 stations and is operated by the University of Alaska and others (Pulpan and Kienle, 1979). For accurately locating earthquakes occurring in subducted lithosphere, this network is better situated than networks in many subduction zones because of its relatively broad aperature across the arc, including stations as much as 100 km behind the volcanic arc.

However, previous published earthquake locations in this region have not utilized the full location potential of the present network of stations in the Cook Inlet area. For example, Pulpan and Kienle (1979) did not incorporate S-wave observations into their locations in this region. Lahr (1975) only used about 15 stations recording in this region to locate events because these were all that were available at that time. At present, Lahr and his coworkers at the USGS do record data from several stations in the Cook Inlet area, and do use these data when preparing catalogs of earthquakes in southwestern Alaska (e.g., see Stephens et al, 1980).

Several recent studies have focused new interest on locations in subducted lithosphere at depths above 200 km, as they have apparently identified a second less active zone of seismicity beneath the planar Benioff zone. The combination of the usual Benioff zone together with this lower, less active zone are known as a "double Benioff zone." This has been clearly observed in Japan (Hawegawa et al., 1979a; 1978b), and observed less clearly in several other places (Reyners and Coles, 1982; Samowitz and Forsyth, 1981; Veith, 1974). In the Cook Inlet area, Lahr (1975) observed a number of well-located events that were about 20 km beneath the main Benioff zone. However, in spite of the fact that these events possessed most of the features we attribute today to double Benioff zones, Lahr's (1975) work is seldom mentioned in most papers concerning double Benioff zones.

Unfortunately, numerous studies have shown that locations of local networks near subduction zones can be influenced by systematic errors. The use of flat-layered velocity models for determining locations in a region of more complex velocity structure can cause the seismicity pattern to exhibit unusual and apparently spurious features. These phenomena have been studied in the greatest detail in the Central Aleutian arc using data from the Adak network. For example, events which seem to cluster at shallow depths near model velocity increases (Engdahl, 1977) have been found to be distributed over a broad range of depths (Laforge and Engdahl, 1979; Frohlich et al., 1982). Increases in the dip of the Benioff zone disappear when ray tracing is used to locate the events (Engdahl et al., 1977), or when the network geometry is augmented using ocean-bottom seismographs. An initial report of the existence of a double Benioff zone in the Adak region (Engdahl and Scholz, 1977) was not confirmed by more detailed research (Topper, 1978). Most recently, a number of other detailed investigations have been performed in Adak.

In the present paper we report relocations of events in the Cook Inlet region using all the available data and the joint hypocenter determination method (Douglas, 1967; Dewey, 1972; Frohlich, 1979). These relocations are among the best published locations available in this region at this time.

METHODS

Selection of Events and Stations

Although the University of Alaska (UA) has operated a network of stations in the Cook Inlet area since 1975, S waves have been reported routinely by UA personnel during only three years, 1978, 1980, and 1981. All of the UA stations are vertical component seismographs only, and so all of the reported S-arrivals were read from vertical component records. Restricting our attention to the region between 59 and 60.5°N and between 151.6 and 154°W, there existed about 400 events for which one or more S waves were available. Of these, 178 had two or more S readings, a gap of 170 degrees or less, and a rms location residual of 0.5 sec or less (Figure III-l).

There exist more than 50 stations operating in southern and southwestern Alaska, however, not all of them are close enough to the Cook Inlet area to improve our locations. For this reason, 30 stations were selected for use in this study. Although phases from all of these stations were used in the locations, in practice only 15 of the stations reported arrivals for more than 30 per cent of the events (Table 111-1).

As an aid in evaluating the quality of the data to be used in the relocation, we graded all of the earthquakes in terms of three parameters (see caption to Figure III-l). These were the number of S observations reported, the "gap" in station coverage (the largest azimuthal gap for which no P-wave observations were available), and the rms residual of the preliminary location reported by the UA network. As noted above, of the approximately 400 events with focal depths beneath 50 km in the study area, there were 72 events with Q = 1, 106 events with Q = 2, and the remainder with Q = 3.

Relocation by the Joint Hypocenter Determination Method

For groups of earthquakes observed by a network of seismic stations the JHD location program used simultaneously determines hypocenters and station corrections so as to minimize the sums of residuals of travel

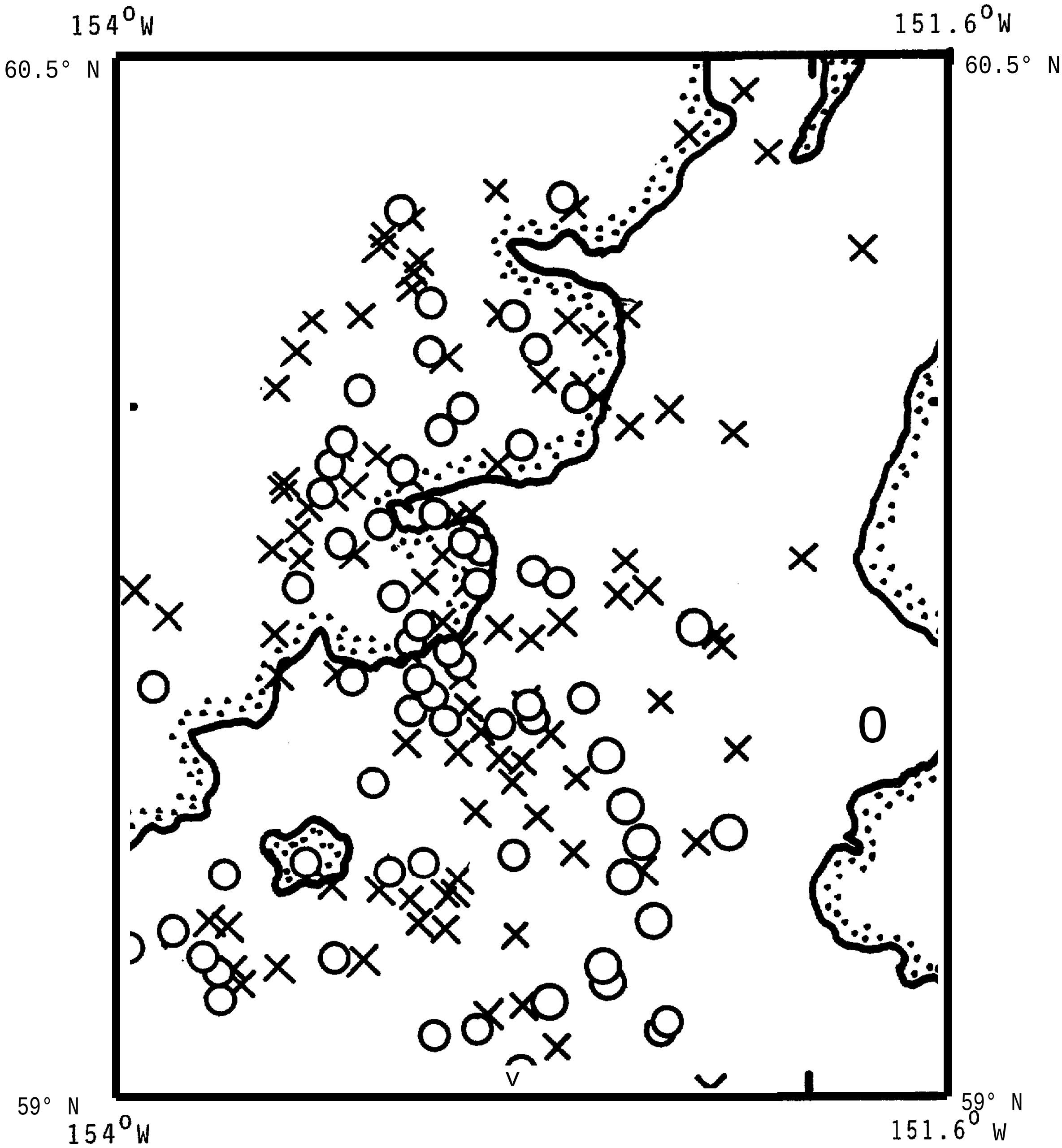

FIGURE III-1 Map of earthquakes locate in this study
Circles are quality Q = 1 events (3 or more S observations, gap of 130° or less, and rms residual of 0.4 sec or less). Crosses are quality Q = 2 events (2 S observations, gap of 170° or less, and rms residual of 0.5 sec or less). Events with one S observation (not shown) were quality Q = 3 events.

TABLE 111-1: Station corrections determined by the JHD relocation process for each of 15 stations that recorded 30 per cent or more of the seismic events. The six relocations were; I - Q = 1 events only (72 earthquakes), P and Swave residuals used to determine station corrections; II - Q = 1 events only (72 earthquakes), Pwaves only used to determine station corrections; III - Q = 1 and Q = 2 events (54 earthquakes), focal depth beneath 100 km in northeast section, Pwaves only used for station corrections; IV - Q = 1&2 events (35 earthquakes),focal depth above 100 km in NE, Pwaves only; V - Q = 1&2 events, (27 earthquakes), focal depth beneath 100 km in SW, P waves only; VI - Q = 1&2 events, (62 earthquakes) Pwaves only, focal depth above 100 km depth.

STATION	RELOCATION I	RELOCATION II	RELOCATION III	RELOCATION IV	RELOCATION V	RELOCATION VI
AU I	0.16 sec	0.15	0.35	0.16	0.10	0.24
AUM	0.31	0.31	0.35	0.31	0.32	0.46
BGM	-0.25	-0.23	-0.06	-0.30	-0.20	-0.42
CDA	-0.20	-0.16	-0.16	-0.32	-0.10	-0.18
CKK	-0.04	-0.04	-0.09	0.11	-0.08	-0.15
HOM	0.42	0.35	0.45	0.36	0.37	0.43
KDC	-0.35	-0.27	-0.56	-0.75	-0.38	-0.14
MCN	0.11	0.15	0.24	-0.07	0.17	0.11
OPT	0.08	0.08	0.20	0.16	-0.01	0.13
PDB	-0.07	-0.10	0.14	-0.06	-0.07	-0.16
RA 1	-0.12	-0.01	-0.28	0.05	-0.13	-0.05
RED	-0.28	-0.33	-0.22	-0.23	-0.47	-0.34
SHU	0.13	0.16	0.29	0.27	0.10	0.04
SLV	-0.19	-0.22	-0.20	-0.35	-0.23	-0.13
SVW	-0.09	-0.14	-0.30	-0.37	-0.15	-0.16

times for a particular flat-layered velocity model. The program used was developed by the author specifically for this project although it should prove generally useful for local network relocations. Unlike the method of, e.g., Spencer and Gubbins (1980), this program does not attempt to adjust the velocity model during relocation. For our relocations we used the velocity model determined by Lahr (see Stephens et al, 1980) which is reproduced in Table III-2.

The JHD program developed for this project is considerably more efficient than the programs used in most previous investigations. During each iteration of a relocation of N events observed at M stations, many JHD programs typically solve a system of 4N + M equations in 4N + M unknowns, This often makes it impractical to relocate jointly more than about 40 events in each group. However, using the method outlined by Frohlich (1979), during each iteration the program used in this project instead solves N systems of 4 equations in 4 unknowns, and thus literally hundreds or thousands of events can be relocated jointly.

As discussed by Douglas (1967) and Frohlich (1979), JHD methods must specify at least one additional constraint equation in addition to the condition that the sum of residuals be minimized. The most common additional equations are to fix one event (the " master event") or alternatively to specify some condition concerning the station corrections, such as making the sum of the station corrections zero. We have used the latter approach in this work. As shown by Frohlich (in preparation), if reasonable care is taken in the selection of the station network, the difference in the relative locations of events by the two methods is negligible.

For most teleseismic JHD relocation schemes, compressional wave observations only are used, and thus a single station correction is determined for each station. Numerous studies have shown that for most networks more reliable locations can be determined if S-wave observations are used (e.g., Buland, 1976). Thus for local network JHD relocations, it would be possible to determine separate station corrections for P and S waves. However, because the majority of the events studied in this project had three or fewer S waves, a single station correction was determined for each station and used to adjust both P and S residuals.

Results

To investigate the dependence of station corrections on the events located, we performed six different JHD relocations on events in the Cook Inlet area (see Table III-1 and Figure III-2). Two JHD relocations concentrated on 72 Q = 1 events, using both P and S residuals to determine station corrections for relocation I, and using P residuals only for relocation II. The station corrections thus determined were remarkably similar, as the station corrections determined in these two trials differed by more than 0.1 sec for only one of the fifteen

TABLE III-2: Velocity model used by Lahr and his coworkers (see Stephens **et al.,** 1980), and **also** used for JHD relocations in this study.

VELOCITY	DEPTH
2.75 km/see	0 - 2 km
5.3	2 - 4
5.6	4 - 1o
6.2	10 - 15
6.9	15 - 20
7.4	20 - 25
7.7	25 - 33
7.9	33- 47
8.1	47 - 65
8.3	65 - --

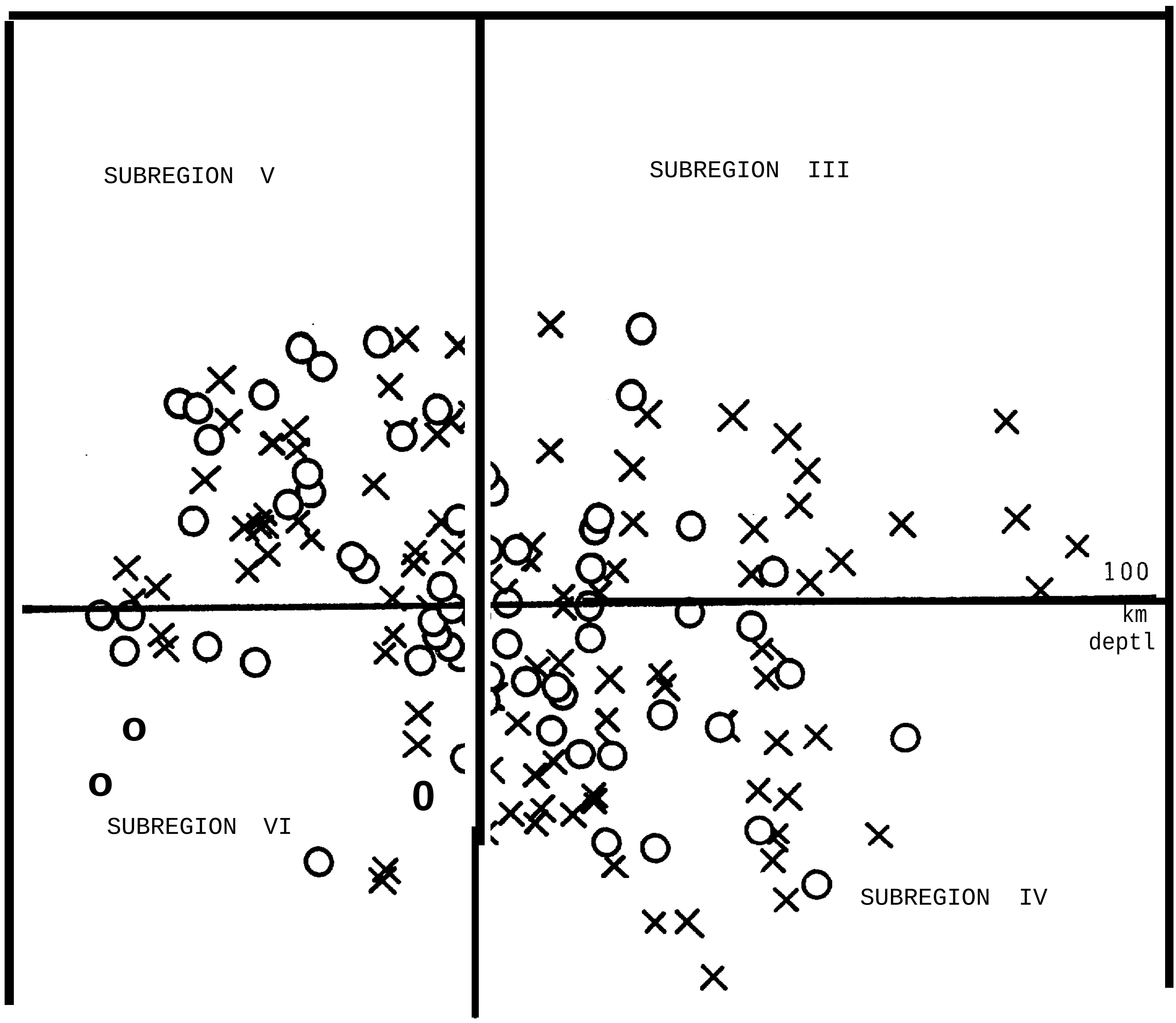

FIGURE III-2 Vertical cross section parallel to the Aleutian trench (azimuthal trend of 32° E of N) showing the events relocated in this study. All Q = 1 events (circles) were relocated in relocations I and II (see table III-1). Both Q = 1 and Q = 2 events (crosses) in the subregions III, IV, V, and VI shown were relocated in relocations III, IV, V, and VI.

stations observing more than fifteen events (Table III-l).

To study the effect of depth and geographic location on the station corrections, we also undertook four additional JHD relocations (Figure III-2) for events with Q = 1 orQ= 2 which occurred above 100 km depth in the southwest, below 100 km depth in the southwest, above 100 km in the northeast, and below 100 km depth in the northeast. Although the station corrections determined in these relocations differed more than those determined in the relocations I and II, for all but three stations the corrections were within 0,2 sec of those determined in the first relocation. Analysis of the station corrections determined in the six JHD relocations revealed that the corrections determined in relocation I were among the median values (the third or fourth largest of the six values determined) for eleven of the fifteen stations. For these reasons, these corrections were used in the relocation process for all of the events that follow, including those of quality Q = 2, and including events at all focal depths deeper than 50 km (Figure III-l).

Cross sections of the relocations of Q = 1 and Q = 2 events clearly delineate the Benioff zone in the Cook Inlet area (Figures III-3, III-4, III-5, III-6 and 111-7), The zone appears to extend to a depth of 160 km, with a dip angle of about 45 degrees and a thickness of 10 to 20 km or less. Two features of the data are worthy of note:

- There appear to be several well located events which lie distinctly beneath the main Benioff zone. For example, in the cross section of Figure 111-5 two events are separated from the main Benioff zone by about 15 - 20 km. Both of these events are well recorded and should be accurately located, in fact, the location of the deeper event used 13 P and 6 S observations. Altogether these events form a so-called "double Benioff zone."

- The cross sections suggest that the dip of the Benioff zone changes between the southwest and the northeast. In these data the dip is slightly larger in the northeast.

These results are comparable to those reported by Lahr (1975) for events in this region. Using a slightly different location technique and a geometrically less extensive network, Lahr (1975) also detected the presence of events beneath and apparently separate from the Benioff zone. He also found a change in the dip from northeast to southwest in the Cook Inlet region, however, unlike the present work Lahr (1975) found that the dip of the Benioff zone was slightly less in the northeast than in the southwest.

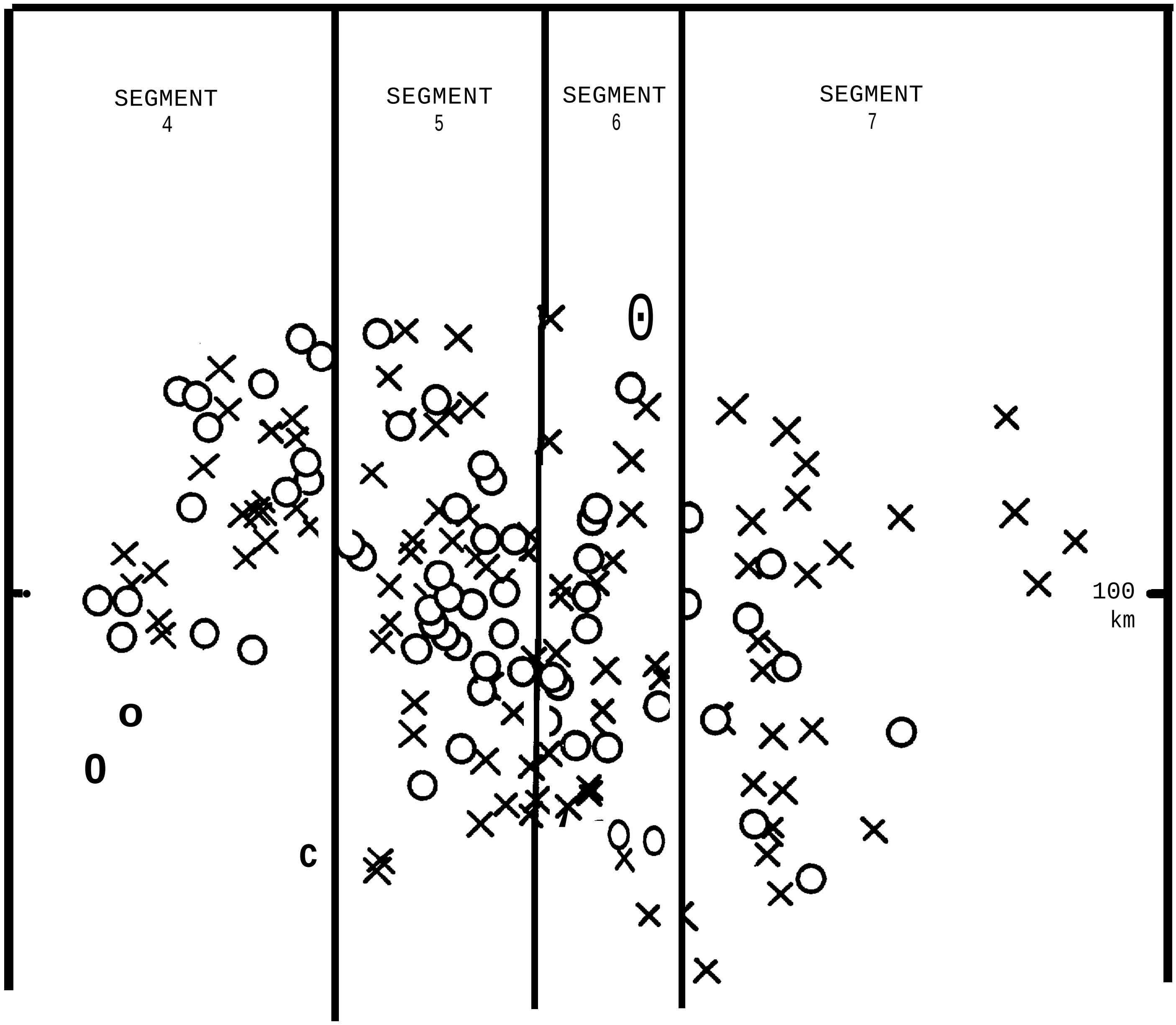

FIGURE III-3. Vertical cross section showing locations of cross sections in subsequent figures. Vertical lines separate segments 4, 5, 6, and 7, showing the events plotted in the cross sections in Figures III-4, III-5, III-6, and III-7.

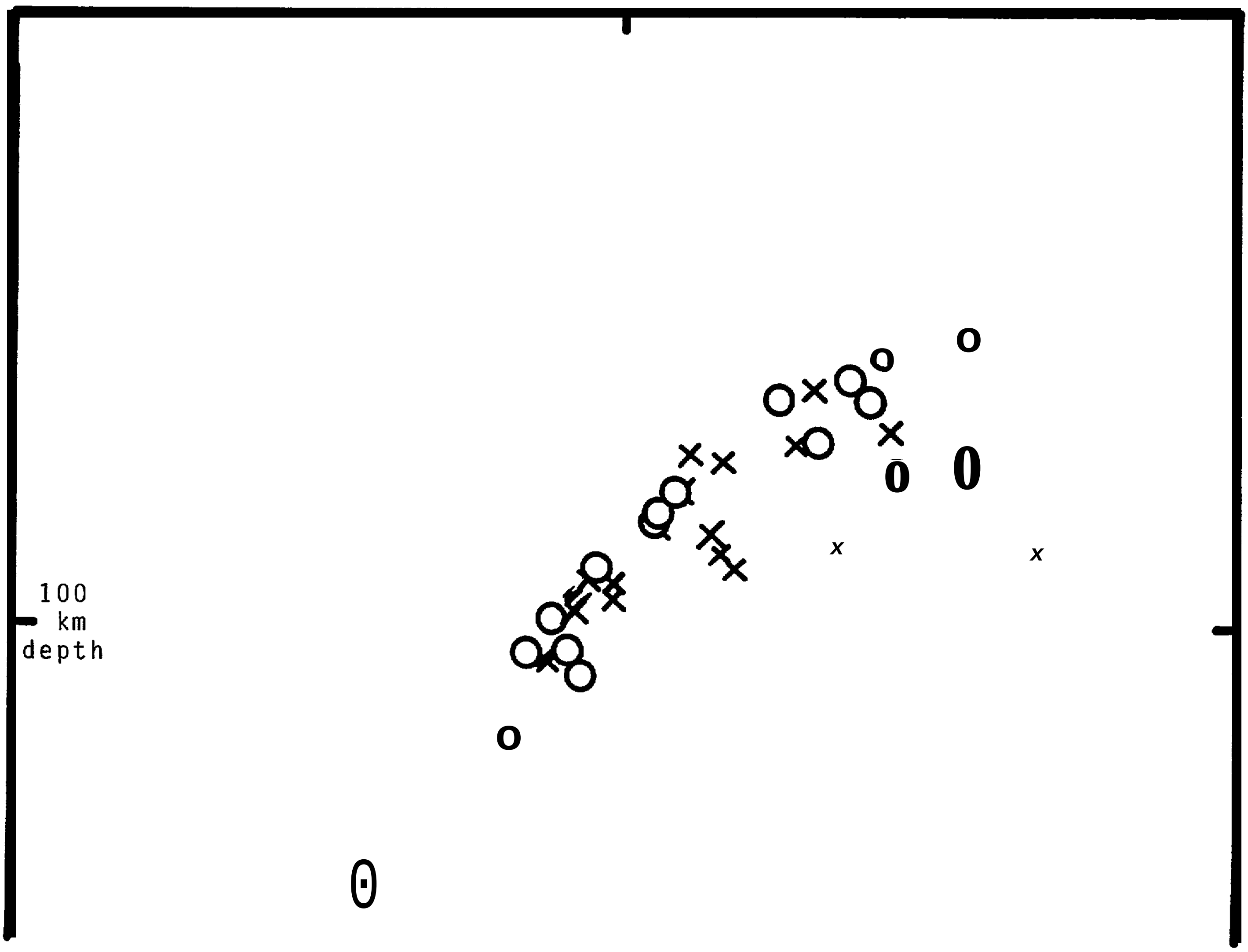

FIGURE III-4 Vertical cross section of segment 4 perpendicular to the Aleutian trench, and perpendicular to the cross sections in Figures III-2 and III-3 (azimuthal trend of 122°E of N) showing the events relocated in this study. See Figure III-3 for location of this cross section, and Figure III-1 for explanation of symbols.

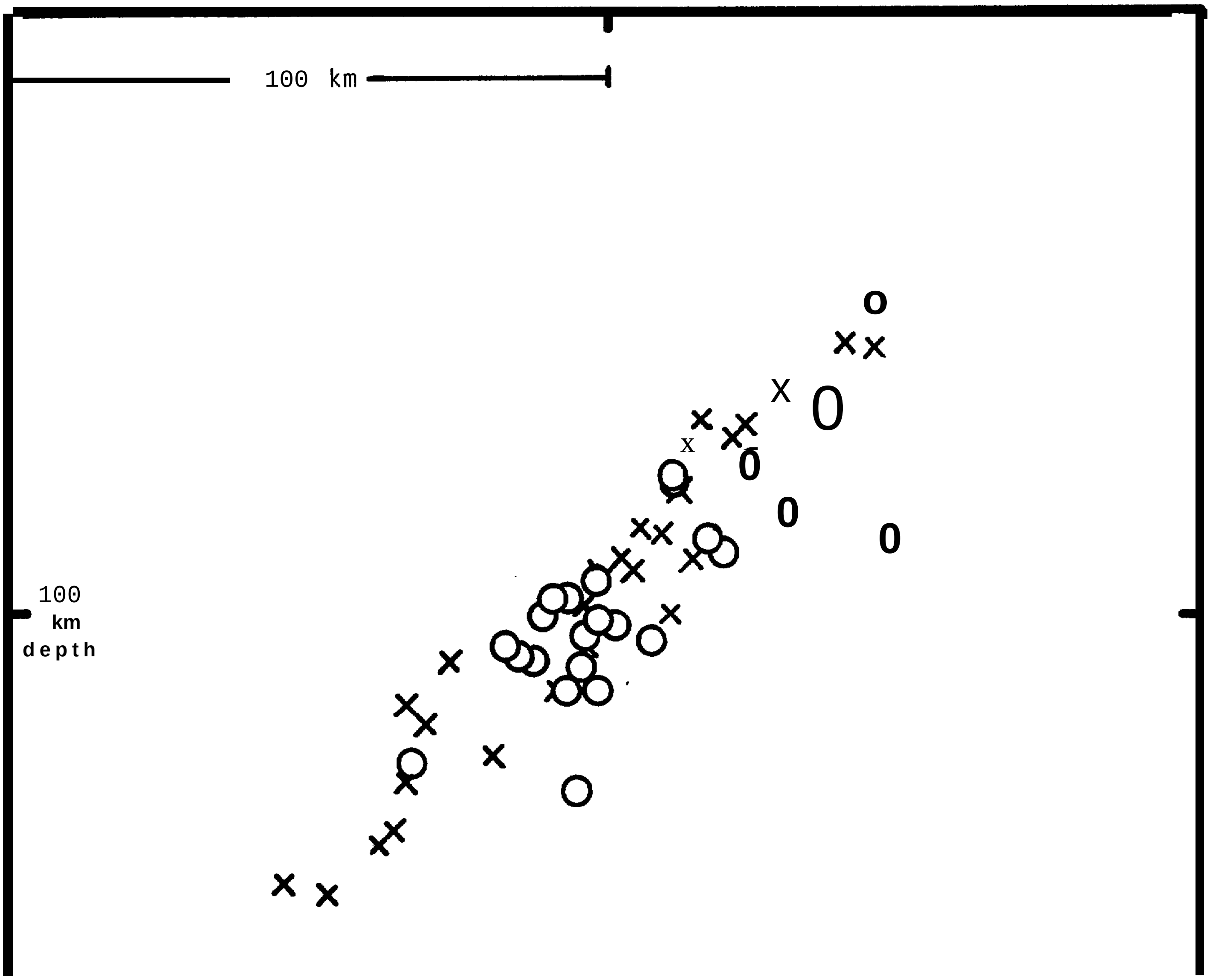

FIGURE III-5 Vertical cross section of segment 5 (see caption to Figure III-4).

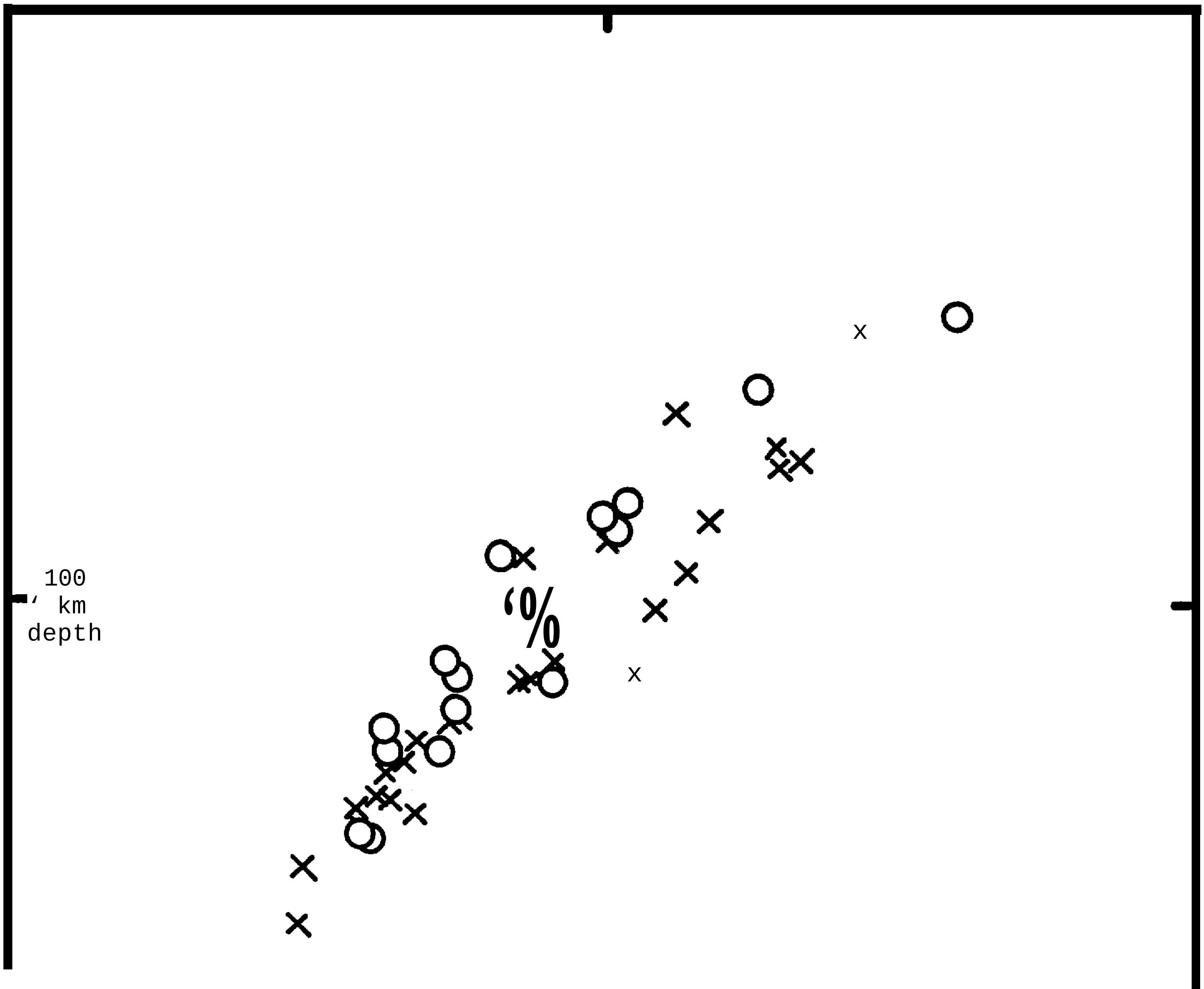

FIGURE III-6 Vertical cross section of segment 6 (see caption to Figure III-4).

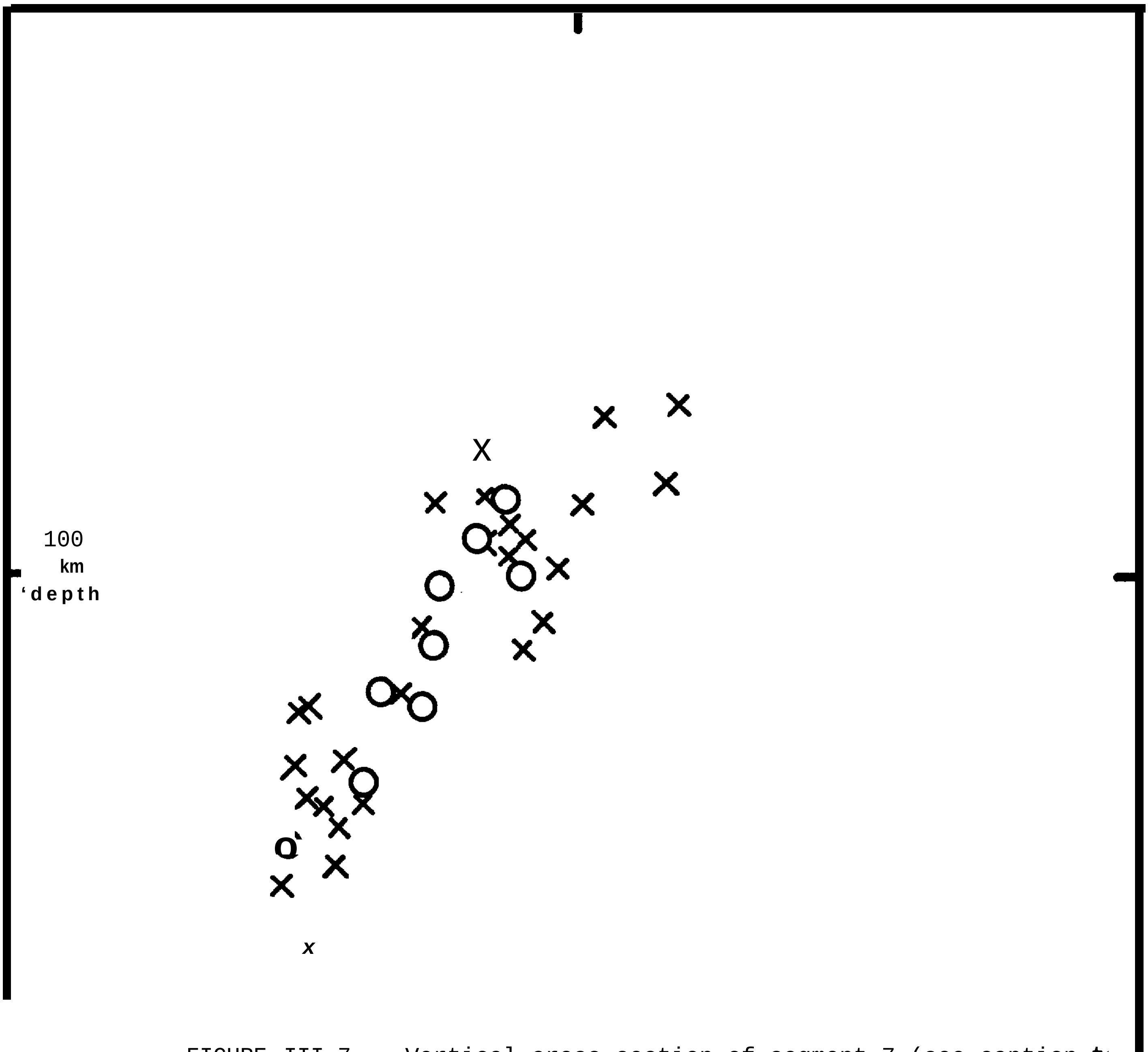

FIGURE III-7 Vertical cross section of segment 7 (see caption t
Figure III-4).

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SECTION IV: JOINT HYPOCENTER DETERMINATION OF TELESEISMICALLY RECORDED EARTHQUAKES ON THE KODIAK SHELF, ALASKA

by

Hans Pulpan and Cliff Frohlich

INTRODUCTION

Recent hydrocarbon exploration offshore of Kodiak Island (Fisher, 1980; Hanley and Wade; 1981) provides new motivation for evaluating the potential for large earthquakes in this area. High seismic risk has been associated with the Kodiak region for at least 200 years (Hansen and Eckel, 1971). In 1964 Kodiak Island was strongly affected by the great Alaskan earthquake (Plafker, 1972), one of the largest earthquakes ever recorded. Because of the spatial limitations of land networks (local and teleseismic) in determining accurate earthquake locations at convergent margin continental shelves, it is desirable to relocate these events using the best methods available.

In the present study we have relocated a number of teleseismically recorded earthquakes that occurred offshore of Kodiak Island since 1964. Most of these events can be considered to be aftershocks of the 1964 Good Friday earthquake. The work reported here is an extension of the work of Lawton et al. (1982), who investigated tha seismicity of this region using data from ocean bottom seismograph stations and the network of land stations operated by the University of Alaska (Pulpan and Kienle, 1979).

METHODS

Difficulties in Relocating Teleseismic Earthquakes

Precise determination of earthquake hypocenters in the Kodiak. shelf area is difficult for several reasons. The seismicity of the area occurs at shallow focal depths, and reliable depth values for shallow events are notoriously difficult to establish unless the depth of the events is comparable to the distance to the closest observing stations. Thus the land based network operated by the University of Alaska on Kodiak Island (Pulpan and Kienle, 1979) will not provide reliable depth estimates for shelf events (Lawton et al., 1982). With the area of interest lying outside that network, the poor azimithal station coverage results sometimes in rather poor constraints on the epicentral parameters.

In addition, teleseismically determined epicentral parameters are affected by the high velocity subducting slab upon rays that travel through it for a considerable distance. A further problem is the uneven azimuthal distribution of recording stations, with the Pacific Ocean producing a large gap. Reasonable depth estimates for shallow events can only be achieved with the help of depth phases, but routinely reported teleseismic depth phases are subject to considerable error for such events (Forsyth, 1982).

Depth Determination

All depth values given by the ISC for the events studied here were based on reported depth phases. The large scatter in depth seen in Figure IV-1, which represents a projection of the events onto a vertical plane striking perpendicular to the trench axis, may be due to phase misidentification.

Mistaking pwP and pP does not constitute a problem in this study, since the results of the relocation indicate that the shallow water depth associated with the events (less than 1 km) causes pP and pwP to arrive at nearly the same time on short period records. However, pP may be buried in the coda of P, especially in the case of larger events. In addition the complexity of upper crustal structure near either the source or the receiver can cause secondary phases easily mistaken for true depth phases. Differentiating pP from SP can also present problems if no additional information about the event is available.

Precise depth determination is of considerable importance for the identification of seismic source zones for purposes of seismic hazards assessment. We therefore scrutinized many teleseismic records for each of the 34 events in search of well defined depth phases. An additional three shallow events (events 35 - 37 of Table IV-1) located somewhat outside the source area of the relocated events were also investigated since bulletin reports indicated consistent depth phases. In the case of 28 of the above events we could not interpret the later arrivals as depth phases in any convincing fashion. Reported late arrivals for two events (numbers 37 and 33) were interpreted as foreshock-mainshock sequences, leaving only eight that permitted depth determination based on secondary arrivals.

Figure IV-2 shows some records for an event that occurred on 22 April 1966 at 57.37 N, 152.27 W (event 37). A very clear phase arrives about seven seconds after P at many US and some European stations. The amplitude ratio between the two phases is very similar for a wide range of azimuths on both short period and long period records, suggesting that the two phases are a foreshock-mainshock sequence. In the absence of a focal mechanism solution for this event the above interpretation is here however to be considered tentative.

Figure IV-3 shows an example of an event that occurred on 22 August 1973 at 57.09 N, 154.12 W. This event is representative of those events where a depth determination could be made with reasonable confidence. A prominent phase arrives at several European stations about 10 seconds after P.. A weaker later phase can also be observed on some of these stations approximately 15 seconds after P. This later phase is very prominent at some US stations and one Asian station, These phases can be interpreted as pP and SP respectively, and correspond to a depth of 36 km for this event.

The depths calculated from later arrivals are based on a

KODIAK SHELF - JHD LOCATIONS

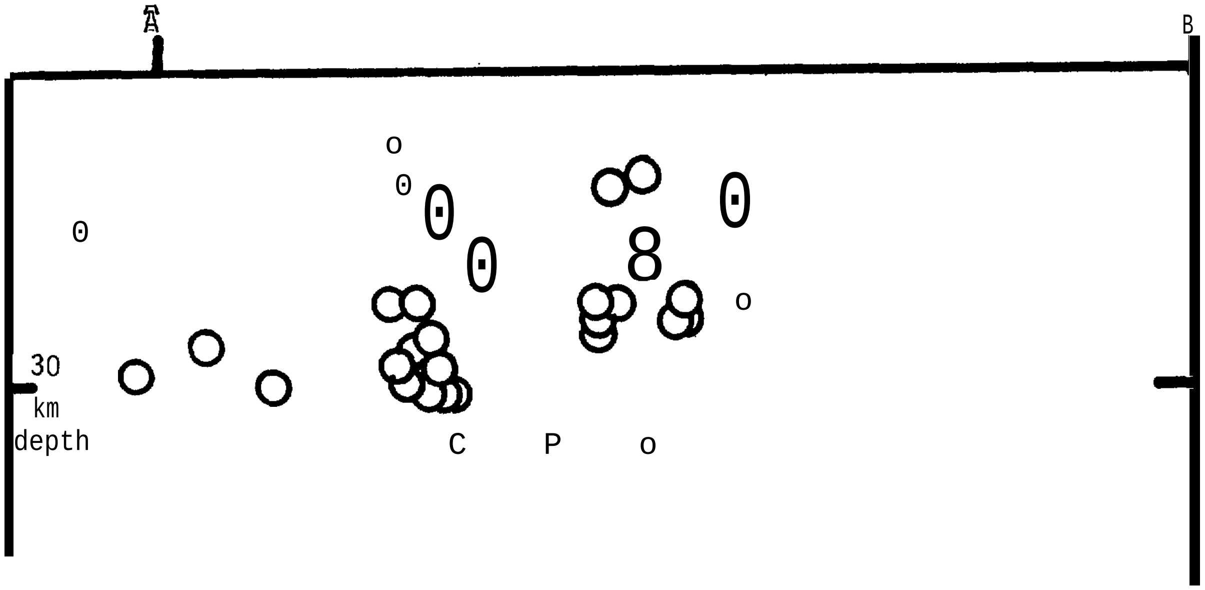


FIGURE IV-1: Vertical cross section of the events relocated by the JHD method in this study. Depths were determined from late arrivals reported in the ISC bulletins, and in some cases from inspection of seismograms recorded at selected WWSSN stations.

TABLE IV-1: Locations determined by the JHD method of 34 events relocated in this study. In addition, the table reports two additional events (#35 - 36) for which reliable depth phases occur as determined from seismograms recorded at WWSSN stations.

NUMBER	DATE	LATITUDE	LONGITUDE	DEPTH	ORIGIN TIME	MAGNITUDE	RELIABLE DEPTH PHASES
1	30 Mar 1964	56.59 N	153.00 w	22.0	0218:05.28	5.8	
2	30 Mar 1964	56.64 N	152.27 W	18.0	1609:26.99	5.7	
3	04 Apr 1964	56.60 N	152.75 W	19.0	0840:29.82	5.3	
4	04 Apr 1964	56.92 N	153.01 w	14.0	0910:54.6	5.8	
5	05 Apr 1964	56.35 N	153.48 W	30.0	0122:13.90	5.6	YES
6	05 Apr 1964	56.28 N	153.60 W	27.0	0141:43.06	5.4	
7	12 Apr 1964	56.67 N	152.33 W	22.0	0124:30.41	5.8	
8	16 Apr 1964	56.51 N	153.04 w	25.0	1926:56.08	5.5	
9	17 Apr 1964	56.53 N	153.00 w	14.0	0449:28.20	5.5	
10	06 May 1964	56.64 N	152.27 W	15.0	1526:35.99	5.5	
11	12 May 1964	56.64 N	152.35 W	11.0	1816:42.10	5.4	
12	27 Sep 1964	56.60 N	152.10 W	21.0	1550:53.41	5*4	
13	23 Jun 1964	56.64 N	152.79 W	31.0	1109:15.10	5.7	
14	22 Jan 1966	56.02 N	153.98 W	34.0	1427:07.09	5*7	YES
15	08 Apr 1966	56.69 N	152.65 W	31.00	2210:56.07	5.0	
16	11 Apr 1966	56.65 N	152.16 W	24.00	2300:22.41	5.0	
17	09 May 1967	56.53 N	152.60 W	22.00	1236:36.24	5.0	YES
18	13 May 1967	56.51 N	152.71 W	23.00	0518:54.21	5.0	

NUMBER	DATE	LATITUDE	LONGITUDE	DEPTH	ORIGIN TIME	MAGNITUDE	RELIABLE DEPTH PHASES
19	29 Jan 1968	56.32 N	153.51 w	6.0	2052:20.75	5.2	
20	22 Dec 1968	56.39 N	154.07 w	29.0	1644:43.81	5.4	YES
21	20 Nov 1969	56.61 N	153.23 W	30.0	2346:10.57	5.2	
22	21 Nov 1969	56.31 N	153.56 W	22.0	0014:10.71	5.1	YES
23	24 Nov 1969	56.17 N	153.77 w	28.0	2251:48.47	5.4	
24	12 Jan 1970	56.71 N	152.13 W	35.0	0454:31.93	5.3	
25	20 Sep 1971	56.45 N	153.15 w	30.0	0644:13.65	5.1	
26	18 Jan 1972	56.75 N	153.12 W	26.0	0017:44.95	5.1	YES
27	01 Aug 1974	56.56 N	152.36 W	23.0	0555:36.11	5.1	
28	01 Aug 1974	56.55 N	152.58 W	22.0	0759:54.92	5.1	
29	01 Aug 1974	56.56 N	152.45 W	10.0	0507:59.81	5.3	
30	07 Aug 1974	56.62 N	152.75 W	35.00	0823:36.26	5.0	
31	22 Ott 1976	56.17 N	153.42 W	24.0	1835:25.19	5.5	
32	10 Aug 1977	56.63 N	152.89 W	28.0	0935:57.55	5.1	
33	12 Apr 1978	56.62 N	152.87 W	10.0	0342:03.51	5.7	
34	12 Apr 1978	56.52 N	152.43 W	22.0	0522:29.90	5.0	
35	11 Mar 1970	57.39 N	153.97 w	42.0	2238:32.4		YES
36	22 Aug 1973	57.09 N	154.12 W	36.0	1814:36.6		YES

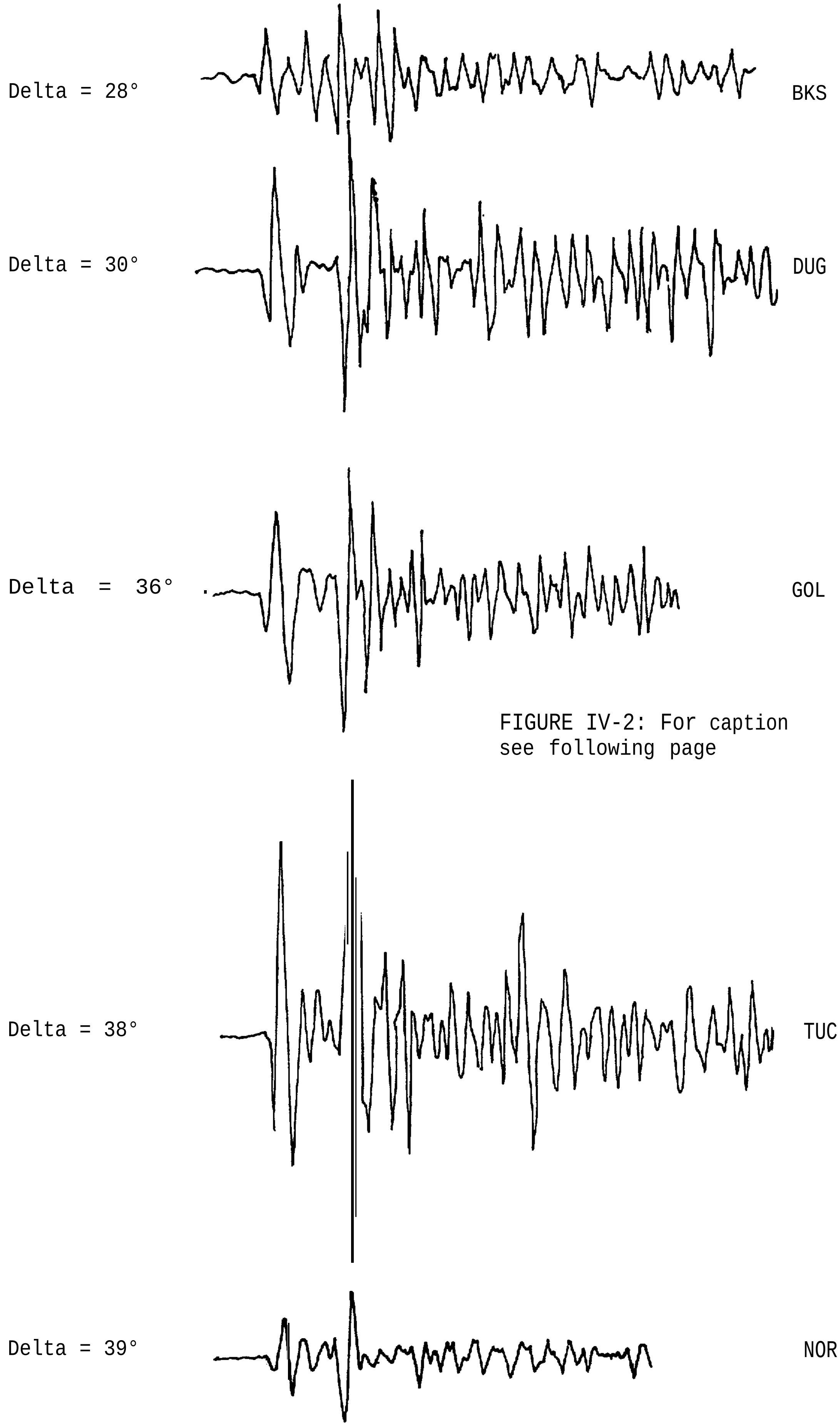


FIGURE IV-2: For caption see following page

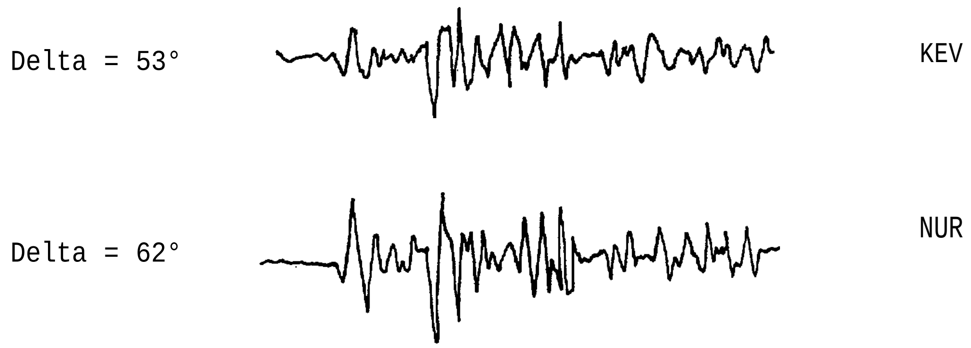


FIGURE IV-2: Short period seismograms (vertical component) at several WWSSN stations for the event which occurred on 22 April 1966 (see Table IV-l). Note the clear phase arriving on most records about 7 sec after the initial P arrival.

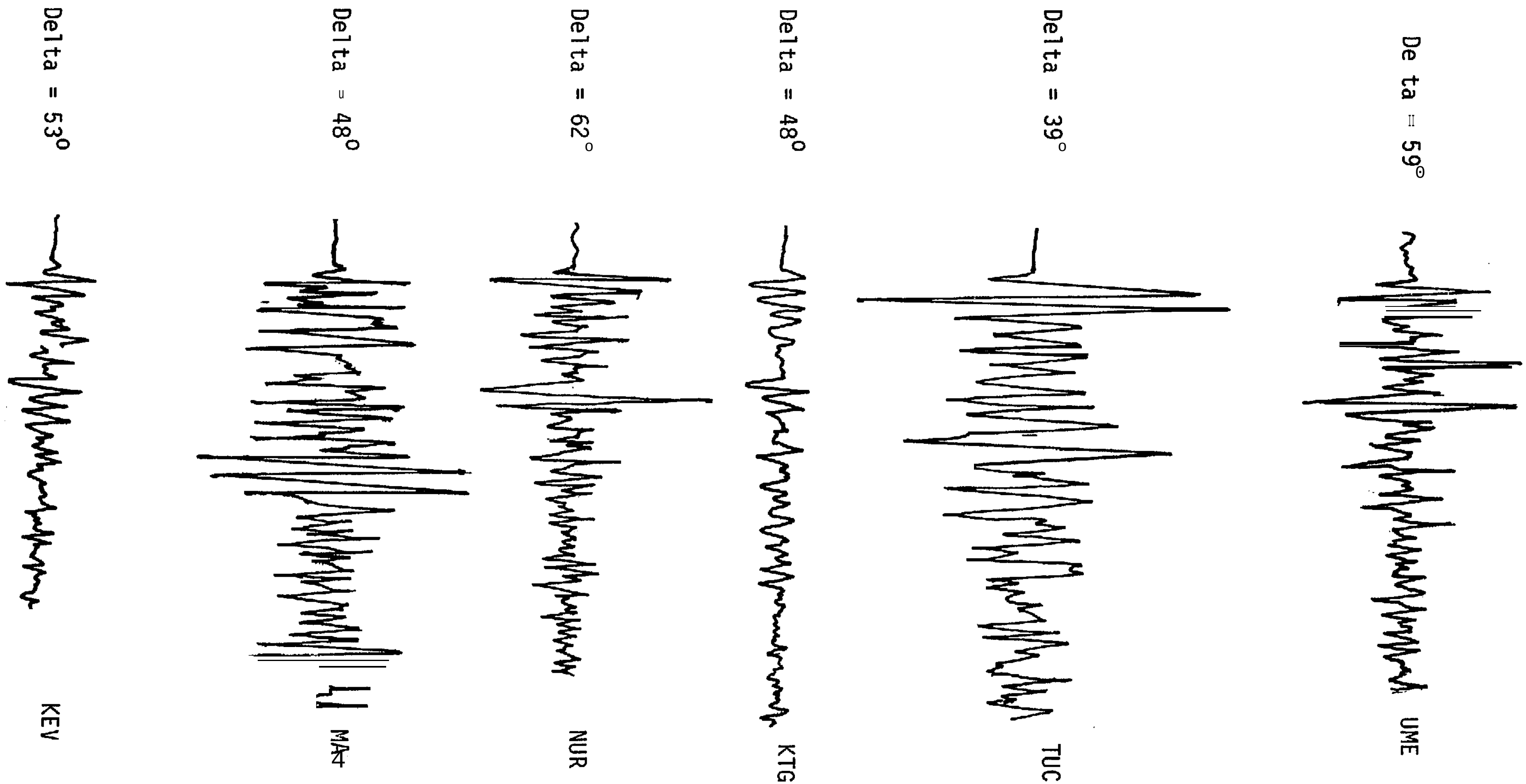


FIGURE IV-3: Short period seismograms (vertical component) at several WWSSN stations for the event which occurred on 22 August 1973 (see Table IV-l).

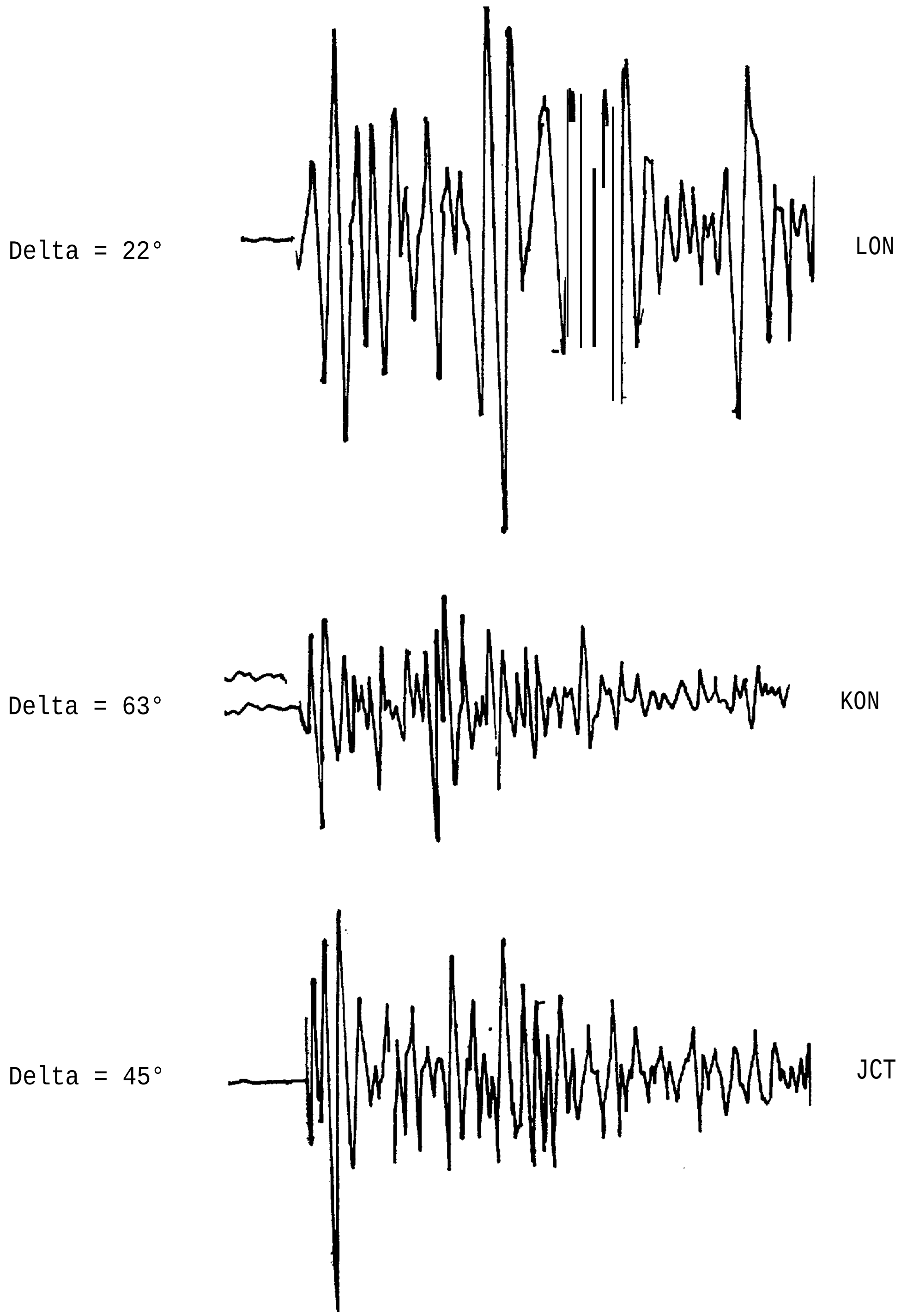
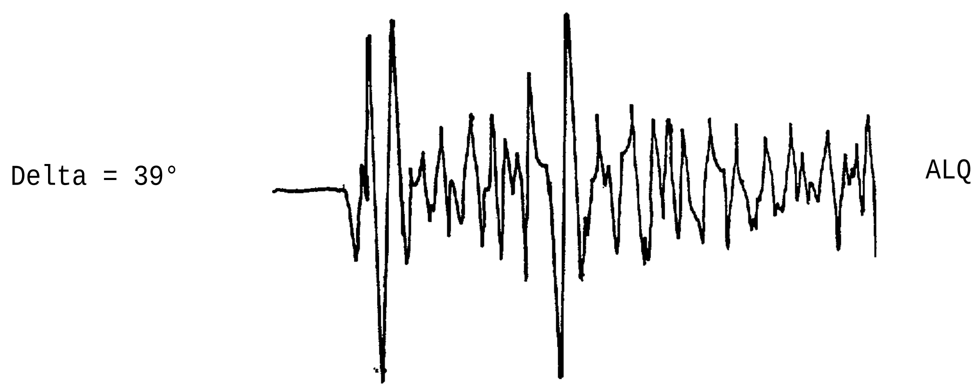


FIGURE IV-3: For caption see preceding page.



Jeffreys-Bullen earth structure and thus do not take into account differences between the J-B structure and the local crustal structure. However, the resulting error is within the limit of accuracy with which later phases can be timed, and the depths would be shifted only a few km if an appropriate local structure were used. The depths derived here are likely to be more reliable than those reported in the ISC bulletins.

If viewed in cross section (Figure IV-4) there is a much smaller scatter apparent in the depth distribution. This is due in part to the smaller number of events, since a few events are considerably shifted from their reported ISC depths. Our locations are the result of a more rigorous approach towards assessing and relocating teleseismic recorded events than has been used routinely in this area previously. Thus the present locations provide a better basis for speculating on the seismotectonics of the area.

Relocation by the Joint Hypocenter Determination Method

There exist several methods which attempt to improve earthquake locations. Three dimensional ray tracing (e.g., Engdahl and Lee, 1976) can directly account for lateral inhomogeneities and provide accurate locations if the three dimensional velocity structure is well known a priori. The large number of calculations associated with this method, however, prevents its extensive use.

The Joint Hypocenter Determination (JHD) method (Douglas, 1967; Dewey, 1972; Frohlich, 1979) solves for hypocentral parameters and station corrections simultaneously for a group of earthquakes under the assumption that the station corrections are the same for each event of the group. This requires the events to be distributed over a limited volume. In this case JHD will generally provide a considerable improvement in the relative location of the events.

We applied the JHD method of Frohlich (1979) to a group of 34 earthquakes covering an approximately 100 X 150 km area (Figure IV-5). Body wave magnitudes of the events ranged from 5.0 to 5.8. Ten WWSSN stations at teleseismic distances were selected to provide good azimuthal coverage (Figure IV-6) and also clear arrivals from the lower magnitude events. The azimuth of one of these, COL, is such that rays to it will travel to a large extent through the high velocity subducting slab. To somewhat offset this the station ADK was used. Rays to ADK also will travel predominantly through the slab which undergoes an approximate 60 degree azimuthal bend in the Kodiak-Lower Cook Inlet area. The station KDC, operated by NOAA, was used as a nearby station (distance about one degree). Thus a total of 12 stations was used for relocation purposes (Table IV-2). About 80 per cent of the arrival times were reread by the authors. Only where copies of the original records were unavailable did we use readings reported from the bulletins.

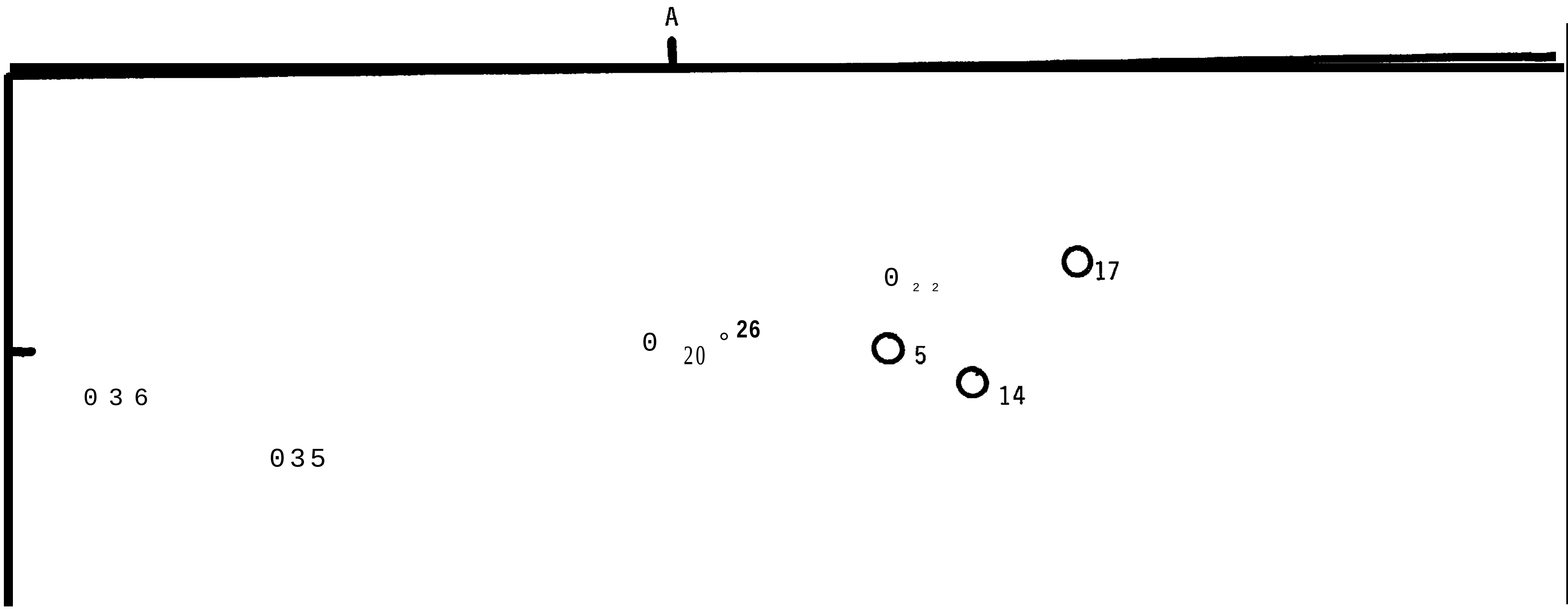


FIGURE IV-4: Vertical cross section of the events for which depths were determined from inspection of seismograms recorded at selected WWSSN stations. Numbers adjacent to each event are the event numbers in Table IV-1. Points A and B are as in Figure IV-1

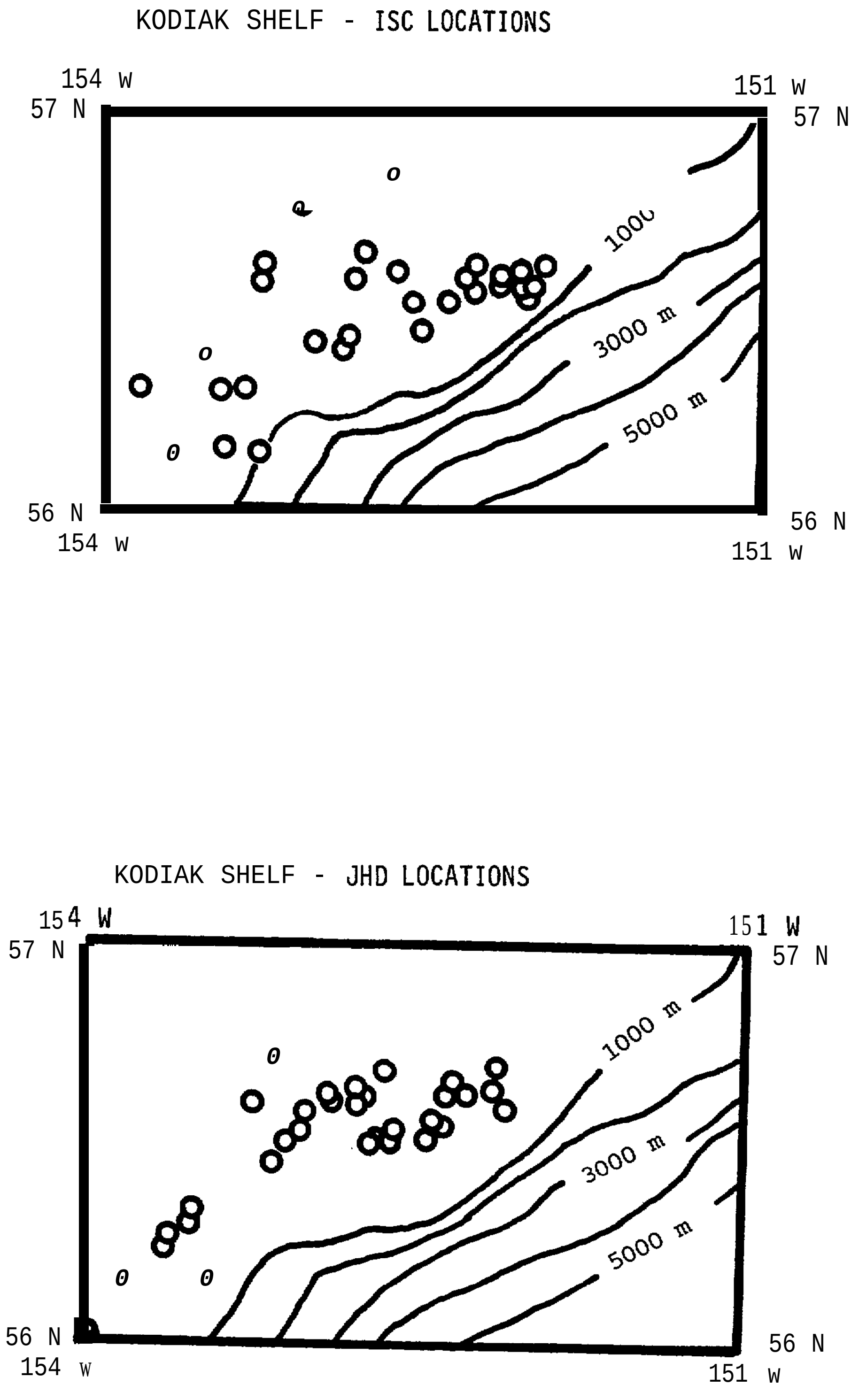


FIGURE IV-5: Map showing the location of 34 earthquakes relocated in this study using the JHD method. Top - locations reported by the ISC; Bottom - same events after relocation by the JHD method.

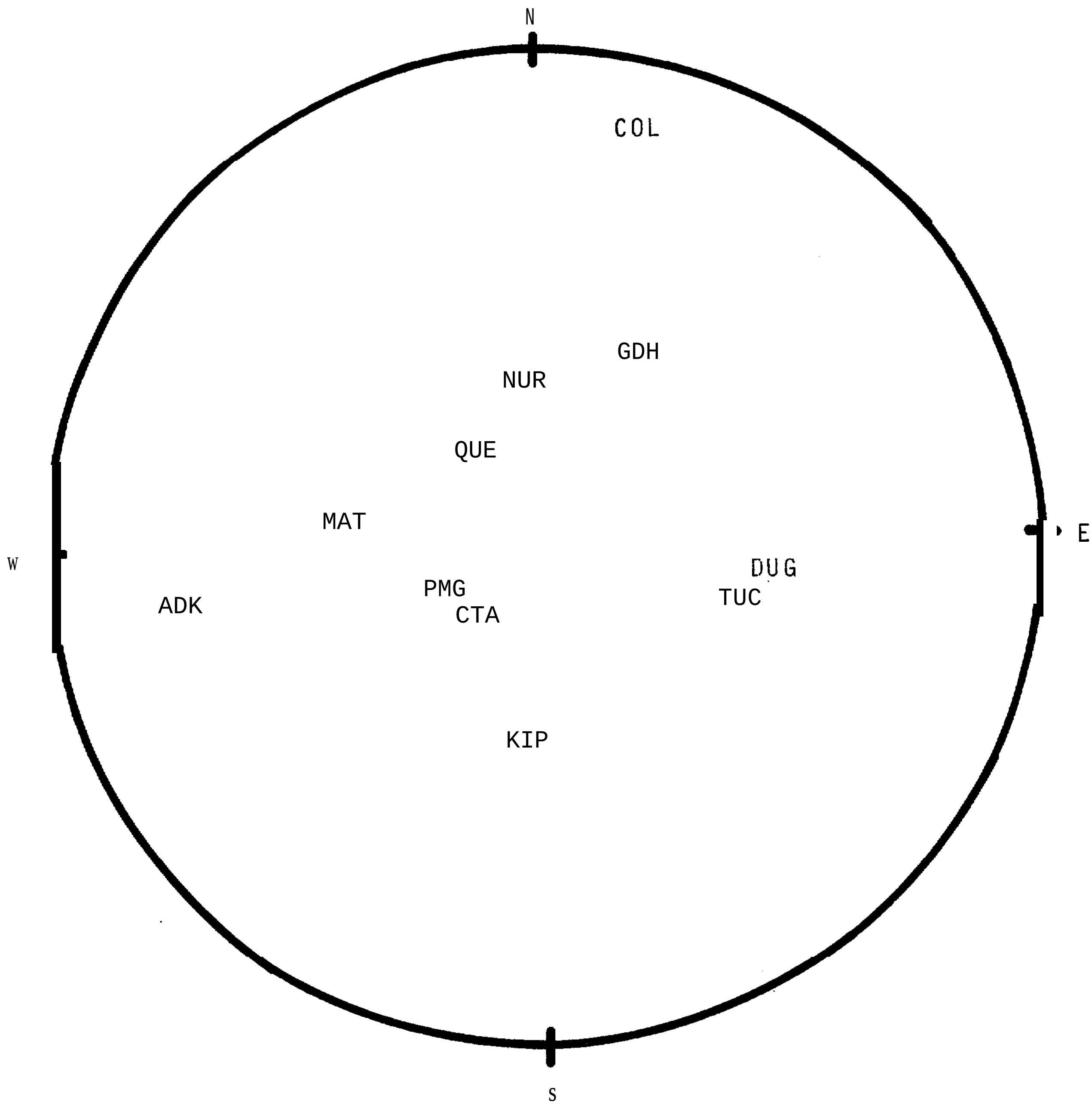


FIGURE IV-6: Distribution on a focal sphere of the stations used in the JHD relocation. One station (KDC) is not shown because its position on the focal sphere depends critically on the velocity model used.

TABLE IV-2: Stations used for JHD relocation of the 34 events in Table IV-1, and stations corrections determined during the JHD relocation.

STATION CODE	CORRECTION
KDC	1.19 sec
COL	-1.81
ADK	-2.73
NUR	0.10
GDH	1.54
TUC	-0.25
DUG	-0.98
KIP	0.06
MAT	-0.33
PMG	0.30
QUE	1.65
CTA	1.27

RESULTS AND DISCUSSION

Upon relocation by the JHD method, the originally diffusely distributed events collapse into two fairly narrow, subparallel groups of events, with only a few events falling outside these groups (Figure IV-4). Before speculating about the implications of this, we will discuss the probably effect of systematic errors in location which affect all the events in the same fashion.

While the relocation provides reliable relative locations for these events, the locations may still be biased as a whole. Four of the relocated events were teleseismically recorded as well as recorded by the UA regional network. The location of these three events is on the average 22 km to the south and 3 km to the west of the corresponding JHD locations. This result is in agreement with the results of Lawton et al. (1982), who compared locations in this area determined by a combined OBS-land-based network with ISC locations.

The OBS study also indicated that well recorded events which are located using stations of the landbased network shift only slightly when data from a combined OBS-land network is used for location. Thus the actual location of the relocated teleseismic events is probably about 20 km south from the locations in Table IV-1. Such a shift puts the events approximately along the shelf break (1000 m bathymetric contour).

Hampton et al. (1979) map a series of faults along the shelf break, some of them offsetting the seafloor by as much as 10 m. There is probably no direct relationship between the relocated events and these faults. The rigidity of the accreted sediments is probably too low to permit brittle fracture at the scale indicated by the magnitude of the events. Focal mechanism solutions available for 11 of the relocated events (Stauder and Bollinger, 1966) all indicate these events to be associated with thrusting on a gently dipping plane. Although the ambiguity in first motion fault plane solutions also permits dip slip faulting on a steeply dipping plane, the direction is opposite to that generally inferred along the imbricate faults of the accretionary wedge, which appears to be aseismic (Chen et al., 1982). However, the relationship between the activity along the main thrust and observed faults might be along the model suggested by Fukao (1979).

The majority of the events relocated in this study are part of the aftershock sequence of the 1964 Good Friday Alaskan earthquake. The relatively narrow belt of aftershock seismicity evident from our relocations probably reflects the release by brittle fracture of stress concentration penetrating the leading seaward edge of the shallow dipping rupture zone associated with the main shock. Since according to Fukao's (1979) model, the main rupture did not propagate further towards the trench along the interface between the overriding and underthrusting plates, the wedge-shaped region of sediments between the trench axis and the leading edge of the rupture zone will be heavily stressed. Thus stresses will be relieved mostly in a ductile manner along slip lines bending from the tip of the rupture zone upwards and emerging at steep dip angles from the ocean floor, producing the mapped seafloor offsets noted by Hampton et al. (1979).

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