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Source Process of Deep and Intermediate Earthquakes as Inferred from Long-Period P and S Waveforms

1. Intermediate-depth Earthquakes in the Southwest Pacific Region

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Abstract

The source process of four intermediate earthquakes with magnitudes of 6.0-6.8 and focal
depths between 100 and 200km has been re-investigated from the analysis of long-period P and S waveforms.
The source process times recovered from the times of the first half cycle of recorded P waves or the group delay times derived from the slopes of equalized phase spectrum show an azimuthal dependence with respect to the orientation of a nodal plane and of the null vector. If we assume shear dislocation models for these earthquakes, the dependence yields a solution to the problem of which of two nodal planes corresponds to the slip plane. Various source parameters have also been estimated from the azimuthal dependence by least squares technique.
The dimension of the slip plane or the fault length and width range in 25-40km and 8-18 km, respectively, and the rise time of dislocation is found to be about 1sec. The rupture velocity might be as low as 3.2km/sec, if two-dimensional propagation is assumed.
The theoretical seismograms of both direct P and S waves appropriate to each recording station have been synthesized on the basis of the estimated source parameters, taking into account the combined effects of wave propagation in the mantle and the crust and of the seismograph response. A good agreement of general features between the observed and synthesized waveforms on the three component seismograms gives support to the above
slip dislocation model. Comparison of the amplitudes on the both kinds of seismograms

yields seismic moment of the order of 1.6-3.0×1026 dyne・cm and the amount of dislocation

of 80-140cm. The stress drop at these earthquakes ranges from 50 to 90 bars but might
exceed 170 bars for one shock. The effective initial stress to produce shear dislocations is
also estimated in relation to the frictional stress on the slip plane.

§1. Introduction

Recent seismic observations covering a low
frequency range and the introduction of elastic
dislocation models have provided a powerful
approach to clarify the dynamical process at
the source not only of large shallow earth-quakes, but also of deep-focus and inter-
mediate-depth earthquakes that occurred be-
tween 100 and 600km (TENG and BEN-MENA-HEM, 1965; BOLLINGER, 1968; BERCKHEMER and JACOB, 1968; MIKUMO, 1969; FUKAO, 1970). It is expected that if the source of all these earthquakes can be interpreted in terms of
shear dislocations, the stress-strain drop at
the source, shear stresses causing these earth quakes, and their variations with depth will be estimated, and that the strength and

thermal state of material in the mantle and their implications to the tectonic structure of a deep seismic zone, such as descending slabs of the lithosphere (ISACKS et al., 1968; MC-KENZIE, 1969), and the cause of deep and
intermediate earthquakes might be brought
into the light.
It appears that there still remain unsolved problems on the observational side in an ap-plication of shear dislocation models to deep earthquakes. The problems are; whether the source is a moving dislocation, or at what
velocity dislocations spread out over a slip
plane formed in the mantle with increased
temperature and pressure; and whether the
radiation pattern of S waves including the
absolute amplitudes and waveforms can be

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explained by a dislocation source with the
same parameters as for P waves.
The purpose of the present paper is to
elucidate more clearly the source process of
intermediate-depth earthquakes, by making
direct approach to the problems. The first step is to make an attempt to estimate the
source parameters including the dimen-
sion, the amount of dislocation and the rup-
ture velocity, from azimuthal variations of
recorded P waveforms. The second approach
is to synthesize theoretical seismograms of
both P and S waves from dislocation sources
with the estimated parameters, taking into
account the combined effects of wave pro-
pagation in the earth and a recording instru-ment, and to compare them with their ob-
served waveforms on the three-component
records, in order to test the appropriateness
of the assumed model. The stress drop at the time of these earthquakes and the initial stress to produce shear dislocations are dis-cussed in some detail.

§2. Data

The earthquakes analyzed here are four intermediate shocks that occurred along island arcs from New Guinea to the Fuji Islands at depths between 100 and 200km, which have been treated in a previous paper (MIKUMO, 1969). Relevant information is again listed in Table 1. Data used are the long-period vertical and two horizontal component seismo-grams recorded at the WWSSN stations.

§3. Analysis

In the previous study (MIKUMO, 1969), theo-retical seismograms of direct P waves have been synthesized from moving dislocation models with variously assumed source para-

meters through a linear filter system com-posed of the earth and instrument, and com-pared with the corresponding records. The comparison has shown that the average wave-form on the records is similar to that from a double-couple point source with a ramp

function time dependence with τ<3sec, and

yielded the seismic moment and probable ranges for some other parameters. Closer examinations reveal, however, some difference in the recorded P waveforms for the first half-period as well as in the slope of the equalized phase spectrum (MIKUMO 1969), depending on the location of recording stations with respect to the source. For this reason, an attempt is made here to evaluate the pos-sible azimuthal dependence of the above two quantities. Although the waveforms are predominantly controlled by the seismo-graph characteristics, we assume that there are no serious differences in its frequency response among the stations. This assumption may be justified by the fact that the shape of calibration pulses placed on seismograms is almost identical at these stations.

(1) The slope of the phase spectrum (or the group delay time)
The equalized source phase spectrum of direct P waves has been given in Fig. 19 in the previous paper, for all stations in the four earthquakes mentioned. It can be seen that there are appreciable differences in the slope of the spectrum, and a possibility has been suggested that the source dimension could in principle be evaluated from the slopes if the initial phase correction is made for a spherical earth. The correction gives a con-stant term independent on frequency on the standpoint of normal modes (BRUNE, 1964; SATO, 1969), and hence yields no effects on the

Table 1. Information of the earthquakes analyzed.

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slope of the phase spectrum. The corrected
source phase φ(ω) in the case of unilateral
faulting may be written (MIKUMO, 1969),

φ(ω)=φU(ω)+φS(ω)=ω τ/2+ω(TL+TW)

and

(1)

where τ is the time constant of dislocation,
L and W are the length ad width of the
slip plane, υp and υ are the compressional
wave velocity and rupture velocity, and αj and
αk are the direction cosines of the relevant
ray relative either to the slip direction or to
the null vector, respectively. The slope of
the phase spectrum dφ/4ω or the group delay
time is then related to the azimuth of the
station ad to the source parameters in the
following form,

(2)

The left-hand side can be determined from
observations. Eq. (2) indicates that L/υp, W/υp
and τ±L/υ may be evaluated directly from
the group delay times at more than three
stations.
In the upper part of Figs. 1(a) (earthquake
No. 23) and 1(b) (earthquake No. 15) are
plotted the group delay times determined from the equalized phase spectra over the
frequency range between 0.01 and 0.15c/s.
There appears to be some linear dependence
of the plots with the azimuth to a possible
slip direction in the earthquakes, but with
rather large scattering. The reason for this
might be be accounted for; i) The assumed
Crustal structures are not always appropriate
for each station (the slopes have been cor-

Fig. 1. Group delay times, observed first half-periods, and source process times. (a) No. 15 (b) No. 23

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rected for a standard structure of HASKELL (1960).). A difference of 10km in the crustal thickness yields the change of about 1.5sec
in the group delay time. ii) The corrected
phase spectra do not change monotonously
with frequency; and iii) The initial time of digitization might not exactly agree with the arrival time of first P waves when they have a blunt beginning. It would therefore be rather difficult to rely on the group delay times in a precision determination of the source parameters.
(b) Source process time
Another method is to infer from observed

records the apparent time duration of energy release at the source region or the total time of faulting process, which is specified by the source finiteness and the rise time of disloca-tion. The displacement of P waves ur0(t) at great distances from dislocation sources may be expressed by (MIKUMO, 1969),

(3)

(4)

where

(5)

(6)

(7)

It is clear from eqs. (5) and (7) that each of
the deconvolved functions, Δu(t), SL(t), and
SW(t) is a rectangular pulse with the width
of τ, 2TL and 2TW respectively. s(t) is a
convolved pulse of SL(t) and SW(t), having a
form of trapezoid specified at the time interval
0〓t〓2 (TL+TW) (HIRASAWA and STAUDER,
1965). ur0(t) is a convolution of Δu(t) with s(t), so that this has such a form as schema-
tically shown in the lower part of Fig. 2.

Fig. 2. Relation between first half-periods and
the source process times.

The width of the base of the pulse, hereafter termed as the source process time T0, is specified by,

(8)

As a next step, we shall determine T0 in
the time domain from observed records.In
the lower half of Figs. 1 (a) and 1 (b), the
times t of the first half-cycle of P waves,
which have been read directly on the vertical
component seismograms at each station, are
plotted by solid circles. We notice that there
are parallel alignments between t and dφ/dω,

and hence this shows that t can be used as a measure of T0. It is now necessary to find a relationship between t and T0. The pro-cedure emloyed here is similar to that used by BOLMNGER (1968, 1970). An empirical rela-tion is established here on a number of hypothetical seismograms, which have been constructed by the convolution of ur0(t) for variously assumed values of T0 with the im-pulse response h(t) of the earth-instrument system. Since the form of ur0(t) is not uni-

quely specified only by the total time of T0
but depends also on the partition of the time
to τ, 2TL and 2TW, so that computations have
been made for various combinations of the

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three parameters within their probable ranges.
Fig. 2 shows an example of the relation for
station HKC (Δ〓40°, Φ=314° in earthquake
No. 15), with a standard single-layered crust,
an average attenuation of T/Q=1.5 and with
instrumental constants in the conventional
system. The hatched part indicates a possi-
ble range of t for an input of T0. Although this kind of relation has to be derived for
each station, the above example may be ap-

plied, as a first approximation, to most of the stations, to recover T0 from t since differences in crustal structures and attenuations do not yield significant discrepancies in the wave-form. Two broken lines B-B and B-S show the relations derived by BOLLINGER (1970) for an incident box-car pulse and half-cosine pulse respectively. However, his relations, do not include the effects of attenuations and of the partition of the total time duration, and are

Table 2. Source process times and direction cosines

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Fig. 3. Source process times plotted against the directions of one nodal plane and of the null vector. (a) No. 12 (b) No. 15 (c) No.6 (d) No. 23

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not used here.
The open circles plotted in the lower part of Figs. 1(a) and 1(b) are the source process times T0 thus estimated from t on the in-
divisual records. It can be seen that the

plotted points show a good consistency with
24φ/dω with less scattering. For this and

the above stated reasons, we use the source process times obtained from this time domain analysis, rather than the group delay times from the frequency domain analysis to evalu-ate the source parameters.

§4. Estimation of Source Parameters

The source process times thus estimated at
a number of stations for the four earthquakes
are tabulated in Table 2, together with the
direction cosines of the station azimuth with
respect to the normals to two nodal planes
and to the null vector respectively. The
uncertainty of the process times due to mea-
surement errors and to allowance in the rela-
tion between t and T0 is of the order of 0.5-
1.0sec. It is found that T0 ranges from almost
zero to about 4sec and sometimes reaches 6
sec, although the average has been estimated
as 2-3 sec (MIKUMO, 1969). Since differences
in local crustal structures and in attenuations
make only a difference of To less than the
above uncertainty, we take the interpretation
that the rather large variations of T0 result
from azimuthally asymmetric source dimen-
sion and rupture propagation.
We shall now examine the possible azi-
muthal dependence of T0. The dependence
has been tested with respect to the three
directions given in the above table, to deter-
mine which of the two nodal planes corre-
sponds to the slip plane. It has been found
for each of the four earthquakes, as shown
in Figs. 3(a)•`3(d), that there are some linear
relations with considerable scatter between
T0 and either of the two directions (α1 or α2)
to the nodal planes and also between T0 and
the direction (α3) to the null vector. We con-
clude from the dependence that the nodal
plane including the two relevant directions
is the slip plane. The slip direction on this
plane can be inferred from the radiation pat-

tern of P wave first motions.
The next step is to evaluate four unknown

source parameters L, W, τ and υ. We also
do not know the direction of rupture propa-

gation at this stage. For this reason, we
assume here that the rupture starts from a
point and the front of it travels two-dimen-
sionally at a constant velocity ƒÒ as illustrated
in Fig. 4(a). This assumption may be under-
stood as bidirectional faulting (OTSUKA, 1964;
HIRASAWA and STAUDER, 1965; FUKAO, private
communication, 1970) in which the rupture
propagates with different velocities along the
longer and and shorter sides of the slip plane.
In this case, eq. (8) may be rewritten in a
more general form,

(9)

where Lj and Lk are the linear source dimen-
sions, and υj and υk are the apparent rupture
velocities, as defined by υj=υ/sinφ and υk=υ/
cosφ, along the j- and k-directions, either of

Fig. 4(a). Geometry for an assumed rupture process.

which corresponds to the slip direction and
the other does to the orientation of the null
vector. Eq. (9) indicates that there could be
some different types of T0～ αj,k diagram, as
can be seen in Figs. 3(a)～3(d).1) In case
of υj,k<υp, the slope ∂T0/∂ αj,k should be
negative if the rupture moves to the positive
direction of the j- or k-axis, and positive for
the rupture moving to the opposite direction.
2) In case of υj,k>υp; for stations with αj,k<
υp/υj,k<1, the slope should be negative as
in 1), but positive for stations with υp/υj,k<
αj,k<1, being symmetric as to the axis of
αj,k=υp/υj,k. The intercept time can take
either positive or negative values.

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Taking into account the above situations

and re-arranging the sign of αj,k for all sta-

tions, we determined three quantities Lj/υ ρ,

Lk/υp and ts=τ ±Lj/υj±Lk/υk by linear least
squares for each earthquake. The computed
results are:

where the j- and k-axes were taken as the
slip direction and the null vector respectively,
and δT0 shows the standard deviation of T0.

From the above, the source dimensions Lj
and Lk can be estimated with rather large
uncertainty by assigning an appropriate value
to the compressional velocity around the
source region. The estimated length L (the
longer one of the two length Lj and Lk) of
the slip plane ranges from 25 to 42 km, while
its width W (the shorter one of Lj and
Lk) is about 6 to 20km, being the order of
1/2•`1/4 of the length, except for earthquake
No. 12. The results show that the slip plane
has a longer dimension in the slip direction
for earthquake No. 15, but it does in the
direction of the null vector for earthquakes
Nos. 6 and 23. In the case of earthquake
No. 12, the dimension has a comparable order
in the two directions.
Once the dimensions were determined, the

amount of dislocation D can also be estimated
from the seismic moment or DLW, which has
been determined in the previous paper (1969;
See also Corrections, 1970) from comparison
between the double amplitudes of the recorded
and synthesized P waves. The estimated
amount is 70•`100cm for three earthquakes
and reaches 140cm for earthquake No. 15.
On the other hand, we cannot uniquely
determine τ and υ by the least squares. It
is possible, however, to estimate the probable
ranges of the two parameters from ts. Here
we suppose two extreme cases: in case (A),
the rupture is propagated along the longer
direction L of the slip plane(φ=π/2, υw=∞);
in case (B) the rupture is assumed to move
along the shorter direction W (φ=0, υL=∞).
Fig. 4(b) shows the relation between τ and
υL,W in the two cases for the four earthquakes

respectively, where two curves from one earthquake give the upper and lower bounds

Fig. 4(b). Relation between the apparent rupture velocity and the rise time of dislocation.

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that come from probable errors in the adopted
dimension. If all the four earthquakes occur
through the dynamical process with similar

time constants and rupture velocities, τ and
υL,W would fall in the hatched region bounded
by the above curves. Case (A) indicates a
region of 0.7<τ<1.4sec and 10<υL<14km/sec.
Case (B) shows rather wide range. If we
take the lowest value of τ as estimated in
Case (A), a region of 0.7<τ<1.8sec and 3.3
<υW<5.5km/sec sec will be bounded, except
for earthquake No.12. The lowest bound of
the relation for the last earthquake does not
go through the hatched region, but υW would
take similar values if τ exceeds 1.7sec. Using
these estimates for the two cases, we have
3.2<υ<5.1 and 13°<φ<29° from the relations
1/υ2=1/υL2+1/υW2 and tan φ=υW/υL, where φ
is measured from the shorter side of the slip
Plane. If we further impose a limitation that

the rupture velocity cannot exceed the shear
wave velocity (MANSINHA, 1964), their possible
values are 3.2<υ<υ8〓4.5km/sec and 13°<φ
<24° respectively.

The stress drop during these earthquakes
has also been estimated by using the obtained
source parameters. For a dip-slip displace-

ment along an infinitely long fault, Δ σ=
(16/3π)μ)D/W (AKI, 1966), while for an infinitely
long vertical fault with strike-slip displace-ment, Δ σ=(2/π)μD/W=(KNOPOFF, 1958), where
D) as defined by D=(π/4)Dm is the average dis-

placement over the entire fault surface. Earth-quake No.15 would correspond to the former case, and the latter may be applied to earth-

quakes Nos. 6, 12 and 23. The above equations assume, however, an infinitely long fault. It
may be more appropriate to consider a cir-

cular crack with a finite radius a. For this
type of dislocation, Δ σ=(7π/16)μD/a (ESHELBY,

Fig. 5. Fault-plane solutions.

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Table 3. Estimated source parameters.

1957; KEYLIS-BOROK, 1959). The stress drop
based on the last assumption (a was taken
here as a=√A/π) are included in Table 3.
The stress drop during earthquake No. 15
might reach 170bar, while the other three
earthquakes show similar values.
All the estimated source parameters of the
four earthquakes are given in Table 3. The
uncertainties in L and W have been reduced
on the condition that τ and υ take similar
values for the four shocks. Fig. 5 illustrates
their fault-plane solutions projected onto the
lower hemisphere of the Wulff net, in which
the slip direction is indicated by arrows.

§5. Comparison between Observed and Syn-
thesized Seismograms of P and S Waves

In the previous study, the synthesized seismograms of P waves have been compared with the vertical component records at 15 stations in each of the four earthquakes ana-lyzed here. There remains, however, an important problem whether observed S wave-forms including the absolute amplitudes could also be explained by the dislocation sources derived from P waves. It has long been recognized from observations that predominant periods of S waves are usually longer than those of P waves, but it does not always appear from theoretical work (HASKELL, 1964; HIRASAWA and STAUDER, 1965) that moving dislocation models could fully account for the observational results. In this section, theoretical seismograms of S waves, as well as of P waves, are synthesized on the basis

of the source parameters estimated in the preceding section, taking into account the effects of attenuation, and compared with their recorded waveforms on the three com-ponents.
The general solutions for the displacement
of P and S waves from moving dislocation sources have been given by MARUYAMA (1963), HASKELL (1964), and BURRIDGE and KNOPOFF (1964), and treated in detail for the far field solutions by HIRASAWA and STAUDER (1965) and SAVAGE (1965). It is more convenient for practical purposes here to write the solutions referring to the geographical coordinate at the source. The m-component of the displacement of P and S waves in an infinite homogeneous medium may be written as, from their solu-tions,

(10)

where the suffixes p and s denotes the quan-tities for P and S waves.The 1- and 2- axes are taken to be perpendicular to the slip plane and parallel to the slip direction respectively. Using the notations in the previous study (MIKUMO, 1969),

(11)

SD(ω)=S1(ω)・S2(ω) and UD(ω)=U(ω)/D.

where

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Here we introduce the geographical coordi-
nate with the origin at the center of the source:
X=sinΘ ・sinΦ, Y=sinΘ ・cosΦ, Z=cosΘ (Θ and
Φ are measured from the zenith and clockwise
from the north). Let the displacement com-
ponents, and the direction cosines of the 1-,
2-, and 3-axes, relative to the new coordinate
be (uX, uY, uZ) and (lm, mm, nm), respectively.
Then we have

(12)

The displacements for the (γ, Θ, Φ)-com-
ponents, which corresponds to those of P, SV
and SH waves, are

(13)

where these components are taken positive
in the directions away from the origin, away

downwards for θ<π/2, and clockwise, respec-
tively. Putting eqs. (10) and (12) into (13),
we obtain,

(14)

where

(5)

cp, cSV, and CSH are the constants specified by
the source mechanism and the location of the
relevant station, and equal to those in the
case of double-couple point source (MIKUMO,
1962). M0 is the seismic source moment
defined as M0=μDLW.
The amplitude, distribution of P, SV and
SH waves can be evaluated from eqs.(14) and
(15), since the orientation of two nodal planes
have been determined from the first motion
pattern of P waves. Figs. 6(a) and 6(b) shows
the distribution of first motion sings of cp,
cSV and cSH for the three kinds of body waves,
together with their nodal curves, in, two
earthquakes Nos. 12 and 15, according to the

Fig, 6. Nodal curves of P, SV and SH waves, (a) No. 12 (b) No. 15

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method of KEYLIS-BOROK (1957). SV waves emitted downward from the source changes
the sign during propagation to the station. This effect is included in the above figures. The above figures Will be useful for the syn-thesis of the three components.
The body waves radiated from the source are suffered from the effects of propagation through the mantle and crust and of the seismograph. The waveforms that would be recorded on seismograms at the earth's sur-face should be derived from the superposition of normal modes in a gravitating spherical earth (for example, SATO et al.,1967;USAMI et al., 1968). It has been shown (SATO, 1969), however, that for direct body waves with periods probably up to about 40sec, the di-vergence effect for a limited but rather wide range of epicentral distances may be approxi-mated With errors less than 5% by the geo-metrical spreading expected from the ray theory. For this reason, the waveforms may be expressed, to a good approximation, by (MIKUMO, 1969)

(16)

where

and

Hp,8(ω)=Pp,8(ω)・Cp,8(ω)・I(ω)

P(ω), C(ω), I(ω) and H(ω) are the mantle transfer function, crustal response, instru-
mental response and the total system response,
respectively. P(ω) for direct S waves has
been computed here by the product of geo-
metrical spreading and attenuation from the
Jeffreys-Bullen Velocity profile and probable
dissipative models, as in the case of direct
P waves. For attenuation effect, we refer
mainly to a Q model (Model 11) of MIKUMO
and KURITA (1968) by reducing the Q Values
by a constant rate of 4/9 at all depths according
to a relation Qα/Qβ=9/4 suggested by ANDER-
SON et al. (1965), and also to their MM8-Qβ

model for the sake of comparison. C(ω) has
also been computed for the vertical and hori-
zontal components of P and SV waves and
for SH waves by the matrix method (HASKELL, 1960, 1962), for the layered structure appro-
priate to each station and for the pertinent
incident angle. when the structure is not
known, the unilayered model of HASKELL has
been applied to the stations.
Now we have the synthesized seismograms
of the Vertical and horizontal components of
P)and SV waves fPu(t), fPw(t), fSVu(t) and
fSVw(t), and of SH waves, fSH(t), through eq.
(16). Fig. 7 shows the observed(full line)
and synthesized (broken line) waveforms on
the Vertical component seismograms for five

Fig. 7. Observed and syntllesized seismograms of earthquake No. 23. (vertical component of P and SV waves)

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Fig. 8. Observed and synthesized seismograms (three components of P and S waves)
(a) earthquake No. 12.

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selected stations in earthquake No. 23. It appears that general features of the wave-forms are in good agreement between the two kinds of seismograms. The S waves recorded at MUN and SPA show rather shorter periods than those predicted. If the short-period oscillations are removed by a proper filtering, the agreement would be much improved.
It is more appropriate to make such a com-parison on the three component seismograms, since a different kind of information is in-

volved in the horizontal records, particularly of S waves. It is, however, usually difficult to see well-isolated direct S waves on all of

the three components at many stations. This is because they are often contaminated by preceding motion of P wave group and later phases such as sS and ScS over a wide range of epicentral distances, and partly because SV waves are sometimes distorted by the crust due to large incident angles. For this reason, the comparison can be made only for a limited number of stations, whereas FUKAO (1970) estimated the fault parameters directly from S waveforms at many stations.
Two horizontal components fN(t) and fE(t) on the synthesized seismograms are derived by the rotation of axis,

(17)

respectively, where φ is the back azimuth
measured from station to epicenter. Figs.
8(a)and 8(b) depict the recorded (full line)
and synthesized broken line three component
seismograms of both P and S waves at four
stations for earthquakes Nos. 12 and 15. All
the seismograms have been normalized to a
seismograph magnification of 1,500.
It may be seen that there is general agree-
ment between the observations and theore-
tical predictions for most of the P and S

waveforms. An introduction of MM8-Qβ
model instead of the Qβ model assumed here
did not make much difference in the S wave
amplitudes. It was also found that an arbi-
trary assumption of Qα=Qβ makes the theore-
tical S-wave amplitude significantly larger
than the recorded amplitudes. An appreci-
able discrepancy lies, however, in the NS-
component of S waves at BAG in earthquake
No. 12 and at COL in earthquake No. 15.
It is likely that the discrepancy comes from
the wave incidence deviating from the great
circle path. There is also a possibility that
different attenuations for SV and SH waves
could reconcile the discrepancy. However,
most acceptable explanations would be that
the S-wave amplitudes at the two stations
are sensitive to the accuracy of the fault
plane solutions, since the stations are located
near the nodal plane of SV waves (Compare
Fig. 5 with Figs. 6(a) and 6(b)). Although
there are also some differences in the ampli-
tudes for specific peaks or troughs, it may
be said that the recorded and synthesized
waveforms of P and S waves, including their
amplitudes, are in good agreement. Also, the
predominant periods of the synthesized S
waves are apparently longer than those of P
waves, and generally agree with the observed
periods. This may be accounted for mainly
by higher attenuations in the mantle and
partly by longer source process times due to
combined effects of the fault length and
width.
It may be concluded from the above com-

parison that the observed S waveforms in-
cluding the absolute amplitudes give support
to the dislocation sources inferred from P
waves.

§6. Discussion

In this paper, we have estimated various
source parameters for shear dislocation
models. There are alternative hypotheses
for deep-focus earthquakes, such as volume
changes due to phase transitions (BENIOFF,
1963; DENNIS and WALKER, 1965; RANDALL,
1966; EVISON, 1967) and tensile fracture (or
extentional dislocation) models (SAVAGR, 1965;

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BURRIDGE, 1968). It appears, however, that azimuthal dependence of the source process times as well as the radiation patterns of the double-couple type, which were found here, favor the shear dislocation models for the four earthquakes analyzed. This does not necessarily mean that the source process of all deep earthquakes can be related to shear dislocations. More deep earthquakes in dif-ferent regions need be analyzed to test the validity of this model.
We have so far assumed bidirectional type of faulting rather than pure unilateral and bilateral, since it may be more realistic to consider that faulting would initiate near the center of the stressed region (SAVAGE, 1966) and proceed rapidly on the fault surface along and across the slip directions, in view of a number of evidence that aftershocks of large shallow earthquakes occur frequently in an area with the main shock at the edge.
If we suppose that unilateral faulting pro-pagates along the longer axis of the slip plane (Case (A)), it would follow that the rupture velocity takes unusually high values. FUKAO (1970)also obtained an extremely high rup-ture velocity for a deep-focus earthquake of Banda Sea from the analysis of S waveforms. In these cases, we would have to take the following interpretations. One would be simultaneous faulting; the source process will be such that body waves-are radiated almost simultaneously over the entire fault surface within the rise time of dislocation (FUKAO, 1970). The source process time in this case is interpreted as the time difference between the arrival time of body waves from the nearest point on the fault surface at the initiation of the rupture process and that from the farthest point at the termination (FUKAO, 1970). Another interpretation would be supersonic dislocations, in which the rup-ture velocity exceeds elastic wave veloci-ties. The possibility has been suggested by ESHELBY (1956) for a part of screw dislocation on a slip plane in crystals, and by WEERTMAN (1969) for slippage for earthquake faults. WEERTMAN (1969) has proved that both screw dislocations and edge dislocations over an

infinite distance could move at a uniform
supersonic velocity on a slip or fault plane,
if the frictional stress decreases with increas-
ing displacement velocity across the slip
plane.
If we assume bilateral faulting, the total
process time is the superposition of the width
of two pulses from two unilateral faulting in
opposite directions. Since a predominant
portion with large amplitude usually lies in
the first-arrived pulse, the synthesized wave-
form is between those from the two uni-
lateral faulting (MIKUMO, 1969). It would be
difficult, therefore, to discriminate bilateral
faulting from unilateral one on the azimuthal
dependence of the source process times.
The concept of bidirectional faulting has
been introduced by FUKAO (private communi-
cation, 1970) in the time domain analysis of a
multiple deep-focus shock in South America.
It has been demonstrated that this type of
faulting could account for our results, parti-
cularly for the rupture velocity. It appears
that the rupture has propagated at veloci-
ties between 3.2 and a shear velocity of 4.5
km/sec in the present case, as has been sug-
gested by a theoretical work (MANSINHA,
1964) and from a number of observations.
The propagating direction makes a small
angle to the shorter axis of the fault plane.
If these interpretations are correct, earth-
quakes Nos. 6 and 23 with predominant
strike slip component would be approximated
by edge dislocations, while earthquake No.
12 having strike-slip with a fraction of dip-
slip component, and earthquake No. 15 charac-
terized by thrust faulting could be associated
with screw dislocations.

We shall next estimate the effective stress
at the initial state to produce slip dislocations
from the results in the preceding sections.
OROWAN (1960) has stated that during the
fault rupture a frictional stress σf acts
to resist the fault slippage, and the initial
shear stress σa decreases to σf when the slip-
page will. cease. Since, in this case, an
amount of the energy σf DA will be lost as
frictional heating, while an energy 1/2 (σa-
σf) DA will go into the generation of seismic

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waves, the stress deduced from seismic waves
is the average effective stress σa-σf (BRUNE,
1970). Although it is difficult to determine
the stresses σa and σf separately from seismic
observations, the effective stress may be in-
ferred from the stress drop and the average
acting stress. We have the following relation
(AKI, 1966; BRUNE, 1968; WYss and BRUNE,
1968) among the average acting stress δ (aver-
age of initial stress σ1 and final stress σ2), the
seismic moment M0, and the strain energy re-
leased E0; σ=(σ1+σ2)/2=E0/DA=μE0/M0, and

ησ=μE8/M0, where E8 and η are the seismic
wave energy and efficiency respectively and
ησ means a lower bound for the actual average
stress. If we assume further that the total
energy release is equal to the sum of the
seismic and frictional energies, ie. E0=E8+Ef,
it follows that σ=ησ+σf (BRUNE et al.,1969).
The stress drop during the faulting process
is Δ σ=σ1-σ2<σ1. Combing these relations, we
have σ1-σf=Δ σ/2+ησ for the effective initial
stress to produce shear dislocations.
The estimated stress drop at the time of
the four intermediate earthquakes, as given
in Table 3, seems slightly larger than that
for shallow earthquakes with similar magni-
tudes (BRUNE and ALLEN, 1967; BRUNE,
1968), and considerably lower than that for
deep-focus earthquakes (BERCKHEMER and
JACOB, 1968; FUKAO, 1970), although their
methods of estimates were different from
ours. on the other hand, the apparent aver-
age stress ησ for the four shocks lies between
8 and 37 bars (MIKUMO, 197O). These values
are of the same order with those for shallow
and deep-focus earthquakes (WYSS and BRUNE,
1968; WYSS, 1970), while WYSS (1970) obtained
270 bars as the mean value for 10 inter-
mediate earthquakes between 45 and 150km
in South America. He stated, however, that
earthquakes at intermediate depths in the
Tonga-Kermadec Trench are not characterized
by such a high apparent average stress. From
these evidence, it may be reasonable to con-
sider that the effective initial shear stress
at intermediate depths will be appreciably
larger than those at the crust and the upper-
most mantle. These high stresses may be

due to high values of σ1, or low values of σf,
or both.
The high initial stress would be associated
with high strength of mater-al, although it
does not appear to reach MCKENZIE's esti-
mate of 2.5kb for the stresses that could be
accumulated in descending slabs of the litho-
sphere in deep seismic zones (MCKENZIE, 1969).
The possible low frictional stress would occur
in relation to the temperature-pressure rela-
tion. It is likely that increasing temperature
at intermediate depths probably with the high
stresses produce partial melting of material
or a well-lubricated state on a slip plane

(OROWAN; 1960: GRIGGS and BAKER, 1969), and
hence this would decrease the dynamic fric-
tional stress at the time of dislocations. The
effective initial stress σ1-σf for the four
earthquakes is estimated to be about 30～120
bars from the above derived relation.
It is to be noted that the above interpreta-
tions should be regarded as only tentative,
and more deep and intermediate earthquakes
need be analyzed.

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References

Aki, K., Generation and propagation of G wave from the Niigata earthquake of June 16, 1964,
Part 2. Estimation of earthquake moment,
released energy and stress-strain drop from the G wave spectrum, Bull. Earthq. Res. Inst., 44, 73-88, 1966.
Anderson, D.L., A. Ben-Menahem and C.B. Archam-beau, Attenuation of seismic energy in the mantle, J. Geophys. Res., 70, 1441-1448, 1965.
Benioff, H., Source wave form of three earthquakes, Bull. Seism. Soc. Amer., 53, 893-904, 1963.

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Berckhemer, H. and K.H. Jacob, Investigation of
the dynamical process in earthquake foci by ana-
lyzing the pulse shape of body waves, Proc. of the
X Assembly of the ESC, Sept., 1968, II, 253-352,
1970.
Bollinger, G.A., Determination of earthquake fault parameters from long-period P waves, J. Geophys. Res., 73, 785-807, 1968.
Bollinger, G.A., Fault length and fracture velocity for the Kyushu, Japan, earthquake of October 3, 1963, J. Geophys. Res., 75, 955-964, 1970.
Brune, J., Travel times, body waves and normal modes of the earth, Bull. Seism. Soc. Amer., 54, 2099-2127, 1964.
Brune, J., Seismic moment, seismicity, and rate of slip along major fault zones, J. Geophys. Res., 73, 777-784, 1968.
Brune, J., Tectonic stress and the spectra of seismic shear waves from earthquakes, J. Geophys. Res., 75, 4997-5009, 1970.
Brune, J. and C.R. Allen, A low stress-drop, low-magnitude earthquake with surface faulting: The Imperial, California, earthquake of March 4, 1966, Bull. Seism. Soc. Amer., 57, 501-514, 1967. Brune, J.N., T.L. Henrey and R.F. Roy, Heat flow, stress and rate of slip along the San Andreas Fault, California, J. Geophys. Res., 74, 3821-3827, 1969.
Burridge, R. and L. Knopoff, Body force equivalents for seismic dislocations, Bull. Seism. Soc. Amer., 54, 1875-1888, 1964.
Burridge, R., Theoretical seismic sources and pro-pagating brittle cracks, J. Phys. Earth, 16, Special Issue, 83-92, 1968.
Dennis, J.G. and C.T. Walker, Earthquakes result-ing from metastable phase transitions, Tectono-physics, 2, 401-405, 1965.
Eshelby, J.D., Supersonic dislocations and disloca-tions in dipersive media, Proc. Phys. Soc. London, B 69, 1013-1019, 1956.
Eshelby, J.D., The determination of the elastic field of an ellipsoidal inclusion and related pro-blems, Proc. Roy Soc. London, A 241, 376-396, 1957.
Evison, F.F., On the occurrence of volume change at the earthquake source, Bull. Seism. Soc. Amer., 57, 9-26, 1967.
Fukao, Y., Focal process of a deep focus earth-quake as deduced from long-period P and S waves, Bull. Earthq. Res. Inst., 48, 707-727, 1970. Griggs, D.T. and D.W. Baker, The origin of deep-focus earthquakes, in Properties of Matter, pp. 23-42, John Wiley & Sons, New York, 1969. Haskell, N.A., Crustal reflection of plane SH waves,

J. Geophys. Res., 65, 4147-4150, 1960.
Haskell, N.A., Crustal reflection of P and SV waves, J. Geophys. Res., 67, 4751-4767, 1962.
Haskell, N.A., Total energy and spectral density of elastic wave radiation from propagating fault, Bull. Seism. Soc. Amer., 54, 1811-1841, 1964. Hirasawa, T. and W. Stauder, On the seismic body waves from a finite moving source, Bull. Seism. Soc. Amer., 55, 237-262, 1965.
Isacks, B,J. Oliver and L.R. Sykes, Seismology and the new global tectonics, J. Geophys. Res., 73, 5855-5899, 1968.
Keylis-Borok, V.I., The determination of earthquake mechanism, using both longitudinal and transverse waves, Ann. Geofis., 10, 105-128, 1957. Keylis-Borok, V.I., On estimation of the displace-ment in an earthquake source and of source di-mensions, Ann. Geofis., 12, 205-214, 1959. Knopoff, L., Energy release in earthquakes, Geophys. J., 1, 44-52, 1958.
Maruyama, T., On the force quivalents of dynamical elastic dislocations with reference to the earth-

quake mechanism, Bull. Earthq. Res. Inst., 41, 467-486, 1963.
Mansinha, L., The velocity of shear fracture, Bull. Seism. Soc. Amer., 54, 369-376, 1964.
McKenzie, D.P, Speculations on the consequences and causes of plate motions, Geophys. J., 18, 1-32, 1969.
Mikumo, T., Mechanism of local earthquakes in Kwanto region, Japan, derived from the amplitude relation of P and S waves, Bull. Earthq. Res. Inst., 40, 399-424, 1962.
Mikumo, T. and T. Kurita, Q distribution for long-period P waves in the mantle, J. Phys. Earth, 16, 11-29, 1968.
Mikumo, T., Long-period P waveforms and the source mechanism of intermediate earthquakes,
J. Phys. Earth, 17, 169-192, 1969; Correction, ibid., 18, 313, 1970.
Orowan, E., Mechanism of seismic faulting in rock deformation, A Symposium, Geol. Soc. Amer. Mem., 79, 323-345, 1960.
Otsuka, M., Study on focal mechanism by the ana-lysis of seismic waves of S-type, Spec. Contr. Geophys. Inst., Kyoto Univ., 4, 37-49, 1964. Randall, M.J., Seismic radiation from a sudden phase transition, J. Geophys. Res., 71, 5297-5302, 1966.
Sato, Y.,T. Usami and M. Landisman, Theoretical seismograms of spheroidal type on the surface of a gravitating elastic sphere, II. Case of Guten-berg-Bullen A' earth model, Bull. Earthq, Res. Inst., 45, 601-624, 1967.

Source Process of Deep and Intermediate Earthquakes 19

Sato, R., Amplitude of body waves in a hetero-geneous sphere, comparison of wave theory and ray theory, Geophys. J., 17, 527-544, 1969.
Savage, J. C., The effect of rupture velocity upon seismic first motions, Bull. Seism. Soc. Amer.,
55, 263-275, 1965.
Savage, J. C., Radiation from a realistic model of faulting, Bull. Seism. Soc. Amer., 56, 577-592, 1966.
Teng, T. L. and A. Ben-Menahem, Mechanism of deep earthquakes from spectrums of isolated body-wave signals, 1. Banda Sea earthquake of March 21, 1964, J. Geophys. Res., 70, 5157-5170, 1965. Usami, T., Y. Sat6 and M. Landisman and T. Odaka, Theoretical seismograms of spheroidal type on the surface of a gravitating elastic sphere, III.

Case of a homogeneous mantle with a liquid core, Bull. Earthq. Res. Inst., 46, 791-819, 1968. Weertman, J., Dislocation motion on an interface with friction that is dependent on sliding velocity, J. Geophys. Res., 74, 6617-6622, 1969.
Wyss, M., and J. N. Brune, Seismic moment, stress, and source dimensions from earthquakes in the California-Nevada region, J. Geophys, Res., 73, 4681-4694, 1968.
Wyss, M., Stress estimates for South American shallow and deep earthquakes, J. Geophys. Res., 75, 1529-1544, 1970.

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