Source mechanisms of deep and intermediate-depth low-frequency earthquakes beneath Iwate volcano, northeastern Japan

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SUMMARY
The 1998–1999 volcanic unrest of Iwate volcano, northeastern Japan, was marked by 350 deep low-frequency earthquakes (DLFs) and 120 intermediate-depth low-frequency earthquakes (ILFs), as well as an intense swarm of shallow volcanic earthquakes that began during 1998 April. The rate of occurrence of the DLFs increased approximately 5 d before that of the shallow volcanic earthquakes increased, and the number of ILFs gradually increased from the middle of 1998. To investigate the relationship between the shallow volcanic activity and the activities of the DLFs and ILFs, we determined their precise hypocentres and source mechanisms by analysing waveform data recorded by a dense seismic network of 47 three-component seismometers located on and around the volcano. The hypocentres of DLFs are concentrated within three regions: the first region is at depths from 31 to 34 km, approximately 10 km south of the summit, the second is at depths from 32 to 36 km approximately 10 km northeast of the summit and the third is at a 37 km depth, approximately 7 km northeast of the summit. In contrast, ILFs are located within a vertical pipe-like region just beneath the summit and sometimes show a vertical migration of the focal depth. Our moment tensor inversion using spectral ratios of body waves indicates that the source mechanisms of the DLFs and ILFs have a significant double couple and a compensated linear-vector dipole component. It is also found that a significant volumetric change is included in the source mechanisms for some DLFs. Such source mechanisms of DLFs and ILFs can be explained by the motion of a tensile crack coupled either with a shear crack or with an oblate spheroid magma chamber. However, the orientations and polarities of the crack motions are not the same in each region of DLFs and ILFs. These results suggest that a complex magma system is present at the source regions of the DLFs and ILFs.

Key words: earthquakes, magma, seismicity, source mechanism, spectral ratio.

1 INTRODUCTION
It is well known that low-frequency earthquakes (LFs) and tremors are observed in the deep crust and mantle beneath volcanoes, although the majority of LFs and tremors originate at depths of less than 3 km (McNutt 1994). For example, LFs at Kilauea volcano, Hawaii, occur at depths from 5 to 60 km (Koyanagi et al. 1987) and LFs beneath Hekla volcano, Iceland are found at depths from 8 to 14 km (Soosalu & Einarsson 1997). LFs occurring at deep portions beneath volcanoes have also been reported at other regions: Long Valley caldera (Pitt & Hill 1994), Spurr volcano, Alaska (Power et al. 1995), Izu-Ooshima volcano, central Japan (Ukawa & Ohtake 1987), and volcanoes in northeastern Japan (Hasegawa et al. 1991; Hasegawa & Yamamoto 1994). Following the study by Koyanagi et al. (1987), these LFs are often classified into two groups: intermediate-depth LF earthquakes (ILFs) that occur at depths from 5 to 15 km and deep LF earthquakes (DLFs) that occur at depths from 30 to 60 km.

To clarify the quantitative relationship between magmatic processes and DLFs and ILFs, previous studies have investigated the source mechanisms of such LFs. For example, Ukawa & Ohtake (1987) explained a DLF beneath Izu-Ooshima volcano by flow in a channel connecting two magma chambers. Nishidomi & Takeo (1996) obtained a focal mechanism of strike-slip type for a DLF near Nikko-Shirane volcano, Japan. Okada & Hasegawa (2000) obtained non-double-couple focal mechanisms for three DLFs near Narugo volcano, Japan. However, the number of estimated focal mechanisms of LFs for each volcanic region is few.

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and no significant common features have been found in the focal mechanisms.

In 1998–1999, we observed many DLFs and ILFs at Iwate volcano (2038 m), one of several active Quaternary stratovolcanoes lying in the northeastern Japan arc (Fig. 1a). A remarkable volcanic tremor of 45 min duration was first observed on 1995 September 15 and initiated intensified ILF activity (Ueki et al. 1996). Although ILF and DLF activities were quite low for the period from 1996 to 1997, DLF activity intensified at the end of 1998 January. At that time, shallow high-frequency (SHF) earthquake swarms (Tanaka et al. 2002), very-long-period (VLP) seismic events (Nishimura et al. 2000) and crustal deformation were observed in the shallow part of the volcano (Sato & Hamaguchi 1999; Ueki et al. 1999; Miura et al. 2000), and these were considered to originate from intrusion of magmatic fluid (Sato & Hamaguchi 1999; Ueki et al. 1999; Miura et al. 2000; Tanaka et al. 2002). After a tectonic earthquake of local magnitude of 6.1 occurred at a shallow location adjacent to the volcano-seismic swarm area on 1998 September 3 (Umino et al. 1998; Nakahara et al. 2002), the ILF and DLF activities gradually weakened as did SHF, VLP and crustal deformation at the shallow location. During the 1998–1999 volcanic activity, we observed in total 350 DLFs and 120 ILFs.

In this study, we examined the DLFs and ILFs of Iwate volcano by analysing high-quality data obtained using our dense seismic network deployed around the volcano. First, we determined spatial-temporal variations of the hypocentre distributions of DLFs and ILFs using the joint hypocentre determination (JHD) method. Subsequently, we examined the source mechanisms of DLFs and ILFs beneath Iwate volcano, applying a spectral ratio method we have recently developed. In this paper, we discuss the implications of our results for the magmatic activity by comparing them with similar results from other volcanoes.

2 SEISMIC OBSERVATIONS ON AND AROUND IWATE VOLCANO

We analysed seismic data from 47 stations consisting of three seismic networks on and around Iwate volcano (Fig. 1): the permanent volcano seismic network of Tohoku University (crosses), a temporary reinforcing seismic network in operation since 1995 (diamonds) and a satellite-based temporary network of the Joint Seismic Observation (JINT, solid squares) operated by a Japanese research group in the period from 1997 October to 1999 May (Hasegawa & Hirata 1999). Seismic signals transmitted from stations (solid circles) of the Japan Meteorological Agency (JMA) by telephone line supplement the data set after 1997 October.

All of the stations in these networks are equipped with three-component seismometers. Most of the seismometers at these stations are short-period instruments with a natural period of $T_0 = 1$ s, but some of them are longer-period instruments with $T_0 = 5$ s (KR B, UEB and YKR), $T_0 = 20$ s (HAN, NSK and TAZ), and one is an STS-2 broad-band seismometer (IWT). The seismometers at the permanent stations are installed in boreholes at depths of either 300 or 557 m from the ground surface (ANS, GNB, HSB, YKB and MTO) together with tiltmeters and strainmeters, or in vaults at the ends of 5–50 m long tunnels. Most of the permanent stations have been in operation for more than 4 yr. The seismometers at the temporary stations are buried in soil or are attached to bedrock. Seismic data from the permanent stations are digitized by a 16 or 22 bit analogue-to-digital converter with a sampling frequency of 100 Hz, and are telemetered to Tohoku University in Sendai by telephone lines. Data from temporary stations IEF, IKG, KNS, OSD, RZT, SDF, UBK, UEB and YKR are telemetered to the university by radio and/or telephone lines with 12-bit resolution, and those from temporary station SZK and the JOINT stations are telemetered by a satellite system with 22-bit resolution (Urabe et al. 1998) to Tohoku University. Seismic signals from temporary stations KHM, KR B, SRY and TKM are continuously stored in a 2 GB hard disc at each site by a digital data recorder with 16-bit resolution and a sampling frequency of 100 Hz. The internal crystal clock of the recorder (precision better than $10^{-4}$ at $-20$ to $+40$ °C) is corrected every hour by a GPS time signal so that timing error is assured to be less than 4 ms.

3 HYPOCENTRE DISTRIBUTIONS OF DEEP AND INTERMEDIATE-DEPTH LOW-FREQUENCY EARTHQUAKES

3.1 Hypocentre distribution of deep low-frequency earthquakes

Recordings of DLFs beneath Iwate volcano generally show clear $P$ and $S$ waves (Figs 2a and b). We measured the arrival times of $P$ and $S$ phases of DLFs on the three-component seismograms and determined their hypocentres whenever the onsets of $P$ and $S$ waves were picked at five or more stations with good azimuthal coverage. The reading accuracies of $P$ and $S$ wave arrival times are better than ±0.2 and ±0.3 s, respectively. We located the hypocentres of 120 DLFs in the period from 1997 October to 2000 June, using the velocity model of Tanaka et al. (2002) and a standard hypocentre determination procedure. The results show that 90 DLFs are concentrated beneath the southern flank of the volcano (south DLFs), and the other 30 DLFs occur beneath the northeastern flank (north-east DLFs). The local magnitudes of the DLFs are distributed in the range 0.5–2.1. Since both DLF groups are concentrated tightly in their respective regions, we apply the JHD method (Douglas 1967; Pujol 1988) separately to the south and the northeast DLFs. Fig. 3 shows hypocentres relocated by the JHD method. The south DLFs are concentrated at depths from 31 to 34 km. Their epicentres form a dense spherical cluster beneath the south flank approximately 10 km from the summit. We call these earthquakes group A. Conversely, hypocentres of the northeast DLFs are divided into two groups, B and C. Group B earthquakes are concentrated at depths from 32 to 36 km approximately 10 km northeast of the summit, while group C earthquakes are located at depths around 37 km approximately 7 km northeast of the summit. Group C earthquakes are significantly deeper than group B earthquakes, because $S-P$ times for group B (4.0 s) are shorter than those for group C (4.7 s) at the nearest station SRY. Taking the reading errors of the arrival times into account, we estimate the typical location errors to be less than approximately 1.5 km in both horizontal and vertical directions. The mean and standard deviations of rms residuals of $P$ and $S$ phases for the relocated northeast DLFs are 0.09 ± 0.05 s and 0.10 ± 0.04 s, respectively, and those for the relocated south DLFs are 0.12 ± 0.07 and 0.13 ± 0.06 s, respectively. Because the rms residuals of $P$ and $S$ phases obtained by the JHD method are reduced to 50 and 70 per cent of those obtained by the ordinary method, respectively, the hypocentres determined by the JHD method are more reliable.

3.2 Hypocentre distribution of intermediate-depth low-frequency earthquakes

Recordings of ILFs beneath Iwate volcano also generally show clear $P$ and $S$ waves (Fig. 2c), hence we determine their hypocentres

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Figure 1. Location map showing the seismic stations. (a) Stations near Iwate volcano and location of Iwate volcano (inset map). (b) Stations around the volcano. The various symbols in the figure denote different types of station: crosses, permanent stations maintained by Tohoku University; diamonds, temporary stations maintained by Tohoku University; squares, temporary stations maintained by the joint seismic observations of Japanese universities (JOINT); solid circles: stations maintained by the Japan Meteorological Agency (JMA). Triangles denote active volcanoes. Dotted lines in (a) show elevation contours at 300 m intervals.
Figure 2. (a) An example of three-component waveforms for the south DLFs from an earthquake which occurred at 23:32 JST on 1998 July 24. This earthquake is characterized by high-frequency onsets and monotonic coda waves. (b) Same as (a) for the earthquake at 01:37 on 1998 August 6. This DLF has the best signal-to-noise ratio we have observed since 1981. The $S$-to-$P$-wave amplitude ratios differ from station to station. (c) An example of three-component waveforms for the ILF earthquake which occurred at 9:01 JST on 1998 October 24. The $S$-to-$P$-wave amplitude ratios differ from station to station. Shaded rectangles on the waveforms of HMK represent the time windows for calculating the spectra shown in Fig. 4.

when the onsets of $P$ and $S$ phases are available for at least four and seven stations, respectively. The reading accuracies of both $P$ and $S$ wave arrival times are better than ±0.1 s. We located 139 ILFs during the period from 1998 April to 2000 June. The local magnitudes of ILFs are distributed in the range 0.0–1.6. Since the hypocentres determined by the standard method are concentrated in a small region, we applied the JHD method to the ILFs, too. Their hypocentres are shown in Fig. 3 together with those of the DLFs. The ILFs are located within a nearly vertical cylindrical or pipe-like region extending from 5 to 12 km in depth, with north–south and east–west dimensions of approximately 1.5 and 1.7 km, respectively. The vertical distribution from 5 to 12 km in Fig. 3 is not due to mislocations, because $S$–$P$ times at station YKB, which is located just above the hypocentres, vary from 0.9 to 1.6 s. The mean

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values and standard deviations of residuals of $P$ and $S$ phases using
the JHD method are $0.09 \pm 0.03$ and $0.10 \pm 0.04$ s, respectively. The
rms residuals of $P$ and $S$ phases obtained by the JHD are reduced
to 70 and 40 per cent of those obtained using the standard method,
respectively. The residuals obtained by the JHD method are almost
within the picking accuracy of the $P$ and $S$ phases.

4 WAVEFORM CHARACTERISTICS OF DEEP AND INTERMEDIATE-DEPTH LOW-FREQUENCY EARTHQUAKES

Fig. 2 shows examples of three-component waveforms of DLFs
and ILFs at five stations. Fig. 2(a) shows an example of DLFs in
group A that have a monotonic S coda with a dominant frequency of approximately 1 Hz. Amplitudes of the 1 Hz monotonic long S coda seem to be similar at all stations so the source mechanism of this DLF cannot consist of a pure double-couple or it is purely isotropic, but it must have some volumetric component. High-frequency waves are clearly seen in the onset of $P$ and $S$ waves at stations located close to the epicentre of this earthquake (e.g. ANS, IWT, GMB, HSB and HMK). Fig. 2(b) shows a DLF ($M = 1.9$) in group A, which occurred at 01:37 on 1998 August 6. This earthquake has the largest signal-to-noise ratio we have observed since 1981. The dominant frequencies of $P$ and $S$ waves for this earthquake are approximately 5 Hz. Amplitudes of $P$ and $S$ waves for this earthquake differ from station to station. These characteristics are quite different from those of the DLF shown in Fig. 2(a). Hence, the waveforms of the DLFs cannot be characterized as a single earthquake family, even though the DLFs in group A are clustered in a small region of approximately 1.5 km in radius. This suggests that the variety of waveforms is caused mainly by a variety of source processes and not by path or site effects.

Fig. 2(c) shows an example of ILFs ($M = 1.6$), which occurred at 9:01 on 1998 October 24. The amplitudes of $S$ waves are larger than those of $P$ waves. As shown in Fig. 2(c), ILFs typically generate significant $S$ waves. The amplitudes of $S$ waves differ from station to station, which implies that the source mechanism is not explained by a pure isotropic explosion or implosion. It should be noted that no ILFs have monotonic long $S$ coda waves.

We examine velocity amplitude spectra for all DLFs and ILFs from vertical and N–S components at station HMK (Fig. 4). The dominant frequencies of $P$ and $S$ waves range from 1 to 8 Hz and from 1 to 6.5 Hz, respectively. Most of the dominant $P$ wave

![Figure 4](image-url)
frequencies are estimated to be 2.5–5.0 Hz. Approximately 30 per cent of DLFs have $S$ waves with a dominant frequency of 1 Hz; the remnants have dominant frequencies from 2 to 6 Hz. Dominant frequencies for ILFs are distributed widely, ranging from 1 to 8 Hz for $P$ waves, and from 1 to 6 Hz for $S$ waves. Only 1 per cent of ILFs have dominant frequencies around 1 Hz, which contrasts with the frequency spectrum found for DLFs.

5 SOURCE MECHANISMS OF DEEP AND INTERMEDIATE-DEPTH LOW-FREQUENCY EARTHQUAKES

5.1 Body wave spectral ratio method

We applied a body wave spectral ratio method to determine the source mechanisms of DLFs and ILFs, because the method has the advantage that a source time function does not need to be postulated before the inversion (e.g. Nishimura et al. 1995). The body wave spectral ratio method used in this study is as follows: we assume that the source sizes of earthquakes are small compared with the wavelength of the seismic waves analysed, allowing a point source approximation (Aki & Richards 2002). Assuming that all components of moment tensors have the same source time function with different amplitudes, the $r$th component of displacement at a location $x$ can be written as

$$u_r(x, t) = \sum_{p, q} M_{pq} S(\tau) * G_{npq}(x, t; \tau). \quad (1)$$

where $M_{pq}$ is the element of the moment tensor, $S(\tau)$ is the source time function and $G_{npq}(x, t; \tau)$ is the $np$ element of the Green function that relates the $r$th component of the displacement at the receiver position $x$ with the $pq$th component of impulsive force acting at the source position at time $t = \tau$. The notation $\tau$ implies spatial differentiation with respect to the $\tau$ coordinate and the asterisk denotes convolution.

Denoting far-field $P$, $SV$ and $SH$ displacements as $u_P^i(t)$, $u_{SV}^i(t)$ and $u_{SH}^i(t)$, respectively, and taking spectral ratios between the three displacements, we can obtain the body wave spectral ratios:

$$R_{SV/P}^{SV}(\omega) = \frac{u_{SV}^i(\omega)}{u_P^i(\omega)} = \frac{\sum_{p, q} M_{pq} G_{npq}^{SV}(\omega)}{\sum_{p, q} M_{pq} G_{npq}^P(\omega),} \quad (2)$$

$$R_{SH/P}^{SH}(\omega) = \frac{u_{SH}^i(\omega)}{u_P^i(\omega)} = \frac{\sum_{p, q} M_{pq} G_{npq}^{SH}(\omega)}{\sum_{p, q} M_{pq} G_{npq}^P(\omega),}$$

$$R_{SH/SV}^{SH}(\omega) = \frac{u_{SH}^i(\omega)}{u_{SV}^i(\omega)} = \frac{\sum_{p, q} M_{pq} G_{npq}^{SH}(\omega)}{\sum_{p, q} M_{pq} G_{npq}^{SV}(\omega)}.$$

where subscripts $v$, $r$ and $t$ denote the vertical, radial and transverse components, respectively. $R_{SV/P}$, $R_{SH/P}$ and $R_{SH/SV}$ are the body wave spectral ratios for the three combinations and are no longer dependent on the source time functions. The body wave spectral ratio $R_{S/P}^{SV}(\omega)$ can be rewritten as

$$R_{S/P}^{SV}(\omega) = A_{S/P}^{SV}(\omega) \exp[\theta_{S/P}^{SV}(\omega)], \quad (3)$$

where $A_{S/P}^{SV}(\omega)$ and $\theta_{S/P}^{SV}(\omega)$ are the amplitude and phase difference, respectively. $l$ and $m$ refer to the body wave type, $P$, $SV$ and $SH$.

We estimate the source mechanisms of DLFs and ILFs by comparing the observed spectral ratios with theoretical ones. First, we explain data processing for the observed DLFs. Observed waveforms are bandpass filtered using the fourth-order filter with a zero-phase shift in the range from 0.7 to 8.0 Hz. Subsequently, we resample the data to a sampling rate of 50 s$^{-1}$, and rotate the data to radial and transverse components. Then, we set a time window with a 10 per cent cosine taper for the three-component data, slightly before the arrivals of the $P$ or $S$ waves. The length of the time window is set to be 0.64, 1.28 or 2.56 s, corresponding to two or three cycles of the $S$ wave. Fig. 5 shows examples of the time windows for vertical, radial and transverse components of DLFs. Finally, we take the Fourier transform of the $P$, $SV$ and $SH$ waves using the fast Fourier transform (FFT) method and calculate the amplitude ratios and the phase differences between the two components.

Next, we calculate the theoretical spectral ratio. We calculate Green’s functions for a given moment tensor, $\mathbf{M} = M_0(m_{1x}, m_{1y}, m_{1z}, m_{2x}, m_{2y}, m_{2z}, m_{3x}, m_{3y}, m_{3z})$, where $M_0$ is the amplitude of seismic moment and the $m_i$ take values from $-1$ to 1, using the discrete wavenumber method (Bouchon 1979, 1981) and reflection and transmission coefficient matrices (Kennett &erry 1979). In the calculation, we assume a layered structure as shown in Nakamichi (2000). This structure is based on the velocity structure of Tanaka et al. (2002) and the attenuation structure of Tsumura et al. (2000). The location of each source is set at the hypocentre determined by the JHD method. Then, we compute synthetic waves in the time domain for an assumed source mechanism by superimposing the Green functions for the six components of the moment tensor. Following the processing procedure for the observed data discussed previously, we obtain the amplitude ratios and the phase differences between the two components.

We represent the observed amplitude ratio and phase difference by $A_{obs}^{SV}(\omega)$ and $\theta_{obs}^{SV}(\omega)$, respectively, and the theoretical equivalents by $A_{th}^{SV}(\omega; \mathbf{M})$ and $\theta_{th}^{SV}(\omega; \mathbf{M})$. The superscript $i$ denotes a station and $j$ indicates $S/P$, $SH/P$ or $SH/SV$. Then, the weighted sum of squared mean residuals of the amplitude ratios and phase differences can be expressed as

$$S(\mathbf{M}) = \sum_{i=1}^{N} \sum_{j=1}^{K} \left[ \log_{10} A_{th}^{SV}(\omega_k; \mathbf{M}) - \log_{10} A_{obs}^{SV}(\omega_k; \mathbf{M}) \right] / 3NK,$$

$$= w_p \sum_{i=1}^{N} \sum_{j=1}^{K} \left[ \theta_{th}^{SV}(\omega_k; \mathbf{M}) - \theta_{obs}^{SV}(\omega_k; \mathbf{M}) \right] / 3NK, \quad (4)$$

where $\omega_k$ is the discretized angular frequency, $N$ is the number of stations and $K$ is the total number of angular frequencies used for the calculation. $w_p$ is the weighting factor and determines the relative importance of the amplitude and phase residuals. We obtain the best-fitting moment tensor solutions that satisfy

$$S(\mathbf{M}) \rightarrow \min.$$

(5)

When the ratio of $P$ and $S$ wave velocities is nearly constant in the crust, amplitude ratios are relatively insensitive to crustal heterogeneities, although seismic-wave amplitudes are strongly distorted by them (e.g. Julian & Foulger 1996). On the other hand, phase differences are sensitive not only to crustal heterogeneities but also to the finite spectral window chosen. Therefore, we set the weighting factor $w_p$ to be small, varying from 0 to 0.05.

The amplitude spectra of the observed $P$, $SV$ and $SH$ waves at each station are normalized and stacked to obtain the average amplitude spectra for $P$, $SV$ and $SH$ waves at each station. We analyse only the spectral ratios in a limited frequency band in which the spectral amplitude is larger than or equal to half the overall maximum of the amplitude spectra of $P$, $SV$ and $SH$ waves. The frequency bands are calculated by this criterion for the $P$, $SV$ and $SH$ waves, independently, and all frequency bands common to the three are used to optimize eq. (4).
Figure 5. Examples of vertical, radial and transverse components of $P$, $SV$ and $SH$ waves. Dotted lines in each trace indicate the time windows for calculating the spectral ratios. (a) DLF at 1:37 1998 on August 6. The window length is 0.64 s. (b) DLF at 19:20 on 1998 August 4. The window length is 1.28 s. (c) DLF at 23:32 on 1998 July 24. The window length is 2.56 s.
To estimate the moment tensor $\mathbf{M}$, we need to minimize $S(\mathbf{M})$. However, eq. (2) has a strong non-linearity and the number of free parameters (six) is too large to use a general grid searching method. We therefore use a genetic algorithm (Holland 1975; Goldberg 1989) that is useful for solving highly non-linear problems, including some geophysical inverse problems (e.g. Sen & Stoffa 1992; Sambridge & Drijkoningen 1992; Sambridge & Gallagher 1993). In our genetic algorithm inversion, each moment tensor component ($m_{ij}$) is allowed to vary from $-1.0$ to $1.0$ and the increment of each component is $0.065$.

We obtain moment tensor solutions for weighting factors, $w_P$, of 0.00, 0.01, 0.02, 0.03, 0.04 and 0.05. Finally, we select the moment tensor solution that satisfies the $P$ wave first-motion polarities as consistently as possible. After making all the calculations, we found that the moment tensor solutions of the DLFs and ILFs do not depend significantly on the weighting factors $w_P$ used. We also estimated the covariance matrix for an estimated moment tensor using an a posteriori Gaussian probability density function obtained from the genetic algorithm inversion. The standard deviations are the square roots of the diagonal terms of the covariance matrix (Tables 1 and 2). Tables 1 and 2 show the best moment tensors selected as above for DLFs and ILFs with the standard deviations of the solutions, respectively.

### 5.2 Source mechanisms of deep low-frequency earthquakes

To obtain a well-constrained source mechanism, we select DLFs with: (1) local magnitudes greater than or equal to 1.0; (2) good signal-to-noise ratios in all three components for at least 16 stations (south DLFs) or at least 11 stations (northeast DLFs); (3) good azimuthal and distance coverage. As a result, five south DLFs and two northeast DLFs were selected out of the total 120 DLFs (see Table 1).

Fig. 6 compares the observed amplitude spectral ratios with the theoretical equivalents for the best-fitting moment tensor solution of DLFA-2 at 21 stations (see Table 1). Figs 7(a)–(g) and Table 1 summarize the moment tensor solutions of the seven DLFs. We do not show the phase difference data because the weights for the phase data are very small compared with the amplitude ratios. As shown, especially in Fig. 7(e), $P$ wave first-motion polarities of the earthquake are very consistent with the best-fitting moment tensor solution. This demonstrates that our moment tensor inversion method is very useful for estimating the moment tensor solutions of DLFs.

The source mechanisms are significantly different from each other (Fig. 7), which is consistent with the result that no earthquake families are observed (Nakamichi 2000). No characteristic orientation of the principal axes of the moment tensors of DLFs is found. To investigate the characteristics of the moment tensors, we decomposed the moment tensors into three elementary force systems that consist of isotropic (volumetric), double-couple (DC) and compensated linear-vector dipole (CLVD) force systems. This decomposition is useful for interpreting DLF source mechanisms: the volumetric component represents a spherical magma oscillation (Croson & Bame 1985; Julian et al. 1998; Miller et al. 1998), the DC component is a manifestation of shear faulting (Natale et al. 1995; Julian et al. 1998; Miller et al. 1998), and the CLVD represents rapid movement of magmatic fluid (Julian & Sipkin 1985; Kanamori et al. 1993; Julian et al. 1998; Miller et al. 1998). The decompositions are performed using the method of Knopoff & Randall (1970), which makes the major axis of the CLVD coincide with the corresponding axis of the DC. Decompositions of the moment tensor components for the DLFs are summarized in Fig. 7 and Table 3. Most of the source mechanisms of DLFs have significant DC (21–80 per cent) and CLVD (18–65 per cent) components. The source mechanisms of DLFA-3 and DLFA-5 have significant volumetric components and the source mechanism of DLFA-2, in particular, has a large volumetric component (31 per cent).

### 5.3 Source mechanisms of intermediate-depth low-frequency earthquakes

We selected four ILFs to estimate their source mechanisms according to the following criteria: (1) local magnitudes greater than or equal to 1.0, (2) good signal-to-noise ratios in all three components at over six stations and (3) a suitable azimuth and distance coverage of stations. The earthquakes selected are listed in Table 2.

We adopted a 0.64 s time window with a 10 per cent cosine taper, which contains one or two cycles of waves at the dominant frequency. The spectral ratios for $P$, $SV$, and $SH$ waves are calculated using the procedure for DLFs. The frequency bands used for the moment...
The residual and the AIC value for the best moment tensor model are 0.294 and 1319, respectively. On the other hand, the residual and AIC value for the normal fault model are 0.405 and 1575. Since the model with the smaller AIC value is statistically better, we conclude that the moment tensor model is a more plausible representation of the source mechanism of DLFA-4 than the normal fault model.

6.2 Source process for low-frequency earthquakes

As mentioned previously, the source mechanisms of DLFs and ILFs contain DC and CLVD components and DLFA-2, DLFA-3 and DLFA-5 include significant volumetric components. For an earthquake having all three components, i.e. DC, CLVD and volumetric components, a tensile–shear coupled source may be an appropriate model. Hill (1977) first presented the basic idea of a fracture network to explain earthquake swarms around volcanoes. Shaw (1980) proposed that magma is injected into the lithosphere by a process analogous to hydraulic fracturing. His model consists of a fracture network of magma-filled tensile cracks coupled with shear cracks, and magma migrates upward through the network of tensile and shear cracks. Shimizu et al. (1987) first proposed a seismic source model of a tensile–shear crack to explain the waveform of volcanic earthquakes observed at Miyake Island, Japan. We expect that a similar phenomenon may occur in the source region of DLFs and ILFs beneath Iwate volcano. That is, when the magma fluid pressure is high enough to produce brittle fractures, both tensile and shear faulting occur simultaneously or successively to generate the DLFs or ILFs. As a result, the source mechanism of the DLFs and ILFs has DC, CLVD and volumetric components. However, some DLFs and ILFs show no significant volumetric component in their source mechanisms. Thus, we must consider the reason for absence of volumetric components, that is, the mechanism of generating CLVD components with magma-filled tensile faulting.

When tensile faulting occurs in a fracture system under high fluid pressure, often associated with dyke intrusion (Julian 1983; Aki 1984; Julian & Sipkin 1985), some other process, such as fluid flow, must compensate the volumetric component expected from a tensile fault. At depths greater than a few kilometres, rocks cannot sustain high tensile stresses, especially those accompanying the creation of a void. Hence, a tensile crack can be created only in the case where a fluid under high pressure flows into it. A tensile crack in an isotropic medium can be represented by a point-force system with three mutually orthogonal dipoles: the diagonal components of the moment are $\lambda + 2\mu$, $\lambda$, and $\lambda$, where $\lambda$ and $\mu$ are Lamé

### Table 2. Relative moment tensor components estimated by the spectral ratio method for intermediate-depth low-frequency events and weights for phase difference.

<table>
<thead>
<tr>
<th>Event ID</th>
<th>Date</th>
<th>Time</th>
<th>$M$</th>
<th>Depth (km)</th>
<th>$N$</th>
<th>$m_{xx}$</th>
<th>$m_{yy}$</th>
<th>$m_{zz}$</th>
<th>$m_{xy}$</th>
<th>$m_{yz}$</th>
<th>$m_{zx}$</th>
<th>$w_P$</th>
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<td>06/28/98</td>
<td>17:46</td>
<td>1.0</td>
<td>5.7</td>
<td>7</td>
<td>0.569</td>
<td>0.503</td>
<td>0.185</td>
<td>0.466</td>
<td>0.691</td>
<td>0.030</td>
<td>0.01</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(0.034)</td>
<td>(0.072)</td>
<td>(0.004)</td>
<td>(0.025)</td>
<td>(0.046)</td>
<td>(0.014)</td>
<td></td>
</tr>
<tr>
<td>ILF-2</td>
<td>08/08/98</td>
<td>10:58</td>
<td>1.5</td>
<td>6.7</td>
<td>9</td>
<td>0.543</td>
<td>0.231</td>
<td>0.047</td>
<td>0.648</td>
<td>0.356</td>
<td>0.528</td>
<td>0.01</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(0.030)</td>
<td>(0.013)</td>
<td>(0.045)</td>
<td>(0.011)</td>
<td>(0.005)</td>
<td>(0.020)</td>
<td></td>
</tr>
<tr>
<td>ILF-3</td>
<td>10/24/98</td>
<td>09:01</td>
<td>1.6</td>
<td>8.2</td>
<td>18</td>
<td>0.623</td>
<td>0.037</td>
<td>0.739</td>
<td>0.577</td>
<td>0.192</td>
<td>0.403</td>
<td>0.02</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(0.049)</td>
<td>(0.033)</td>
<td>(0.041)</td>
<td>(0.020)</td>
<td>(0.009)</td>
<td>(0.032)</td>
<td></td>
</tr>
<tr>
<td>ILF-4</td>
<td>12/25/99</td>
<td>19:34</td>
<td>1.5</td>
<td>8.4</td>
<td>11</td>
<td>0.463</td>
<td>0.732</td>
<td>0.282</td>
<td>0.589</td>
<td>0.056</td>
<td>0.485</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(0.055)</td>
<td>(0.026)</td>
<td>(0.233)</td>
<td>(0.093)</td>
<td>(0.235)</td>
<td>(0.126)</td>
<td></td>
</tr>
</tbody>
</table>

$x$-axis directed north, $y$-axis directed east, $z$-axis directed down. Each moment tensor component is normalized with respect to the scalar moment defined by Dahlen & Tromp (1998). The numerical values in parentheses indicate the estimated standard errors for the moment components. $M$, magnitude; $N$, number of stations used for the inversion. MT, moment tensor inversion; SF, single-force inversion.
Figure 6. Comparisons of the observed amplitude spectral ratios (thin lines) to the theoretical amplitude spectral ratios (thick lines) of the best-fitting moment tensor solution for DLFA-2. The frequency ranges used for inversion are denoted by a range of theoretical ratios. On the whole, the theoretical amplitude ratios fit the observed ones closely.
Figure 7. Source mechanisms estimated for seven DLFs and four ILFs (left). Decompositions into volumetric, DC and CLVD parts of their moment tensors (right). The area of each plot is proportional to the largest principal moment. Lower focal hemispheres are shown in equal-area projection, with compressional fields shaded. The location of $P$ and $T$ axes are indicated with crosses. (a) DLFA-1. (b) DLFA-2. (c) DLFA-3. (d) DLFA-4. (e) DLFA-5. (f) DLFB-1. (g) DLFC-1. (h) ILF-1. (i) ILF-2. (j) ILF-3. (k) ILF-4. Most of the $P$ polarities are consistent with the best source mechanisms.

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The volumetric components in the source mechanisms and the long barrier between the crack and the chamber when magma starts to open are interpreted to being generated by the breaking of a liquid body is determined by the impedance contrast between the solid and the fluid. Aki et al. (1977) showed that the damping of vibrations of the fundamental mode of a liquid body is determined by the impedance contrast between the solid and the fluid. They obtained the formula

\[ Q = \ln[(Z + 1)/(Z - 1)] \]

where \( Z \) is the impedance contrast: \( Z = \rho_s \alpha/\rho_l \gamma \); \( \rho_s \) and \( \rho_l \) are the densities of the solid and fluid, respectively; \( \alpha \) is the P wave velocity, and \( \gamma \) is the acoustic wave velocity in magma. Here, we adopt \( \rho_l = 2.7 \text{ g cm}^{-3} \) and \( \gamma = 2.2 \text{ km s}^{-1} \), which are appropriate for basaltic magma (Murase & McBirney 1973). \( \rho_s = 3.0 \text{ g cm}^{-3} \) and \( \alpha = 7.8 \text{ km s}^{-1} \) are also adopted following the structure at a depth of 33 km in Tanaka et al. (2002). Then, we obtain \( Z \approx 4.0 \) and \( Q \approx 6.2 \) from eq. (6). The \( Q \) value for the fundamental mode is obtained from observed spectra using the standard relation \( Q = f/\Delta f \), where \( f \) is the frequency of the spectral peak and \( \Delta f \) is the width of the peak at its half-energy level (Aki & Richards 2002).

### Table 3. Percentage of volumetric, double-couple and CLVD components, residuals and AIC values for moment tensor solutions and single-force solutions of intermediate-depth low-frequency events.

<table>
<thead>
<tr>
<th>Event ID</th>
<th>Volumetric</th>
<th>DC</th>
<th>CLVD</th>
<th>Residual</th>
<th>MT</th>
<th>SF</th>
<th>MT</th>
<th>SF</th>
</tr>
</thead>
<tbody>
<tr>
<td>DLFA-1</td>
<td>11</td>
<td>66</td>
<td>23</td>
<td>0.3397</td>
<td>0.3635</td>
<td>1985</td>
<td>2055</td>
<td></td>
</tr>
<tr>
<td>DLFA-2</td>
<td>31</td>
<td>23</td>
<td>46</td>
<td>0.4179</td>
<td>0.4897</td>
<td>1427</td>
<td>1535</td>
<td></td>
</tr>
<tr>
<td>DLFA-3</td>
<td>23</td>
<td>21</td>
<td>56</td>
<td>0.2800</td>
<td>0.3853</td>
<td>2115</td>
<td>2538</td>
<td></td>
</tr>
<tr>
<td>DLFA-4</td>
<td>1</td>
<td>34</td>
<td>65</td>
<td>0.2940</td>
<td>0.3255</td>
<td>1319</td>
<td>1396</td>
<td></td>
</tr>
<tr>
<td>DLFA-5</td>
<td>18</td>
<td>61</td>
<td>21</td>
<td>0.3818</td>
<td>0.3993</td>
<td>2240</td>
<td>2287</td>
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</tr>
<tr>
<td>DLFB-1</td>
<td>6</td>
<td>49</td>
<td>45</td>
<td>0.1743</td>
<td>0.2592</td>
<td>326</td>
<td>435</td>
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<tr>
<td>DLFB-2</td>
<td>2</td>
<td>80</td>
<td>18</td>
<td>0.4532</td>
<td>0.4791</td>
<td>1363</td>
<td>1393</td>
<td></td>
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<tr>
<td>ILF-1</td>
<td>3</td>
<td>52</td>
<td>45</td>
<td>0.1407</td>
<td>0.2476</td>
<td>599</td>
<td>654</td>
<td></td>
</tr>
<tr>
<td>ILF-2</td>
<td>7</td>
<td>21</td>
<td>72</td>
<td>0.2764</td>
<td>0.3247</td>
<td>987</td>
<td>931</td>
<td></td>
</tr>
<tr>
<td>ILF-3</td>
<td>4</td>
<td>41</td>
<td>55</td>
<td>0.3325</td>
<td>0.3666</td>
<td>887</td>
<td>931</td>
<td></td>
</tr>
<tr>
<td>ILF-4</td>
<td>0</td>
<td>97</td>
<td>3</td>
<td>0.4220</td>
<td>0.5067</td>
<td>1055</td>
<td>1145</td>
<td></td>
</tr>
</tbody>
</table>

MT, moment tensor inversion; SF, single-force inversion.

As mentioned previously, the source mechanisms of DLFs and ILFs have DC and CLVD components and those of some DLFs also have a significant volumetric component. Although a scalar seismic moment for LF is required to account for a volume change at the source of LF, we have not obtained the scalar seismic moments by the spectral ratio method. Here, we estimate the volume change at the source for the event DLFA-2 for which the local magnitude is 1.4. The scalar seismic moment of this event is approximately \( 1.3 \times 10^{11} \text{ N m} \). We consider only a volumetric (isotropic) component for estimating a volume change at the source of LF, since CLVD sources produce no volume change. The scalar moment of the volumetric component is approximately \( 4.1 \times 10^{10} \text{ N m} \), which is 31 per cent of the total scalar seismic moment. For \( \lambda = \mu = 1.7 \times 10^{10} \text{ Pa} \), the two volume-change values \( \Delta V_{\text{vol}} = M_{\text{vol}}/\lambda (\lambda + 2\mu) \) are \( 1.45 \text{ m}^3 \) and \( \Delta V_{\text{vol}} = M_{\text{vol}}/(\lambda + 2\mu) = 0.80 \text{ m}^3 \) are obtained for a crack and a general explosion or implosion, respectively (Müller 2001). The
former probably gives an upper limit for the true volume change $\Delta V$, causing $M_{\text{vol}}$, whereas the latter may represent a lower limit (Müller 2001).

### 6.3 Activity of deep and intermediate-depth low-frequency earthquakes

In 1995 September, the first significant volcanic tremor and ILFs were observed at a depth of 8 km beneath Iwate volcano (Ueki et al. 1996). Hence, to examine the temporal change in the seismic activity from 1995 September, we relocate the hypocentres of DLFs and ILFs only from the data of the permanent stations that operated at that time. To reduce the errors in the hypocentral locations, we first calculate the station corrections for the stations installed before 1998. We use the residuals of the arrival times for the DLFs and ILFs located by the JHD method in the previous section.

We plot focal depths of ILFs in the period from 1995 September to 2000 June in Fig. 10(a). We find that remarkable temporal changes occur in focal depth from 1998 April to 2000 June. The hypocentres were mainly distributed in the depth range 7–10 km from 1995 September to 1998 April, whereas the hypocentral range expanded to 5–12 km depth from 1998 April to October. Most of the ILFs occurred at depths of 8–12 km until 1999 September; then the depth range changed again to 6–11 km. As suggested for the cases of vertical migration of seismicity at Mt St Helens and Mt Spurr (Shemeta & Weaver 1986; Power et al. 1995), here the migrations are considered to be a manifestation of recharge and migration of magma and/or readjustments of stress to magma movement around the hypocentral region.
Low-frequency earthquakes at Iwate volcano

For example, during the 1991 catastrophic eruption of Pinatubo volcano, Philippines, the largest eruption in the 20th century, a striking temporal correlation between DLF activity and geological and seismological changes at the surface was observed (White 1996). At Mount Spurr, Alaska, DLF activity began to increase 4 months before the 1992 eruptions and peaked after the eruptions (Power et al. 1995). Four DLF earthquakes beneath Miyake Island, Japan were observed in the period from 1999 January to 2000 April before the 2000 eruption (Fujita & Ukawa 2000). Although the activity of Iwate volcano has gradually decreased and no eruption has been reported as of 2003 June, the temporal correlation of DLFs and shallow activity is probably a representation of physical connections between the deep and shallow regions. However, our analyses show that no systematic features are found in the source mechanisms of DLFs. These results suggest that the magmatic system connecting the deep region (that is the source region of DLFs) to the shallow part of the volcano is not simple and a complex magma plumbing system has been activated in both deep and shallow crust during the volcanic unrest of Iwate volcano.

7 CONCLUSIONS

We have determined the hypocentre distributions and source mechanisms of the DLFs and ILFs by analysing data from a dense seismic network in and around Iwate volcano. The main results are summarized as follows: (1) DLFs are located within three concentrated regions: the first group is located at 32 km depth, approximately 10 km south of the summit, the second is at 33 km depth, approximately 10 km northeast of the summit and the third is at 37 km depth approximately 7 km northeast of the summit. ILFs are located within a nearly vertical pipe-like region extending in depth from 5 to 12 km beneath the volcano. (2) Source mechanisms of DLFs and ILFs consist of CLVD and DC components. Some DLFs show a significant volumetric component in their source mechanism. (3) Opening of a tensile crack coupled with a shear crack or with the volume change of an oblate spheroid magma chamber can generate DLFs and ILFs of volumetric, CLVD and DC mechanisms. (4) No characteristic orientations of the principal axes of the moment tensors of DLFs and ILFs are recognized. The above results suggest the existence of a complex magma system at the concentrated focal regions of DLFs and ILFs.

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Figure 10. Depth–time diagram of (a) ILFs and (b) DLFs in the period from 1995 September to 2000 June. The grey squares and diamonds show the depths of ILFs and the south DLFs (group A), respectively. The grey stars show the depths of the northeast DLFs (groups B and C).

REFERENCES


Figure 11. (a) Cumulative frequencies of SHFs, ILFs and DLFs plotted against time during 1998 and 1999. Thin, thick and very thick lines show the cumulative numbers of SHFs, ILFs and DLFs, respectively. An arrow indicates the $M = 6.1$ earthquake of 1998 September 3. The SHFs are located in the region bounded by the latitudes of 39.8° N and 39.9° N and longitudes of 140.85° E and 141.10° E. (b) Cumulative frequencies of SHFs and DLFs plotted versus time from 1998 March to May. Arrows indicate the onsets of the abrupt increases of SHFs.


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