Sensitivity of Boreal Forest Carbon Balance to Soil Thaw

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We used eddy covariance; gas-exchange chambers; radiocarbon analysis; wood, moss, and soil inventories; and laboratory incubations to measure the carbon balance of a 120-year-old black spruce forest in Manitoba, Canada. The site lost 0.3 ± 0.5 metric ton of carbon per hectare per year (ton C ha^{-1} year^{-1}) from 1994 to 1997, with a gain of 0.6 ± 0.2 ton C ha^{-1} year^{-1} in moss and wood offset by a loss of 0.8 ± 0.5 ton C ha^{-1} year^{-1} from the soil. The soil remained frozen most of the year, and the decomposition of organic matter in the soil increased 10-fold upon thawing. The stability of the soil carbon pool ($\sim$150 tons C ha^{-1}) appears sensitive to the depth and duration of thaw, and climatic changes that promote thaw are likely to cause a net efflux of carbon dioxide from the site.

The climate of the boreal zone has warmed significantly in this century and is predicted to warm further in the next century [1]. The seasonally and perennially frozen soils of boreal forests contain one of the largest pools of carbon in the terrestrial biosphere, 200 to 500 gigatons of carbon (1 Gt = 10^9 metric tons) [2], an amount that could increase the concentration of CO_2 in the atmosphere by as much as 50% if it were released by climatic warming [3, 4]. However, the sensitivity of boreal carbon balance to temperature, and hence the magnitude and even direction of such a feedback, is unknown.

We used the eddy covariance technique [5], soil gas-exchange chambers [6], natural abundance radiocarbon analysis [7], and wood, moss, and soil inventories [8, 9] to investigate the carbon balance of a 120-year-old black spruce forest in Manitoba, Canada, from 1994 to 1997 [10]. The site was typical of much of the North American boreal forest [11]. Natural vegetation extended for several kilometers in all directions across a flat landscape. Dense black spruce (10 m tall) and feather moss grew in slightly higher, better drained areas, and sparse spruce (1 to 6 m tall) and sphagnum

moss grew in lower, wetter locations. About 45% of the area within 500 m of the eddy flux tower was covered by feather moss, 45% was covered by sphagnum moss, and the remaining 10% was covered by fen [8].

The winters were harsh. The minimum temperature observed was $-52°C$ and 3 or 4 months of each year had daily highs of $-10°$ to $-25°C$ (Fig. 1C). The growing season began and ended with abrupt temperature transitions, usually in May and October. Daytime temperatures in the growing season were 15° to 25°C, and July and August were nearly frost free. The soil

and lower moss layers warmed gradually in summer and stayed below air temperature until October (Fig. 1C) [12]. Patches of soil remained frozen in late summer beneath sphagnum hummocks and beneath feather moss at depths of 75 to 100 cm, a pattern of discontinuous permafrost that is common in central Canada [13].

The site contained 40 ± 20 tons C ha^{-1} in aboveground live and dead spruce [9] and 45 ± 13 tons C ha^{-1} in live and dead moss [7, 8]. The largest pool of carbon was in the soil—200 ± 50 tons C ha^{-1} in the sphagnum areas and 90 ± 20 tons C ha^{-1} in the feather moss areas [8]. These stocks are typical of northern forests [2]. Most of the soil carbon was in a humified layer we refer to as the deep pool, which was $\sim$40 to 80 cm beneath the surface and just below the depth penetrated by fires. The deep carbon had bulk ages of several hundred to 7000 ^{14}C years before present [7] and is considered to be nearly inert in most biogeochemical models.

We made eddy covariance measurements of whole forest CO_2 exchange for 22,000 hours from March 1994 to October 1997 [5]. Daily net uptake started in early May, peaked at 10 to 15 kg C ha^{-1} day^{-1} from late May to early July, and declined to near zero in August and September (Fig. 1A). The forest lost 6 to 8 kg C ha^{-1} day^{-1} in October and 2 to 3 kg C ha^{-1} day^{-1} from December to April. The absolute rates and seasonal course of exchange were markedly different from those ob-

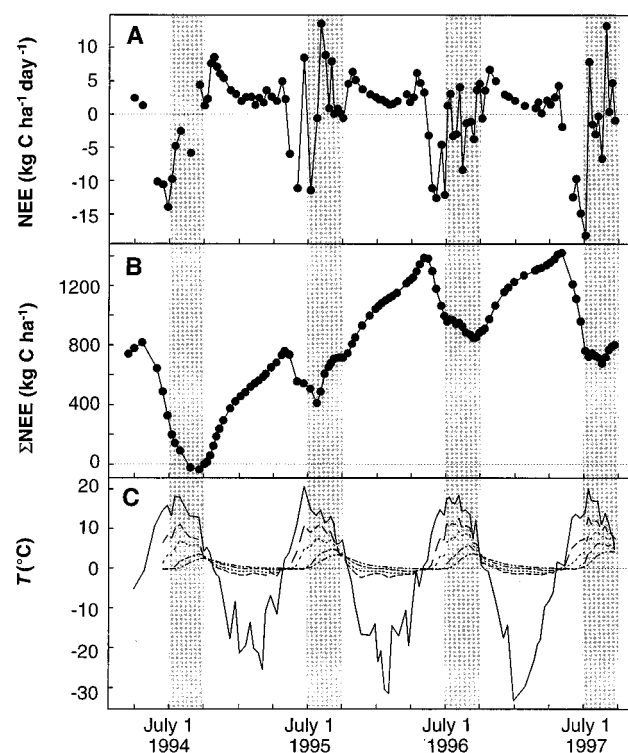M. L. Goulden, S. C. Wofsy, B. C. Daube, S.-M. Fan, D. J. Sutton, A. Bazzaz, J. W. Munger, Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA 02138, USA.
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Fig. 1. (**A**) Daily carbon balance measured by eddy correlation (NEE). (**B**) Accumulated CO_2 exchange (ΣNEE) starting at zero on 1 October 1994. (**C**) Daily mean temperatures at a height of 30 m (——), 10 cm beneath the sphagnum moss surface (– – –), 20 cm beneath the surface (---), 50 cm beneath the surface (T_{50}) (– - - - –), and 100 cm beneath the surface (– -- –). Lines connect means for 7 days with complete observations; unconnected points in (A) indicate a gap in the observations of at least 13 days. Vertical bands begin 1 July and end 30 September. Positive fluxes in (A) indicate a net efflux of CO_2 from the forest [5, 14, 15].

served above a deciduous oak and maple stand at the Harvard Forest in central Massachusetts (*14*). Carbon accumulation at Harvard Forest began in late May and continued at a comparatively high rate into September. Net uptake at Harvard Forest lagged that at the evergreen boreal site by a month or more even though Harvard Forest is 13° farther south.

The integrated eddy flux measurements at the boreal site showed a loss of 0.7 ± 0.5 ton C ha^{-1} from October 1994 to October 1995, a loss of 0.2 ± 0.5 ton C ha^{-1} from October 1995 to October 1996, and a gain of 0.1 ± 0.5 ton C ha^{-1} from October 1996 to October 1997 (Fig. 1B) (*15, 16*). We found no evidence of high rates of carbon accumulation, possibly associated with nitrogen deposition, elevated CO_2, or climatic warming, as has been suggested is occurring in the boreal zone on the basis of analyses of atmospheric CO_2 (*17*). We emphasize, however, that the forest is relatively old, and stands in the area that are younger or dominated by deciduous vegetation may be accumulating carbon at high rates. The annual gross production remained consistent from year to year at ~8.0 tons C ha^{-1} year^{-1}, whereas annual respiration ranged from 8.9 tons C ha^{-1} year^{-1} in 1995 to 7.9 tons C ha^{-1} year^{-1} in 1997 (*18*). The carbon balance was the small residual of two much larger fluxes, a pattern that may make it extremely sensitive to climate.

Analyses of forest recovery after fire indicated that, on average, 120-year-old black spruce stands in the area accumulate 0.2 to 0.4 ton C ha^{-1} year^{-1} in the moss layer (*8*). Measurements of stem increment at the site showed that 0.2 to 0.4 ton C ha^{-1} year^{-1} has been stored in woody biomass during the past 10 years (*9*). This combined gain (0.6 ± 0.2 ton C ha^{-1} year^{-1}), coupled with the loss measured by eddy covariance (0.3 ± 0.5 ton C ha^{-1} year^{-1}) (Fig. 1B), implies that 0.9 ± 0.6 ton C ha^{-1} year^{-1} was lost from the deep pool.

A direct measurement of the loss of soil carbon was not feasible because of the heterogeneity, although several lines of indirect evidence support this conclusion. The midsummer decline in net uptake (Fig. 1A) was associated with an increase in respiration. Respiration increased by ~10 kg C ha^{-1} day^{-1} from July to August, whereas photosynthesis remained relatively constant except for a decline of ~5 kg C ha^{-1} day^{-1} caused by decreasing day length. The midsummer increase in respiration was not caused by warmer air temperatures (Fig. 1C); whole-forest respiration, after controlling for air temperature by considering only 8° to 12°C windy nights, increased from 41 kg C ha^{-1} day^{-1} in July to 51 kg C ha^{-1} day^{-1} in August (Fig. 2). Rather, the midsummer in-

crease in respiration coincided with warming at depths of 20 to 100 cm (Fig. 1C).

The nocturnal CO_2 efflux at the moss surface measured with automated chambers (*6*) showed a midsummer increase comparable to that observed by eddy covariance. From June to August moss surface respiration increased by 8 to 15 kg C ha^{-1} day^{-1} at all three sphagnum sites measured independent of moss temperature and by 10 kg C ha^{-1} day^{-1} at one of two feather sites (Fig. 3A). This seasonal increase in respiration (R_{deep}) (*19*) was correlated with the soil temperature at a depth of 50 cm (T_{50}) (Fig. 3B). These observations eliminate the possibilities that aboveground or moss-layer sources were responsible for the midsummer increase in respiration because the increase was insensitive to diurnal and synoptic variations in air temperature. The measurements are also inconsistent with an increase in root growth being responsible because the increase was smallest beneath feather moss, where root respiration appears to be greatest (*20*). Instead, the observations imply that the decomposition of deep carbon increased in midsummer by ~10 kg C ha^{-1} day^{-1}.

Analyses of $^{14}CO_2$ in the pore space of the moss layer provided further evidence of the decomposition of deep carbon (*7*). The $\Delta^{14}C$ of moss-layer CO_2 in late fall and winter, after correction for atmospheric dilution, was −80 to −150‰, corresponding to a bulk source age of 600 to 1300 ^{14}C years. This depletion of $^{14}CO_2$ indicates that autotrophic respiration was minor and that deep decomposition was the main source of the winter efflux observed by eddy covariance (2 to 3 kg C ha^{-1} day^{-1}) and the similar efflux observed from the snowpack with chambers (*7*).

The sensitivity of R_{deep} to T_{50} (Fig. 3B) was much greater than is considered typical for soil respiration (*21*), which suggests that deep decomposition was controlled by a mechanism other than the direct effect of temperature on enzyme activity. Laboratory and field studies have shown that anoxia
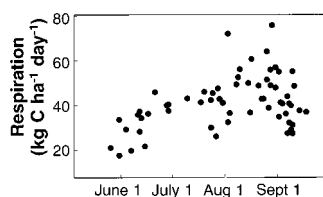
due to flooding controls decomposition of organic matter in tundra soils (*3, 4*), and the apparent high sensitivity of R_{deep} to T_{50} could have resulted from changes in watertable depth that correlated with T_{50}. However, soil at the site did not dry out in midsummer (*22*), and R_{deep} did not vary with rainfall (Fig. 3B).

An alternative explanation is that T_{50} was related to the depth of the active layer and that the midsummer increase in respiration was a result of the increasing volume of unfrozen soil. The CO_2 efflux from deep soil incubated in the laboratory increased from 0.016 kg C per ton C day^{-1} at −2°C to 0.15 kg C per ton C day^{-1} at 5°C (*23*). These observations correspond to an increase at the site from 1.5 to 3 kg C ha^{-1} day^{-1} in completely frozen soils to 15 to 30 kg C ha^{-1} day^{-1} in thawed soils. These rates are consistent with the field observations (Figs. 1 to 3) and provide evidence that soil thaw directly controlled the seasonal cycle of deep decomposition. In contrast, chamber measurements in nearby wetlands do not show a seasonal increase in deep decomposition (R_{deep}), apparently because anoxia at those sites limits decomposition even after thaw (*24*).
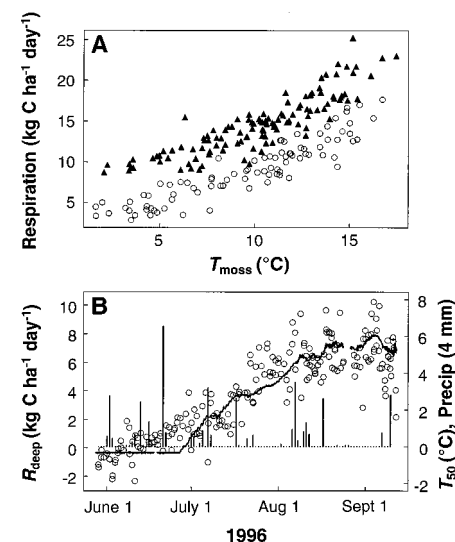
Fig. 2. Nocturnal CO_2 efflux (whole forest respiration) on windy nights with air temperatures of 8° to 12°C. Points are means for individual nights with at least four valid 30-min observations (*5, 14*). Both respiration and photosynthesis were reduced during periods with freezing nights (*5*), typically before mid-June and after early September, apparently because of changes in leaf activity.

Fig. 3. (**A**) Nocturnal respiration at a sphagnum moss surface as a function of moss temperature (T_{moss}) before 18 July 1996 (open circles), and after 19 July 1996 (filled triangles). The efflux of CO_2, which reflects the combined rate of moss, root, and microbial metabolism, was measured every 3 hours with an automated chamber (*6*). (**B**) Seasonal increase in respiration (R_{deep}) (open circles) [see (*19*)], soil temperature 50 cm beneath a sphagnum surface (T_{50}) (solid line), and rainfall (millimeters divided by 4) (vertical lines). The least squares fit in (B) was R_{deep} (kg C ha^{-1} day^{-1}) = 0.76 + 1.09 × T_{50} (°C) ($n = 207$, $r^2 = 0.69$). The slopes were 1.7 and 1.3 at the other sphagnum chambers and 1.5 and −0.4 at the feather moss chambers.

The eddy covariance, chamber, and laboratory observations all indicate a deep decomposition in late summer of ~10 kg C ha^{-1} day^{-1} and an annual deep decomposition of 0.5 to 1.5 tons C ha^{-1} (25). The corresponding turnover time for soil carbon, ~150 years, is shorter than the mean age measured with ^{14}C, which implies that the deep pool was not in steady state. The long-term input of carbon to the deep pool, based on ^{14}C measurements, is ~0.2 ton C ha^{-1} year^{-1} (7), which indicates a net loss during the study of 0.3 to 1.3 tons C ha^{-1} year^{-1}. This confirms the interpretation that a loss of carbon from the soil offsets the gains in moss and wood. A similar loss of carbon has been reported for tundra in Alaska (4) and attributed to climatic warming. It is possible that the loss of carbon from the black spruce site likewise resulted from warming. The mean air temperature and the diurnal range, an indicator of sunlight (12), were generally above average during the early summers of the study (26), possibly resulting in increased soil thaw. Moreover, the interannual variability in carbon balance was correlated with the early summer air temperatures (Fig. 1) (27).

The global mean temperature is predicted to increase ~2°C by the year 2100 (1). Warming of this magnitude would likely eliminate permafrost at the site (13) and, assuming sufficient drainage, significantly increase the decomposition of deep carbon. Warming may also stimulate plant production, although we expect this effect would be modest. The air temperatures in midsummer are already optimal for photosynthesis (5), and the rapid changes in temperature in spring and fall (Fig. 1C) limit the associated increase in length of the growing season (28). The soil currently contains more carbon than is stored in the vegetation of a mature temperate deciduous or boreal forest (2). An indirect stimulation of production, caused either by increased nutrient mineralization or by an invasion of deciduous trees (3), would have to be large to offset the expected loss of soil carbon. Changes in climate that promote thaw are therefore likely to cause a net loss of carbon from evergreen boreal ecosystems of the type studied.

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9. S. T. Gower *et al.*, *J. Geophys. Res.*, in press. The 1985 to 1994 wood increment in four 15 × 15 m plots on feather moss was calculated from ring width by using site-specific allometric equations. Sequential inventories in 1993 and 1994 found no spruce mortality. The sphagnum areas were not sampled and we assume that the wood increment there was negligible.
10. The northern study area Old Black Spruce site (NOBS) of the Boreal Ecosystem–Atmosphere Study (BOREAS) [P. J. Sellers *et al.*, *Bull. Am. Met. Soc.* **76**, 1549 (1995)] is 40 km west of Thompson, 25 km east of Nelson House, and 4 km south of Route 391 (55.879°N, 98.484°W).
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15. We calculated the daily and annual carbon balances as described in (14) and with a 7-day averaging interval and a minimum friction velocity of 0.2 m/s (5). An analysis of the local energy budget indicated an underestimation of turbulent flux by up to 20% (5), but this would have little effect on the calculated annual carbon balance provided that it was a uniform bias (14). The largest uncertainty in daily and annual carbon balances was associated with selective biases from day to night (14). We attempted to correct for these (5) but note that the treatment of calm nights is controversial, and therefore we attach an uncertainty of ±0.5 ton C ha^{-1} year^{-1} to account for remaining errors.
16. Similarly, A. Grelle [thesis, Swedish University of Agricultural Sciences, Uppsala (1997)], in a long-term record of CO_2 exchange above a 50- to 100-year-old spruce and pine boreal forest in Sweden, observed an annual loss of CO_2.
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18. The observations of net exchange were separated into respiratory and photosynthetic fluxes by estimating daytime respiration based on a series of regressions between CO_2 efflux on windy nights and air temperature (5, 14).
19. Shallow respiration ($R_{shallow}$) was calculated from the exponential regression between measured respiration and the temperature at a depth of 5 cm (T_{moss}) before deep soil thaw (24 May to 23 June). Deep respiration (R_{deep}) was calculated as the difference between measured respiration and $R_{shallow}$. $R_{shallow}$ is the component of respiration that responds to diurnal and synoptic variation in temperature; R_{deep} is the component that is insensitive to short-term temperature variation. Diurnal changes in air temperature propagated to a depth of about 20 cm, implying that $R_{shallow}$ is mostly live moss respiration and R_{deep} is root respiration and deep decomposition.
20. The chamber measurements integrated from May to October 1996 show a net efflux of 4 to 6 tons C ha^{-1} at the feather moss sites and no net exchange at the sphagnum sites. This pattern apparently reflects a large difference in root respiration and litter input.
21. J. W. Raich and W. H. Schlesinger, *Tellus Ser. B Chem. Phys. Meteorol.* **44**, 81 (1992); but see J. Lloyd and J. A. Taylor [*Funct. Ecol.* **8**, 315 (1994)] and M. Kirschbaum [*Soil Biol. Biochem.* **27**, 753 (1995)].
22. As measured by time-domain reflectometry within 100 m of the tower [R. H. Cuenca, S. F. Kelly, D. Gochis, *Eos*, Nov. 12 Suppl. (1996)].
23. Soil from the humified layer was excavated and shipped to Menlo Park with minimum disturbance. Soil samples were placed in air-tight, foil-wrapped containers with only the top and bottom surfaces of the cores exposed and held at −2° and +5°C. The CO_2 in the headspace was determined periodically over 100 days and the linear fit of CO_2 versus time was used to calculate respiration. We do not know whether all of the deep carbon is labile and note that some of it is charcoal and probably recalcitrant. See also J. S. Clein and J. P. Schimel [*Soil Biol. Biochem.* **27**, 1231 (1995)].
24. J. Bubier, unpublished material.
25. The soil at a depth of 50 to 100 cm thawed for 100 to 50 days each year (Fig. 1C), implying the decomposition of 0.5 to 1.0 ton C ha^{-1} from midsummer to early fall. The CO_2 efflux during intervals with freezing soil temperatures was ~0.5 ton C ha^{-1} year^{-1}, leading us to estimate an annual deep decomposition of 1.0 ± 0.5 ton C ha^{-1} year^{-1}. Regression analyses (Fig. 3B) also imply ~1.0 ton C ha^{-1} year^{-1}, whereas the laboratory incubations (23) suggest a somewhat higher rate.
26. The mean temperature at the Thompson Airport (40 km east) from 1 June to 15 July was 13.6°C in 1994, 15.0°C in 1995, and 15.2°C in 1996. The comparable mean from 1969 to 1996 was 13.1°C, and the intervals in 1995 and 1996 were among the five warmest on record. The mean temperature at the site was 1°C cooler in 1997 than in previous years, indicating near-normal temperatures in 1997. The diurnal range from 1 June to 15 July was 15.2°C in 1994, 15.5°C in 1995, 14.3°C in 1996, and 14.1°C from 1969 to 1996. The range in 1995 was the second largest on record and implies that solar radiation was 8% above average (12). The annual mean temperatures during the study were close to the 1969–1996 average. See also J. Overpeck *et al.*, *Science* **278**, 1251 (1997) and W. L. Chapman and J. E. Walsh, *Bull. Am. Meteorol. Soc.* **74**, 33 (1993).
27. The interannual trends in carbon balance were not obviously related to variation in deep soil temperature (Fig. 1). The reliability of these measurements as

an indicator of interannual variability in thaw depth is uncertain because placement of the probes may have disturbed the local conditions, one of the sites was disturbed by the installation of ancillary equipment in 1995, and the horizontal distribution of soil frost may have varied from year to year.

28. A uniform 2.5°C increase in temperature would in-

crease the annual number of sunlit hours on days with mean temperatures above freezing by ~8%. The length of the growing season at Harvard Forest, where warming in spring is more gradual and the growing season is shorter because the forest is deciduous, is more sensitive to warming (5, 14).

29. This work was supported by the U.S. National Aero-

Evidence of Shallow Fault Zone Strengthening After the 1992 *M*7.5 Landers, California, Earthquake

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Repeated seismic surveys of the Landers, California, fault zone that ruptured in the magnitude (*M*) 7.5 earthquake of 1992 reveal an increase in seismic velocity with time. *P*, *S*, and fault zone trapped waves were excited by near-surface explosions in two locations in 1994 and 1996, and were recorded on two linear, three-component seismic arrays deployed across the Johnson Valley fault trace. The travel times of *P* and *S* waves for identical shot-receiver pairs decreased by 0.5 to 1.5 percent from 1994 to 1996, with the larger changes at stations located within the fault zone. These observations indicate that the shallow Johnson Valley fault is strengthening after the main shock, most likely because of closure of cracks that were opened by the 1992 earthquake. The increase in velocity is consistent with the prevalence of dry over wet cracks and with a reduction in the apparent crack density near the fault zone by approximately 1.0 percent from 1994 to 1996.

A fault plane undergoes sudden stresses, shaking, and cracking during an earthquake. Extensive research has been directed toward understanding this phenomenon (*1*, *2*), but many facets remain obscure.

We focus on the rate at which a fault regains its strength following a large earthquake. This rate is needed to understand how fault zones strengthen or "heal" after an earthquake, but so far, only simple laws have been assumed based on laboratory experiments rather than direct observations in the field. In addition, the healing rate may affect the probability of another earthquake in a fault zone. Experimental studies (*3*) indicate that a longer interval since the previous episode of faulting correlates with higher stress drop in the subsequent rupture. Studies of repeated earthquakes along a fault (*4*) show trends that are consistent with state- and rate-dependent healing models (*5*).

We had the particularly favorable situation of probing the evolution of a shallow

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fault that had recently undergone large displacements. Earlier efforts to identify dilatancy that might be detectable near fault zones failed because of low sensitivity coupled with what we now know to be subtle precursory changes due to dilatancy (*2*, *6*). Repeated surveys near Parkfield (*7*) is showing small changes in velocity over time, but in the absence of a large earthquake and with uncertainty about the precise location of the velocity change, its significance is hard to assess. A comparison of earthquakes before and after a specific large event showed a small coseismic reduction in wave velocity at stations with unconsolidated sedimentary rocks that were strongly shaken (*8*), suggesting the temporal change was a shallow effect of shaking rather than a physical change in the bedrock. One study (*9*) suggests changes in scattering of *P* waves with a 1-s period around the time of large earthquakes, but the pattern is not yet well established.

We conducted two identical seismic experiments on 2 November 1994 and 6 August 1996 (Fig. 1) to monitor the change of fault zone physical properties after the 1992 M7.5 Landers earthquake. A pair of explosions, or "shots," in 30-m boreholes along the Johnson Valley fault segment of the Landers fault zone were detonated in each experiment, using 400 to 700 pounds of chemical emulsions for each. A pair of lin-

ear three-component seismic arrays recorded the arrivals of seismic waves for each explosion. The arrays were 3-km in length and aligned perpendicular to the fault. The two arrays were separated by 13 km, and the explosions were located between the arrays (Fig. 1). The array along line 1 had 36 stations and the array along line 3 had 21 stations.

Line 1 is centered at the region that experienced the maximum amount of slip—about 3 m—on the Johnson Valley fault during the Landers earthquake (Fig. 1). Slip is smaller near line 3, and also diminished to the north until the surface rupture connected with the Homestead Valley fault. Fault slip at depth is more difficult to determine, but seems to resemble the slip at the surface (*10*). The recurrence of faulting on the Johnson Valley fault is estimated to exceed 1000 years (*11*, *12*).

The data from line 3, collected in 1996, had *P* waves visible on all traces near 1 s. The *S* waves had a longer period and were more prominent on the horizontal components near 2 s, and the fault zone trapped modes appeared from 3 to 8 s (Fig. 2). The trapped waves showed larger amplitudes,
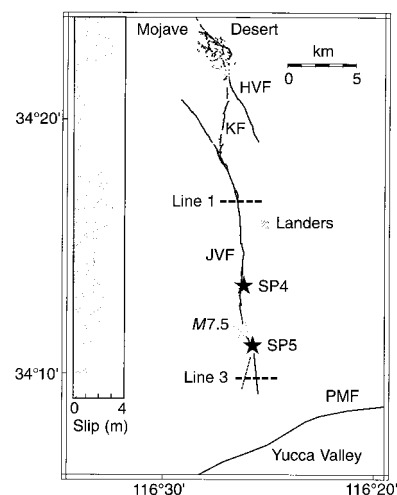

Fig. 1. Map of the study region showing locations of two seismic arrays at line 1 and line 3 and two explosions SP4 and SP5 in the fault zone of the 1992 Landers, California, earthquake. Only the southern half of the Landers rupture lies within this map, and the dextral surface-fault slip profile is shown (inset) to the left of the map (*12*). JVF, Johnson Valley fault; KF, Kickapoo fault; HVF, Homestead Valley fault; and PMF, Pinto Mountain fault.