tionary events occurred outside the peat and alluvial depositional areas in which plants are most often preserved.

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Cumulate Xenolith in Oahu, Hawaii: Implications for Deep Magma Chambers and Hawaiian Volcanism

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The maximum depth at which large (>1000 km^3) terrestrial mafic magma chambers can form has generally been thought to be the Moho, which occurs at a mean depth of about 35 kilometers beneath the continents and 8 kilometers beneath ocean basins. However, the presence of layers of cumulus magnesium-rich spinel and olivine and intercumulus garnet in an unusual mantle xenolith from Oahu, Hawaii, suggests that this rock is a fragment of a large magma chamber that formed at a depth of about 90 kilometers; Hawaiian shield-building magmas may pond and fractionate in such magma chambers before continuing their ascent. This depth is at or near the base of the 90-million-year-old lithosphere beneath Oahu; thus, rejuvenated stage alkalic magmas containing mantle xenoliths evidently also originate below the lithosphere.

THE DEEPEST LEVEL AT WHICH A large mafic magma chamber can form depends upon the extent of dynamic equilibrium maintained between magma in the chamber and the wall rock and the balance between the deviatoric stress and lithostatic pressure in the mantle where the magma chamber may be located. Magma chambers beneath the ocean ridges are not thought to form below the petrologic Moho, that is, 8 to 10 km below sea level (*1–3*). Seismic studies have been able to detect shallow magma chambers and magma pathways (*4*), but so far have failed to detect deep (~100 km), large magma chambers beneath the ocean basins, perhaps because of resolution problems. The deepest continental magma chambers are generally thought to occur near the Moho [mean depth ~35 km (*1*)]. Inferences on physical and chemical attributes of terrestrial magma chambers have emerged largely through studies of very large (≥1000 km^3) layered intrusions, which represent frozen magma chambers exposed at the surface (*2*). These chambers contain a characteristic rock type, cumulates, which formed by crystallization, accumulation, and segregation of early crystallizing minerals (cumulus) from the magmas (*1*). In this report we suggest that a 90-km-deep magma chamber was present beneath the island of Oahu on the basis of a garnet-bearing cumulate xenolith with layered texture.

The evolution of a typical Hawaiian volcano is marked by four different stages (*5*). In the initial preshield stage, diverse lava types (basanite to tholeiite) erupt. Next, in the shield stage (the main stage), voluminous eruptions of dominantly tholeiitic lavas occur and a caldera develops. The shield stage is usually followed by postshield stage when alkalic basalt (and its differentiates) erupts, fills the collapsed caldera, and forms a thin cap over the shield. In the rejuvenated stage, typically following a period of quiescence and erosion, small volumes of alkalic magmas, many carrying mantle xenoliths, erupt through small vents scattered across the shield.

The eastern part of Oahu is built of a large, tholeiitic shield volcano, known as the Koolau volcano (*5*), that formed 1.8 to 2.7 million years ago. Following cessation of tholeiitic basalt eruption, rejuvenated stage alkalic lavas of the Honolulu Volcanics (HV) erupted 0.3 to 0.6 million years ago through several vents that cut across the volcano (*6*). The HV lavas contain a large number of mantle and crustal xenoliths from a range of depths (*7*); spinel lherzolites are the most abundant upper mantle xenoliths. Garnet-bearing pyroxenites, wehrlites, websterites, and lherzolites are much less common and occur only at the Salt Lake crater at the southwestern flank of the Koolau shield volcano (*8*). The spinel lherzolites are samples of variably enriched mantle lithosphere, and the garnet-bearing rocks are fragments of 60- to 80-km-deep dikes (*9*). Dunite xenoliths are also common, and they represent cumulates that formed in crustal magma chambers (maximum depth ~15 km) of the Koolau volcano (*10*).

We identified a garnet- and spinel-bearing cumulate dunite [sample no. NMNH 114881-8, Dale Jackson Collection, Smithsonian Institution] that was found to occur as an inclusion (xenolith) in HV lavas from the Salt Lake crater. In this rock, discontinuous layers of cumulus spinel grains are interlayered with cumulus olivine grains. Garnet occurs as an intercumulus phase in both layers and in the transition zone (mixed olivine and spinel layer) between adjacent spinel and olivine layers (Fig. 1). Grain sizes of the cumulus phases vary considerably even in a single layer (Fig. 1A). Small (0.25 mm^2) round spinel inclusions are common in olivine grains of the mixed olivine and spinel layer, but olivine grains of the olivine-rich layers are generally free of spinel inclusions. Spinel grains in the spinel layers are coarse to medium size (2 to 7 mm^2) and euhedral to round. Cumulus olivine crystals are mostly rounded, a texture suggesting that they were a reaction relation with the magma. Orthopyroxene forms a thin rim between spinel and garnet, and it may have formed by some late reaction between garnet, spinel, and the magma, such as spinel + magma = garnet + orthopyroxene. Large (2 mm) phlogopite grains also occur interstitially in one area of the xenolith. Whether the phlogopite crystallized from the interstitial melt (that is, primary) or was a later product of metasomatic crystallization is not clear. The petrographic observations suggest that (i) initially spinel and olivine crystallized together as cumulus phases (spinel may have started earlier) and (ii) intercumulus garnet later formed interstitially. The layering may have formed by in situ oscillatory crystallization processes of the type proposed for layered intrusions (*1*). In several continental layered intrusions, such as the Stillwater intrusion, alternate layers of spinel

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and olivine occur, however, in these the spinel is a chromite (*1*) and not a pleonaste spinel of the type found in the xenolith (Table 1). The most significant difference between cumulates of layered intrusions and this xenolith is the presence of garnet in the xenolith, which indicates that it formed within the uppermost mantle. On the basis of the textural similarities between this xenolith and those in continental layered intrusions and cumulate sections of ophiolites, we thus argue that this xenolith originally formed by cumulus processes similar to those that have been inferred for crustal layered intrusions.

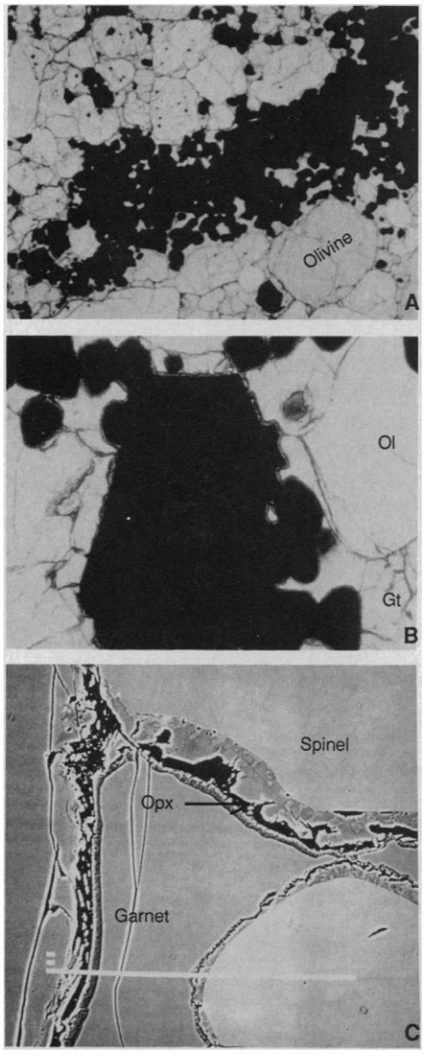

Fig. 1. (**A**) Layered cumulate texture of xenolith NMNH# 114881-8. The dark layer is dominantly composed of euhedral cumulus grains of spinel. The colorless layers are olivine (plane polarized light; 70 mm lengthwise). (**B**) A close-up of a euhedral spinel and small round spinel grains enclosed in intercumulus garnet (plane polarized light; 9 mm lengthwise). (**C**) Back-scattered electron photograph of resorbed spinel, intercumulus garnet, and orthopyroxene rim containing stringers of broken down spinel (length of the long bar is 100 μm). Sp, spinel; Ol, olivine; Cpx, clinopyroxene.

Although the size or shape of the magma chamber in which this cumulate rock formed are matters of speculation, it is unlikely that it formed in a small dike or sill-like chamber. This conclusion is based on the observation that garnet pyroxenite xenoliths of Oahu that have been inferred to represent deep-seated (75 km) veins, dikes, and sills (*9*) do not have a cumulate texture. Therefore, by analogy with crustal layered intrusions, we hypothesize that the cumulate xenolith formed in a relatively large (perhaps ~1000 km^3) magma chamber.

We determined the compositions of the minerals of this xenolith with a Cameca electron microprobe using the techniques of Sen and Jones (*11*). Individual grains of all phases are compositionally homogeneous within analytical uncertainty. Olivine is Fo_{84} (Table 1), a composition that overlaps with that of olivines in the dunite-suite xenoliths of Oahu that have been interpreted as deformed cumulates (*10*). Spinel is highly aluminous and Ti- and Cr-poor and is similar to the spinel in pyroxenite-suite xenoliths of Oahu, which have been interpreted to be of magmatic origin (*9*). There is significant difference in Cr/Al ratios between large euhedral spinel grains and small irregular stringers that are dispersed throughout the orthopyroxene (Table 1). Garnet ($Py_{64}Gr_{14}Alm_{22}$) is also compositionally similar to that of the pyroxenite-suite xenoliths of Oahu (*9*). Orthopyroxene is strongly aluminous but has a variable Al content (Table 1 presents two extreme compositions), which suggests disequilibrium.

Liquidus phase relationships on the join diopside-enstatite-pyrope-grossular in the system $CaO-MgO-Al_2O_3-SiO_2$ (CMAS) have direct relevance to the conditions of crystallization of this cumulate rock. In the phase diagram of Kushiro and Yoder (*12*), a piercing point (**u** in Fig. 2) representing the isobaric univariant curve fo + sp + gt + l (*13*) occurs on this join at 26 kbar. We infer that initial crystallization of cumulus olivine and spinel occurred in a magma chamber along the divariant surface ol + sp + l. Intercumulus crystallization of garnet occurred when the composition of the interstitial magma reached the univariant line ol + sp + gt + l. Furthermore, the resorption texture shown by the olivine crystals substantiates the experimental determination that the piercing point **u** is indeed a tributary reaction point (*12*) where olivine is consumed. This piercing point vanishes at a pressure between 26 and 20 kbar because of the expansion of the spinel liquidus field at the expense of garnet liquidus field at lower pressures and replacement of the piercing points gt + fo + di + l and sp + gt + fo + l with two new piercing points, namely, gt + di + sp + l and sp + fo + di + l (Fig. 2). Therefore, the minimum pressure at which the observed cumulus and intercumulus phase assemblages of the xenolith could have formed is between 20 and 26 kbar because the univariant line (and the piercing point) along which such an assemblage would crystallize does not occur at a lower pressure. At a pressure of about 28 kb, or the forsterite liquidus field collapses en-

Table 1. Composition of constituent minerals of 114881-8. Eu, big euhedral spinel grains; Irr, small irregular spinel grains in association with orthopyroxene reaction rim between garnet and spinel; Oxy, number of oxgens per formula unit. The number in parentheses is total number of grains analyzed. High cation totals reflect the presence of Fe^{3+}.

	Ol (7)	Gt (12)	Sp (eu)	Sp (Irr)	Opx[a]*	Opx[b]†
SiO_2	40.01	42.18	0.09	0.82	50.99	53.43
TiO_2	0.00	0.35	0.60	0.07	0.17	0.12
Al_2O_3	0.00	23.49	60.16	64.19	8.19	5.43
Cr_2O_3	0.00	0.14	1.55	0.21	0.17	0.09
FeO‡	15.77	11.33	17.77	15.33	12.04	11.51
MnO	0.19	0.40	0.08	0.25	0.43	0.40
NiO	0.32	0.00	0.00	0.00	0.00	0.00
MgO	45.16	18.32	18.73	19.07	27.03	27.72
CaO	0.07	5.47	0.02	0.09	1.66	1.97
Total	101.52	101.68	99.00	100.03	100.68	100.67
Oxy	4	12	4	4	6	6
Si	0.994	2.985	0.002	0.021	1.806	1.884
Ti	0.000	0.019	0.012	0.001	0.005	0.003
Al	0.000	1.959	1.864	1.927	0.342	0.226
Cr	0.000	0.008	0.032	0.004	0.005	0.003
Fe	0.328	0.670	0.391	0.327	0.356	0.340
Mn	0.004	0.024	0.002	0.005	0.013	0.012
Mg	1.672	1.932	0.734	0.724	1.427	1.457
Ca	0.002	0.415	0.000	0.002	0.063	0.074
Total	3.004	8.014	3.038	3.012	4.016	3.999

*High-Al orthopyroxene. †Low-Al othopyroxene ‡Total Fe.

tirely and thus the tributary piercing point gt + sp + fo + l vanishes. Therefore, on the basis of the phase diagram of Kushiro and Yoder, the maximum pressure at which the xenolith may have crystallized is between 26 and 28 kbar. In Kushiro and Yoder's diagram the garnet compositional field is continuous between grossular and pyrope. On the basis of new experimental evidence, Milholland and Presnall (*14*) pointed out that Kushiro and Yoder's diagram is incorrect and suggested that the garnet compositional field is discontinuous and that there are two separate primary phase fields—one for pyrope-rich garnet and the other for the grossular-rich garnet. The garnet in the xenolith is pyrope-rich. The univariant line of interest in the revised CMAS system is thus fo + py + sp + l, and it occurs only at pressures above 30 kbar. Although in the CMAS system the liquid coexisting with the olivine, spinel, and pyrope along the univariant line is tholeiitic, Milholland and Presnall (*14*) suggested that the liquid could be alkalic in a natural magmatic system at 30 kbar. Overall, therefore, the mineralogy and texture of the xenolith and the revised phase relations suggest that it crystallized at a pressure of at least 30 kbar. This pressure corresponds to an approximate depth of about 90 km, which is at or near the base of the lithosphere beneath Oahu (*9*). The above estimation assumes that the phase relations in the CMAS system closely approximate those in natural mafic magmas, as has been argued by many workers (*12*).

In summary, the garnet-bearing cumulate may have formed in a large mafic magma chamber at or near the base of the lithosphere beneath Oahu. The parent magma from which this cumulate crystallized may have been tholeiitic or alkalic (if other components do affect the CMAS phase relations). The occurrence of interstitial phlogopite may not be taken as a clue to the alkalic or tholeiitic parentage inasmuch as it is not certain whether this phlogopite is metasomatic or primary. Furthermore, the lack of appropriate experimental data concerning phlogopite stability in a differentiated mafic liquid at high pressure prohibits speculation on the magmatic parentage of the rock on the basis of phlogopite alone.

The parent magma of this cumulate xenolith was most likely not related to rejuvenated stage alkalic lavas for the following reasons. First, during crystallization of a mantle-derived primary basaltic magma at high pressure, a highly forsteritic olivine crystallizes at a higher temperature than clinopyroxene, and as the magma cools, the olivine becomes more Fe rich. Therefore, the olivine crystallizing with clinopyroxene will have a higher Fe content than the olivine (without cpx) that would crystallize at a higher temperature from the same magma. The olivine in the cumulate xenolith (Fo_{84}) is less forsteritic (more Fe-rich) than the most Mg-rich olivine (Fo_{85}) of the pyroxenite suite xenoliths, which contain a large amount of clinopyroxene and have been interpreted as high-pressure precipitates from rejuvenated stage HV parent magmas (*8*). Hence, the olivine compositions rule out any petrogenetic relation between the cumulate and pyroxenite suite xenoliths [and rejuvenated stage HV magmas parental to the pyroxenites (*8*)]. Second, the field where clinopyroxene is on the liquidus in the basalt tetrahedron expands at higher pressures, and therefore, a primary basalt magma will crystallize greater amounts of clinopyroxene at a higher pressure [(*8*) and references therein]. Pyroxenite suite xenoliths, which contain similarly forsteritic olivine as the layered cumulate, appear to have crystallized at a maximum pressure of 25

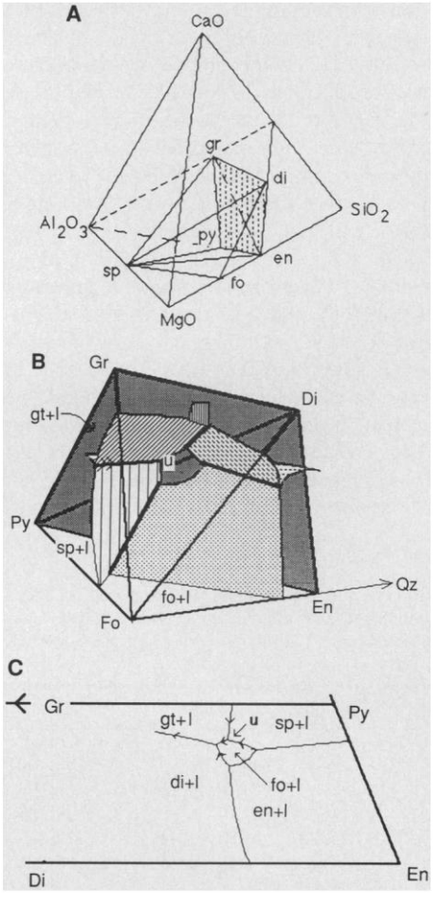

Fig. 2. Schematic isobaric liquidus phase diagram of the system CMAS (**A**) at 26 kbar (*12*) (di, diopside; fo, forsterite; sp, spinel; py, pyrope; gr, grossular; en, enstatite). (**B**) Schematic isobaric univariant curves (thick lines) in the Gr-Py-Fo-Qz part of the system. The ruled and dotted planes represent divariant surfaces, and the shaded plane is the join Gr-Di-En-Py. **u** represents a piercing point made by the univariant curve l + fo + sp + gt on the join Gr-Di-Py-En (l, liquid). Liquidus volumes of sp, gt, and fo are shown. (**C**) The piercing point **u** and the three other piercing points and the fo + l field on the join Gr-Di-En-Py, as determined by Kushiro and Yoder (*12*), are shown [drawn after (*12*)].

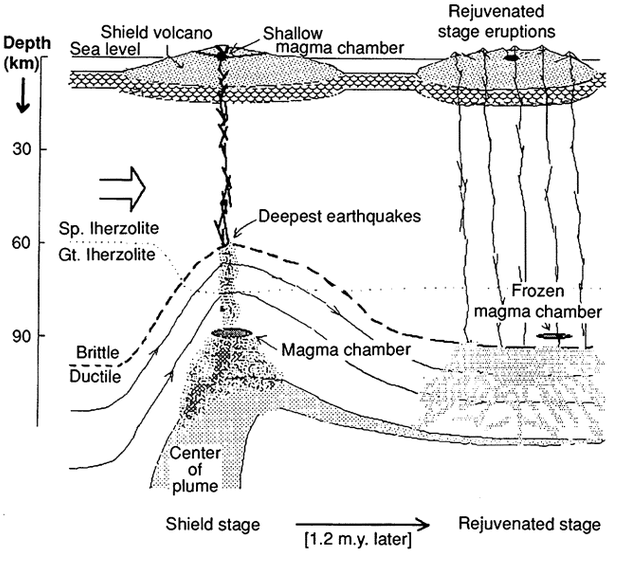

Fig. 3. A schematic general model [modified after (*8*)] illustrating possible lithospheric processes and magmatic evolution in Hawaiian volcanoes. In the shield forming stage (left) magmas segregate from the plume and possibly from the surrounding asthenosphere and collect in deep magma chambers at ~90 km. The garnet dunite xenolith is a fragment of a cumulate pile from a magma chamber beneath the Koolau shield volcano. The number, size, and shape of magma chambers beneath any volcano are a matter of speculation, and the ones shown are schematic only. Heating of the lithosphere and probable Cr-enrichment and Fe-depletion of the surrounding lithosphere by reactive passing magmas likely cause deepening of the transition from spinel lherzolite to garnet lherzolite down to a depth of about 75 km (*8*). The rejuvenated stage is represented by the eruption of xenolith-bearing alkalic Honolulu Volcanics on Oahu, 1.2 million years after the cessation of tholeiitic volcanism (Koolau). Because the system has now moved off the hot spot, the lithosphere has cooled and is deeper. In sharp contrast to the tholeiites, which pond at various depths during their ascent, the xenolith-bearing alkalic lavas probably rise rapidly and continuously from the mantle (*7*). The conduit system (*4, 7*) used by the tholeiites (which is preserved as frozen intrusions) is not shown in the right half of the diagram in order to maintain clarity. See text for further details.

kbar (*8*), which is 5 kbar lower than the likely pressure of cumulate crystallization. If the layered cumulate were a precipitate from the parent magma of the pyroxenites (that is, HV parent magmas), then it should have had a higher content of clinopyroxene and certainly not be devoid of this phase. Therefore, we consider it unlikely that the parental magma of the xenolith was related to the rejuvenated stage alkalic HV lavas. On the other hand, the magma from which the xenolith crystallized could represent the ultimate parent of the Koolau tholeiites because post-shield alkalic magmas did not erupt on Koolau. These magmas may have been generated at a minimum depth of about 90 km and underwent ponding and fractionation in deep magma chambers before ascending and fractionating further in shallow crustal chambers and then finally erupting through the shield volcano.

Seismic studies of the currently active Kilauea volcano on Hawaii suggest that earthquakes record the ascent of magma through the lithosphere (*4*). The foci of the majority of the deep earthquakes are about 60 km, which has been suggested to be the depth at which magma accumulates before its explosive ascent (*4*). The xenolith that we have described, however, indicates that a large magma chamber was present at significantly greater depth beneath Koolau volcano. Data from Nd, Sr, and He isotopes indicate that in general, Hawaiian tholeiites carry a chemical signature of a deep plume in the mantle beneath Hawaii (*8, 15*). Thus, Hawaiian shield-building magmas may originate at deeper levels and pond at about 90 km. The general absence of earthquakes below 60 km beneath Kilauea indicates that the upper mantle below that depth behaves in a ductile manner and that magma rises diapirically rather than by fracturing (*16*). As the magma reaches 60 km depth, however, it enters a more brittle lithosphere and continues its ascent by way of fractures. The occurrence of veined xenoliths, which represent magma-filled fractures, from 75 to 80 km (*9*) and of the cumulate xenolith in the Honolulu Volcanics of Oahu suggests that about 1.2 million years after the Koolau tholeiitic magmatism stopped and the volcano moved away from the hot plume, the brittle-ductile transition beneath the Koolau volcano descended to about 90 km depth.

Another important implication of this xenolith is concerned with the generation of the post-shield Honolulu Volcanics in which this xenolith occurs as an inclusion. On the basis of Nd and Sr isotopic similarity (that is, depleted character) between HV and mid-ocean ridge lavas (which are generally believed to have been extracted from the lithospheric mantle), Chen and Frey suggested that HV magmas were generated by partial fusion of the lithosphere (*15*). However, in order for a HV magma to entrain this xenolith, at least some of the magma must have formed at or near the base of the lithosphere or within the asthenosphere.

A petrologic model of the Hawaiian lithosphere was presented by Sen (*7*) on the basis of mantle xenoliths from Oahu; the inferences made above necessitate a revision of this model (Fig. 3). As the lithosphere moves over the hot plume, shield-building tholeiite magmas separating from the plume (and surrounding asthenosphere) initially accumulate and fractionate in chambers near the base of the lithosphere. Fractionation of olivine, spinel, and garnet from these magmas leads to residual basaltic magmas with higher absolute abundances of rare earth elements (REE) and relatively light REE-enriched (chondrite-normalized) character. These differentiated magmas then leave the chambers and rise first through a ductile region, and then above ~60 km through fractures in the brittle lithosphere. In the rejuvenated (or HV type) stage, descent of the brittle-ductile transition to about 90 km beneath Oahu allows post-shield magmas, generated from the asthenosphere, to carry xenoliths from this depth.

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17. We thank the late E. D. Jackson who collected this xenolith and identified it as a garnet dunite in his field notes, and S. Sorensen (Smithsonian Institution) for providing access to the Dale Jackson Collection. We also thank D. Clague, H. S. Yoder, Jr., D. C. Presnall, G. Draper, and an anonymous reviewer for comments on different versions of this manuscript. Discussions with D. Clague and D. C. Presnall led to significant improvement of the presentation. D. Clague provided a slide of the original sample, photomicrographs, and a thin section of the xenolith that contained phlogopite. This research was supported by National Science Foundation grants EAR-8815858 and EAR-8903879 to G.S.

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Redox Regulation of Fos and Jun DNA-Binding Activity in Vitro

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The proto-oncogenes c-*fos* and c-*jun* function cooperatively as inducible transcription factors in signal transduction processes. Their protein products, Fos and Jun, form a heterodimeric complex that interacts with the DNA regulatory element known as the activator protein–1 (AP-1) binding site. Dimerization occurs via interaction between leucine zipper domains and serves to bring into proper juxtaposition a region in each protein that is rich in basic amino acids and that forms a DNA-binding domain. DNA binding of the Fos-Jun heterodimer was modulated by reduction-oxidation (redox) of a single conserved cysteine residue in the DNA-binding domains of the two proteins. Furthermore, a nuclear protein was identified that reduced Fos and Jun and stimulated DNA-binding activity in vitro. These results suggest that transcriptional activity mediated by AP-1 binding factors may be regulated by a redox mechanism.

A VARIETY OF SYSTEMS HAS EVOLVED in prokaryotic and eukaryotic cells for coupling environmental signals to the selective regulation of gene expression. These have in common the ultimate formation of specific protein-DNA complexes (*1*). A class of genes, including c-*fos* and c-*jun*, encodes inducible transcription factors that function as intermediary transcriptional regulators in signal transduction