Late Cretaceous Oceans and the Cool Tropic Paradox

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Oxygen isotopic proxies of paleo–sea surface temperatures (SSTs) suggest that Maastrichtian (about 66 million years ago) tropical SSTs were lower than those of today. They also demonstrate that Maastrichtian latitudinal SST gradients were much lower than those of the present. The low Maastrichtian SST gradients indicate that meridional heat transport was much greater or latitudinal differences in the balance of radiation to and from the sea surface were much less extreme during the latest Cretaceous than they are today, or that both conditions were true. These findings challenge traditional interpretations of "greenhouse" Late Cretaceous climates.

Oxygen isotopic ($\delta^{18}O$) studies of marine fossils have been interpreted to indicate that tropical oceans were characterized by SSTs well below the present-day mean value of 27.5°C during most of the Late Cretaceous (*1*) and Paleogene [89 to 23 million years ago (Ma)] (*2*, *3*). Isotopically based estimates of latest Cretaceous (Maastrichtian) tropical SSTs are particularly low (18° to 21°C) (*1*). In contrast, fossil assemblages have often been interpreted to indicate that low-latitude Late Cretaceous and Paleogene SSTs approximated or exceeded those of the present day (*4*, *5*). Studies of high-latitude fossil assemblages (*5*), terrigenous weathering products (*6*), and $\delta^{18}O$ signals of marine carbonates (*2*, *7*) generally agree that high-latitude regions were substantially warmer than at present throughout those intervals.

The isotopic interpretation of low tropical paleo-SSTs directly challenges the applicability of standard "greenhouse" simulations of Late Cretaceous and Paleogene climate conditions (*8*, *9*). Such simulations predict that high atmospheric concentrations of greenhouse gases (such as CO_2) warmed both low-latitude and high-latitude regions. The apparent incompatibility of isotopic interpretations with greenhouse climate simulations and some fossil interpretations constitutes a "cool tropic paradox."

Isotopically based estimates of Late Cretaceous tropical SSTs have been questioned on several grounds (*4*, *5*): (i) Few isotopic data have been available for such estimates (*1*). (ii) Few tropical sites have been analyzed. (iii) Chronostratigraphic resolution has generally been poor [±11 million years in one recent study (*10*)]. (iv) SST estimates derived from isotopic data are contingent on local effects of precipitation and evaporation. (v) Previous studies have relied on the skeletal carbonate of subsurface or cool-season organisms for estimation of SSTs. (vi) Changes in the $\delta^{18}O$ value of the terrestrial surface hydrosphere are rarely considered.

In order to better resolve the Late Cretaceous cool tropic paradox, we determined the oxygen isotopic signals ($\delta^{18}O_c$) of Maastrichtian near-surface planktic foraminifera from 13 Deep Sea Drilling Project (DSDP) and Ocean Drilling Project (ODP) sites in the Atlantic, Indo-Pacific, and Southern oceans (Fig. 1) (*11*). We analyzed 1 to 14 samples of well-preserved ^{18}O-depleted (near-surface, warm season) taxa from each site (Figs. 2 and 3). We analyzed *Rugoglobigerina rotundata* at all sites except southern high-latitude ODP sites 689B and 690C, where we analyzed *Globigerinelloides multispinus* (*Rugoglobigerina* species do not occur at those sites). Recognition of these species as near-surface warm season taxa is based on isotopic study of 39 planktic foraminiferal species from representative mid- and high-latitude sites (*12*). In order to estimate the maximum seasonal SST gradient, we also analyzed samples of *Heterohelix globulosa* from ODP site 690C (70°S). This species exhibits the highest $\delta^{18}O_c$ values of planktic foraminifera at that site (*12*); consequently, we interpret it to have been either a winter sea surface species or a subthermocline species. In either case, its $\delta^{18}O_c$ signal should correspond to a cooler season SST at 70°S.

The sites ranged in absolute paleolatitude from ~8° to ~70°, and in paleodepth from about 1 to 3.5 km (*13*). The sampled sequences are centered on a horizon midway between the 30N/29R paleomagnetic reversal [which occurred ~350,000 years before the Cretaceous-Tertiary boundary (65 Ma)] and the first occurrence of *Micula murus* (400,000 to 600,000 years before the 30N/29R reversal) (*14*). The samples are thus age-equivalent on the scale of a few hundred thousand years or less.

The mean $\delta^{18}O_c$ values of the surface-dwelling, warm season, late Maastrichtian planktic foraminifera range from 0.2 per mil at high southern latitude sites to −1.4 per mil at some lower latitude sites, resulting in a latitudinal $\delta^{18}O_c$ difference of 1.6 per mil. In contrast, the $\delta^{18}O_c$ of modern surface-dwelling planktic foraminifera changes by 6.0 per mil over 60° of latitude, ranging from −2.0 per mil at 0°N to 4.0 per mil at 60°S (*2*). Because we analyzed the most ^{18}O-depleted taxa, the low Maastrichtian gradient has not been biased by analysis of deep-water or cool season plankton. Furthermore, our sampling of a narrow stratigraphic window and our analysis of multiple samples per site preclude interpretation of the low gradient as an artifact of poor chronostratigraphic resolution (*15*).

The $\delta^{18}O_c$ values of sea surface–dwelling planktic foraminiferal tests depend on the mean oceanic $\delta^{18}O_w$ (seawater $\delta^{18}O$ on standard mean ocean water scale) value, the temperature of calcification, the net evaporation or precipitation in the ambient water mass, and the calcification rate. In the modern ocean, geographic variation in planktic foraminiferal $\delta^{18}O_c$ values is controlled primarily by ambient temperature. For example, the modern 30°C range of SSTs corresponds to a $\delta^{18}O_c$ range of almost 7.0 per mil. The modern $\delta^{18}O_c$ gradient is slightly lower because net evapora-

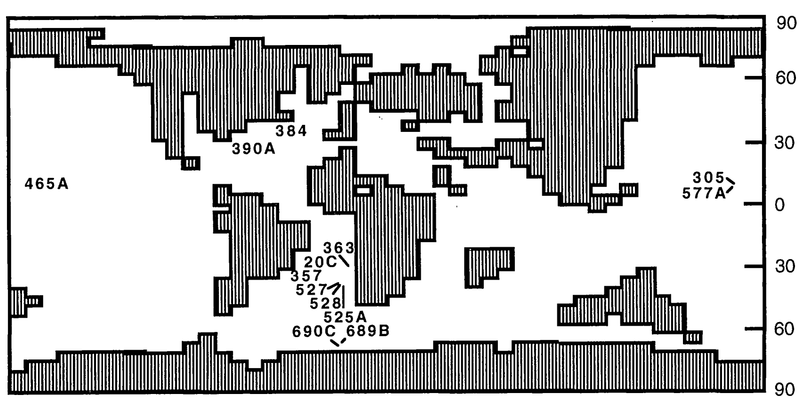

Fig. 1. DSDP sites and ODP sites sampled for analysis of Maastrichtian sea surface properties. The paleogeography was adapted from the University of Chicago Paleogeographic Atlas Project (*29*).

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tion increases modern tropical $\delta^{18}O_w$ values by approximately 1.0 per mil relative to the mean ocean $\delta^{18}O_w$ (*2*).

We can derive a lower bound estimate for late Maastrichtian SSTs by assuming that the late Maastrichtian ocean was characterized by modern summer and winter sea surface salinity (SSS) gradients and a mean $\delta^{18}O_w$ value of −1.0 per mil (*16*). In this case, our data imply that latest Cretaceous summer SST estimates were lower than those of the present day at low latitudes (13.7° to 16.9°C at 8° to 10°N in the central Pacific, 20.5°C at 26°N in the western North Atlantic gyre-margin, and 20°C at 33°S in the western South Atlantic gyre-margin) and higher than those of the present at high latitudes (10°C at 70°S) (Fig. 4A). Based on the $\delta^{18}O_c$ of planktic *H. globulosa* at ODP site 690C, high-latitude cool season SSTs were also relatively high during the late Maastrichtian (7.4°C at 70°S).

One can derive upper bound SST estimates by estimating mean deep-water temperature and salinity, assuming that surface waters cannot have been denser than mean deep water, and calculating the paleotemperature and paleosalinity values that satisfy the $\delta^{18}O$-temperature equation at an estimated surface-water paleodensity that is equal to the estimated mean deep-water paleodensity.

If we assume that mean deep water was characterized by paleosalinity of 34 practical salinity units (psu) and $\delta^{18}O_w$ of −1.0 per mil, the mean $\delta^{18}O_c$ value of benthic foraminiferal tests corresponds to a mean deep-water paleodensity of 1026.13 kg m^{-3} (*17*). That deep-water paleodensity estimate plus planktic foraminiferal $\delta^{18}O_c$ values allow summer sea surface paleotemperatures and paleosalinities as high as 21.5°C and 37.3 psu at ~8°N in the east-central Pacific (DSDP sites 305 and 577A) (Fig. 4). This is ~6.5°C cooler than the modern east-central Pacific sea surface at 8°N (*18*). However, similar assumptions allow near-surface paleotemperatures and paleosalinities of 26.6°C and 39.3 psu at ~10°N (DSDP site 465A), which is only ~1.4°C cooler than the summer sea surface in the modern east-central Pacific.

Our deep-water paleodensity estimate allows summer sea surface paleotemperatures and paleosalinities as high as 29.2°C and 40.5 psu at 33°S in the western gyre-margin of the Maastrichtian South Atlantic (DSDP site 357). It allows surface-water paleotemperature and paleosalinity values as high as 29.8°C and 40.8 psu at 26°N in the western gyre-margin of the Maastrichtian North Atlantic (DSDP site 390A).

The preceding estimates suggest that Maastrichtian SSTs of low-latitude sites 305 and 577A were, respectively, more than 6.5° and 10°C lower than modern tropical SSTs. These estimates also suggest that Maastrichtian summer SSTs of sites 357, 390A, and 465A approached modern tropical SSTs only if those sites attained SSS values of 39 to 41 psu (assuming mean oceanic paleosalinity of 34 psu). Today, such deviations from mean oceanic salinity are limited to semienclosed evaporative basins (such as the Red Sea). In contrast, planktic foraminiferal $\delta^{18}O_c$ values and estimated deep-water paleodensity allow a maximum summer sea surface paleotemperature and paleosalinity of 12.2°C and 34.4 psu at 70°S (ODP sites 690C and 689B). Similarly, the $\delta^{18}O_c$ values of planktic *H. globulosa* allow late Maastrichtian winter SST and SSS values of 8.3°C and 33.6 psu at the same latitude (site 690C). These summer and winter high-latitude SST estimates require little deviation from mean oceanic salinity.

Regardless of sea surface evaporation-precipitation balances, isotopically derived SST estimates are sensitive to the assumed $\delta^{18}O_w$ value of the mean ocean (Fig. 5). For example, if the late Maastrichtian ocean was characterized by modern SSS gradients and mean $\delta^{18}O_w$ of 1.0

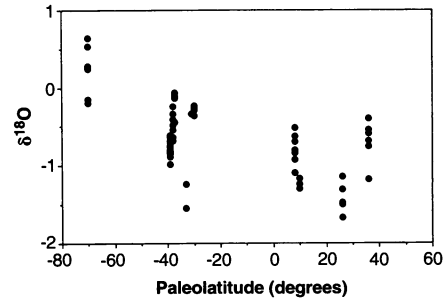

Fig. 3. Oxygen isotopic ($\delta^{18}O_c$) values of the late Maastrichtian (upper chron C30N) surface-dwelling planktic foraminifera *R. rotundata* and *G. multispinus* (*n* = 68).

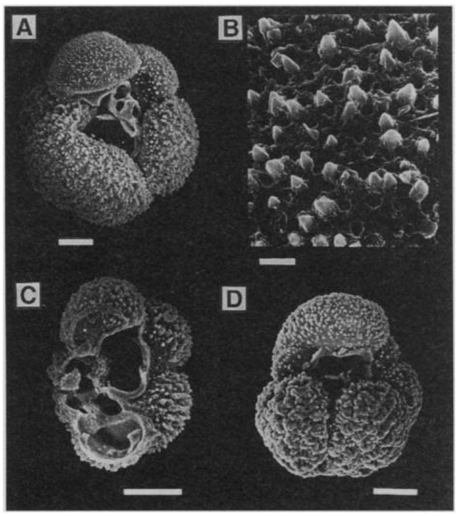

Fig. 2. Scanning electron photomicrographs of representative *R. rotundata* specimens. (**A**) Specimen from DSDP site 305. (**B**) Close view of specimen from DSDP site 305 (scale bar, 10 μm). (**C**) Specimen from DSDP site 357 (specimen has been broken to illustrate the lack of calcitic overgrowth and the good preservation of the inner and outer surfaces). (**D**) Specimen from DSDP site 390A. Specimens from all three sites are in a good state of preservation. Scale bars in (A), (C), and (D), 100 mm.

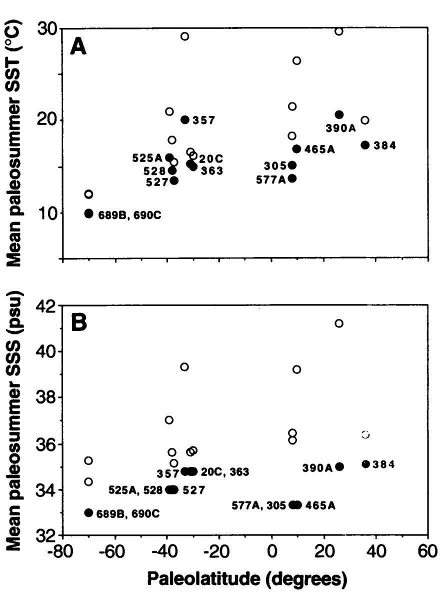

Fig. 4. (**A**) Late Maastrichtian summer SST estimates [made with the assumption that mean oceanic $\delta^{18}O_w$ was −1.0 per mil (*16*) and the average change of $\delta^{18}O_w$ of surface seawater with a 1-psu change in salinity ($\Delta\delta w/\delta S$) was 0.35 (*30*)]. (**B**) Late Maastrichtian summer SSS estimates. Lower bound estimates (solid circles) assume absolute variation in salinity gradients equivalent to that of modern summer sea surfaces [after (*18*)]. Upper bound estimates (open circles) assume mean oceanic salinity of 34 psu and summer surface-ocean density of 1026.12 kg m^{-3} (*17*). Site locations are indicated by marking of each minimum estimate with one or more DSDP or ODP site designations. The coolest low-latitude estimates are from the eastern equatorial Pacific (DSDP sites 305, 465A, and 577); the relative coolness of lower bound estimates from these sites may largely reflect paleoequatorial upwelling. Furthermore, Late Cretaceous foraminifera of site 577A are heavily dissolved (*31*) and their $\delta^{18}O_c$ values may have been skewed by preferential dissolution of more ^{18}O-depleted (warmer water) specimens (*2*). Warm mid-latitude estimates are from western gyre-margins of the Atlantic (DSDP sites 357 and 390A); these provide the highest tropical paleotemperature estimates. In the modern ocean, summer SSTs of western gyre-margins (at 30° of latitude) are within 1°C of the mean equatorial SST (*18*). Given similar gyral modification of Maastrichtian sea surface properties, the $\delta^{18}O_c$ values of summer surface-dwelling foraminiferal tests from sites 357 and 390A closely resembled those of the mean equatorial sea surface.

per mil, tropical SSTs of the eastern equatorial Pacific would have approached modern values (26°C) and summer SSTs would have been ~19°C at 70°S (ODP sites 689B and 690C) (Fig. 5). Consequently, the magnitude of latitudinal $\delta^{18}O_c$ and SST gradients is unaffected by the assumed $\delta^{18}O_w$ value of the mean ocean.

Our isotopic data indicate that southern high-latitude SSTs were at least 10°C higher in the Maastrichtian than they are at present. This is consistent with Maastrichtian reconstructions of high-latitude terrestrial plant assemblages (*5*). Given constant insolation at the top of the atmosphere, relatively warm high-latitude temperatures could result from increased heat transport from low to high latitudes by atmospheric or oceanic processes or both, or from increased trapping of heat at high latitudes (*19*). Estimates of tropical SSTs are critical for evaluation of the relative importance of these mechanisms. Increased atmospheric or oceanic heat transport from low to high latitudes should slightly decrease low-latitude SSTs. In contrast, increased atmospheric concentrations of greenhouse gases (such as CO_2) alone would result in surface warming at all latitudes (including the tropics) (*20*). Finally, increased low-altitude cloud cover at high latitudes could warm high-latitude regions while leaving tropical surface temperatures unaffected (*21*).

Our isotopic results demonstrate that a wide range of tropical SST and SSS estimates are compatible with Maastrichtian planktic $\delta^{18}O$ values and a density-stratified ocean. These estimates can be used to define three extreme tropical scenarios. Under the first scenario, Maastrichtian tropical SSS gradients closely resembled modern gradients, mean oceanic $\delta^{18}O$ was −1.0 per mil, and the mean tropical Maastrichtian SST was 20° to 21°C (~6°C below the modern value). The second scenario also assumes that mean oceanic $\delta^{18}O$ was −1.0 per mil, but that tropical paleo-SSTs locally approached 30°C and tropical paleo-SSS values were as much as 7 psu greater than mean oceanic values. In the third scenario, tropical paleo-SSTs approached 30°C, and the late Maastrichtian tropical SSS gradient was similar to that of the modern ocean, but mean oceanic $\delta^{18}O$ was 1.0 per mil.

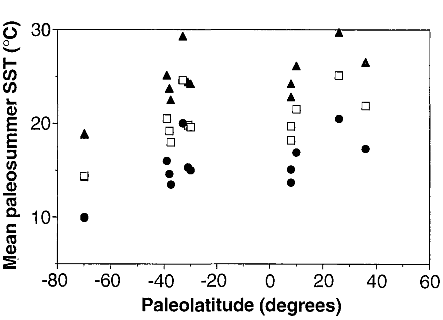

Fig. 5. Late Maastrichtian summer SST estimates, made with the assumption that mean oceanic $\delta^{18}O_w$ was −1.0 per mil (circles), 0.0 per mil (squares), and 1.0 per mil (triangles). These estimates assume $\Delta\delta w/\delta S$ of 0.35 (*30*) and absolute variation in salinity gradients equivalent to that of the modern summer sea surface.

Although relatively cool (20° to 21°C), the mean tropical paleo-SST estimate of the first scenario appears to be compatible with fossil assemblage data (*22*). The second and third scenarios avoid low tropical paleo-SST estimates but appear unlikely. The second scenario is not supported by GENESIS atmospheric General Circulation Model (GCM) experiments, which retrodict net annual precipitation in the near-equatorial intertropic convergence zone to have been even higher during the Maastrichtian than at present (*23*). The third scenario is problematic because it is extremely unlikely that the mean $\delta^{18}O$ value of Late Cretaceous oceans could have approached 1.0 per mil (*24*).

Given the problems associated with the second and third scenarios, it appears likely that Maastrichtian tropical SSTs were lower than those of the present. Whether Maastrichtian SST and SSS are best approximated by the first scenario or by some less extreme combination of the three scenarios, the latitudinal gradient in SST was much lower during the Maastrichtian than it is today. The first and third scenarios allow Maastrichtian SST gradients of only 11°C in summer and ~13°C in winter. The second scenario allows Maastrichtian summer and winter SST gradients of ~20°C and ~22°C from tropical regions to high latitudes. In contrast, modern SSTs vary by as much as 32°C between the thermal extremes of the western equatorial Pacific (29.6°C) and the Southern Ocean (−1.7°C); from 0°N to 70°S, modern mean SST gradients are ~28°C in summer and slightly greater than 27°C in winter (*18*).

The low Maastrichtian SST gradients would have required very high levels of heat transport from low to high latitudes or a latitudinally more equable surface radiation balance than exists at present. Possible mechanisms for increased latitudinal heat transport include increased low-latitude deep-water formation (*25*), extension of shallow circulation to depths greater than those of the present day (if a less strongly stratified Cretaceous ocean is assumed) (*9*), and increased atmospheric mass (*26*). Mechanisms for maintaining a more equable surface radiation balance include lower surface albedo at high latitudes (due to absence of ice), increased atmospheric albedo at low latitudes (due to high-altitude clouds), and increased trapping of heat at high latitudes (due to low-altitude clouds) (*9, 21, 27*). The relative importance of these various mechanisms to past and present climate conditions is actively debated (*9, 28*). Nonetheless, it appears likely that one or more of these mechanisms must be invoked to explain the low latitudinal SST gradients and low tropical SSTs of the Late Cretaceous.

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sites spanning a wide range of late Maastrichtian paleodepths (1 to 3.5 km) and paleolatitudes (36°N to 70°S). If these tests were in isotopic equilibrium with their paleoenvironments, the mean $\delta^{18}O_c$ value of benthic foraminifera from all 16 sites (0.51 per mil) approximates that of benthic foraminifera in equilibrium with mean late Maastrichtian deep water. Given a mean $\delta^{18}O_w$ value of -1.0 per mil for late Maastrichtian seawater (*16*), a mean benthic $\delta^{18}O_c$ value of 0.51 per mil indicates a mean deep-ocean paleotemperature of 10.2°C. Estimates of the mean $\delta^{18}O_w$ of Late Cretaceous seawater typically assume that there was no globally significant Late Cretaceous ice volume and that the mean $\delta^{18}O$ value of the terrestrial surface hydrosphere has not changed over the past 100 million years (*16*). The mean salinity of Late Cretaceous seawater can be similarly estimated. Given a lack of substantial ice volume and a constant salt balance in the surface hydrosphere, Late Cretaceous oceans were characterized by a mean salinity of ~34 psu. If the late Maastrichtian deep ocean was characterized by a mean temperature of 10.2°C and a mean salinity of 34 psu, its mean density was 1026.13 kg m^{-3} [density equation is from F. J. Millero and A. Poisson, *Deep-Sea Res.* **28A**, 625 (1981)].

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Alpe Arami: A Peridotite Massif from Depths of More Than 300 Kilometers

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The abundance of $FeTiO_3$ and chromite precipitates in olivine of the Alpe Arami peridotite massif, Switzerland, requires a much higher solubility for highly charged cations than is found in mantle xenoliths from depths of 250 kilometers. Three previously unknown crystal structures of $FeTiO_3$ were identified that indicate that the originally exsolved phase was the high-pressure perovskite polymorph of ilmenite, implying a minimum depth of origin of 300 kilometers. These observations, coupled with the unique lattice preferred orientation of olivine, further suggest that the original composition of olivine may reflect that of a precursor phase, perhaps wadsleyite, that is stable only at depths greater than 400 kilometers.

In recent years, rocks have been discovered that were brought to Earth's surface by a poorly understood process that can accompany or follow subduction of shallow material to great depths. These bodies, some of which contain diamonds (*1*), provide a window into processes active at depth in subduction zones (*2*). Such rocks can potentially provide information about the composition and phase relations in the mantle wedge overlying subduction zones, and they may cast light on the problem of recycling of volatiles into the deep interior. We now report observations that suggest that the Alpe Arami peridotite massif, Switzerland, has come to the surface from depths of 400 to 670 km, the mantle transition zone.

The Alpe Arami massif, which yields consistent mineral ages of ~40 million years (*3*), is situated within Lepontine gneisses of the Ticinese root zone of the Pennine Alps. It consists of a small wedge of garnet lherzolite bordered by kyanite eclogite, all showing variable degrees of low-pressure hydrous alteration (*4*). The maximum pressure of metamorphism in the Lepontine nappe increases toward the south; the Alpe Arami massif, located near the southern margin, records the highest pressures, with garnet-orthopyroxene-clinopyroxene equilibration at approximately 1200 to 1300 K and 4 to 5 GPa (*3*, *5*, *6*). These results establish a minimum depth of origin of 120 to 150 km.

Like most peridotite massifs, Alpe Arami has experienced considerable subsolidus deformation and partial recrystallization of olivine, with consequent development of foliation and lineation. The oldest generation of olivine consists of larger, plastically deformed, porphyroclasts that exhibit an unexplained pattern of lattice preferred orientation (LPO) that is different from that displayed by any other peridotite (*4*, *7*) and is inconsistent with the crystal plasticity of olivine (*8*). This pattern must have formed by a different, unknown, mechanism.

We analyzed fresh peridotite that exhibited large (1 to 2 cm), deep red–violet garnets, which commonly displayed bright apple-green clinopyroxenes (1 to 3 mm), both as inclusions and as clusters around their margins (Fig. 1A). Similar textures have been reported from mantle xenoliths and were interpreted as evidence of an ultradeep (>300 km) origin (*9*). The garnets of Alpe Arami also exhibited abundant exsolved needles of rutile and small, oriented, euhedral-to-subhedral crystals of pargasite-edenite amphibole, which were always accompanied by an orthopyroxene-spinel symplectite and were usually accompanied by magnesite, apatite, and an Fe-Ni sulfide (Fig. 1, B and C). These multiphase inclusions appear to represent a complex reaction between an originally exsolved phase, probably clinopyroxene (plus olivine?), and the host garnet during decompression of the massif.

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