H2@C60 are also exactly the same as those of Bu4NBF4 for reduction and 0.5 mM in o-dichlorobenzene with 0.05 M Bu4NBF4 for oxidation. The voltammogram of H2@C60 exhibited four reversible reduction waves and one irreversible oxidation peak at the same potentials as C60 as well as an experimental error of ±0.01 V.

In order to clarify the reactivity of H2@C60, the solid-state mechanochemical [2+2] dimerization reaction (23) was conducted. A mixture of H2@C60 and 1 molar equivalent of 4-aminopyridine as the catalyst (24) was vigorously shaken by the use of a high-speed vibration mill for 30 min under N2, according to our previous procedure (23, 24). The 1H NMR spectrum of the product mixture exhibited a signal at δ = −4.04 ppm of the [2+2] dimer, (H2@C60)2, and a signal of unchanged H2@C60 at δ = −1.44 ppm, in an integrated ratio of 3:7. This result indicates that the dumbbell-shaped dimer of H2@C60 is formed in the same yield as that for the reaction of empty C60 (24) (Fig. 4). No effect of the encapsulated H2 was observed upon reactivity of the C60 cage. The extent of the uptake shift of the 1H NMR signal (2.60 ppm) observed for the dimer (H2@C60)2 was similar to that observed upon the same dimerization reaction in 3He NMR for 3He@C60 (6.36 ppm) (9, 10).

The endohedral fullerene H2@C60 is nearly as stable as C60 itself. For example, the encapsulated H2 does not escape even when heated at 500°C for 10 min. Thus, H2@C60 can be viewed as a stable hydrocarbon molecule that has neither C-H covalent bonds nor C=C=H interactions. It is likely that our method could be used to synthesize endohedral fullerenes such as D2@C60 and HD@C60 as well as the homologous series with C60. Our work here complements the total chemical synthesis of C60 recently achieved by Scott and co-workers (25) and implies that organic synthesis can be a powerful means for the production of yet unknown classes of endohedral fullerenes.

References and Notes
20. Materials and methods are available as supporting material on Science Online.

Corrected Late Triassic Latitudes for Continents Adjacent to the North Atlantic

Dennis V. Kent1,2* and Lisa Tauxe3

We use a method based on a statistical geomagnetic field model to recognize and correct for inclination error in sedimentary rocks from early Mesozoic rift basins in North America, Greenland, and Europe. The congruence of the corrected sedimentary results and independent data from igneous rocks on a regional scale indicates that a geocentric axial dipole field operated in the Late Triassic. The corrected paleolatitudes indicate a faster poleward drift of ∼0.6 degrees per million years for this part of Pangea and suggest that the equatorial humid belt in the Late Triassic was about as wide as it is today.

Paleomagnetism is used to determine ancient latitude, but its reliability depends on two assumptions: (i) that the time-averaged geomagnetic field is closely approximated by that of a geocentric axial dipole (GAD), and (ii) that there is no systematic bias in how the geomagnetic field is imprinted in rocks. Although the GAD hypothesis (I) is supported by paleomagnetic data for the past few million years (2, 3), departures from the GAD model
have been invoked to explain anomalously shallow directions observed in some older rocks (4–6). On the other hand, a shallow bias or inclination error (7) has been found in laboratory redeposition experiments and in some modern natural sediments (8). A good example of the ambiguity in distinguishing between a non-GAD field and recorder bias is found in the paleomagnetic record from continental basins that developed during rifting of the Pangea supercontinent in the early Mesozoic and are now distributed along the margins of the North Atlantic (Fig. 1). One of the largest and best studied of the rift basins is the Newark basin in eastern North America, where more than 5000 m of strata (mainly continental redbeds) were recovered in scientific drilling (9) and which yielded a 35 million year (My)–long record of latitudinal drift calibrated by an astronomically tuned geomagnetic polarity time scale (10, 11). The average paleomagnetic inclinations from the Dan River basin (12) and the Fundy basin (13) indicate low paleolatitudes that are consistent with the Newark basin data. However, paleomagnetic directions from the more distant Jameson Land basin in Greenland (Fig. 1) are anomalously shallow and imply a paleolatitude \( \sim 10^\circ \) lower than predicted from coeval sections in North America (14). This discrepancy is too large to be explained by uncertainties in the reconstruction of Greenland to North America (15, 16). Therefore, either the magnetizations of the sedimentary rocks are biased by inclination error or the Late Triassic time-averaged field included large nondipole (axial octupole) contributions (4).

Fig. 1. Paleomagnetic sampling localities of key Late Triassic and earliest Jurassic sedimentary and igneous rock units plotted on a Pangea reconstruction (16). For reference, present-day latitude/longitude grids (5° by 5°) are shown for each continent. Insets give the paleomagnetic direction data (solid and open symbols are on the lower and upper hemisphere, respectively, of equal-area projections) and the implied paleolatitudes (\( \lambda \)) from each section: DR, Dan River basin (12); NB, Newark basin with data from seven drill cores labeled with the first initial of the drill core (M, W, S, R, T, N, and P) (Table 1) (10); JL, Jameson Land basin (14); StA, St. Audrie’s Bay section (19). Other data discussed in the text are from F, the Fundy basin (13, 22); Mn, the Manicouagan impact structure (24, 25); and Ro, the Rochechouart impact structure (23). The data are summarized in Table 1.

Fig. 2. E/I analysis of sample characteristic magnetization directions from Fleming Fjord Formation, Jameson Land, Greenland (12). (A) The trajectory (solid red line) of mean inclination versus elongation calculated for the Jameson Land data shown in the inset to Fig. 1, as the data are inverted with values for \( f \) ranging from 0.3 to 1.0. Yellow lines are examples of bootstrapped trajectories. The predicted E/I trend with latitude of the TK03.GAD model (17) is shown as the blue dashed line; the E/I of the data consistent with the model is circled. (B) A histogram of 1000 bootstrapped intersections of the kind shown in (A) from bootstrapped curves. The 95% confidence bounds on the corrected inclination are also shown (Table 1).
Fig. 3. Paleolatitude nomogram for the Late Triassic and earliest Jurassic of a portion of Pangea. Paleolatitude contours are based on a smoothed progression of latitudes that are calculated from corrected mean inclinations, according to the E/I method, from seven drill cores from the Newark basin (NB, with corrected paleolatitudes adjacent to solid lines indicating the age ranges of the cores) and take into account present geographic relationships. Mean paleomagnetic declinations for the Late Triassic of eastern America are typically within a few degrees of north–south (Table 1), and hence differences in present-day latitudes of the sites in North American coordinates closely approximate differences in paleolatitude. Site latitudes for the Jameson Land (JL) section in Greenland (14), the St. Audrie’s Bay (STA) section (19), and the 214-Ma-old (27) Rochechouart impact structure (Ro) (23) in Europe were transferred to North American coordinates according to reconstruction parameters from Bullard et al. (16) (Table 1). The geomagnetic polarity time scale from the NB cores (11) was used as the basis of magnetostratigraphic correlation (solid and open bars denote normal and reverse polarity, respectively) and age control. Mean paleolatitudes from corrected inclinations are indicated for JL, STA, and the Dan River basin (DR) (Table 1). Paleolatitudes with asterisks are for igneous rocks from earliest Jurassic (~200 Ma) Central Atlantic Magmatic Province lavas in the NB (21) and Fundy (F) basins (22) and from the 214-Ma-old (26) Manicouagan (Mn) impact structure in Quebec (24, 25) and Ro. Climate lithofacies are shown for the DR, NB, and F (35) and JL (37) sections where c is coal, s is saline minerals, and e is eolian deposits; light to dark shading in lithology columns ranges from fine-grained redbeds to black shales, with stippling indicating sandstones. The inset is (left) a Landsat image of Africa (44) and (right) the global zonal mean profile of evaporation minus precipitation (E-P) (45). Green and yellow colors indicate more humid and arid conditions, respectively, the present-day latitudinal ranges of which are shown on the paleolatitude nomogram for comparison.
Although the data set from St. Audrie’s Bay in Somerset, United Kingdom, is rather small (27 sites), the results are included because they are possibly the first for the Late Triassic of Europe to be supported by modern demagnetization techniques and the site-mean directions are available (19).

We compared the corrected results, which differ in age as well as location, by constructing a paleolatitude versus age nomogram (13) by fitting a second-order polynomial curve to the corrected paleolatitude versus age progression determined from the Newark basin cores (Table 1). We then used that relationship to predict paleolatitudes over the same 35-My interval for tectonically contiguous or reconstructed areas. The corrections for inclination error bring the paleolatitude data for the Triassic basins into agreement (Fig. 3). In particular, the corrected paleolatitude for the previously discrepant result from Jameson Land is now consistent with the paleolatitude predicted from the corrected Newark basin data. The data set from St. Audrie's Bay also falls into line after correction.

As an independent check on the validity of the corrections for inclination error, we can use paleomagnetic data from approximately coeval igneous rock units, which are not subject to inclination error. Lava flows of the Central Atlantic Magmatic Province of the earliest (~200 Ma) Jurassic age (20) occur in several of the rift basins in eastern North America. For comparison with the Newark basin sedimentary results, we used a compilation of the most reliable data as determined by Prevoit and McWilliams (21) for the three major extrusive units in the Newark and nearby Hartford basins. These lava flow data indicate a paleolatitude of 17 ± 5° for the Newark basin (Table 1), which is in much better agreement with the paleolatitude after correction for inclination error (19°) than without correction (~9°), as determined from the immediately underlying (Upper Triassic and lowermost Jurassic) sedimentary rocks of the Passaic Formation in the Newark basin Martinsville core (Fig. 3 and Table 1). Paleomagnetic results from the Upper Triassic and lowermost Jurassic Blomidon Formation in the Fundy basin (13) give an anomalously low paleolatitude (~12°), but these inclination-only sedimentary data could not be tested or corrected with the E/I method. Nevertheless, paleomagnetic data (22) for the overlying North Mountain Basalt indicate a paleolatitude of 26°, which is consistent with the paleolatitude predicted from the corrected Newark data (Fig. 3). Paleolatitudes from two older igneous rock units also agree with the paleolatitude-age matrix predicted from the corrected Newark basin data (Fig. 3): Rochechouart (23) in France and Manicouagan (24, 25) in Quebec, which are impact structures dated at ~214 Ma (26, 27).

With the caveat that these igneous rocks may not adequately average secular variation because of the paucity of cooling units, the overall agreement of the igneous data with the predictions based on the E/I method supports the hypothesis that inclination error pervasively affects the sedimentary rocks we studied. An early study that found no substantial difference in paleomagnetic directions, all of normal polarity, from sediments and igneous rocks in the Newark basin (28) was before routine use of progressive thermal demagnetization, which has revealed that a depositional component of normal and reverse polarity with shallow directions in these red beds is typically masked by a steeper normal polarity thermochemical overprint (29). On the other hand, an early-noticed (30) discordance in presumed Triassic latitudes between North America and Europe in a Pangea configuration is likely to be an artifact of age differences and inclination error, because it was largely based on a comparison between data from Jurassic igneous rock units in eastern North America and Triassic sedimentary rocks in western Europe. The corrected St. Audrie’s Bay data show that Late Triassic latitudes of Europe are consistent with those of North America in a Pangea (16) fit.

Slow poleward motion of ~0.3° per My for North America was inferred from the original Newark basin paleomagnetic data (10), but these are biased by inclination error, which underestimates paleolatitudinal change. The corrected inclinations indicate a much faster rate of poleward motion of ~0.6° per My from ~235 to 200 Ma, which emphasizes the need for precise spatiotemporal registry of climate-sensitive lithofacies in paleoclimate studies. The overall distribution of such facies within the cyclic successions suggests that generally moist conditions extended from the coal-bearing and black shale units in the Dan River basin (31) near the paleoequator and black shales of the Lockatong Formation in the Newark basin at 5 to 10° to where eolian deposits in the Fundy basin (32) are encountered at ~15° paleolatitude. The width of the equatorial belt of the characteristic magnetization data; λ is paleolatitude calculated from the mean inclination; f is the flattening factor determined from E/I analysis, f is the corrected mean inclination and ±f is its 95% confidence interval; λ is the corresponding corrected paleolatitude and ±λ is its 95% confidence interval; and Ref. is the literature source for the age and paleomagnetic data.

Table 1. Summary of paleomagnetic data from Late Triassic and earliest Jurassic rocks. Slat and Slon are the latitude and longitude of the sampling localities; entries in parentheses are site locations in Greenland and Europe transferred to North American coordinates according to Bullard et al. (16). Age is the mean age of the sampled rocks; n is the number of data included in the mean values; D and I are the mean declination and mean inclination of the characteristic magnetization data; λ is paleolatitude calculated from the mean inclination; f is the flattening factor determined from E/I analysis, f is the corrected mean inclination and ±f is its 95% confidence interval; λ is the corresponding corrected paleolatitude and ±λ is its 95% confidence interval; and Ref. is the literature source for the age and paleomagnetic data.

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<th>Locality</th>
<th>Slat (°N)</th>
<th>Slon (°E)</th>
<th>Age (Ma)</th>
<th>n</th>
<th>D (°)</th>
<th>I(°)</th>
<th>λ (°N)</th>
<th>±λ (°N)</th>
<th>f</th>
<th>f (°)</th>
<th>±f (°)</th>
<th>λ (°N)</th>
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<td>Dan River</td>
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<td></td>
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<td>333</td>
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<td>217</td>
<td>308</td>
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<td>—</td>
<td>—</td>
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<td>(22)</td>
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<td>214</td>
<td>11</td>
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<td>±5.8</td>
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<td>—</td>
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<td>—</td>
<td>(24–26)</td>
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<td>19.2</td>
<td>±2.6</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
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torial humid belt in the Late Triassic was comparable to today’s (Fig. 3), a conclusion that contrasts with some previous suggestions of a much more restricted or even dry equatorial belt in the Triassic (33, 34). Poleward motion can explain the generally drier northwest and up-section facies pattern in the Mesozoic rift basins of eastern North America (32, 35) as this part of Pangea drifted out of the equatorial humid belt. At the same time, the up-section progression to more humid facies in the Fleming Fjord Formation (36, 37) and the overlying plant-bearing Kap Stewart Formation of latest Triassic and earliest Jurassic age (38) in the Jameson Land basin would reflect the drift of this area into the temperate humid belt.

We conclude that the congruence of the corrected paleomagnetic data from sedimentary rocks and independent data from igneous rocks ranging over thousands of kilometers and tens of millions of years indicates that a GAD field similar to that of the past 5 My was operative at least 200 Ma in the Late Triassic and earliest Jurassic. In particular, we see no evidence for a major octupole contribution in either the shapes of the distributions of directions in the sedimentary units or in the geographic distribution of site paleolatitudes. As indicated by other recent studies (17, 39–41), there is thus little empirical basis to invoke inclination error in sedimentary units or in the geographic distribution of site paleolatitudes. Instead, our results suggest that inclination error in sedimentary rocks may be more prevalent than has been supposed, perhaps especially in cases where the magnetizations that have been isolated are most likely to represent a depositional remanence carried by hematite. The success of the E/I method (17) to determine the degree of flattening and to correct any bias in inclinations from the distribution of directions should provide motivation for more intensive sampling of sedimentary rock units and for making detailed data more accessible.

References and Notes
32. J. F. Hubert, K. A. Mertz, Geology 8, 516 (1980).
46. We thank P. Olsen for many discussions of the Triassic-Jurassic world and the reviewers for insightful comments. Supported by National Science Foundation grant nos. EAR-0310240 (D.V.K.) and EAR-0003395 (L.T.). This is LDEO contribution #6700.

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Fig. S1
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An Astronomical 2175 Å Feature in Interplanetary Dust Particles
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The 2175 angstrom extinction feature is the strongest (visible-ultraviolet) spectral signature of dust in the interstellar medium. Forty years after its discovery, the origin of the feature and the nature of the carrier(s) remain controversial. Using a transmission electron microscope, we detected a 5.7-electron volt (2175 angstrom) feature in interstellar grains embedded within interplanetary dust particles (IDPs). The carriers are organic carbon and amorphous silicates that are abundant in IDPs and in the interstellar medium. These multiple carriers may explain the enigmatic invariant central wavelength and variable bandwidth of the astronomical 2175 angstrom feature.

Much of what is known about grains in space comes from spectral features observed in emission, polarization, and absorption (1–7). The 2175 Å peak is by far the strongest feature observed in the ultraviolet (UV)–

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visible wavelength range along most lines of sight for which it can be measured (Fig. 1, A and B) (4–7). The feature is enigmatic: Its central wavelength is almost invariant, but its bandwidth varies from one line of sight to another, suggesting multiple carriers or a single carrier with variable properties. From interstellar abundances of the elements and typical UV transition strengths, the carrier is either oxygen-rich (e.g., oxides or silicates) or carbon-rich (e.g., graphite or organic compounds) (1–3). We searched UV spectra of chondritic IDPs for an extinction feature near the ~2175 Å interstellar feature (Fig. 1). Materials similar to the two most abundant grain types seen in the