GLOBAL WINDS AND AEROSOL UPDRAFTS CREATED BY THE CHICXULUB IMPACT EVENT.
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Introduction: The iridium-rich layer at the Cretaceous-Tertiary boundary indicates high-energy ejecta from the Chicxulub crater rained through the atmosphere after the impact. The energy deposited by the debris warmed the atmosphere over continent-sized areas near the impact site and antipode. The thermal contrast with the surrounding atmosphere generated winds aloft with speeds of ~80 $msec^{-1}$ to near-supersonic values of 255 $msec^{-1}$. These winds, in turn, drove surface winds that could have reached hurricane strength over broad regions. Within those regions, the tropopause was temporarily destroyed and aerosols (sea salt, soil dust, soot from fires) were swept from the troposphere into the stratosphere.

Thermal Pulse: The most significant global environmental effects of the Chicxulub impact event were the direct and indirect effect of ejected debris that rained through the atmosphere. The impact ejecta was distributed globally, but concentrated around the impact site and antipode [1-3], producing peak atmospheric heating and pressure differences in those areas. The Earth's surface may have received up to one-third of the downward-radiated infrared flux and reached temperatures >650 K in some areas [2], and the upper stratosphere, where the primary heating is taking place, may have attained ~1000 K or more [4]. The energy was deposited on a timescale ($\lesssim$4 days [3]) that is considerably shorter than the radiative time constant for the bulk atmosphere (~20 days), implying the power absorbed by the atmosphere as a whole was retained for timescales much longer than the impact plume reentry timescale. A vertically averaged temperature increase caused by the impact can thus be calculated by simply time-integrating the power per area experienced at any given location on the globe (*e.g.*, the time integral of the data in Fig. 9 of [3]). If this time-integrated energy per area is E, then the mean temperature increase (ΔT), assuming the energy homogenized through the atmospheric column, was $\Delta T \approx (Eg)/(c_p \Delta p)$, where g is gravity, c_p is specific heat at constant pressure, and Δp is the pressure difference of the atmospheric layer. This temperature increase applies to the troposphere and lower stratosphere, which contain most (~95%) of the atmospheric mass and will, thus, be substantially less than surface and upper stratosphere temperatures quoted above.

The amount of heating depends on the ejected mass and the speed distribution in the plume, the latter of which can be modeled as a power law with a poorly constrained exponent. Thus far, exponent values of 0, −2, and −5 have been explored [2,3]. The results indicate that ~4000 km-diameter air columns at the impact site and antipode may have reached temperatures up to ~100 to 150 K warmer; the characteristic T spatially averaged over the heated areas are ~50 and 80 K warmer for the $e_v = -5$ and -2 cases, respectively. The temperature increase away from the impact site and antipode are ~20 K, leading to peak horizontal temperature contrasts up to 80 to 130 K. In the $e_v = -5$ case, the principal heating occurs at the antipode, with a smaller temperature increase near the impact site. For $e_v = -2$ or larger values, the opposite is true. In an intermediate case, hot atmospheric regions in both hemispheres may have been produced.

Winds: We expect that a heated region, which has low density, formed a thermal plume (Fig. 1a) that ascended into the stratosphere and expanded laterally (Fig. 1b); cooler surrounding air then converged at the surface beneath the plume (Fig. 1c). The wind speed generated in the plume can be estimated using energetic arguments. The air ascended to its level of neutral buoyancy, which was probably in the low- to mid-stratosphere depending on the exact value of ΔT and the extent to which ΔT was diluted by mixing with surrounding colder air. This released a potential energy per mass of ~$\Delta T(gd/T)$, where d is the vertical distance traveled, which was converted into kinetic energy ~u^2, where u is the mean wind speed. Thus, u ~ $(\Delta Tgd/T)^{1/2}$. As we discussed previously [3], the speed distribution in the plume is still uncertain, which affects ΔT. We consider a conservative range from 10 to 100K. Assuming d ~ 20 km (putting the level of neutral buoyancy in the lower stratosphere) and ambient T $\approx$ 300 K, then a peak ΔT of 10 to 100 K implies that u ~80 to 255 $msec^{-1}$. These speeds correspond to 290 to 920 $kmhr^{-1}$ (180 to 570 mph), the upper range of which approaches the speed of sound (295 $msec^{-1}$ or 660 mph) in the upper troposphere and lower stratosphere. We note that these winds correspond to mass-averaged speeds of the entire ~10 km tall, ~4000 km-wide heated regions; localized areas might have reached speeds several times greater.

As heated air rose into the stratosphere, inward flow occurred along the surface of the Earth to make up for the mass deficit in the troposphere (Fig. 1b,c). The characteristic expansion timescale of the thermal plume in the stratosphere was ~L/u ~ 2 to 4 × 10^4 sec, where L ~ 4000 km and u ~ 100 to 200 $msec^{-1}$ are the characteristic plume width and expansion speed, respectively. Inflowing winds must have filled the gap over a comparable timescale, implying tropospheric winds also of ~100 to 200 $msec^{-1}$. Assuming the impact generated a turbulent, well-mixed troposphere, a standard boundary equation can be used to calculate surface winds, $u \sim u_* \ln(z/z_*)$, where z is the altitude, z_* is the roughness length, and u_* is a speed that characterizes how turbulent friction affects the winds aloft. We assume winds at a 10-km altitude of 100 to 200 $msec^{-1}$. Shrubby ecosystems would have had roughness lengths of $z_* \sim 0.1$ m [5], implying winds at 2 and 10-m heights of 25 to 50 and 40 to 80 $msec^{-1}$, respectively. Forests, for which $z_* \sim 1$ m, would have experienced 10-m-height winds (near the treetops) of 25 to 50 $msec^{-1}$. Similar roughness lengths, hence speeds, would have been relevant over rough seas. These gale-force winds, which would have occurred over continent-size areas at the impact site and antipode, would extensively damage continental and marine (e.g., coral-reef) ecosystems.

Aerosols: Within the thermal plume, the tropopause was temporarily destroyed. As heated air rose towards its level of neutral buoyancy and flowed outward (away from the impact site and antipode) in the stratosphere, the plume swept up the aerosols (dominantly sea salt and dust) from the troposphere and deposited them into the stratosphere. Assuming troposphere aerosol abundances similar to that over remote continental sites and marine sites today [6], this implies 4.4 × 10^{11} g were swept into the stratosphere, a factor of 2 to perhaps an order of magnitude larger than the total mass of H_2O aerosols currently in the stratosphere (2 × 10^{10} to 2 × 10^{11} g [7]). Soot from fires generated by the impact could have also been swept up into the stratosphere.

Discussion: The wind bursts aloft and subsequent surface winds are distinct from the impact airblast generated by the expanding shock wave around the impact site. Airblast wind speeds, in excess of 500 $msec^{-1}$ near the impact site, diminished with distance, but were still capable damaging the landscape within 500 to 1000 km of the impact site [8]. Any winds generated by the atmospheric thermal plume would have occurred after the airblast passed near the impact site; furthermore, the thermal plume described here would have been the principal source of winds elsewhere in the world.

It is not yet clear if these winds would have stoked any impact-generated firestorm created by the re-entering debris or whether associated rain may have helped suppress the fires. It is also unclear how the weather generated by the firestorms interacted with the weather generated by the thermal plumes. Nonetheless, the previously unrecognized phenomena of global wind storms may have been an important environmental effect of the Chicxulub impact event. It changed, at least temporarily, the coupling between the atmosphere, hydrosphere, and surface ecosystems. The extensive surface winds would have damaged continental, near-shore, and shallow marine ecosystems, not only around the impact site but on the opposite side of the world, and longer-term climatic changes would have been compounded by additional aerosols swept up into the stratosphere. Atmospheric thermal plumes can also be produced by smaller and, thus, more frequent types of impact events on Earth and other planetary bodies with atmospheres (Venus, Mars, Titan). The Manicouagan (~214 Ma) and Popigai (~36 Ma) impact events may have generated significant atmospheric heating [9] and are potential sources of the same type of wind storm.

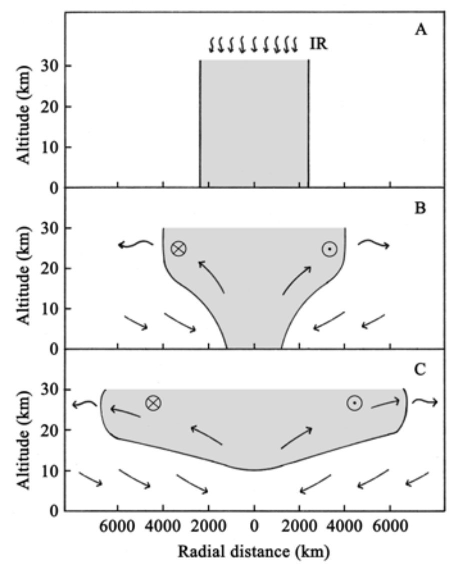

Fig. 1. Schematic time evolution of thermal plume.

References: [1] Argyle E. (1989) *Icarus*, *77*, 220-222. [2] Melosh H.J. et al. (1990) *Nature*, *343*, 251-254. [3] Kring D.A. and Durda D.D. (2002) JGR, 107(E8), 22p. [4] Zahnle K.J. (1990) *GSA Sp. Pap. 247*, 271-288. [5] Seinfeld J.H. and Pandis S.N. (1998) *Atm. Chem. & Physics*, Wiley, NY. [6] Jaenicke R. (1993) in *Aersol-Cloud-Climate Interactions*, Academic, San Diego, 1-31. [7] Turco R.P. et al. (1982) *Rev. Geophys. Space Phys.*, *20*, 233-279. [8] Emiliani C. et al. (1981) *EPSL*, *55*, 317-334. [9] Durda D.D. and Kring D.A. (2004) *JGR*, *109(E08004)*, 14p.