Convection in the Atmosphere and in the Tank Laboratory

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Abstract

Convection was examined through incompressible tank experiments and observations of the compressible atmosphere. In the laboratory, the temperature evolution and heating processes were recorded and analyzed for two initally stable, stratified configurations: one with a two-layer density gradient, and the other with six distinct temperature layers. Temperature and the height of the convecting layer varied as $\sqrt{t}$. Convection in the atmosphere, behaves much the same way, despite being compressible and requiring slight modifications to the physics of its processes. Dry convection in the desert of Yuma, Arizona, and moist convection in a Charleston, South Carolina thunderstorm were examined for the stability of their temperature profiles with height.

1 Introduction

Convection is the transfer of heat and mass in a fluid. When considering weather and the atmosphere, it is an especially important process. When warm moist air rises from the surface up to a pressure where it cools, the water vapor condenses and creating cumulus clouds. Convection is the process that is responsible for much of the heat transport in the atmosphere.

In our exploration of convection, we examined convection in the tank laboratory with the incompressible fluid of water. We examined the similarities between the tank experiments to the atmosphere by looking at the similarities between convection in incompressible and compressible fluids. In the atmosphere, moisture has a huge effect on convection, and our examination of convectino in the atmosphere included both dry and moist convection.

2 Convection in the Fluids Laboratory

2.1 Overview

The goal of this experiment was to examine how convection takes place in the simpler setting of the tank experiment. Convection in an incompressible fluid is driven by differences in density and temperature, and as shall be explained, the density in an incompressible fluid is dependent only on temperature. The heating pad in the bottom of our tanks provided the energy that drove convection. There were two different convection experiments that both started as stable situations, until the heating pad created instability. The first was the stratified experiment, where there was a stable set-up of six layers of water, each with a different temperature, the coldest at the bottom, and the warmest at the top. The second, the two-layer experiment, had a high salinity (dense) layer at the bottom, and a plain water (less dense) layer at the top. Thermometers at different levels in the tank recorded the temperature once per second for the duration of the experiment so that the temperature profile could be examined.

2.2 Theory of Convection in an Incompressible Fluid

Generally, buoyancy is the important subject to consider. If one fluid is positively buoyant with respect to another fluid, it will rise, and if a fluid is negatively buoyant with respect to another, it will sink. If the two fluids are neutrally buoyant, neither will rise or sink with respect to the other (Marshall et al., 2004b).

Where do we begin when examining buoyancy in water? For any incompressible fluid, density is a function only of temperature,

$$\rho = \rho_{ref}(1 - \alpha(T - T_{ref})), \tag{1}$$

where ρ is density, ρ_{ref} is the reference density, α is the coefficient of thermal expansion, T is temperature, and T_{ref} is reference temperature. This density-temperature relation can be examined in terms of pressure and height through the hydrostatic relation,

$$p = g \int_z^\infty \rho z. \tag{2}$$

If we examine a stable setup (as in Figure 1) with fluid parcels that have been perturbed from their environment, we discover that the environment's density at (for instance) height 2 is

$$\rho(z_2) \approx \rho_1 + \frac{\partial \rho}{\partial z}_E \delta z \tag{3}$$

where $\frac{\partial \rho}{\partial z}_E$ is how density varies with height, and δz is the height difference between heights 1 and 2 (Marshall et al., 2004b).

The buoyancy of the parcel is dependent on the environmental density profile or how $\frac{\partial \rho}{\partial z}_E$ changes. The buoyancy relation is described by

positively buoyant if

$$\left(\frac{\partial \rho}{\partial z}\right)_E > 0; \tag{4}$$

neutrally buoyant if

$$\left(\frac{\partial \rho}{\partial z}\right)_E = 0; \tag{5}$$

negatively buoyant if

$$\left(\frac{\partial \rho}{\partial z}\right)_E < 0. \tag{6}$$

This translates directly to the temperature profile as

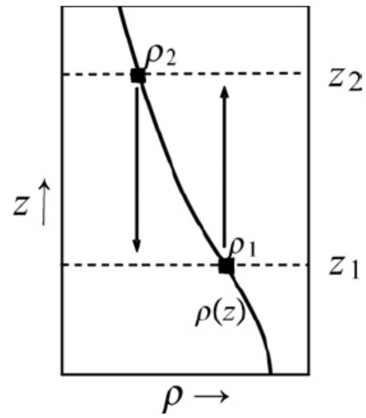

Figure 1: Perturbed fluid parcels in a height dependent situation. As parcels at densities ρ_1 and ρ_2 are perturbed from their environments their density against their new environment's density will either sink or rise based upon its buoyancy. Image from (Marshall et al., 2004b).

stable:

$$\left(\frac{\partial T}{\partial z}\right)_E > 0; \tag{7}$$

neutral:

$$\left(\frac{\partial T}{\partial z}\right)_E = 0; \tag{8}$$

unstable:

$$\left(\frac{\partial T}{\partial z}\right)_E < 0. \tag{9}$$

What this states is that if temperature increases with height, then the situation is stable to perturbations, but if it decreases with height, then the incompressible situation is not stable and convection will occur (Marshall et al., 2004b).

When considering the energy in this system, the underlying principle is that energy will always be minimized. If our initial potential energy for the setup in Figure 2.2 is $PE_i = \rho_1 g z_1 + \rho_2 g z_2$, then after the perturbation our final energy is $PE_f = \rho_1 g z_2 + \rho_2 g z_1$. If the change in potential energy is less than zero, then conservation of energy implies that the new setup is more stable and that convection will not occur. If the change in potential energy is greater than zero, then the new set up is less stable than the previous setup, and convection will occur to minimize potential energy once again (Marshall et al., 2004b).

In the tank, the environment's temperature can be described by

$$T = \Gamma z, \tag{10}$$

where Γ is $\frac{\partial T}{\partial z}$, the lapse rate in the water, and z is the height. From thermodynamics,

$$H = z \rho c_w \Gamma \frac{\partial z}{\partial t}. \tag{11}$$

Here, H is the heat flux, t is time, and c_w is the heat capacity of water (4.18 J/g).

From here using some calculus identities we see that

$$\frac{1}{2}\frac{\partial^2 z}{\partial t^2} = \frac{H}{\rho c_w \Gamma}, \tag{12}$$

and from there,

$$z^2 = \frac{2H}{\rho c_w \Gamma} t + K. \tag{13}$$

Here, K is a constant of integration, and can be set to zero for our initial parameters when the height of the convecting layer is zero. We can see that the height of the convecting layer increases as $\sqrt{t}$, and by Equation 10, temperature must also increase like $\sqrt{t}$ (Tang).

It is more useful to write Equation 13 as

$$z = \sqrt{\frac{2H}{\rho c_w \Gamma} t} \tag{14}$$

where we assume the initial constant is equal to zero. If we know the height of the convecting layer with time, then we can use a linear fit to find the lapse rate, Γ by taking the natural log of both sides of Equation 14. For instance,

$$ln(z) = \frac{1}{2} ln(t) + \frac{1}{2} ln\left(\frac{2H}{\rho c_w \Gamma}\right) \tag{15}$$

and finally,

$$\Gamma = \frac{2H}{\rho c_w e^{2(ln(z) - ln(t))}}. \tag{16}$$

Calculating the lapse rate from Equation 16, will be very useful in our analysis of the convecting tank results.

2.3 Procedure and Results in the Stratified Experiment

The stratified convection experiment consisted of setting up a stable situation in the tank where we had five to six layers of water, each with a separate temperature (see Figure 2. We set up five temperature probes on the side of the tank at 1.6, 2.6, 4.9, 7.9 and 10.7 inches from the bottom of the tank. Placing in the different layers of water had to be done very carefully, such that the different temperatures didn't mix and ruin the stratification. A floating device consisting of steel and sponges was placed on the surface, and allowed from water to flow out of the sides instead of down, which kept the layers separate. Our layers were initally at 23°C (from 0-3 inches), 29°C (from 3-5.4 inches), 32°C (from 5.4-7.9 inches), 35°C (from 7.9-9.9), 38°C (from 9.9-12.7 inches), and 45°C (from 12.7-15 inches). Each thermometer was placed in a separate layer of water. The top layer did not have a thermometer in it and was used as a hot lid to keep heat from escaping out of the main tank into the surroundings. Once the tank was full, the heating pad (with an area of 18 by 18 inches and energy of 1620 watts) was switched on. A small bit of potassium permanganate was added in so that we could observe the convecting layer as it rose up.

From the thermometers, we were able to plot the thermometers versus time of the experiment as seen in Figure 3. It should be noted that the initial temperatures read

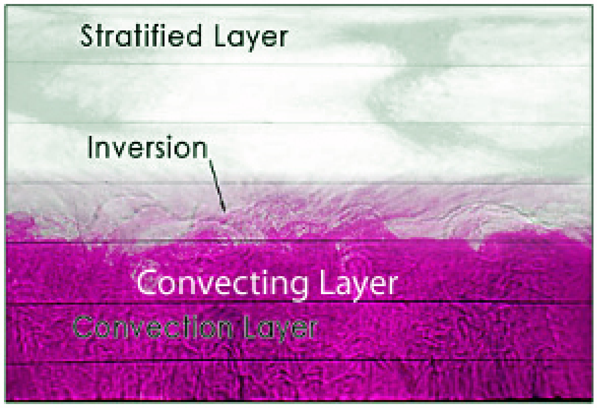

Figure 2: Setup for the Stratified Experiment. Five or six layers of increasingly warmer water were layered carefully (to avoid mixing). After a heating pad was switched on beneath the tank, crystals of permanganate showed the increase in heigh of the convecting layer and revealed as the convecting layer broke through each layer of warmer water. Image from Weather in a Tank: http://www-paoc.mit.edu/labguide/convect.html

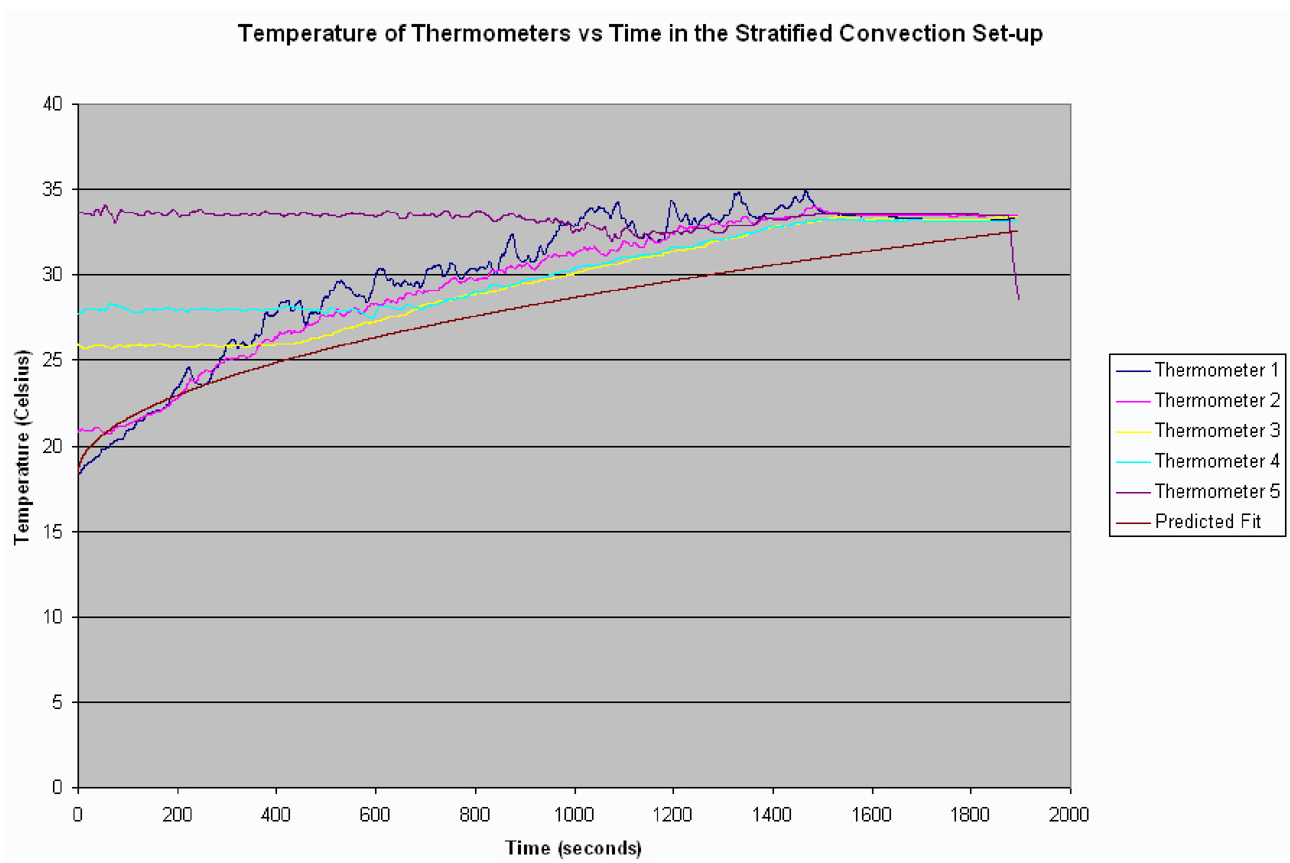

Figure 3: Temperature of the Thermometers with Time. As the convecting layer reaches each thermometer, the thermometer joined along the main trend, which increases approximately as $\sqrt{t}$. The predicted value (the sixth line) is calculated from the lapse rate Γ which was determined using the height of convecting layer as it changed with time.

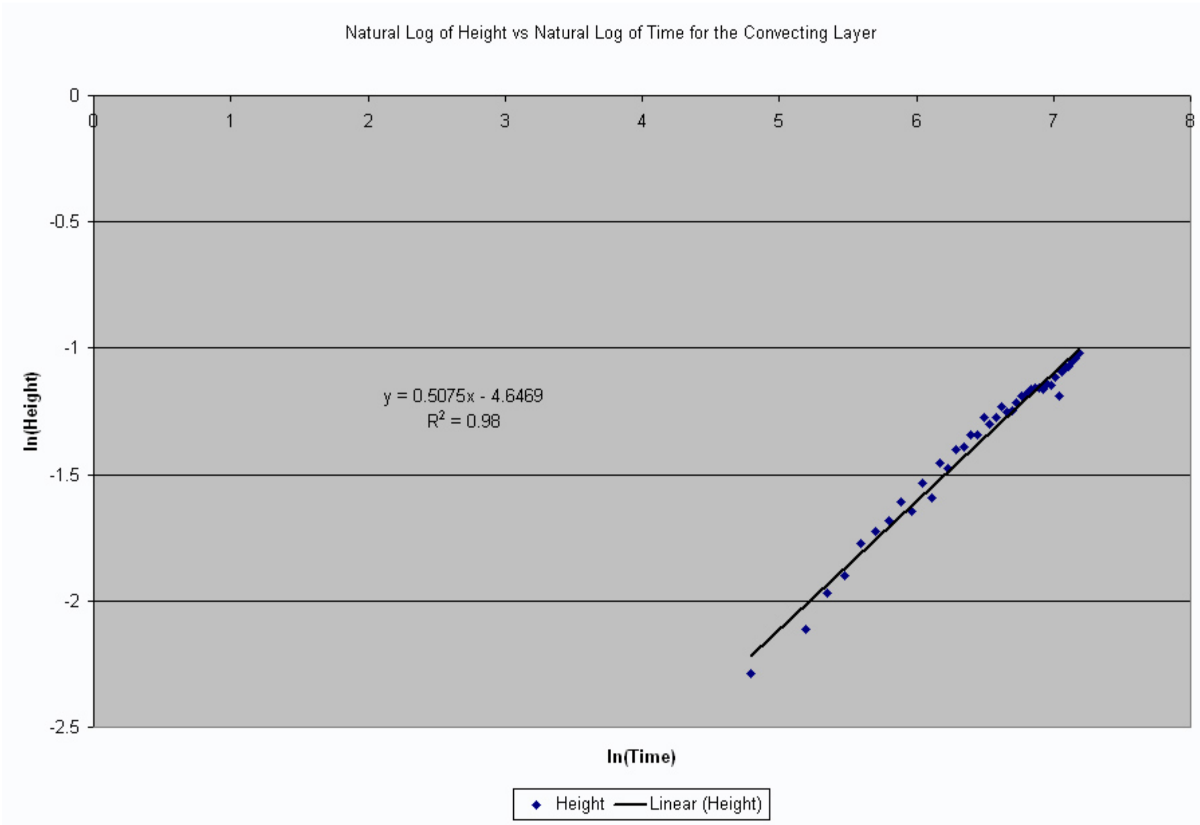

Figure 4: Natural Log Plot of Convecting Layer Height versus Time for the Stratified Experiment. As expected in Equation 15 we see that this equation's line of best fit ($y = 0.5075x - 4.6469$) has a slope near 1/2. From the y-intercept we can calculate the lapse rate, Γ from Equation 16. We found Γ to be 39.76°C/meter.

from the water source do not match the initial temperatures as recorded by these thermometers. This is likely because the water cooled a bit on its way from the source to the tank. As time increased, each thermometer joined in the main temperature trend (which increases as $\sqrt{t}$) as the convecting layer reached each thermometer. We determined the lapse rate of the system by our observations of the height of the convecting layer with time; height should also have a $\sqrt{t}$ relation. Using Equation 16, we were able to determine Γ to be 39.76°C/meter. This was determined from the best fit seen in Figure 4, where the slope is seen to be very close to 1/2 (as expected in Equation 15) The y-intercept of this best fit directly led to the best fit for the lapse rate from Equation 16. Once we had Γ we were able to visually compare our observed height of the convecting layer with the height we would expect as in Figure 5 and predict a best fit temperature curve for Figure 3. For the height of our calculated convection layer (which uses the best-fit lapse rate), the calculated height fits the observations (in Figure 5) very well, since the lapse rate was calculated as a best fit from this data. When we move this lapse rate to predict the temperature curve as a function of time (in Figure 3), the predicted curve is a bit lower than the observed curves from the thermometers. This could be due to inaccurate observations of the convecting-layer height (which would change Γ) and to the original set-up of the stratified experiment, which had near-discontinuities between the layers. Our model assumes that the lapse

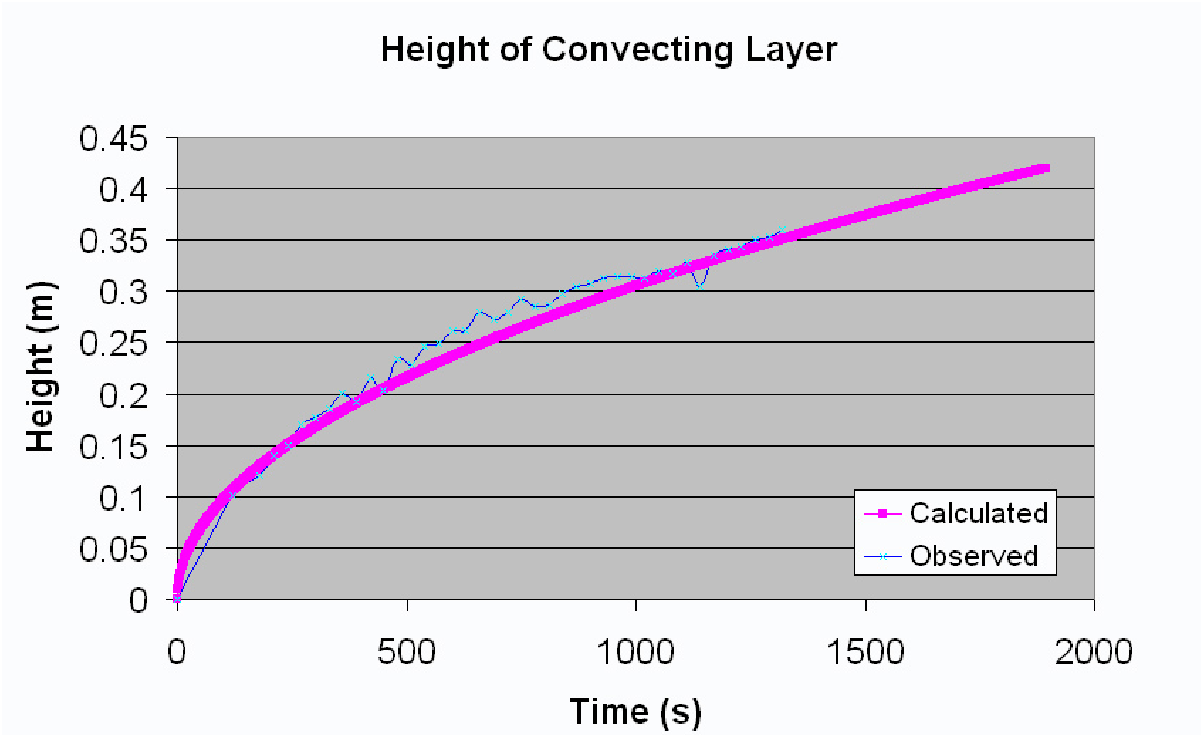

Figure 5: Height of the Convecting Layer with Time for the Stratified Experiment. Notice that the calculated curve fits very well (since it is calculated from these observed convecting layer heights).

rate would create a linear increase of temperature with height, which is the eventual outcome after convection. Creating such a linear set-up initially is difficult to achieve in the lab.

2.4 Procedure and Results in the Two-Layer Experiment

The two-layer experiment was conducted with the same premises as the stratified experiment (see Figure 6). With five-thermometers placed at heights of 2, 4, 6, 8 and 10 inches, the tank consisted of a dense (high-salinity) section of water underneath a less dense, fresh water layer; the density of the lower section was 1020 kg/m^3. Again, water was carefully added to avoid mixing before the experiment began. The initial inversion (the interface between dense and light fluid) was at 6 inches, the same height as the third thermometer. The heating pad in this experiment was smaller with an initial area of 10 x 10 inches and a power of 500 watts.

Figure 7 shows the thermometer's results from the two-layer experiment. The bottom two thermometers show a very similar linear temperature profile, as they are both in the convecting layer. The third (red) thermometer is at the interface between the two layers and is increasing slightly as time increases. This result is rather expected considering that the bottom layer should be becoming more energetic with time, and should eventually push the convecting layer past the interface. For now though, we see that the temperature near the interface is increasing; had there been an extra hour for our experiment, it is likely that we would have seen the interface be crossed. The top thermometer, in the fresh layer, showed a slight temperature decrease, likely

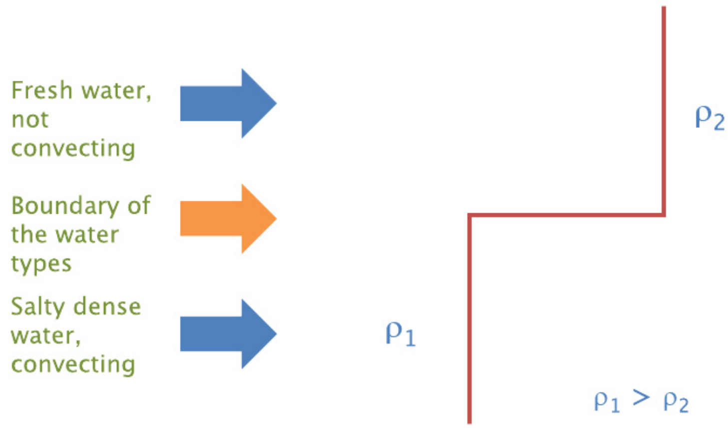

Figure 6: Set-up for the Two-Layer Tank Experiment. Dense, high-salinity water was below less dense, fresh water, creating a stable situation as by Equation 4. Once the heating pad is turned on, the hot water at the bottom will convect up until it hits the inversion between the two layers. If the dense layer gets hot enough that it can convect into the fresh water, then both layers will begin to mix.

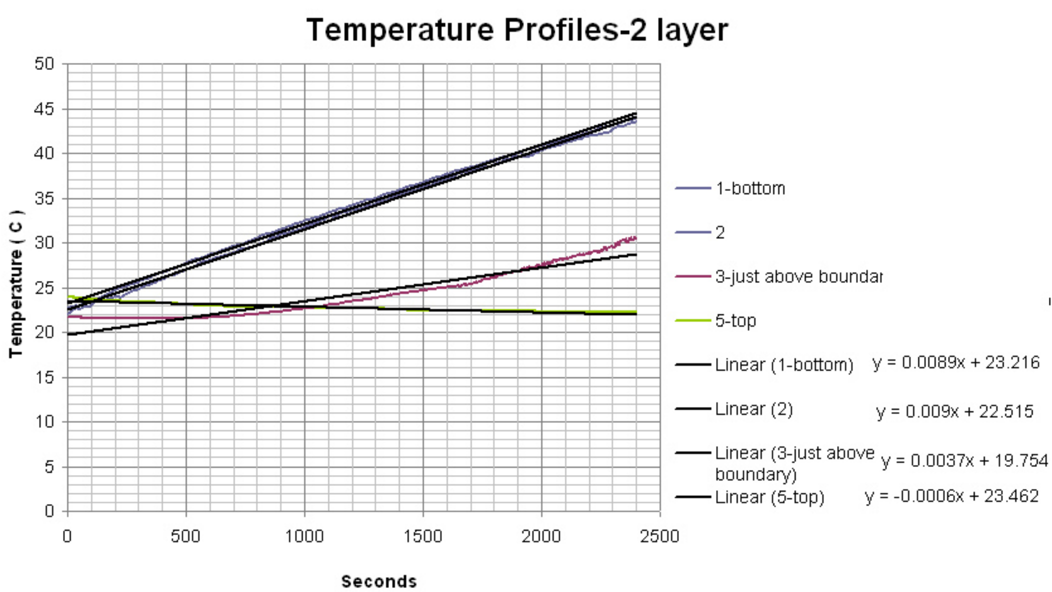

Figure 7: Temperature versus Time in the 2-layer Tank Experiment. Four thermometers recorded the temperatures of the various layers. The two thermometers that increased linearly were in the convecting, bottom layer. The thermometer that decreases slightly was in the top, fresh water layer. The red thermometer was at the interface between the two layers. Linear fits are shown for each thermometer's data.

because this layer had no convection and was actually losing heat to the surrounding environment.

When observing the convection in both the temperature and stratified experiments, the height of the convecting layer is not constant; in fact, there a generally plumes that break through the interface, creating waves and an undulating bound to the mixed layer. The thermometers record these fluctuations in temperature from rising and falling plumes. The variance for each thermometer was calculated by subtracting the linear fit from the observed temperature and then taking the square root of its magnitude. For the lowest thermometer, this average variance was found to be around 0.72°C.

3 Convection in the Atmosphere

3.1 Overview

While the tank experiments are examining convection, their convection is in an incompressible fluid. The atmosphere, composed of compressible air, has different thermodynamics involved that must be taken into account. Still, the main ideas behind convection are the same; an atmosphere with a stable profile will not convect, while one that is unstable will convect.

3.2 Theory of Convection in Dry Compressible Fluids

Unlike incompressible fluids, compressible fluids show density, ρ as a function of both temperature and pressure, or $\rho = \rho(p, T)$. In the atmosphere as pressure decreases (and height from the surface increases), temperature and density decrease rapidly. We will need to assume a few new pieces of information about our fluid. First we will assume that it follows the ideal gas law,

$$p = \rho R T, \tag{17}$$

where R is the universal gas constant for dry air, a mixture of nitrogen, oxygen and other gases, neglecting water vapor. The second assumption is that any movements of air are done adiabatically. An adiabat is a path along which entropy is conserved and no work is done; for our purposes, adiabatic movements imply that no heat is transferred from the air parcel to its surroundings. As motions of convection in the atmosphere is generally faster than radiation, adiabatic motions are a safe assumption.(Marshall et al., 2004a)

For a unit parcel of mass one, the first law of thermodynamics tells us that

$$\delta Q = c_v dT + p dV, \tag{18}$$

where c_v is the heat capacity at constant volume and V is volume. δQ, change in heat, is zero since all movements are adiabatic. For our unit mass, $\rho V = 1$ so, $dV = d(1/\rho)$. From there we know that

$$dV = -\frac{1}{\rho^2} d\rho \tag{19}$$

and with the Ideal Gas Law (Equation 17),

$$pdV = -\frac{RT}{p}d\rho = -\frac{dp}{\rho} + RdT. \tag{20}$$

Knowing that $R + c_v = c_p$, where c_p is the heat capacity at constant pressure, then the 1st Law of Thermodynamics (Equation 18) can be rewritten as,

$$\delta Q = (R + c_v)dT - \frac{dp}{\rho} = c_p dT - \frac{dp}{\rho}. \tag{21}$$

We know that all motions are adiabatic, so $\delta Q = 0$, and hydrostatic balance states that $dp = -\rho_E g dz$ where ρ_E is the density of the environment. We finally see that our lapse rate becomes

$$\frac{dT}{dz} = -\frac{g}{c_p} = -\Gamma_d \tag{22}$$

the dry adiabatic lapse rate. In the atmosphere this is approximately 10K/km. (Marshall et al., 2004a)

We then find that our stability requirements have changed slightly to show such that

stable:

$$\left(\frac{\partial T}{\partial z}\right)_E > -\Gamma_d; \tag{23}$$

neutral:

$$\left(\frac{\partial T}{\partial z}\right)_E = -\Gamma_d; \tag{24}$$

unstable:

$$\left(\frac{\partial T}{\partial z}\right)_E < -\Gamma_d. \tag{25}$$

If the temperature is decreasing with height faster than the adiabatic lapse rate, then the situation is unstable. The atmosphere is almost always stable to dry motions, but convection can occur, especially in deserts. Sometimes, superadiabatic processes can occur, when convection cannot occur fast enough to move heat away from the surface.

If a process is adiabatic, we can define a quantity, potential temperature (θ), where for adiabatic motions (constant entropy), $\delta\theta = 0$. Potential temperature,

$$\theta = T\left(\frac{p_0}{p}\right)^{\kappa}, \tag{26}$$

is the temperature that a parcel would have when moved adiabatically to a reference pressure, p_0 (κ is R/c_p). For adiabatic movements, potential temperature is conserved. If a parcel is moved along a surface of constant potential temperature, then the parcel is neutral with respect to the atmosphere. If it increases its potential temperature with height, it is in a stable configuration, whereas if it decreases its θ with height, it is unstable. To restate,

stable:

$$\left(\frac{\partial \theta}{\partial z}\right)_E > 0; \tag{27}$$

neutral:

$$\left(\frac{\partial \theta}{\partial z}\right)_E = 0; \tag{28}$$

unstable:

$$\left(\frac{\partial \theta}{\partial z}\right)_E < 0. \tag{29}$$

Seeing stability and instability on temperature profiles from radiosondes as they travel through the troposphere is easy with potential temperature. For example, if the slope of the potential temperature profile is positive, then the atmosphere in that region is stable and will not convect (Marshall et al., 2004a).

With the tools of dry compressible convection and potential temperature in hand, it is time to examine convection in the Earth's atmosphere.

3.3 Dry Convection Observations and Discussion

With the relative simplicity of dry, compressible convection in mind, we first examined hourly radiosonde profiles from Yuma, Arizona. Radiosondes are balloons that are launched and record temperature, pressure, moisture and other variables as they travel up into the atmosphere. As Yuma is in a desert (see Figure 8), we could be assured that there was little moisture affecting the convection that we would hopefully observe in the data. In the case of Yuma, temperature and pressure were recorded. With a reference pressure at the surface, these observations were used to calculate the potential temperature at every available pressure (height) in the atmosphere (see Figure 9).

In these potential temperature profiles, it is important to note that for the first time's profile (blue) is very stable. The slope of this line is positive for all pressures/heights. As the day heats up, we see the overall profiles shifting to the warmer right. There is also a vertical section near the surface that begins to appear after around 10:45 am (the cyan profile, #4). Here is where the atmosphere is neutral and this shows the height of the convecting layer; the height of the convecting (mixed) layer is increasing as the day continues. It is not common for the dry atmosphere to have an unstable profile as it is usually able to convect heat away fast enough, which results in the neutral, convecting signature, a straight line. We also see some instability at the very surface, where the negative slope is an indication of a superadiabatic lapse rate. Here, heat cannot be convected away fast enough to return the profile of the atmosphere to a neutral or stable situation.

As seen before in the tank experiments, convection transports heat. In Yuma, we can estimate the amount heat per square meter from the sun (as recorded by station surface data) and see how much is convected into the atmosphere (as recorded by radiosondes (see Figure 10). This heat per area (energy density) can be calculated from the radiosondes by an equation related to Equation 13 from the tank experiments,

$$H\Delta t = \frac{c_p}{g}\sum_{i=1}^{N} \Delta T_i \delta p_i \tag{30}$$

(Illari, 2009).

We see that the amount of incoming solar radiation is greater than the amount of heat per square meter as recorded by the radiosondes. However, the most important

Figure 8: Surface Conditions of Yuma, Arizona. Yuma, Arizona, near the Mexican border is located in a desert, making it the perfect location to observe moisture-free, dry convection. Image from http://platial.com/img/2007/02/0/Yuma_AZ.jpg.

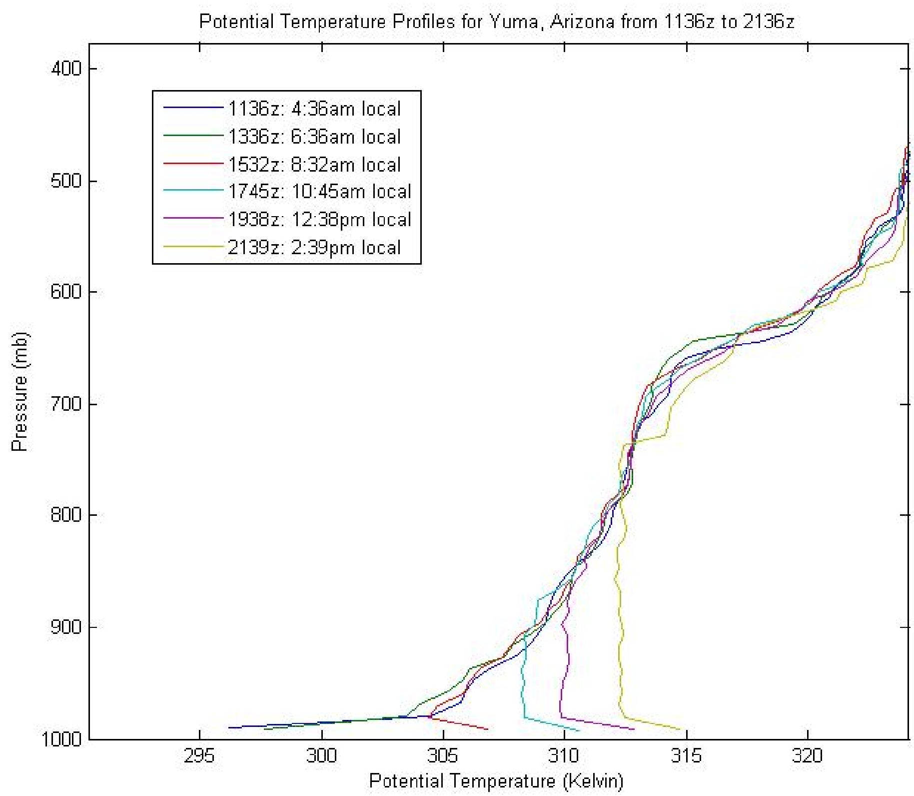

Figure 9: Potential Temperature Profiles for Yuma, Arizona from 4am to 2pm local time. As the sun warms the surface, we can see the potential temperature profiles generally moving to the right; as the day continues, we can also see a small, superadiabatic section at the surface where the heat cannot be convected away fast enough. There is also a vertical, convecting region that has an increasing height as time increases.

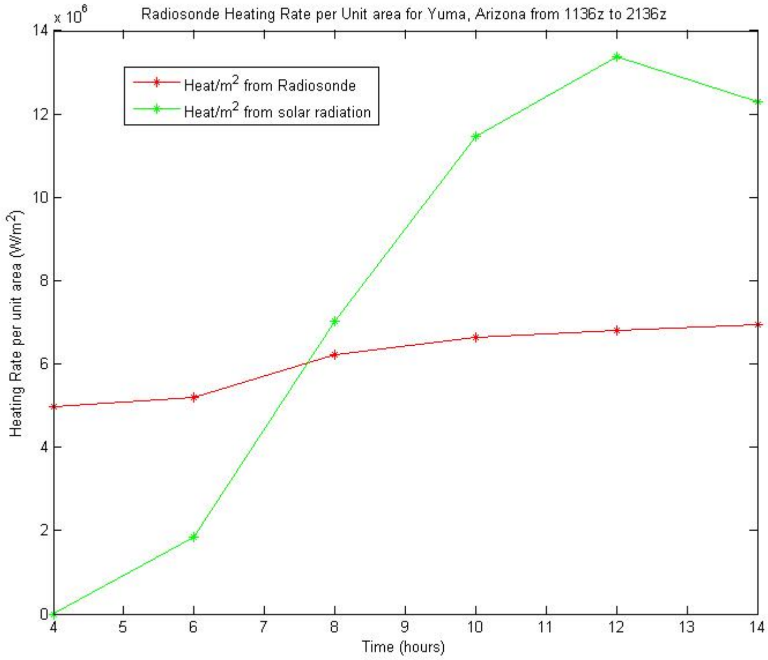

Figure 10: Heat per square meter as recorded by surface data and radiosondes in Yuma, Arizona. Notice that energy density for incoming solar radiation is much greater than that recorded by the radiosondes, which is assumed to be convecting. A good portion of the incoming solar radiation was likely reflected.

feature is that both are of the same order of magnitude, $10^6 W/m^2$, indicating that convection is an important process in the heat transport through the atmosphere.

3.4 Theory of Convection in Moist Compressible Fluids

Unfortunately, the atmosphere is rarely dry; most of the time, it has a non-negligible amount of water vapor present. This water vapor imparts a specific humidity q which equals ρ_v/ρ, where ρ_v is the density of water and ρ, the density of all air is approximately ρ_d, the density of dry air. When the air is holding the maximum amount of water vapor, then its saturated specific humidity $q^* = \frac{e_s/RT}{p/R_vT} = \frac{e_s}{p}\frac{R_v}{R}$; e_s is the partial pressure due to water vapor and R_v is the gas constant for water. The partial pressure of water vapor is a function only of temperature. As a moist parcel rises, while it cools, its q^* is decreasing; when the temperature of the parcel is such that $q = q^*$, then water vapor begins to condense after this point (Marshall et al., 2004a).

Before we assumed that dry convection was adiabatic, but in the case of moist convection, when water vapor condenses, a latent heat is released, and δQ is not zero. We consider such motions to be moist adiabatic, although they are not strictly adiabatic as they release heat into the surroundings. Latent heat can be described as $\delta Q = -L\delta q$, where L is the latent heat released and dq is the differential amount of specific humidity. Equation 21 must be revised as

$$c_p dT = \frac{dp}{\rho} - L dq^*. \tag{31}$$

From here, since the saturated vapor pressure of water e_s is an exponential-like function of temperature only, we can express $\frac{de_s}{dT} = \beta e_s$. With hydrostatic balance and some rearrangements of the differential form of q^*, we'll find that the moist adiabatic lapse rate, Γ_s, can be described by,

$$-\frac{dT}{dz} = \Gamma_s = \Gamma_d \left[\frac{1 + Lq^*/RT}{1 + \beta Lq^*/c_p}\right]. \tag{32}$$

It is important to note that $\Gamma_s < \Gamma_d$. In the atmosphere, Γ_s is approximately $4 - 7°\mathrm{C/km}$. The stability conditions in Equations 23,24 and 25 can easily be transformed into conditions for moist convection by simply replacing Γ_d with Γ_s. Also analogously, moist potential temperature (derived from the 1st Law of Thermodynamics, and κ) is defined as

$$\theta_e = \theta \exp \frac{Lq}{c_p T} \tag{33}$$

This potential temperature behaves the same way as normal potential temperature does in Equations 27, 28, and 29 as it is conserved in dry and moist processes. For dry convection, $q = 0$, so $\theta_e = \theta$ (Marshall et al., 2004a).

3.5 Moist Convection Observations and Discussion

Although dry convection is simpler, it ignores the ubiquitous water vapor that has a great effect on heat transport. As water vapor will produce a latent when it condenses, it will make a difference in convection through its smaller lapse rate and its moist adiabatic lapse rate (which is not truly adiabatic).

During the period in which we studied convection, there were major thunderstorms that went through the Southeastern states in the United States. Thunderstorms involve deep moist convection and were certain to produce intriguing radiosonde profiles. (One caveat however is that radiosondes were likely sent out either before or after the thunderstorm passed given these storm's inherent lightning dangers.) Charleston, South Carolina was hit by significant rain and convection on late, April 13 and early, April 14, 2009 as can be seen in the radar image at the time (in Figure 11).

Using the equations for potential temperature and moist potential temperature (26 and 33), a short time series for South Carolina was plotted to see how the forecasted profiles changed as the convective system approached (see Figure 12). For the entire time series, the dry potential temperature showed a stable atmosphere where the temperature was increasing for all heights. However, observationally, we know that convection was occurring: South Carolina was hit by a thunderstorm. Dry potential temperature is not sufficient; moist potential temperature must be used. The saturated moist potential temperature (also on Figure 12) shows that the temperature throughout the day increased and then when the storm hit at 3 UTC (cyan), we notice that the moist profile becomes neutral for much of the lower to middle troposphere. Before the storm hit (the two previous profiles, red and green), we also notice conditional instability, the negative profiles at the surface, which indicate that if moisture were present, then this location would have convection occuring. Moist potential temperature reveals that moist convection has occurred, while dry potential temperature does not see the possibility for this convection.

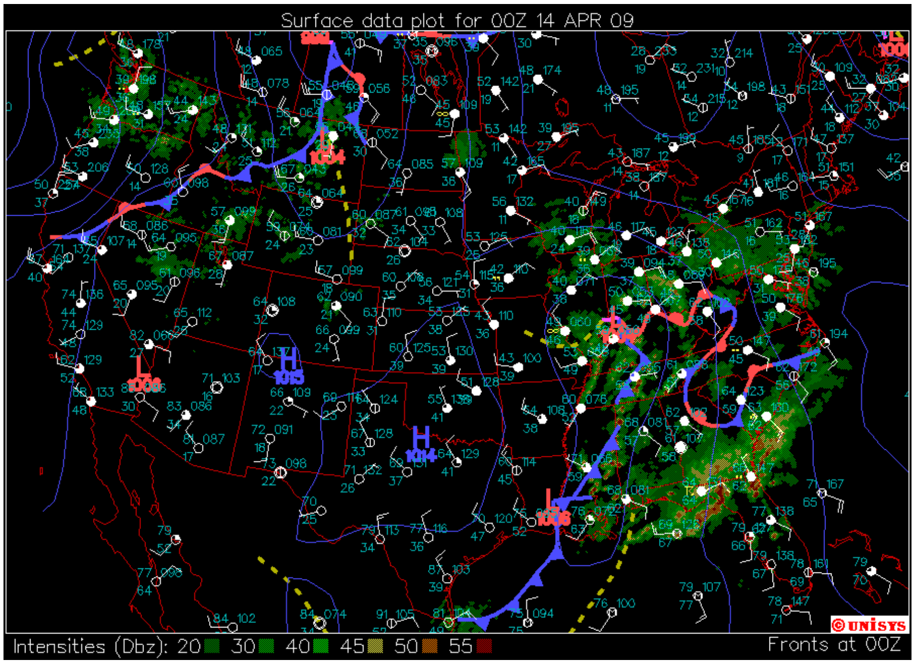

Figure 11: Archived Surface Analysis of the Continental United States on April 14, 2009 at 00 UTC. Charleston, South Carolina (although through the brunt of the storm at this point) has been hit by a major thunderstorm.

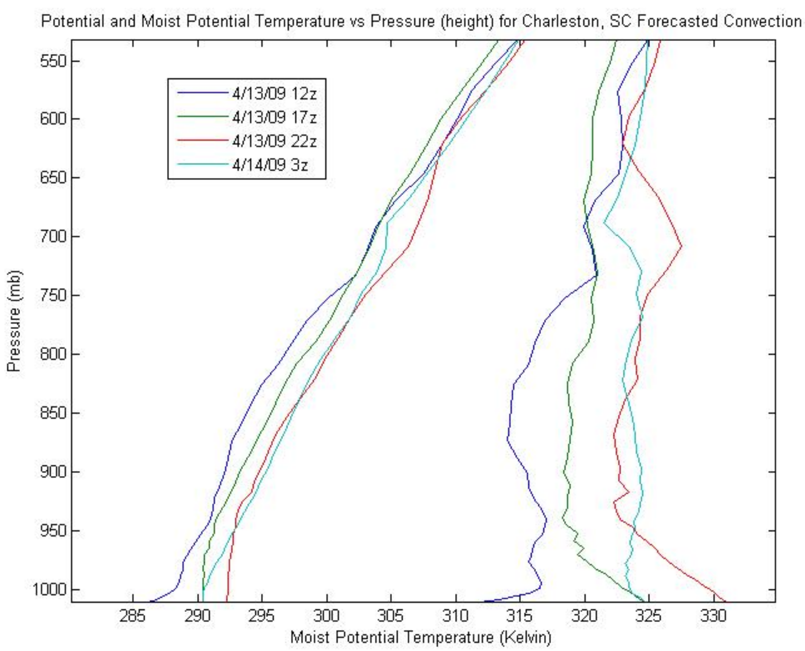

Figure 12: Dry and Moist Potential Temperatures over Charleston, South Carolina. While the dry potential temperature shows a stable atmosphere, the moist potential temperature reveals the atmosphere's actual susceptibility to moist convection.

4 Conclusions

4.1 Connections between the Laboratory and Atmosphere

Although there is a major difference between the laboratory tank convection and atmospheric convection, namely the degree of compressibility (or incompressibility), the similarities between the two processes are significant: instability produces convection. For the tank experiments this instability meant that a warmer or lighter fluid was underneath a cooler or heavier fluid. Although heavy and light, cool and warm, are terms that cannot be used for the atmosphere, the idea is the same. If a fluid has a higher potential temperature and it is below one with a lower potential temperature, convection will occur. The stability criteria for all three cases (incompressible, dry compressible and moist compressible) of convection are similar when explained in terms of their appropriate conserved quantities.

4.2 Concluding Remarks

Convection, as the atmosphere's main process for transporting heat and mass, depends on the stability of the environment. To understand how this process worked, an incompressible analogue was examined in the laboratory with stable configurations created with temperature and density stratifications. For the stratified-temperature, experiment, the convecting layer eventually reached the top of the tank and we found a height and temperature profiles for the convecting layer followed $\sqrt{t}$. The 2-layer, density-stratified configuration was more stable than the other. Convection engulfed the lower denser layer almost immediately after the heating pad was started, but it never broke through into the lighter layer on top. It is likely that our high density fluid was too salty to observe a complete mixing of the tank during the allotted lab period time. In the atmosphere, we saw dry convection occured in the mixed layer over deserts where moisture was not available. The mixed (convecting) layer's maximum height increases through the day, as seen as the neutral portion of the potential temperature plots increases through the day. When the potential temperature is kept constant, the situation is neutral, and convection has occurred. In the moist atmosphere, we found that potential temperature shows a stable configuration for the atmosphere, and does not take into account the major effect of water vapor's release of latent heat. Moist potential temperature, which does take this latent heat into account, is used to show conditional convection, which is dependent on the availability of water vapor; when convection has happened, the moist potential temperature becomes constant with height, indicating that neutrality in the atmosphere has been achieved. Convection, in all of its facets, effectively transports heat and mass to minimize the energy of the atmosphere system.

4.3 Future Work

Although fluctuations in the temperature data from the tank experiments is inherently chaotic, it would be interesting to examine it for any frequencies that appear to dominate more strongly, and to see if this could explain anything about the convection in the system. Examining moist convection more fully with cases where the saturation specific humidity was not reached would be interesting. Comparing the continum of atmospheric convection from dry to moist to saturated moist would likely point out interesting moisture-dependent features in the evolution of convection.

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