UPWELLING FLOW RATES AND WATER MIXING STUDY.

1. Flow Rate Data.

This review seeks to clarify the expected upwelling flow rates produced by wave-powered devices such as offered by Ocean-based Climate Solutions, Inc., including data and reports compiled since 2005 by its parent company Atmocean, Inc.

As a chronological summary, Atmocean, Inc. was founded in 2006 to commercialize the wave-powered upwelling technology designed by its CEO Philip Kithil and first tested in October 2005 off Corpus Christi TX and then in December 2005 deployed from the Weatherbird II ocean research vessel operated by Bermuda Biological Station for Research (BBSR, now BIOS). The latter design (10” diameter and 500’ deep) showed cold water reaching the surface in about 60 minutes. Assuming 100% efficiency and tube volume of ~273 ft³ (7.7 m³) gives flow rate of 7.7 m³/hour. Significant wave height was estimated at 1.8 m and wave average period of 8 s.

In a subsequent test in March 2007 off San Diego of 30” diameter by 500’ deep pump instrumented with thermistors top and bottom and triaxial accelerometers on the bottom-mounted valve flappers (to verify pump start time), Atmocean found the deep cold water reached the surface in about 11 minutes after first pump stroke, as seen in Figure 6 from http://www.ocean-based.com/wp-content/uploads/2020/06/white-paper.pdf. Again assuming 100% efficiency, with the tube volume of 2,450 ft³ (69.5 m³) gives flow rate of 379 m³/hour. Significant wave height from NDBC data buoy 46258 was 1.3 m and wave average period of 7 s.

In 2008, Atmocean supplied University of Hawaii with several 30” diameter by 1,000’ deep pumps for testing filmed by Discovery Channel for its program “Hungry Oceans”. Manufacturing constraints required two 30” by 500’ tubes connected end-to-end using a steel cylinder “coupler”. This test was documented in the paper by Angelique White et.al. “An Open Ocean Trial of Controlled Upwelling Using Wave Pump Technology” published by American Meteorological Society in volume 27 of its JOURNAL OF ATMOSPHERIC AND OCEANIC TECHNOLOGY, which has been widely cited as source data for wave-driven upwelling flow rates.

In constructing the 1,000’ tube, UH reconfigured the design with mid-couplers at the top of the tube on three of the four pumps, and spooled-up both top and bottom sections onto the mid-coupler. This caused excess strain on the plastic tube material as it became the load-bearing element, rather than the inner steel cable.

During deployment this package flipped causing the tube to become twisted. This twisting prevented upwelling until the twists resolved, reducing the volume of water forced up the tube and the corresponding flow rate. One can see the twisted tube in this photo from the White et.al. paper.
The retarded flow is evidenced by the lengthy time (3.1 hours) documented in the White paper for cold water from 1,000 feet deep to reach the top thermistor. White et.al. further report “Within an hour after the pump had reached its target depth, both midcoupler-mounted thermistors recorded temperatures 2.8°C below expected for water at 165 m. In another two hours time, temperatures at the top coupler were depressed by greater than 1.8°C relative to the surrounding water. This difference in time of one hour between the lower tube flow rate and upper tube flow rate is not discussed but suggests more twisting in the upper tube.

Again assuming 100% efficiency, White et.al. report states “Using the volume of the single pump (136 m³) and the elapsed time required for cold water to reach the top coupler…we estimate the rate of artificial upwelling over the operational phase of deployment to be approximately 45 m³ per [hour]”. Significant wave height from NDBC data buoy 51001 was 1.4 m and wave average period of 6 s.

Our flow rate from our March 2007 test of 379 m³/hour is thus 8.4 greater than the flow rate of 45 m³/hour reported by White et.al.

2. Water Mixing Study.

In 2006, Atmocean, Inc. retained Professor Isaac Ginis from University of Rhode Island Graduate School of Oceanography to evaluate the mixing effects induced by our wave-driven upwelling pumps.

The key questions were:

A. Being colder and therefore denser, will the upwelled water sink back down or will it remain above the thermocline?
B. If it remains above the thermocline, what are the physical mixing processes and what are the resulting changes in upper ocean temperature?
C. Will this cooling of the upper ocean reduce hurricane intensity, and if so by what magnitude?

The full study is copied below.
I. Theoretical and Numerical Investigations of Upper Ocean Cooling by Wave-Driven Pumps

Introduction

Atmocean, Inc. is developing wave-activated pumps that bring cold water from the deep ocean to the surface to reduce the upper ocean layer heat content. The efficiency of this method can be demonstrated by the following simple estimations. Let’s say that a 1-m$^3$ water parcel located at 500-m depth is 10°C cooler than a 1-m$^3$ water parcel located at the surface. The temperature difference causes the parcel at 500-m depth to be 2.5 kg heavier than the surface parcel. To pump the parcel at 500-m depth up to the surface would require about 1250 J/m$^3$ of energy. On the other hand, cooling the surface parcel 10°C by other means would require 4x10$^7$ J/m$^3$ of energy - 30,000 times more! While this new technology of cooling the ocean surface seems to be efficient, there are significant scientific and technological questions that have to be addressed before it may be considered for the hurricane mitigation problem.

Ginis Consulting, LLC has undertaken a theoretical investigation sought to answer the following questions:

1. What processes are responsible for the spatial distribution of cool water in the upper ocean layer produced by a single pump?
2. What are the quantitative estimates of the temperature anomalies, depth of penetration, horizontal size, and descend rate of the cooled area generated by a single pump?
3. What are the effects of stratification due to salinity in the upper ocean mixed layer?
4. What processes are responsible for dispersion of the cold water intrusion after it is released by a single pump? What are the qualitative estimates of the spatial and time scales involved?
5. What processes are responsible for merging the cold water intrusions produced by multiple pumps? What are the qualitative estimates of the spatial and time scales involved?
1. Plume

Cold water lifted by a wave-activated pump to the sea surface is heavier than surrounding water and therefore must fall due to gravity. A simple theoretical model of this process can be described as follows. Continuous pumping of cold water to the surface is equivalent to the action of a localized sink of heat (negative buoyancy flux) under which a stationary convective flow of cold water develops. The resulting buoyant region can be described as a “plume” - a column of one fluid moving through another. Usually, a plume widens as it descends because of entrainment of the surrounding fluid at its edges. Density stratification of the ambient fluid influences the buoyancy within the plume. In a stably stratified environment, buoyancy of the descending plume increases with depth and will eventually become neutral or positive, stopping the plume’s descent.

Typically, self-induced turbulence in a plume far exceeds the ambient turbulence. For the purpose of this investigation, we initially assume that the surrounding fluid is at rest. There is a great deal of literature about turbulent convective plumes. Theoretical methods describing such plumes are well developed in the literature and are usually based on similarity theory and dimensional analysis. Most previous theoretical and experimental studies have been conducted for ascending convective plumes in the atmosphere. In this section, we use theoretical methods for studying the plume of cold water generated by a single wave-activated pump.

Let’s assume that the pump brings cold water to the sea surface with the temperature anomaly $\Delta T$ [K] at the rate $P$ [m$^3$/s]. Then, the associated negative buoyancy flux can be expressed as

$$Q = c\rho_0 P \Delta T$$

(1.1)

where $\rho_0 = 1027$ kg m$^{-3}$ is the average density of seawater and $c = 4185$ J kg$^{-1}$K$^{-1}$ is heat capacity of water. If $P = 1$ m$^3$/s and $\Delta T = 10$ K, then $Q \approx 4 \times 10^7$ W = 40 MW. This is an enormous power source! By comparison, the world's largest turbine (manufactured by the German companies Enercon and REpower) delivers up to 6 MW and has an overall height of 186 m and a diameter of 114 m (source Wikipedia).
Figure 1. A schematic diagram of a buoyant plume generated by a single wave-activated pump.

We first consider a case in which we neglect the effect of salinity on density stratification. In this case, density anomaly $\Delta \rho$ is uniquely determined by the temperature anomaly $\theta$ through a linearized equation of state:

$$\Delta \rho = -\rho_0 \alpha \theta,$$  \hspace{1cm} (1.2)

where $\alpha$ = thermal expansion coefficient.

Assuming the flow is stationary and the radial profiles of mean vertical velocity $w$ and mean buoyancy are similar at all heights, the governing equations can be written as:

$$\frac{d}{dz} \left( w R \right)^2 = \frac{a_2}{a_1} \alpha g \theta R^2,$$  \hspace{1cm} (1.3)

$$\frac{d}{dz} \left( w \theta R \right)^2 = \frac{a_4}{a_3} \gamma w R^2.$$  \hspace{1cm} (1.4)

Here, $z$ = vertical coordinate directed downward, $R(z)$ = radius of the plume, $g$ = acceleration due to gravity, $\gamma(z)$ = vertical temperature gradient ($\gamma < 0$ corresponds

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to stable stratification), and \( a_i \equiv \text{nondimensional coefficients (according to [7]},
\)
\( a_i = a_1 = 1/4 \) and \( a_2 = a_3 = 1/3 \). Here and below, \( \theta \) corresponds to its absolute value, i.e. \(|\theta|\).

We also assume that the mean radial (inflow) velocity is proportional to the vertical velocity:

\[
R(z) = \beta z,
\]

where nondimensional coefficient \( \beta \) represents the entrainment rate of the plume into the ambient fluid. According to [2, 7], \( \beta \sim 0.15 \).

We first consider neutrally stratified fluid \((\nu = 0)\), which corresponds to a well-mixed layer in the upper ocean. The solutions of the above system of equations can be written as:

\[
w \approx \left( \frac{3a_2}{8\pi a_1 a_3} \frac{\alpha g P \Delta T}{\beta^2 z} \right)^{1/3}, \quad \theta \approx \left( \frac{a_1}{3\pi^2 a_2 a_3^2} \frac{P^2 \Delta T^2}{\alpha g \beta^4 z^5} \right)^{1/3}
\]  

(1.6)

Considering the values of the nondimensional parameters specified above, we obtain:

\[
w \approx \left( \frac{2\alpha g P \Delta T}{\pi \beta^2 z} \right)^{1/3}, \quad \theta \approx \left( \frac{4P^2 \Delta T^2}{\pi^2 \alpha g \beta^4 z^5} \right)^{1/3}
\]  

(1.7)

Let’s do some estimates.

If \( P = 1 \text{ m}^3/\text{s}, \Delta T = 10 \text{ K}, \) and \( \alpha = 3 \times 10^{-4} \text{ K}^{-1} \), then at a depth \( z = 30 \text{ m} \) (typical depth of the mixed layer in the Gulf of Mexico in summer), we obtain \( w \sim 0.3 \text{ m/s and } \theta \sim 1 \text{ K} \). If \( P = 0.33 \text{ m}^3/\text{s}, \) then \( w \sim 0.2 \text{ m/s and } \theta \sim 0.48 \text{ K}. \)

Note that according to (1.5), the radius \( R \) is dependent only on depth and entrainment rate \( \beta \) (\( \beta = 0.15 \) in this case), and \( R = 4.5 \text{ m at } z = 30 \text{ m.} \)

Physical interpretation of the above solution is as follows: The pump-generated cold plume widens and its temperature increases (i.e. its temperature anomaly decreases) with depth by entraining the warmer ambient fluid. The temperature anomaly is reduced by more that 10 times when the plume reaches the bottom of the mixed layer. If there is a large temperature gradient below the mixed layer, which is typical for the Gulf of Mexico in summer, the cold plume will not be able to descend further and will remain in the

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mixed layer. The heat content of the mixed layer will thus be reduced. It should be noted that the cold water in this case will “concentrate” near the bottom of the mixed layer and will spread horizontally with time. This process will be discussed below.

We now estimate the descent time of each volume of cold water. Let \( Z \) be the level at which the volume is located at a given time (i.e. its Lagrangian coordinate). Considering \( w \) as a derivative \( dZ / dt \) in (1.7), we obtain the following nonlinear differential equation:

\[
\frac{dZ}{dt} = \left( \frac{2\alpha g P \Delta T}{\pi \beta^2 Z} \right)^{1/3},
\]

where \( t \equiv \) time. Integrating this equation, we obtain

\[
Z^{4/3} = \frac{4}{3} \left( \frac{2\alpha g P \Delta T}{\pi \beta^2} \right)^{1/3} t; \quad Z = \frac{4}{3} \left( \frac{2\alpha g P \Delta T}{\pi \beta^2} \right)^{1/4} t^{3/4},
\]

\[
t = \frac{3}{4} \left( \frac{2\alpha g P \Delta T Z^4}{\pi \beta^2} \right)^{1/3} = \frac{3}{4} \frac{Z}{w}.
\]

We can now readily estimate that a cold volume released at the sea surface will reach the depth of \( Z = 30 \text{ m} \) in \(~75 \text{ s}\) if the pump rate \( P = 1 \text{ m}^3/\text{s} \) and in \(~110 \text{ s}\) if \( P = 0.33 \text{ m}^3/\text{s} \).

Fig. 2 shows the vertical profiles of \( \Theta(z) \) [K] and \( w(z) \) [m/s] in the case of \( P = 1 \text{ m}^3/\text{s} \). We can see that the temperature anomaly decreases with depth faster than the vertical velocity does.
Figure 2. Vertical profiles of (1) $10 \, w(z) \, [m/s]$ and (2) $\theta(z) \, [K]$ in the plume case.

The above solution is in good agreement with results of laboratory experiments (see for example, equation (6.1.3) in [2]).

We should note that the above solution is strongly dependent on the following three parameters: $\beta \equiv$ entrainment rate of the plume into the ambient fluid, $P \equiv$ the pump rate, and $\Delta T \equiv$ temperature deference between the cold water brought to the sea surface and the ambient water. These parameters can be more precisely determined from future field experiments.

We now estimate the turbulent exchange coefficient $K$ in the plume. According to classical semi-empirical turbulence theory,

$$K = l^2 \left| \frac{\partial w}{\partial r} \right|, \quad (1.10)$$

where $l \equiv$ turbulence scale, which is usually assumed to be proportional to the cross-stream size $l = aR$, where $a \equiv$ nondimensional parameter. Rewriting (1.10) using a finite-difference approximation, we obtain
\[
K \sim a^2 R \omega = a^2 \left| \frac{2 \alpha g \beta P \Delta T z^2}{\pi} \right|^{1/3}.
\]

If \( a = 0.5 \), \( w = 0.3 \text{ m/s} \), and \( R = 5 \text{ m} \), then \( K \sim 0.5 \text{ m}^2/\text{s} \). This value of \( K \) is much larger than typical values of the background turbulence in the mixed layer.

2. Thermals

Let’s now consider another theoretical model when cold water is not released from the pump continuously, but by discrete volumes, so-called “thermals” (Fig. 3). A thermal can be defined as “discrete buoyant element in which the buoyancy is confined to a limited volume of fluid” [8]. Thermodynamics of thermals is extensively investigated in the literature [1, 2, 6, 9].

An important property of a thermal is that its linear dimension \( R \) in the first approximation increases proportionally to the vertical distance [1, 2, 6]:

\[
R = \beta_1 z.
\]  

(2.1)

This equation is analogous to plumes considered above, but the nondimensional entrainment coefficient, \( \beta_1 \), is much larger than \( \beta \) in plumes, according to experimental data. For example, according to [1], \( \beta_1 \approx 0.25 \). This implies significantly more efficient turbulent mixing with the ambient fluid than in the case of plume.

The shape of thermals is generally not spherical. According to experimental data in [1], the volume of a single thermal can be estimated as

\[
V \approx 3 R^3.
\]  

(2.2)

Let’s denote the initial volume of a thermal released near the sea surface as \( V_0 \), with a temperature anomaly from the ambient fluid as \( \Delta T \). Then, the excessive mass of the thermal due to its cooler temperature is

\[
\Delta M = V_0 \Delta \rho = V_0 \rho_0 \alpha \Delta T.
\]  

(2.3)

Conservation of heat requires

\[
V(z) \theta(z) = V_0 \Delta T,
\]  

(2.4)

where \( \theta(z) \) is the absolute value of temperature anomaly from the ambient fluid. From (2.4), (2.2), and (2.1), we obtain:
\[ \theta(z) = \Delta T \frac{V_0}{V(z)} = \frac{V_0 \Delta T}{3R^3(z)} = \frac{V_0 \Delta T}{3 \beta_1^3 z^3}. \] (2.5)

Figure 3. A sketch of descending thermals released near the sea surface.

We can see immediately that, as expected, the thermal is mixed with the surrounding water much faster than the plume: \( \theta \sim z^{-3} \) instead of \( z^{-5/3} \).

Since the wave pump ejects cold water on wave upslope, each release may be considered to be a thermal of a volume \( V_0 \), where \( V_0 = PT_w \), \( P \) is the pump rate, and \( T_w \) is the wave period.

Let's do some estimates: If \( P = 1 \text{ m}^3/\text{s} \) and \( T_w = 10 \text{ s} \), then \( V_0 = 10 \text{ m}^3 \). If we also assume that \( \Delta T = 10 \text{ K} \) and \( \beta_1 = 0.25 \), then at a depth of \( z = 30 \text{ m} \), the temperature anomaly \( \theta < 0.1 \text{ K} \), which is 10 times smaller than the temperature anomaly of the plume at the same depth!

If \( P = 0.33 \text{ m}^3/\text{s} \) and \( T_w = 15 \text{ s} \), then \( V_0 \sim 5 \text{ m}^3 \). In this case, the temperature anomaly at the bottom of mixed layer is \( \theta < 0.05 \text{ K} \).

The vertical velocity of the thermal is calculated using the following approximation [1, 2, 6]:

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\[ w \sim \sqrt{\frac{gR \Delta \rho}{\rho_0}}. \]  

(2.6)

Using (2.3) and (2.4), we obtain:

\[ w \sim \sqrt{\frac{gR \alpha V_0 \Delta T}{V}} = \sqrt{\frac{gR \alpha V_0 \Delta T}{3R^3}} = \frac{1}{R} \sqrt{\frac{\alpha gV_0 \Delta T}{3}} = \frac{1}{\beta_1 z} \sqrt{\frac{\alpha gV_0 \Delta T}{3}} \]  

(2.7)

We can see that the vertical velocity decreases with depth much faster than in the case of plume \((w \sim z^{-1} \text{ instead of } z^{-1/3})\). For \(V_0 = 10 \text{ m}^3\), \(\Delta T = 10 \text{ K}\), and \(z = 30 \text{ m}\), \(w \sim 0.04 \text{ m/s}\), which is about 10 times smaller than that in the plume! For \(V_0 = 5 \text{ m}^3\), \(w \sim 0.03 \text{ m/s}\) at \(z = 30 \text{ m}\).

Fig. 4 shows 10 \(w(z)\) [m/s] and \(\theta(z)\) [K] in this case. It is evident that the temperature anomaly decreases with depth faster than the vertical velocity does. This setup indicates a very efficient heat exchange between the thermal and the ambient fluid.

We now estimate the descent time of each thermal. Let’s denote by \(Z\) the level at which the thermal is located at a given time (i.e. its Lagrangian coordinate). Considering \(w\) as a derivative \(dZ / dt\) in (2.7), we get the following nonlinear differential equation:

\[ \frac{dZ}{dt} = \frac{1}{\beta_1 Z} \sqrt{\frac{\alpha gV_0 \Delta T}{3}}, \]  

(2.8)

where \(t = \text{time}\). Integrating the above, we obtain

\[ Z \sim \left( \frac{4}{3 \beta_1^2} \alpha gV_0 \Delta T \right)^{1/4} t^{1/2}; \quad t \sim \frac{\beta_1 Z^2}{2} \left( \frac{3}{\alpha gV_0 \Delta T} \right)^{1/2} = \frac{1}{2} Z W \]  

(2.9)
From these equations, we can estimate that it takes ~6 min for a thermal of volume $V_0 = 10 \text{ m}^3$ (~8 min for $V_0 = 5 \text{ m}^3$) to reach the bottom of the mixed layer ($Z = 30 \text{ m}$). This result can be compared to ~1 min in the case of plume.

Note that the above results are strongly dependent on the value $\beta_1$, which is usually determined experimentally. According to previous laboratory studies, this parameter cannot be much smaller than the value used in our analysis. Therefore, our conclusion about the very efficient mixing of a thermal with the ambient water is likely to be robust.

In summary, we have shown that releasing cold water near the sea surface by discrete volumes (thermals) is a very efficient method for keeping it within the upper part of the mixed layer. Other factors, such as the background turbulence and current, will increase further the efficiency of this method. The background turbulence will increase mixing of the thermal with the ambient fluid. The current will move the thermal horizontally and therefore the next thermal will be released in a new environment, making horizontal mixing even more efficient.

3. Effect of stratification: thermals

Vertical temperature gradients in the mixed layer are small and typically can be neglected. Salinity within the mixed layer, however, may increase with depth, making the

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$Figure 4. Vertical profiles of (1) 10 w(z) [\text{m/s}] and (2) \theta(z) in the thermal case.$

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density stratification stable. In this section, we consider this situation, which commonly occurs in the Gulf of Mexico.

The density anomaly of each thermal released by the pump decreases with depth as the thermal descends due a reduction of its temperature anomaly: \( \Delta \rho(z) = \rho_0 \alpha \theta(z) \).

Applying (2.5), we obtain

\[
\Delta \rho(z) = \rho_0 \alpha \theta(z) = \frac{\alpha \rho_0 V_0 \Delta T}{3 \beta_1^3 z^3}.
\]  

(3.1)

Because of the salinity-regulated stable stratification, the background density increases with depth as

\[
\Delta \rho_b(z) = \rho_0 \alpha_1 \gamma_s z,
\]  

(3.2)

where \( \alpha_1 \equiv \) coefficient of seawater expansion due to salinity and \( \gamma_s \equiv \) background vertical gradient of salinity. By equating (3.1) and (3.2), we obtain a depth, \( z_\ast \), which corresponds to zero (neutral) buoyancy of the thermal:

\[
z_\ast = \left( \frac{\alpha V_0 \Delta T}{3 \beta_1^3 \alpha_1 \gamma_s} \right)^{1/4} \quad \text{or} \quad z_\ast = \left( \frac{\alpha g V_0 \Delta T}{3 \beta_1^3 N^2} \right)^{1/4},
\]  

(3.3)

(3.4)

where \( N \) is the Brunt-Väisälä or buoyancy frequency.

If \( V_0 = 10 \, \text{m}^3, \Delta T = 10 \, \text{K}, \beta_1 = 0.25, \alpha = 3 \times 10^{-4} \, \text{K}^{-1}, \) and \( N \sim 2 \times 10^{-2} \, \text{s}^{-1} \) (typical value in the Gulf of Mexico), then from (3.4) it follows that \( z_\ast \approx 10 \, \text{m} \).

Thus, for the considered set of parameters, a thermal will not fall below a depth of \( \sim 10 \, \text{m} \) (even in a thermally unstratified mixed layer!) due to the effect of salinity stratification.

If the pump rate is \( P = 0.33 \, \text{m}^3/\text{s} \), we can estimate that \( V_0 = 5 \, \text{m}^3 \) and that \( z_\ast \approx 7 \, \text{m} \).

4. Effect of stratification: plumes

Considering the effect of salinity on plumes, we first modify the equation of state for the density anomaly as

\[
\Delta \rho = \rho_0 (\alpha \theta + \alpha_1 s),
\]  

(4.1)
where $\alpha$ and $\alpha_1 \equiv$ expansion coefficients of seawater due to temperature and salinity, respectively, and $S(z) \equiv$ salinity anomaly in the plume. Governing equations (1.3)-(1.4) can be modified as:

\[
\frac{d}{dz} \left( wR^2 \right) = \frac{a_2}{a_1} g(\alpha \theta + \alpha_1 s) R^2, \quad (4.2)
\]

\[
\frac{d}{dz} \left( w\theta R^2 \right) = 0, \quad (4.3)
\]

\[
\frac{d}{dz} \left( w_s R^2 \right) = -\frac{a_4}{a_3} \gamma_s wR^2. \quad (4.4)
\]

Here, $\gamma_s(z) \equiv$ vertical salinity gradient as a function of depth ($\gamma_s > 0$ corresponds to stable stratification).

Let’s now introduce a new variable

\[
b(z) = \alpha \theta(z) + \alpha_1 s(z). \quad (4.5)
\]

Using (4.5) and combining (4.3) and (4.4), we obtain

\[
\frac{d}{dz} \left( w b R^2 \right) = -\frac{a_4}{a_3} \alpha_1 \gamma_s wR^2. \quad (4.6)
\]

Equation (4.2) can be rewritten as

\[
\frac{d}{dz} \left( wR \right)^2 = \frac{a_2}{a_1} gbR^2. \quad (4.7)
\]

Assuming as before that

\[
R(z) = \beta z., \quad (4.8)
\]

equations (4.6) and (4.7) now have two unknowns: $w(z)$ and $b(z)$. These equations are nonlinear and cannot be solved analytically. We therefore find the solutions using numerical methods. Once $w(z)$ and $b(z)$ are calculated, we can readily calculate $\theta(z)$ and $s(z)$ from (4.3)-(4.4).
Figs. 5 and 6 show the numerically-calculated profiles of $w(z)$ and $\theta(z)$ for neutral and stable stratifications using the parameters defined in Section 1.

![Figure 5](image1.png)

**Figure 5.** Profiles of $10 \, w(z) \, [m/s, dashed \, line]$ and $\theta(z) \, [K]$ in the mixed layer for neutral stratification.

![Figure 6](image2.png)

**Figure 6.** The same as in Fig.5, but for stable salinity stratification, corresponding to $N=2 \times 10^{-2} \, s^{-1}$.

We can see that stable stratification due to salinity leads to some decrease in the vertical velocity and some increase in temperature anomalies, but this effect is considerably smaller than in the case of thermals.

5. **Horizontal dispersion of the cold-water intrusion**

As discussed above, a cold water parcel ejected by the wave pump efficiently mixes with the surrounding water while descending through the mixed layer. With a neutral density stratification, the parcel can fall to the bottom of the mixed layer, forming a volume of water that is colder and denser than the background water but less cold and dense than the water below. Hereafter, this volume of water will be defined as a “cold-water intrusion”.

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Once it is formed, the intrusion spreads horizontally under its own weight. This process is known as a gravity current and is a quite common and essential mechanism for moving fluid with larger density horizontally in the oceans and atmosphere. Below, we will analyze the intrusion at the bottom of the mixed layer following the theoretical framework of [10, 11]. We consider two cases: 1) an isolated intrusion produced by a thermal (instantaneous source of cold water), and 2) an intrusion produced by a plume (continuous source of cold water).

a) **Intrusion produced by a thermal** (Fig. 7). We assume here that the horizontal scale of the intrusion is much larger that the vertical scale and the horizontal gradient of pressure is counterbalanced by the viscous stress. Let’s denote the background density in the mixed layer, the density below the mixed layer, and the density in the intrusion as $\rho_1, \rho_2, \rho_3$, respectively ($\rho_1 < \rho_3 < \rho_2$). We will use the same indices for corresponding temperatures ($T_1 > T_3 > T_2$). The intrusion with temperature $T_3$ has negative buoyancy relative to the ambient water in the mixed layer, but it does not sink because of a small deformation of the bottom of the mixed layer (Fig. 7). The fraction of the intrusion that penetrates below the mixed layer is

$$\varepsilon = \frac{h_2}{h} = \frac{\rho_3 - \rho_1}{\rho_2 - \rho_1} = \frac{T_1 - T_3}{T_1 - T_2}, \quad 0 \leq \varepsilon \leq 1 \quad (5.1)$$

where $h$ $\equiv$ height of the intrusion and $h_2$ $\equiv$ depth of the intrusion below the mixed layer. As discussed above, temperature anomaly in the thermal when it reaches the bottom of the mixed layer, $\theta \equiv T_1 - T_3$, is much less than $T_1 - T_2$. The ratio $\varepsilon$ is small in this case, and thus the intrusion will spread as on a solid surface. From the hydrostatic relationship, the pressure in the intrusion can be estimated as

$$p = \frac{gh(\rho_3 - \rho_1)(\rho_2 - \rho_3)}{2(\rho_2 - \rho_1)} = \frac{gh \alpha \rho_0 (T_1 - T_3)(T_3 - T_2)}{2(T_1 - T_2)}, \quad (5.2)$$

**Figure 7. Schematic of an isolated cold-water intrusion at the bottom of the mixed layer.**
where $\rho_0 \equiv$ average density and $\alpha \equiv$ thermal expansion coefficient.

In our case, $T_3 - T_2 \approx T_1 - T_2$, so

$$p \approx gh(\rho_3 - \rho_1)/2 = gh\alpha \rho_0 \theta/2.$$ (5.3)

The pressure gradient force responsible for dispersion of the intrusion is

$$-h\nabla p = -\frac{g\alpha \rho_0 (T_1 - T_3)(T_3 - T_2)}{2(T_1 - T_2)} h \nabla h.$$ (5.4)

It is balanced by the viscous friction force

$$c\rho_0 \nu (u/h).$$ (5.5)

Here $u \equiv$ the horizontal velocity vector, $\nu \equiv$ kinematic viscous coefficient, and $c \equiv$ nondimensional constant (according to [10], $c=12$). We now use the following conservation law in the hydraulic approximation [10]

$$\frac{\partial h}{\partial t} = -\text{div}(h u),$$ (5.6)

where $t \equiv$ time. Note, that here we neglect the heat exchange between the intrusion and the environment. Equating (5.4) and (5.5) we can derive an equation for $u$ and substitute it in (5.6). As a result, we obtain the following nonlinear equation for $h$:

$$\frac{\partial h}{\partial t} = k \Delta h^4.$$ (5.7)

Here,

$$k = \frac{g(\rho_2 - \rho_3)(\rho_3 - \rho_1)}{8c \nu \rho_0 (\rho_2 - \rho_1)} = \frac{g\alpha (T_1 - T_3)(T_3 - T_2)}{8c \nu (T_1 - T_2)} \approx \frac{g\alpha \theta}{8c \nu}.$$ (5.8)

The nonlinear equation (5.7) allows automodelling solutions [10, 11]. The automodelling solution is sought in the following form:

$$h(r, t) = A t^\sigma F\left(\frac{r}{B t^{\lambda}}\right),$$ (5.9)
where $F \equiv$ dimensionless function, $(\sigma, \lambda) \equiv$ dimensionless parameters, and $(A, B) \equiv$ constants of corresponding dimensions. The products $At^\sigma$ and $Bt^\lambda$ represent the characteristic thickness, $h(t)$, and the horizontal radius, $R(t)$, respectively of the intrusion.

The total volume of the intrusion, $hR^2$, where $hR^2 \sim AB^2 t^{\sigma+2\lambda}$, must be conserved and equal to its initial value $V_0$. This implies that $\sigma + 2\lambda = 0$. Another relationship between $\sigma$ and $\lambda$ can be obtained by equating the horizontal gradient force, which is proportional to $h/R$, and the viscous stress, which is proportional to $u/h^2 \sim R/th^2$, where $u \sim R/t$ = the characteristic speed of horizontal dispersion. From here, it follows that $3\sigma = 2\lambda - 1$. From the two above equations, we find that $\sigma = -1/4$ and $\lambda = 1/8$. The coefficients $A$ and $B$ can be found from dimensional analysis. There are two dimensional parameters in this problem: $V_0$ [m$^3$] and $k$ [m$^{-1}$ s$^{-1}$].

From the dimensional analysis, we can obtain $A \sim (V_0/k)^{1/4}$ and $B \sim (kV_0^3)^{1/8}$. Substituting (5.9) in (5.7), we obtain a nonlinear differential equation for function $F$, which allows an analytical solution:

$$h(r,t) = \frac{1}{2} \left[ \frac{V_0}{\pi kt} \right]^{1/4} F(\xi), \quad (5.10)$$

$$F(\xi) = \begin{cases} 
(1 - \xi^2)^{1/3}, & 0 \leq \xi \leq 1, \\
0, & \xi \geq 1, 
\end{cases}$$

$$\xi = \frac{r}{\left[ \frac{8}{3} \left( \frac{kV_0^3}{\pi^3} \right)^4 \right]^{1/8}}.$$ 

The universal dimensionless radial profile $F(\xi)$, applied for any moment of time, is shown in Fig. 8. Since the equation (5.7) is nonlinear, its solution (5.10) is qualitatively different from standard parabolic equations. First, the intrusion has a sharp boundary (for linear parabolic equations, the solutions have typically no well-defined boundary). Another difference is that the speed of dispersion depends on the initial volume $V_0$ – it increases with an increase of $V_0$. Also, the dispersion speed quickly decreases with time.
From (5.9) and (5.10) we obtain equations for the radius and thickness of intrusion as functions of $V_0$ and $t$:

$$R = \left[\left(\frac{8}{3}\right)^4 \frac{kV_0^3 t}{\pi^3}\right]^{1/8}$$

$$h|_{r=0} = \frac{1}{2} \left[\frac{V_0}{\pi kt}\right]^{1/4}.$$  \hspace{1cm} (5.11)

**Figure 8. Universal dimensionless radial profile of $h$ during horizontal dispersion of the intrusion.**

If we assume: $\nu = 10^{-3}$ m$^2$/s, $\theta = 0.1$ K (for $P = 1$ m$^3$/s), $\alpha = 3 \times 10^{-4}$ K$^{-1}$, and $c = 12$ [1], then $k \approx 3 \times 10^{-3}$ m$^{-1}$s$^{-1}$. Fig. 9 illustrates the evolution of two different intrusions with volumes $V_0$ equal to 100 m$^3$ (1,3) and 1000 m$^3$ (2,4). It is evident that the time dependence of the radius and thickness of the intrusion is rather weak. This indicates that the horizontal dispersion of an isolated intrusion under its own weight is a slow process.
Figure 9. Maximum thickness (1,2) and radius (3, 4) of isolated cold-water intrusions with volume \( V_0 \) equal to 100 m\(^3\) (1,3) and 1000 m\(^3\) (2,4) as function of time.

b) Intrusion produced by a plume. We considered above the dispersion of an isolated intrusion. Now we consider a similar problem, but here the intrusion is generated by a plume when it reaches the bottom of the mixed layer and spreads across horizontally (Fig. 10). In contrast to the isolated intrusion case, the spreading intrusion in this case is constantly replenished with descending cold water. We assume that the main part of the plume is already in a stationary state and "delivers" cold water to the bottom of the mixed layer with a constant intensity flux \( Q \sim \pi R^2 w \) (m\(^3\)/s), where the values of \( w \) and \( R \) refer to the level where the plume ends. We consider the non-stationary process of further dispersion of cold water horizontally along the bottom of the mixed layer.

As in the previous case, we neglect the effect of mixing with the environment and the background stratification. We can therefore still use the equation (5.7) and consider a solution diminishing for \( r \rightarrow \infty \). In this case, however, we specify a stationary source \( Q \) at the axis, \( r = 0 \).
Figure 10. Schematic of the horizontal dispersion of the intrusion generated by a plume when it reaches the bottom of the mixed layer.

As in the previous problem, we search for an automodelling solution (5.9). In this case, the volume of the intrusion is proportional to $hR^2 \sim t^{\sigma + 2\lambda}$ and is equal to $Qt$. We can thus obtain the following relationship: $\sigma + 2\lambda = 1$. Another relation, which is based on the balance of the pressure gradient and viscous forces, is the same as in the previous problem. From these two equations, we find $\sigma = 0$ and $\lambda = 1/2$. The coefficients $A$ and $B$ can be determined from a simple dimensional analysis. There are two dimensional parameters in this problem: $Q$ [m$^3$/s] and $k$ [m$^{-1}$ s$^{-1}$]. From dimensional analysis, we obtain $A = (Q/k)^{1/4}$ and $B = (kQ^3)^{1/8}$. Thus,
\[ h(r,t) = \left( \frac{Q}{k} \right)^{1/4} F_1 \left( \frac{r}{\left( kQ^3 \right)^{1/8} t^{1/2}} \right). \]  

(5.12)

The problem is thus reduced to an ordinary differential equation for function \( F_1 \). From the above consideration, we can see that horizontal dispersion of cold water in this case is analogous to the process of horizontal diffusion, with the “effective” coefficient

\[ D = \left( kQ^3 \right)^{1/4} \left( \frac{g \alpha \theta Q^3}{8c \nu} \right)^{1/4}. \]  

(5.13)

We can now estimate the radius the intrusion, \( L \):

\[ L \sim (Dt)^{1/2} \left( \frac{g \alpha \theta Q^3}{8c \nu} \right)^{1/4} t^{1/4}; \]  

(5.14)

and its volume

\[ L^2 h \sim h \left( kQ^3 \right)^{1/4} t. \]  

(5.15)

Taking into account that the volume (5.15) must be equal to the volume of water coming from the plume \( Qt \), we obtain the characteristic thickness of the intrusion

\[ h \sim \left( \frac{Q}{k} \right)^{1/4}. \]  

(5.16)

Let’s now do some estimates. As discussed in Section 1, at a depth of \( z = 30 \) m, \( R = 4.5 \) m. For \( P = 1 \) m\(^3\)/s, \( w \sim 0.3 \) m/s, \( \theta \sim 1 \) K, and the mass flux \( Q \sim \pi R^2 w \sim 20 \) m\(^3\)/s. For \( P = 0.33 \) m\(^3\)/s, \( w \sim 0.2 \) m/s, \( \theta \sim 0.48 \) K, and \( Q \sim 13.3 \) m\(^3\)/s.

If we assume as before that \( \nu = 10^{-3} \) m\(^2\)/s, \( \alpha = 3 \times 10^{-4} \) K\(^{-1}\), and \( c = 12 \), then \( k \approx 3 \times 10^{-2} \) m\(^{-1}\)s\(^{-1}\) (for \( P = 1 \) m\(^3\)/s) and \( k \approx 1.5 \times 10^{-2} \) m\(^{-1}\)s\(^{-1}\) (for \( P = 0.33 \) m\(^3\)/s). These values are an order of magnitude larger than in the case of an isolated intrusion. This is because the temperature anomaly \( \theta \) is in order of magnitude larger in this case. From (5.13), we estimate the “effective” coefficient of horizontal diffusion \( D \sim 4 \) m\(^2\)/s (for \( P = 1 \) m\(^3\)/s) and \( D \sim 2.4 \) m\(^2\)/s (for \( P = 0.33 \) m\(^3\)/s) and use these values to estimate the radius of intrusion from (5.14). Fig. 11 shows the dependence of radius of the intrusion as function of time for different parameters of the plume. We can see the horizontal
dispersion of the intrusion is very efficient and much faster than that in the isolated intrusion case. In about one hour, the radius of the intrusion $L$ can reach $\sim 120$ m for ($P = 1$ m$^3$/s) and $\sim 93$ m for ($P = 0.33$ m$^3$/s).

![Figure 11](image)

**Figure 11.** Radius of the intrusion as function of time for different parameters of the plume. (1) $Q = 20$ m$^3$/s and $k \approx 3 \times 10^{-2}$ m$^{-1}$s$^{-1}$; (2) $Q = 10$ m$^3$/s and $k \approx 3 \times 10^{-2}$ m$^{-1}$s$^{-1}$; (3) $Q = 20$ m$^3$/s and $k \approx 10^{-2}$ m$^{-1}$s$^{-1}$.

We should note that spatial dispersion of a cold-water intrusion, generally speaking, is a rather complex process. In addition to the gravity force considered here, the dispersion is also affected by the background turbulent diffusion and stratification. However, for typical oceanic conditions in the Gulf of Mexico during summer, both of these latter effects are rather small. Therefore, our theoretical results are likely to be robust.

In summary, we have considered spatial and temporal evolutions of isolated and plume-like intrusions of cold water when they reached the bottom of mixed layer. We concluded that in the former case, the horizontal dispersion is very slow, while in the latter case, it is much faster. How does this relate to the evaluation of cold water generated by a wave pump? As discussed above, the pump ejects cold water at the surface by finite volumes (thermals) on wave upslope. Each thermal falls to the bottom of the mixed layer ($Z = 30$ m) in about 6-8 minutes (assuming that density stratification is neutral). In both cases,
the fall time of a thermal is considerably smaller that the time scale of the horizontal dispersion (and the time scale difference in the isolated intrusion case is even greater than in the plume-like intrusion case). This implies that the thermals will “pile-up” on top of each other and will form a plume-like intrusion. Therefore, the above estimates for the plume-like intrusion are probably more applicable than those for an isolated intrusion.

6. Interaction between multiple pumps

In an array of pumps, the cold-water intrusions will interact with each other if the pumps are placed at reasonably close distances to each other. The intrusions will eventually merge and create a uniform cold-water “pool”. There are three main mechanisms by which the pump-induced cold-water intrusions can interact: 1) horizontal turbulent diffusion, 2) gravity currents, and 3) advection by background current.

Simple estimations show that horizontal turbulent diffusion is not very effective. The characteristic time scale of horizontal diffusion is \( \tau_K = \frac{L_0^2}{K} \), where \( L_0 \equiv \) the distance between pumps, and \( K \equiv \) turbulent exchange coefficient. If \( L_0 = 100 \text{ m} \) and \( K = 0.1 \text{ m}^2/\text{s} \) (a typical ocean value for large-scale horizontal diffusion), we obtain that the time scale \( \tau_K \) is more than a day.

The effect of background horizontal current is generally dependent on the current strength and mutual orientation of the pumps and the current. In the case shown schematically in Fig. 12, the time of “merger” can be estimated as \( \tau = \frac{L_0}{u} \). If \( L_0 = 100 \text{ m} \) and \( u = 0.05 \text{ m/s} \) (a typical value in common ocean waters away from oceanic fronts and eddies), then the cold water generated by one pump will reach the next pump along the current in about half an hour. In the above configuration, horizontal jets of cold water will initially form along each chain of pumps in the direction of the current, and it will take some time until these jets merge.

![Figure 12. Schematic diagram of a multiple pump array.](image)

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Finally, the merger time due to gravity currents can be estimated as \( \tau_K = L_0^2 / D \), where \( D \) is the “effective” coefficient of horizontal diffusion given by (5.13). According to the estimates made in section 5, the value of \( D \) is \( \sim 4 \text{ m}^2/\text{s} \) for the set of parameters discussed in that section. If \( L_0 = 100 \text{ m} \), then \( \tau_K \) is less than an hour.

References

II. Numerical investigation of the impact of the upper ocean layer cooling on hurricanes using a coupled hurricane-ocean model.

Introduction

Tropical cyclones represent an extreme case of air-sea interaction. The effect of this interaction as a negative feedback on tropical cyclone development and intensity has been well established. It is known that strong surface winds in a tropical cyclone induce turbulent mixing in the upper ocean and entrainment of the underlying cold water into the ocean mixed layer, which cools and deepens (e.g., Bender et al. 1993; Ginis 2002). Both observational and real case numerical studies (e.g., Black, 1983; Bender and Ginis 2000) showed that the SST anomalies induced by tropical cyclones can reach up to 5-6 °C. Studies also showed that tropical cyclone intensity is more sensitive to the local SST changes under the hurricane core than to those beyond the core area (e.g., Emanuel, 1999; Shen et al., 2000). Therefore it can be expected that cooling of the ocean area underneath the hurricane core may reduce its intensity. As discussed in Part 1 of this report, wave-activated pumps can reduce the upper ocean heat content by bringing much colder water from ocean depths to the surface. An array of these pumps can be used to create a cooled region in front of the hurricane to mitigate its intensity.

Ginis Consulting, LLC has undertaken a study to address the following questions:

1. How sensitive is the hurricane intensity to the cooling of various size and magnitude ocean regions induced by an array of wave-activated pumps?

2. What are the optimal size and magnitude of the cooled region in order to reduce the hurricane intensity by 10-20%?

The study has been conducted for idealized and real-hurricane simulations using a fully coupled hurricane ocean model described next.

1. The hurricane-ocean coupled model

The NOAA Geophysical Fluid Dynamics Laboratory (GFDL) hurricane model coupled with the Princeton ocean model (Bender et al. 2006) is used in this study with idealized oceanic and atmospheric conditions. The GFDL hurricane prediction system became operational in 1995 as the U.S. National Weather Service’s (NWS) official hurricane model. Since that time, it has provided forecast guidance to forecasters at the NWS’s Tropical Prediction Center (TPC) and has been the most reliable forecast model for track error during the past decade (Table 1, courtesy of James Franklin, TPC). In addition, a version of the GFDL model (GFDN) was imported to the United States Navy in 1996 (Rennick 1999) to provide forecasted guidance in the Western Pacific. In the following year, the GFDN model started to provide forecasts in the other ocean basins including those in the Southern Hemisphere.
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**TABLE 1** Average track errors in nautical miles [nm] for all forecasts run in the Atlantic between 1996 and 2005 for the GFDL, NCEP’s GFS, UKMET and Navy’s NOGAPS models. Results are for the time interpolated models. Since numerical weather prediction models are generally not available to the forecasters in time to make their forecasts, a simple technique exists to take the model forecasted position and intensity, and adjust the forecast to apply to the current synoptic time and initial conditions. This adjustment is usually 6 or 12 hours, depending on the availability of the last model guidance. These adjusted versions are known for historical reasons as interpolated models which are generally indicated by the letter “I” at the end of the name (e.g., GFDI for the GFDL interpolated model).

**Figure 1.** Atlantic hurricane season intensity forecast errors for the GFDL model (solid line), the interpolated GFDI model (dot-dashed line), compared to the official TPC forecast (doted line)) and DECAY SHIPS (dashed line) for the years 2000 through 2006.
Important model upgrades have been introduced over the years to the physics and spatial and vertical resolutions. These upgrades led to significant increase in GFDL model intensity forecast skill as shown in Fig. 1. The operational GFDL model presently has ~9 km spatial resolution in the innermost movable mesh. For this study we increased the finest resolution to ~ 4.5 km to improve further its intensity forecast skill.

Figure 2. SST anomalies at 120 h in the control experiment.

1. Idealized experiments

For the idealized experiments in this study the coupled model is integrated for 120 hours starting with a normal size initial vortex embedded in specified initial, horizontally uniform environmental conditions. A GATE (Global Atlantic Tropical Experiment) III condition in the tropics is used for the atmospheric environmental thermal profile, which has air temperature of 27°C and relative humidity of 84% at the lowest model level (~ 35 m). A steady easterly environmental wind of -5 m/s is used. In the control experiment the ocean is initially horizontally uniform and motionless with the sea surface temperature of 28.5°C. The initial vertical temperature profile with a mixed layer depth of 30 m is typical for the northern Gulf of Mexico in September. In this study, we use a 5-m vertical resolution within the upper 100 m ocean depth to accurately resolve the upper ocean response to hurricane forcing. Simulation of the SST response in the control experiment is shown in Fig. 2. A typical SST cooling pattern is generated due to hurricane-ocean interaction. The largest cooling is on the right sight of the hurricane track, consistent with observations and other modeling studies (Ginis, 2002).
In two sensitivity experiments cooled swaths were placed in front of the hurricane as shown in Fig. 3. The size of each swath was about 400 km in the along-track direction and covered the entire computational domain in the cross-track direction. The temperature anomalies within the mixed layer in the swaths are 1°C and 2°C, correspondingly. Temperature profiles in the control and two sensitivity experiments at the beginning and the end of integrations at the location of maximum cooling are shown in Fig. 4. Note the differences in the initial profiles. At the end of the integration, the profiles look similar, with somewhat larger cooling in the swaths where the temperature anomalies were introduced. This is not surprising. This is because the entrainment rate of cooler water from thermocline into the mixed layer is dependent on the temperature difference between the mixed layer and the thermocline below. The entrainment rate is always reduced in the cooled swaths because of the reduced temperature differences (the larger the initial cooling of the mixed layer, the less the entrainment rate). However, the final temperature profiles after the hurricane passage are not that important for the hurricane intensity. The temperature anomalies (SSTA) underneath the hurricane core are more important. That is shown in Fig. 5. We can see that the SSTA are greatly reduced when the hurricane crosses the cooled swaths.

Evolution of hurricane central pressure and maximum winds in the numerical experiments are shown in Fig. 6. In both sensitivity experiments the hurricane intensity is reduced after the storm encounters the cooled swath. The maximum winds are reduced from ~145 kts to ~135 kts (6% reduction) in the 1°C swath experiment and to ~130 kts (10% reduction) in the 2°C swath experiment. No tangible impact on the hurricane tracks is found in these experiments, as shown in Fig. 7.
Figure 4. Temperature profiles at the beginning of the model integration (green) and at 120 h (red) in the control experiment (left), the 1°C swath experiment (center) and 2°C swath experiment.

An additional experiment was run in which the size of the 2°C swath was doubled along the track direction. As result, the hurricane intensity was further reduced from ~145 kts to ~120 kts (22% reduction). These sensitivity experiments clearly indicate that both the size and magnitude of the cooled area encountered by a moving hurricane make important impact on the hurricane intensity reduction.

Figure 5. SST anomalies (SSTA) underneath the hurricane core region, defined as a circular area around the storm center with r=100 km, as function of time.
Figure 6. Evolution of central pressure and maximum winds in three numerical experiments: control (blue line), 1°C anomaly (left panel, red) and 2°C anomaly (right panel, red).

Figure 7. Hurricane tracks in the control experiment (blue), the 1°C swath experiment (left, red) and 2°C swath experiment (right, red).

In order to investigate a potential impact of a cooled swath on a real hurricane in the Gulf of Mexico, we simulated Hurricane Ivan that crossed the Gulf of Mexico in 2004 and made landfall in Florida panhandle.

Hurricane Ivan (2004)

Figure 9. Observed track of Hurricane Ivan (2004).
Figure 10. The SST field and the 48-h track forecast of Hurricane Ivan (initial forecast time: 00 UTC 14 September 2004) using the GFDL hurricane forecast system – control case.

The simulation of Hurricane Ivan was conducted in a semi-operational mode, in which we use the operational GFDL hurricane and ocean initialization procedures which utilize all available atmospheric and oceanic observations available at the time of the forecast. The details of the model initialization can be found in Bender et al. (2006). For this study we have chosen one forecast case with the initial forecast time: 00 UTC 14 September 2004. That was the time when Ivan entered the Gulf of Mexico from the Caribbean Sea as a category 5 hurricane. Figure 10 shows the SST and forecast track of Hurricane Ivan for this case, which we refer to as the control case. Ivan created a large SST cooling before making landfall which likely led to its weakening. In two sensitivity experiments we placed 2°C cooled swaths as shown in Fig. 11. The locations of these swaths were chosen for consistency with Liz Richie’s simulations of Hurricane Ivan. The initial temperature profiles in the swaths are similar to those shown in Fig. 4. for the 2°C idealized experiment. Nevertheless, the size of the each swath in the along track direction was about two times smaller than those in the idealized experiments.

The SST fields at the end of the 48-h forecasts in these experiments are shown in Fig. 12 and the corresponding central pressure and maximum winds forecasts are shown in Fig. 13. In the case of one swath the central pressure and maximum winds were reduced by ~5 mb and ~5 kts (~3.5%), respectively. In the case of two swaths the intensity was further reduced by not significantly. Judging from the idealized experiments, doubling the size of the swath along the track (i.e. making it the same as in the idealized experiments) would likely reduced the intensity further. This evidently demonstrates that the size of the swath in the along-track direction should an important consideration for mitigation planning.
Figure 11. Locations and size of a 2°C cooled swath in the first sensitivity experiment and two 2°C cooled swaths in the second sensitivity experiment.

3. Recommendations for the size and magnitude of upper ocean cooling that can be generated by wave-activated swaths and required number of pumps

a) Magnitude of cooling

Our idealized hurricane experiments show that the hurricane intensity can be reduced as much as 20% if the averaged temperature anomaly in the mixed layer is equal to 2°C. This probably should be considered as a minimum requirement.

b) Along-track size of the swath

The along-track size should be dependent on the hurricane forward speed. For a speed of 5 m/s and mixed layer cooling of 2°C we found that the intensity can be reduced by ~10% for a 400-km size and by ~20% for an 800-km size.

Figure 12. The SST fields at 48 h in the first (left) and second sensitivity experiment (right).
c) Cross-track size of the swath

The minimum size of the cooled swath in the cross-track direction should be similar to the size of the hurricane core region where the air-sea fluxes that provide energy supply to the hurricane are the largest. A typical size of the hurricane core region is in order of 100 km. Another determining factor that needs to be considered is the track forecast errors by the National Hurricane Center. The cooled swath must obviously be positioned in front of the projected path. Fig. 14 shows the NHC track forecast errors in the very busy 2005 hurricane season calculated by two methods: using the NHC methodology (absolute errors) and based on the closest approach (the shortest distance between the observed track and the forecast track). The latter is a more relevant parameter that should be used for estimating the maximum size of the swath in the cross track direction. As seen in Fig. 14, the closest approach track errors are more than twice smaller than the absolute errors. A sum of the track forecast errors at a specified lead time and the hurricane core region is a reasonably good approximation of the required size of the cooled swath in the cross-track direction. For example, with a 24 h lead time, the total size of the cross-track swath should be at least ~150 km (50 km – track error and 100 km – hurricane core size). The corresponding sizes for 48 h and 72 h lead times will be ~200 km and ~250 km.

![Figure 13. Evolution of central pressure and maximum winds in three numerical experiments for Hurricane Ivan: control (blue line), one 2°C swath (left panel, red) and two 2°C swaths (right panel, red), which are located as shown in Fig. 11.](image)

d) Estimation of the size of the required pump array

Our theoretical analysis of cold water generated by a single pump (with the rate of 0.33-1.0 m³/s) indicates that it is probably feasible to generate average temperature anomalies of
at least 2°C in the mixed layer of a depth of 30 m. This, of course, is greatly dependent on the pump rate, which will be more accurately determined during future field tests.

Our theoretical analysis of spatial and temporal evolution of the cold water intrusions generated by a single and multiple pumps indicates that the intrusions will probably merge and create a continues region in not more than several hours if they are separated by ~100 m. Thus, in order to cover a region of 100 km x 100 km, $10^6$ pumps will be required.

Figure 14. Average track forecast errors in km of the National Hurricane Center in the 2005 hurricane season. The errors calculated using the NHC methodology (absolute errors) are shown in blue. The errors calculated based on the closest approach (the shortest distance between the observed track and the forecast track) are shown in red.

Let’s estimate how many pumps would be required to reduce hurricane intensity by 10% and 20%. We should note that since hurricane damage is roughly proportional to the wind speed cubed, 10% and 20% reduction in wind speed corresponds to ~30% and ~50% reduction in damage. Also note that one category on the Saffir-Simpson scale roughly corresponds to a 15% change in hurricane intensity (this number varies for different categories).

In our estimations below, we assume that a hurricane moves with a 5 m/s translation speed and the pumps generate a region of a uniform 2°C cooling.

1. In order to reduce the hurricane intensity by 10% with a 24 hour lead time, we will need to cool an area of the 150 km x 400 km, $6\times10^6$ pumps will be required. A 48 lead time would require $8\times10^6$ pumps, correspondingly.
2. In order to reduce the hurricane intensity by 20% with a 24 hour lead time, we will need to cool an area of the 150 km x 800 km, 12x10^6 pumps will be required. A 48 lead time would require 16x10^6 pumps, correspondingly.

References:


